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Tomographic imaging of multiple mantle plumes in the uppermost lower mantle

Mark Rhodes* and J. Huw Davies†

Department of Earth Sciences, University of Liverpool, Liverpool, L69 3GP. *E-mail: davies@liverpool.ac.uk*

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SUMMARY

A high-resolution global tomographic inversion of *P*-wave traveltimes has been undertaken utilizing an *a priori* model. The results image cylindrical slow velocities in the upper and lower mantle beneath many current hotspot locations. Locations at which such plume-like features are imaged passing from the uppermost lower mantle to the upper mantle include, Afar, Society Islands, Crozet, Kerguelen, Iceland, Hawaii, East Africa, Cape Verde and the Canary Islands. The validity of these images has been investigated with synthetic recovery tests. These images suggest that these plumes could be from the lower mantle and therefore are not hindered in crossing the upper/lower mantle boundary.

Key words: hotspots, mantle, mantle convection, plumes, seismic structure, tomography.

INTRODUCTION

Morgan (1971) postulated the presence of mantle plumes as part of the upwelling nature of Earth's convective regime but until recently few images of these phenomena existed. Tryggvason *et al.* (1983) produced some of the first seismic images of the Icelandic hot spot using *P*-wave traveltime data. Employing temporary seismometer arrays VanDecar *et al.* (1995) used teleseismic traveltime residuals to image a fossil mantle plume beneath South America and Wolfe *et al.* (1997) used regional and teleseismic data to illuminate the upper mantle seismic structure of the Iceland mantle plume. Nataf & VanDecar (1993) detected the Bowie mantle plume to the west of Canada from its delay of the traveltime of seismic phases. Bijwaard *et al.* (1998) have recently imaged the Yellowstone, Hawaii, Iceland and East African plumes crossing from the lower to the upper mantle, in their high resolution tomographic study. An indirect demonstration of the possible link between low-velocity seismic heterogeneity in the lower mantle and hotspots was made in the 'remarkable correlation' of Cazenave *et al.* (1989).

METHOD AND RESULTS

Earthquakes with teleseismic phases from 28° to 98° were carefully selected from the Engdahl *et al.* (1998) event catalogue negating the need for generating summary rays (Hager & Clayton 1989; Vasco *et al.* 1994). The resulting 775 000 *P*-wave phases covered the widest azimuth-epicentral distance range (Rhodes & Davies, in preparation). The model space was parametrized by a local basis function into $5^\circ \times 5^\circ$ equal area cells, i.e. approximately 550 km × 550 km at the surface. Radially the mantle was subdivided into 29 layers, each 100 km thick.

To counter the corrupting influence of the crustal and uppermost mantle structure which can be poorly resolved in teleseismic body wave tomography, a 3-D *a priori* model of upper mantle velocity heterogeneity was used (Rhodes 1998; Rhodes and Davies, in preparation). The model contains crustal thickness, age of oceanic lithosphere, and subducting slab information, but no *a priori* hot spot information. The traveltime residuals were inverted for slowness perturbations around the reference velocity model using the SIRT iterative row-action sparse matrix inversion technique (Gilbert 1972; Clayton & Comer 1983). The inversion was adapted to simultaneously solve for earthquake relocation and slowness perturbations (Spakman 1988). Since we utilized a 3-D *a-priori* model and undertook earthquake relocation, neither station nor earthquake static corrections (Hager & Clayton 1989) were evaluated. The linear set of equations were augmented to introduce explicit damping and smoothing operators (Pulliam *et al.* 1993; Lees & Crosson 1990), these better constrain the poorly resolved regions. The 2-D ray-tracing scheme of Comer (1984) was adapted to allow individual 1-D velocity models for each ray, derived from the great circle path through the 3-D *a priori* model. The radial reference model was *iasp91* (Kennett & Engdahl 1991). Inversions incorporating the 3-D *a priori* model resulted in an increased traveltime variance reduction over an inversion utilizing only *iasp91* (33 per cent as compared to 18 per cent). Fig. 1 shows layer slices produced using the *a priori* model.

