GLOBAL MANTLE TOMOGRAPHY:
RECENT ACHIEVEMENTS AND FUTURE CHALLENGES

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Abstract

We discuss the present status of global mantle tomography and distinguish two classes of models: the P velocity models based on large datasets of travel times from the International Seismological Centre bulletins, often referred to as "high resolution" models, and the S velocity models based on a combination of surface wave and hand picked body wave travel times, or waveforms, referred to as "long wavelength" models. We discuss their respective strengths and weaknesses and present the view that future improvements in resolution require the utilization of the rich information contained in complete broadband seismic waveforms. This is presently within our reach owing to theoretical progress as well as the increase in computational power in recent years.

Introduction

Global seismic tomography of the earth's mantle is a powerful tool for the investigation of the dynamics of the earth's interior. When combined with geodynamical simulations and mineral physics experiments that help in their interpretation, tomographic images allow us to address such outstanding questions as: What is the fate of lithospheric slabs as they descend into the deep mantle? What is the morphology and significance of rising plumes and in what regions of the mantle do they originate? Are there distinct geochemical reservoirs in the mantle? What is the scale and nature of heterogeneity in the boundary layer at the base of the mantle? Geophysicists are still debating these and other fundamental questions (e.g. Albarède and van der Hilst, 1999) while seismologists strive to improve the lateral and depth resolution of their tomographic models.

There has been much progress in seismic tomography since its introduction in the late 1970's (e. g. Aki et al., 1977; Sengupta and Toksöz, 1977; Dziewonski et al., 1977). The first models were very smooth, resolving wavelengths on the order of 3000-4000 km through parametrizations in terms of spherical harmonics up to degree 6-8. They also focused separately on the lower mantle, using P wave travel time data (e.g. Dziewonski et al., 1977; Dziewonski, 1984) or on the upper mantle, based on surface wave dispersion data (e. g. Woodhouse and Dziewonski, 1984; Nataf et al., 1986; Montagner and Tanimoto, 1991). In parallel, many regional tomographic studies of the upper mantle were conducted, focusing either on subduction zones, which correspond to a favorable geometry for P travel time tomography (e.g. Kamiya et al., 1988; van der Hilst and Spakman, 1989) or oceanic areas largely inaccessible to body waves, but well sampled by surface waves (e.g. Montagner, 1986). A review of mantle tomography of the 1980's can be found in Romanowicz (1991).

The last decade has seen a spectacular increase in the number of regions studied by tomographic methods. Also, global models now span the entire depth range of the mantle, and achieve lateral resolutions corresponding to wavelengths of less than 1000 km, and in some places, 500km. In addition to solving for volumetric heterogeneity, tomographers are also starting to address issues such as topography of upper mantle discontinuities (e.g. Gu et al., 2001). It is not the purpose of this paper to review these studies exhaustively, but rather to focus on the achievements and limitations of the present generation of global tomographic models and discuss the challenges for the future. We first discuss the two rather distinct classes of present global tomographic models: the so-called "high resolution" models and "long wavelength" models, their respective strengths and weaknesses, and then describe recent theoretical progress, and how this can help us move forward towards the next generation of waveform based, finer scale, global models.

"High resolution" models

Most P velocity models to date have been based on the extensive dataset of first arrival travel time picks provided over the last 30 years in the bulletins of the International Seismological Centre (ISC). In using this dataset, the large quantity of global data available (hundreds of thousands to millions of travel time picks) compensates for the absence of control of the user on the quality of individual picks. The tomographic mantle models thus derived (e.g. van der Hilst et al., 1997; Obaysashii and Fukao, 1997; Vasco and Johnson, 1998; Bijwaard et al., 1998; Obayashii and Fukao, 1997; Karason and van der Hilst, 2001) are parametrized in local basis functions, typically blocks of 3 to 5^o in latitude and longitude. They have provided us with sharp images of fast velocity slabs under the western Pacific and the Americas, and clarified the interaction of these slabs with the mantle transition zone, and in particular the 670 km discontinuity (Figure 1). It has been shown that some slabs lie flat and do not penetrate far into the lower mantle, while some others appear to continue to great depths (e.g. vander Hilst et al., 1997). Recently, Fukao et al. (2001) have suggested that the barrier to slab penetration into the lower mantle is not at the 670 km discontinuity but rather at the bottom of the transition zone as defined by Bullen (1963), around a depth of 1000 km. Other studies suggest that coherent structure related to slabs reaches at least 1700 km (van der Hilst and Karason, 1999).