* Now at: Statoil, Stavanger 4035, Norway; mrh@statoil.com
† Corresponding author. Now at: Department of Earth Sciences, Cardiff University, PO Box 914, Cardiff, CF10 3YE.

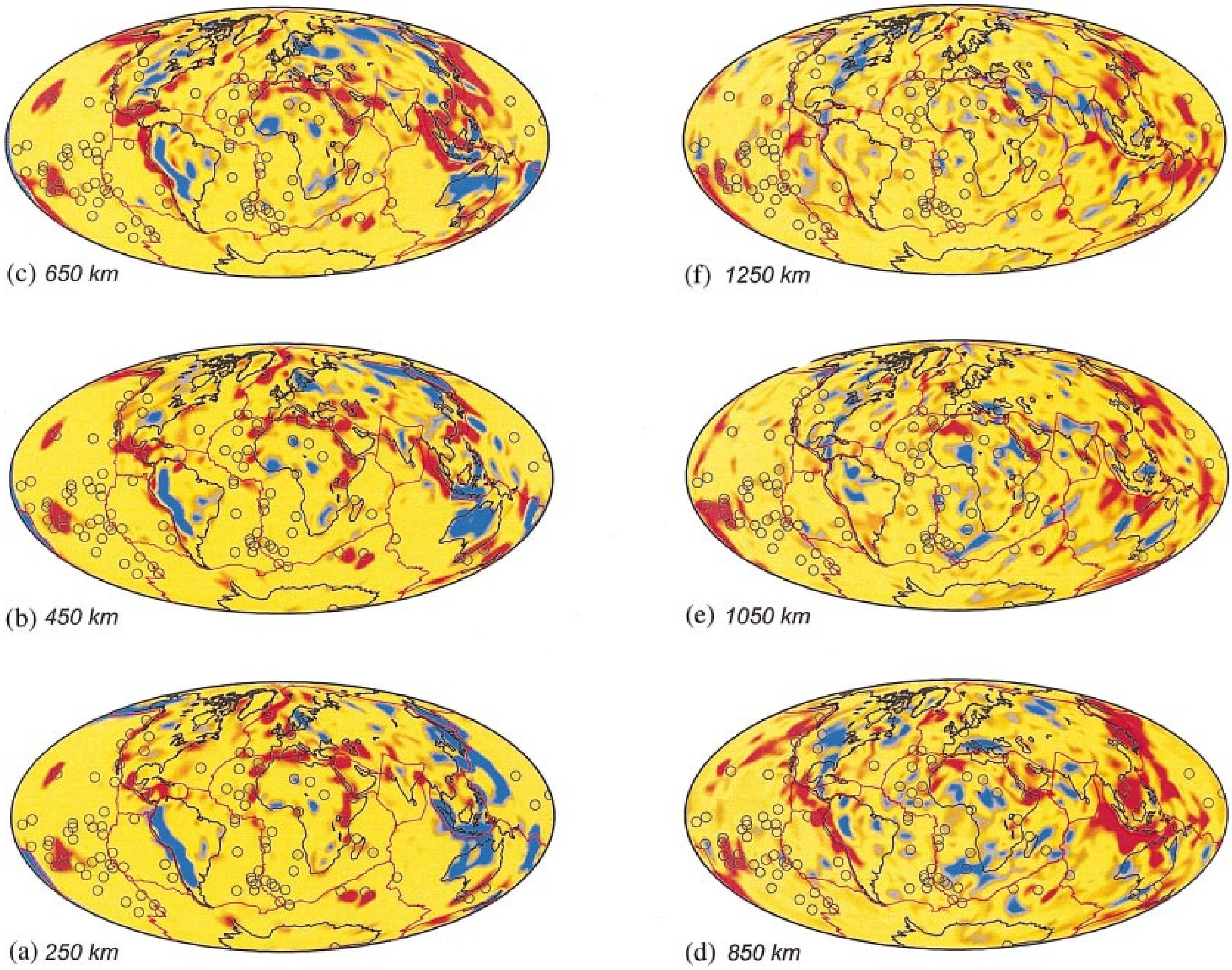


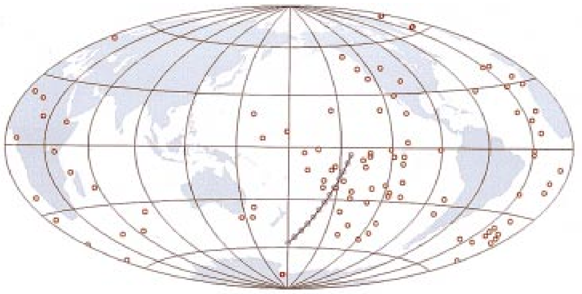
Figure 1. Horizontal layer slices through a *P*-wave seismic velocity tomographic model derived on an equal area grid. The inversion utilized a 3-D *a priori* model of upper mantle velocity heterogeneity. The colour legend for the per centage velocity perturbation is given in Fig. 2. Red regions are slower than average, while blue regions are faster than average. The open black circles represent the 99 hot spots included as part of the 3SMAC model (Nataf & Ricard 1996). Note slow features extending down from the surface to 850 km beneath many significant hotspots, e.g. Afar, Kerguelen, Iceland, Society Islands, Hawaii, etc.

Localized cylindrical slow velocities were imaged vertically beneath postulated hotspots. Such features are observed to extend down to at least 850 km beneath Iceland, Hawaii, Afar, Crozet, Kerguelen, East Africa, Cape Verde and Canary Islands. Geometrically similar but weaker features are seen down to 650 km beneath Comoros, Chubb/Bowie, Tasmania, Guyana and Cameroon. The simplest interpretation is that these are all images of mantle plumes. A vertical section through the Society Islands Superswell (Cazenave *et al.* 1989; McNutt *et al.* 1996) (Fig. 2), shows the presence of slow material extending from the surface through the 660 km discontinuity, eventually fading at some 1200 km depth.

DISCUSSION

Most global high-resolution seismic tomography studies to date have removed signal common to a station (station correction), with the aim of preventing the smearing of very shallow unresolvable structure beneath stations in the interior of the model. In contrast it is possible that these studies have removed signal from the interior of the model and absorbed it into corrections. This might be a part explanation why many of the global studies, in contrast to regional studies, have not imaged as many plumes and subducting slabs. Given the near vertical nature of teleseismic rays in the uppermost mantle in global studies, there is a strong trade-off for the length of shallow vertical structures. This study shows that using an *a priori* model and no explicit station corrections leads to a model that generates plume-like features beneath many hotspots. We note that most of the plume-like structures fade out at depth rather than spreading out at depth, which is what would be expected if vertical smearing was dominant. We note though that Iceland does show broadening at depth, suggesting that its image might be reflecting ray smearing.

To investigate this point further, the resolution of these hotspots was analysed via synthetic recovery tests. Nineteen



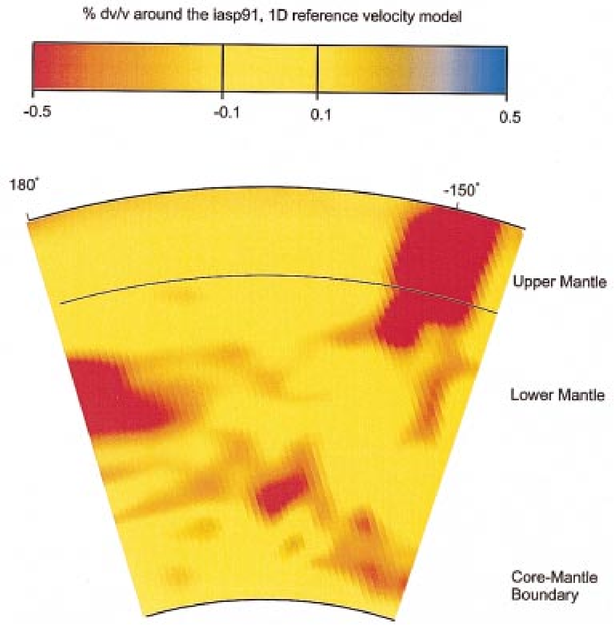


Figure 2. A vertical section through the Society Island hotspot showing the presence of a continuous vertical structure of slow material in both the upper and lower mantle.

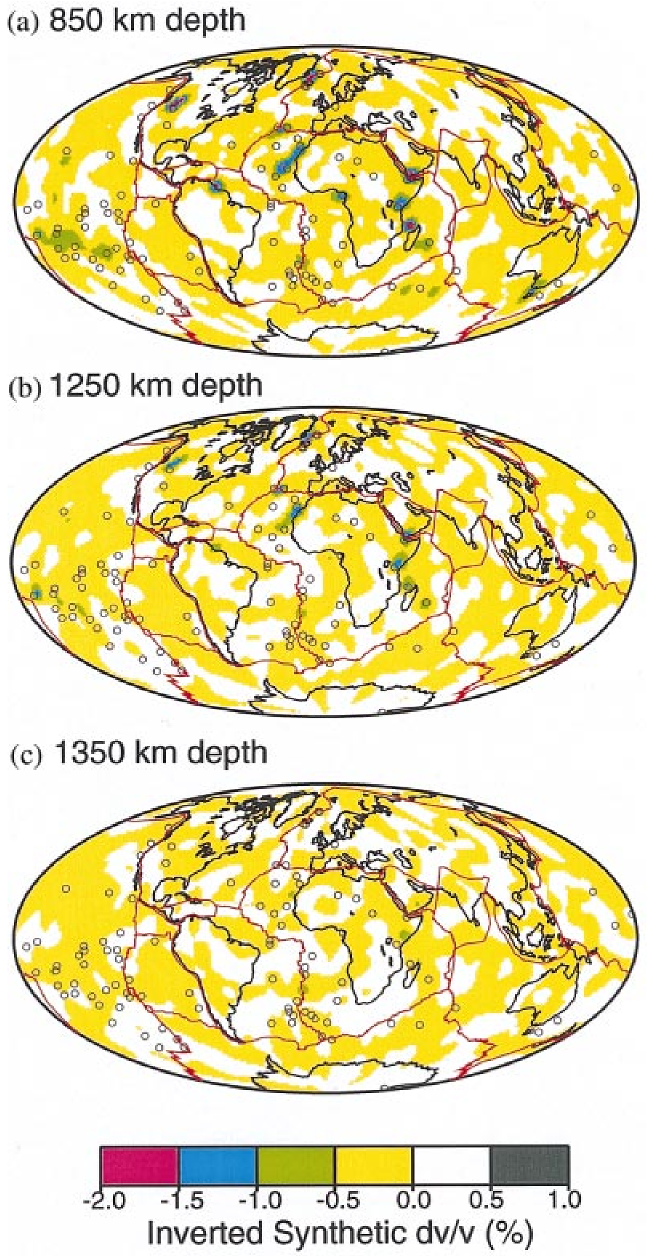


Figure 3. Resolution recovery tests conducted using the same ray dataset used in the production of the tomographic layer slices of Fig. 1. Slow features (−4 per cent velocity perturbation) were placed vertically beneath 19 'hotspots' from the surface down to 1300 km depth. Fig. 3(a) shows the velocity perturbations recovered at 850 km depth. It shows that all features could be recovered to some extent at this depth, though some, e.g. Hawaii and Samoa, are very weak. Fig. 3(b) shows the velocity perturbations recovered at 1250 km depth which formed the deepest layer in the synthetic recovery test. It shows that many of the features could be recovered at this depth, but not all, e.g. Kerguelen and Crozet. Fig. 3(c) shows that there is little vertical smearing beneath many hotspots into the deeper layer at 1350 km depth—but this is not true for all, e.g. East African Rift.