Because of the particular distribution of sources (concentrated around subduction zones) and stations (mostly on continents), the first arrival travel time data provide good resolution under subduction zones, but lack coverage under the oceans, both in the upper mantle and in the lower mantle, even when secondary phases available from the ISC bulletins, are included (e.g. Vasco and Johnson, 1998; Figure 2). In order to improve resolution in the oceans, it is necessary to add other types of waves that are not available (or not as reliable) from the ISC bulletins. such as, for the lowermost mantle, PcP (e.g. Obayashi and Fukao, 1997), and Pdiff (Karason and vanderHilst, 2001). Still, many areas of the ocean (both spatially and in depth) remain unsampled or poorly resolved. Therefore these models tell us little about the low velocity regions in the mantle, which are preferentially located under the oceans.

From the theoretical point of view, these tomographic inversions are performed in the framework of ray theory, and the block inversions, which comprise a large number of parameters, employ iterative approximative techniques such as LSQR (e.g. Spakman and Nolet, 1988). It has been shown, however, that these approximations do not result in a strong bias in the tomographic models produced (e.g. Boschi and Dziewonski, 2000).

"Long wavelength" models

The second type of global tomographic models are primarily S velocity models (although some P models have also been derived with similar techniques and geometry, e.g. Masters et al., 2000). These models rely primarily on data collected by individual scientists from global databases of long period and broadband seismic waveforms collected by members of the Federation of Global Seismic Networks (FDSN). The data considered include normal mode spectral measurements (i.e. splitting functions, Giardini et al., 1988; Resovsky and Ritzwoller, 1999), which provide constraints on the longest wavelengths of lateral heterogeneity, and surface wave dispersion or waveform data, which provide sampling in the upper mantle, in particular across the oceans. To attain resolution in the lower mantle, travel times of S, SS, SSS, ScS, ScS_n, and SKS are generally measured using cross correlation techniques: thi includes differential travel times, as well absolute travel times, the latter measured by comparison with synthetic seismograms computed in spherically symmetric earth models, accounting for earth's ellipticity. These inversions use a parametrization of the mantle in global basis functions in the horizontal dimension (spherical harmonics) and a polynomial or cubic spline representation in depth (e.g. Masters et al., 1996; Su et al., 1994; Grand et al., 1997; Liu and Dziewonski, 1998; Ritsema et al., 1999).

The theoretical framework is generally that of zeroth order asymptotic theory for normal modes and surface waves (the "Path Average Approximation", PAVA, e.g. Woodhouse and Dziewonski, 1984), in which it is assumed that sensitivity is confined to the great circle path between the epicenter and the station, and, along this path, varies only with depth (1D kernels). For body waves, ray theory is assumed, an infinite frequency approximation in which the sensitivity of a given phase is confined to the infinitesimal ray path and is uniformely distributed along the ray.

Because of the sampling provided by surface waves and body waves included in the inversions, these models have the advantage of allowing good constraints on the longest wavelengths (i.e. the lowest spherical hamonic degrees) features of mantle heterogeneity at the global scale, and in particular better resolution in the oceans. Gradually refined since the mid-1980's, at which time the maximum degree of spherical harmonics expansion was 6-8, models are now exhibiting details corresponding to spherical harmonics degree 24 or larger (i.e. spatial resolution of under 1000 km at the surface). They have revealed several distinctive and stable large scale features of mantle structure, which provide strong constraints on mantle circulation models. First, the spectrum of lateral heterogeneity varies with depth in the mantle. Dominated by low degrees (up to degree 8) in the uppermost mantle and by degree 2 in the transition zone (e.g. Li and Romanowicz, 1996), the spectrum becomes approximately white at midmantle depths, before changing again to a red character, dominated by degrees 2 and 3, in the lowermost $\sim$ 500 km of the mantle. The rms heterogeneity follows these changes, being strong in the topmost mantle, weak in the midmantle and increasing again at the base of the mantle (Figure 3). S velocity structure is correlated with tectonics in the first 200-300km of the upper mantle, with low velocities associated with mid-ocean ridges and back arcs, and fast velocities in continental shield and platform areas (Figure 4). S velocity gradually increases as a function of age in the oceans. In the transition zone (400-1200km), the patterns are most clearly dominated by fast velocities associated with subduction zones, while between 1200 and 2000 km it is more difficult to recognize any coherent patterns. Towards the bottom of the mantle, the structure becomes again more organized, and reveals two distinctive large low velocity regions, generally referred to as "plumes", under the central Pacific and under Africa. The central Pacific low velocity region is surrounded by a ring of fast velocities, commonly associated with fossil subduction (Figure 4).