columns with a hot spot at the surface were assigned a −4 per cent velocity perturbation down to 1300 km depth, all other cells had a zero velocity perturbation (Hawaii, Samoa, Society Islands, Gambier, Guyana, Yellowstone, Tristan da Cunha, Cameroon, Cape Verde, Azores, Canary Islands, Iceland, Afar, East African Rift, Comoros, Reunion, Crozet, Kerguelen, Tasmania). Inversion of the residuals predicted by the synthetic plume model allows one to evaluate the recovery of such structures and the degree to which they are smeared. Fig. 3 illustrates that all the selected synthetic plumes could be resolved at 850 km depth to some degree; though some are very weak, e.g. Hawaii and Samoa. Fig. 3(b) shows that a subset can be imaged down to the 1250 km depth, while others are absent such as Kerguelen and Crozet. Fig. 3(c) shows that very few are smeared down, though Eastern Africa Rift and Tristan da Cunha are two of the more prominent exceptions. Therefore many structures extending to such depths would be well resolved. The results demonstrate that we can expect to recover up to 85 per cent of the input velocity perturbation, and no more than 50 per cent in the lower mantle. This poor amplitude reconstruction is common in body-wave tomography and reflects the effect of damping (regularization) arising from the inhomogeneous ray distribution.

The results presented here suggest that some localized slow seismic velocity features can be detected down to at least 1200 km depth using the arrival times of short period *P*-phases; and that they are not masked by wavefront healing phenomena. We do not detect the plumes much below 1300 km. This could either reflect wavefront healing or that plumes do not come from deeper due to the presence of a mid–mantle boundary. We will briefly expand on these alternatives in the following paragraphs.

Wavefront healing would result if the width of the Fresnel volumes at such depths could become greater than the plume diameters. A crude estimate of the radius of the quarter-wavelength Fresnel zone (R), for a ray of length L, wavelength λ, is approximately $[(L/2+\lambda/4)^2-(L/2)^2]^{1/2}=(L\lambda/4)^{1/2}$. For a ray bottoming at around 1300 km depth, L is approximately 4000 km; while for short-period P-waves of period 1 s, the wavelength at this depth is approximately 10 km $=>R$ approximately $(4000*10/4)^{1/2}=100$ km. This is of the order of the radius of plumes, e.g. Shen *et al.* (1998) constrain the Iceland plume to have a diameter of 200 km or less.

Montagner (1994) and Wen & Anderson (1995) have argued for a boundary in the uppermost lower mantle. A boundary deeper in the interior of the mantle has been suggested in the work of Kellogg *et al.* (1999), van der Hilst & Karason (1999), and Davaille (1999). If such a boundary prevents the passage of material then one would expect it to (i) have a large thermal boundary layer across it, leading to high levels of seismic velocity heterogeneity (ii) produce seismic reflections (iii) lead to a triplication in the traveltime curve. None of these have been convincingly seen in this or any other seismic data. We note that Kawakatsu & Niu (1994) and Kaneshima & Helffrich (1999) have observed reflections at around 1000 km and 1200 km depth, respectively, but they seem to be localized and not global.

The fact that we do not image some plumes extending into the lower mantle, which we should be able to resolve with our ray set, could result from a variety of reasons. (i) The Fresnel zone for these plumes could be wider since the sampling rays could be longer ray paths and/or they are detected by longer period seismometers; (ii) the relevant plume is thinner than estimated; (iii) such plumes are shallow features and do not extend into the lower mantle, or (iv) they occur within such a close distance of fast features that they cannot be resolved apart at the 5° resolution of this inversion, this is unlikely. Given the 5° resolution we cannot constrain the minimum radius of the plumes, or therefore the maximum magnitude of the seismic anomaly of the plumes.