All current models show many compatible features at long wavelength, even though details vary significantly from one to the other. Correlation between models is very strong at least up to degree 4 throughout the mantle.

When comparing these long wavelength models to the "high-resolution" P models discussed previously, it is encouraging to note that, in most subduction zone areas, they are very compatible, the former providing a "low-pass filtered" image of the latter, as recently illustrated by Fukao et al. (2001) (Figure 5). While the long wavelength models do not yet attain images in subduction zones that are as "crisp" as those of high resolution P models, they, in turn, provide better resolution in regions of low velocity under the oceans, and in particular the two "super-plume" regions in the lowermost mantle. While there are still many issues to be resolved regarding the nature and role of these plumes in the general mantle circulation, it is clear from a comparison of different models that they represent major features which rise high above the core-mantle boundary (CMB), possibly connecting with low velocity structures in the upper mantle (Figure 6).

The waveform approach: Towards improved global resolution

Most long wavelength tomographic models rely on phase and travel time data, and therefore are restricted to phases (especially for body waves) that are well separated on seismograms. Furthermore, the use of ray theory prevents the utilization of diffracted waves, whereas these phases provide critical sampling of the D" region immediately above the CMB. In order to more fully exploit information contained in the seismic wavefield, even at relatively long periods, full waveform inversion capabilities are required.

Methods have been designed for waveform inversion of surface waves, including higher mode wavepackets, such as the waveform partition inversion method of Nolet (1990). While suitable for surface waves, such methods however still rely on the high frequency PAVA approximation and are thus not appropriate for body waves.

Physical insight on what is needed to compute "broadband" sensitivity kernels for body waves is obtained from the results of normal model perturbation theory. In the zeroth order approximation (PAVA), asymptotic coupling along a single mode branch provides the ability to distinguish heterogeneity located on the minor and major arc (odd degree heterogeneity, eg. Mochizuki, 1986; Romanowicz, 1987), but the corresponding kernels are 1D, i.e. they vary uniformely with depth along the entire source-station path (Figure 7). In order to reproduce the variation of sensitivity along the ray path, across-branch coupling needs to be included (Li and Tanimoto, 1993; Li and Romanowicz, 1995; Marquering and Snieder, 1995).

To zeroth order asymptotically, this brings out some key features of body wave broadband sensitivity kernels, namely that the sensitivity is concentrated along the ray, but is non negligible across some width depending on the frequency content of the waves considered, and also that it varies with position along the ray, being stronger near the source and receiver as well as around the location of surface bounce points and near the bottoming points of the ray (Figure 8).

To zeroth order in the asymptotic approximation, kernels are 2D, that is, only the effects of structure in the vertical plane containing the source and the receiver are taken into account. To obtain 3D kernels, which incorporate the effects of out of plane focusing effects, it is necessary to include higher order terms in the asymptotic expansions (e.g. Romanowicz, 1987), or to numerically compute integrals involving products of spherical harmonics. Such theoretical 3D kernels based on normal mode or ray theoretical approaches have been developed, resulting in "banana doughnut" shaped kernels (e.g Zhao et al., 2000; Hung et al., 2000), but so far have not widely applied in global tomography, mostly because of their heavy computational needs.

In the past 6 years, we have developed and applied a non linear asymptotic normal mode approach (NACT, non-linear asymptotic coupling theory, Li and Romanowicz, 1995), in which we derive global tomographic S velocity models by inverting waveform data exclusively, using 2D asymptotic kernels as described above. While we do not yet account for off-great circle path propagation due to elastic effects, this approach allows us to implement a waveform inversion approach for both surface waves and body waves, and in particular, to include Sdiff and wavepackets containing contributions from several phases that sample the mantle very differently (such as ScS and SSS, Figure 8).