There could be many reasons as to why no plumes are detected beneath many hotspots; (i) in some cases it is possible that a plume is not the cause of some hotspots; there is even no agreement on a the definitive list of hotspots; (ii) the ray set has no resolution beneath certain hotspots; (iii) the signal of some hotspots is only weakly captured since there are insufficient seismic stations that suffer the plume delay, because there are none, or too few stations directly above the hotspot, and therefore due to the high noise level of the data set a convincing image is not reconstructed; (iv) the plume has no seismic anomaly, possible because it is a 'wetspot' rather than a 'hotspot' (Neumann 1994); if the 'wetspot' is just hydrated mantle and there is no free water, it is not clear that the seismic velocity of such a 'wetspot' would be very different from ambient.

In addition to the plumes mentioned above, further plume-like features are found (i) off the Antarctic Peninsula down to a depth of 650 km, (ii) near Deccan down to 650 km, (iii) beneath the Caspian Sea down to 650 km, (iv) beneath the Hindu Kush down to 650 km depth, (v) at a bend in the N. Mid-Atlantic ridge, down to 650 or 1050 km depth, (vi) off New England down to 650 km depth, (vii) beneath Novaya Zemlya down to 650 km depth, (viii) beneath the Baffin basin to 650 km depth, (ix) the Tasman Sea (down to 850 km or maybe 1250 km depth), and (x) beneath central Mexico to 850 km. Broader slow features are observed beneath the North Sea down to 250 km depth, beneath most of Mediterranean down to 650 km depth and beneath S. E. China coming up under the South China Sea, from a depth of 1250 km. Evidence is also found of a low-velocity layer beneath S. Africa at mid-mantle depths which seems to link up with the features beneath the East Africa rift in the shallower lower mantle, and upper mantle, as seen for example in the work of Grand *et al.* (1997). The resolution of none of the above have been tested, and we would like to repeat that the nature of a teleseismic ray-set could exaggerate the actual depth extent of some of these features.

Some of the most recent research has shown a correlation between local ultra-low velocities at the base of the core mantle boundary and their surface hot spot location (Williams *et al.* 1998). Beneath Iceland a localized patch of ultra-low seismic wave speed material has been located at the core–mantle boundary through the modelling of seismic waveforms (Helmberger *et al.* 1998). This agrees with the interpretation of receiver function work done over Iceland, that the mantle plume must come through from the lower mantle (Shen *et al.* 1998), and the imaging of a broad plume from the core–mantle boundary to the surface beneath Iceland (Bijwaard & Spakman 1999). There is also work that argues that the plume beneath Iceland is restricted to the upper mantle (Foulger *et al.* 2000). In related research, Osmium isotopic analysis suggests that the Hawaiian plume originates at the core mantle boundary (Brandon *et al.* 1998). These results suggest connectivity between plume material that is derived at the core mantle boundary and eventually appears at Earth's surface. The images presented in this paper could be the expression of this material upwelling through the shallowest layers of Earth and reveal material transfer by plumes across the upper mantle/lower mantle transition. Bijwaard *et al.* (1998) have also imaged many of the same plumes passing through from the lower to the upper mantle. These results would support the suggestion that plumes are the active upwellings throughout the whole mantle.

If the reason that no narrow plumes are imaged below 1300 km is wavefront healing, then it is tempting to take the facts quoted above to suggest the strong possibility that these plumes pass from D″ to the surface. This together with the observation of Farallon and Tethys subducting slabs in the lower mantle, and of other slabs reaching D″ (Grand *et al.* 1997), strongly supports a form of whole mantle convection. We note that the Farallon beneath eastern North America and the Tethys beneath southern Asia are also imaged in this work (Fig. 1f). In contrast the alternative interpretation for the plumes not being imaged beneath around 1300 km, i.e. two families of plumes due to a boundary, clearly requires layered convection.

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