Another important feature of our approach is that we separate the seismogram into energy wavepackets that are weighted according to criteria based on sampling of the mantle as well as amplitude (Figure 8). This allows us, in particular, to emphasize wavepackets, such as Sdiff, (or the Sdiff/SnKS wavegroup in longitudinal component seismograms), which have good signal to noise ratio, but can be dwarfed by later arriving multiple S phases (or phases like PS,SP..) that typically have much larger amplitudes. We have shown that the wavepacket weighting is particularly important to improve resolution near the bottom of the mantle, whereas the more accurate 2D kernels (as compared to PAVA) are critical in the transition zone and mid-mantle (Figure 9; Mégnin and Romanowicz, 1998). Our global 3D SH mantle models thus derived (Li and Romanowicz, 1996; Mégnin and Romanowicz, 2000), and more recently SV models (Panning et al., 2000), open the way towards exploiting more fully the rich information contained in seismic waveforms for the purpose of global tomography. In particular, both elastic and anelastic structure can thus be retrieved (e.g. Gung et al., 2001).

Waveform inversions are non-linear, and at each iteration, involve the computation of synthetic seismograms in the updated 3D model. Whereas using approximate partial derivatives to minimize computational time may be sufficient, accurate 3D seismograms represent the key to progress in waveform tomography of the future. These may be computed using higher order normal mode theoretical approaches (Clévédé and Lognonné, 1996) or, alternatively, numerical methods, such as the recently developed spectral element method (e.g. Komatitsch and Vilotte, 1998; Komatitsch and Tromp, 1999; Capdeville et al., 2001), as our computational power rapidly improves.

Forward modelling: a complementary approach to tomography

Uneven spatial coverage, particularly at large depths in the mantle, combined with errors in the data and imperfections in the theory impose damping conditions in tomographic inversions that prevent accurate recovery of the amplitudes of lateral heterogeneity. Insight into both the strength of heterogeneity as well as the true length scales can be obtained through the use of forward modelling on specific well sampled "corridors" through the mantle.

One such well sampled region, from the point of view of lower mantle structure, is the central Pacific ocean, which lies in a favorable geometry for paths between the highly seismic Fiji-Tonga subduction zone and north America, populated by numerous broadband stations at appropriate epicentral distances for the observation of diffracted waves interacting with the Pacific superplume. Forward modelling of S, Sdiff and ScS travel times (relative to reference phases such as SKS, in order to minimize contributions from upper mantle structure) has shown that existing tomographic models are able to successfully reproduce the observed trends of travel times (as referenced to a spherically symmetric model) with distance, but they fail to provide good estimates of the amplitudes of the variations observed (e.g. Bréger et al., 1998). In fact, the tomographic models underestimate these variations by a factor of two or three. By starting with a global tomographic model and keeping the shape of the anomalies approximately fixed, and simply modifying the amplitudes in specific subregions in and immediately above D", it is possible to obtain a much better fit to the observed travel times (e.g. Bréger and Romanowicz, 1998). The resulting "modified" models exhibit lateral variations in S velocity at the base of the mantle that reach on the order of 4-5% over distances of several hundred kilometers. There is an indication that, while the strongly reduced velocities associated with the superplume continue to large heights above the CMB (possibly 1000 km, Figure 10), this plume is characterized by a broad base which becomes narrower as it rises into the lower mantle (Bréger et al., 2001). Similar results have been found under Africa, in the region of the african superplume, along profiles made possible by the deployment of the temporary Tanzania PASSCAL array (e.g. Ritsema et al., 1998; Ni and Helmberger, 1999).

Other studies have shown evidence for complexity at the border of the two large "superplumes" at the base of the mantle, such as the existence of ultra-low velocity zones (e.g. Garnero and Helmberger, 1996), or anisotropy (Vinnik et al., 1998; Russell et al., 1998).

Discussion and Conclusions

Current P tomographic models provide good insight into the variety of behaviors of subducted slabs as they descend into the lower mantle, even though it is still not clear whether any of them reach all the way to the core mantle boundary. On the other hand, S tomographic models provides constraints on the long wavelength structure as well as on the morphology of low velocity regions associated with rising currents. The current images of the latter are not as sharp as those of the subduction zones, because of the limited lateral resolution and possibly problems related to wavefront healing that hinder our abilities to "see" low velocity regions. However, S tomographic models indicate that these plumes rise high above the core-mantle boundary and into the upper mantle, where they may connect laterally to the shallow low velocity zones associated with ridges (e.g. Mégnin and Romanowicz, 2000). Clarifying their importance in the global circulation will necessitate improved resolution under oceanic areas, as can be achieved through waveform tomography.

Waveform tomography will also help better address the issue of the relation between the variations of P and S velocity in the mantle (e.g. Robertson and Woodhouse, 1996; Masters et al., 2000). However, major efforts to instrument the ocean floor with long term broadband seismic observatories must also continue as vast areas of the globe, especially in the southern oceans remain poorly covered.

In addition, as discussed above, forward modelling of travel times indicates that large lateral variations of structure at the base of the mantle occur on lengths scales of a few hundred kilometers, at least in the vicinity of the two large "superplumes". Such strong variations, as well as those known to occur in the crust and uppermost mantle, have so far been beyond the reach of tomography, in particular due to the limitations of the theoretical framework employed: ray theory does not handle diffraction and frequency dependence, whereas normal mode perturbation theory requires weak and smooth lateral variations of structure. However, we are currently reaching a turning point in seismology, where beyond forward modelling of travel times, it is becoming practical to compute full synthetic seismograms in a global spherical earth, in as complex 3D structures as necessary, and to periods of interest (10 sec or less), using numerical methods (e.g. Komatitch and Tromp, 1999; Capdeville et al., 2001). These can provide valuable insights and better physical understanding on the interaction between heterogeneity and seismic wave propagation as expressed in observed waveforms and should lead to theoretical improvements that will enable us to take into account the key effects at relatively low computational costs.

Clearly, the full level of understanding of the complexity of structure in the earth's mantle and in particular at its base, will only be achieved by a combination of precise forward modelling and tomographic inversion approaches, and sustained efforts in global observations.

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Figure Captions

Figure 1. Examples of depth cross sections in several subduction zone areas showing fast anomalies associated with subducted slabs as revealed by P travel time tomography using ISC data (from Karason and vanderHilst, 2000)

Figure 2. Examples of coverage obtained at different depths in the mantle using compressional wave phases from the ISC, for the time period 1964-1987 (Vasco and Johnson, 1998). The phases included are P, PP, PKPab, PKPbc, PKPdf.

Figure 3. Comparison of rms S velocity variation with depth for four recent global tomographic models of the mantle: SAW24B16 (Mégnin and Romanowicz, 2000); SB4L18 (Masters et al., 1999); S362D1 (Gu et al., 2001); S20RTS (Ritsema et al., 1999).

Figure 4. Comparison of maps of lateral variations in S velocity at 3 sample depths in the upper, mid and lowermost mantle, for the same four recent models as shown in Figure 3.

Figure 5. Maps of the subduction zone areas around Japan at different depths in the transition zone showing lateral variations in velocity obtained in a "high resolution" P velocity model (top, WEPP2, Obayashi et al., 1997) compared to those obtained in a "long wavelength S model (bottom, SAW12D, Li and Romanowicz, 1996).

Figure 6. Comparison of cross-sections across the african "superplume" for the four S velocity models discussed in Figures 4 and 5.

Figure 7. Comparison of sensitivity kernels for SS (left) and Sdiff (right), using the surface wave PAVA approximation (Top) and the NACT approximation (bottom) (e.g. Li and Romanowicz, 1995).

Figure 8. Illustration of wavepacket selection procedure. (a) Low-pass filtered ($f < 1/32Hz$) transverse seismogram of 01/19/1988 Chile earthquake recorded at GEOSCOPE station CAN. (b) Low-pass filtered ($f < 1/32Hz$) longitudinal seismogram of 12/03/1983 Chagos Archipelago earthquake recorded at GDSN station KONO. In both cases, top trace is the observed seismogram and bottom trace is the synthetic seismogram calculated for the spherical reference model PREM (Dziewonski and Anderson, 1981). Only the waveforms within the indicated windows are used in the inversions. Although the identification of seismic phases is not required in the inversion, major seismic phases are labelled for reference. Note the energy wavepackets which contain more than one phase with different sampling in the mantle.

Figure 9. Correlation as a function of depth of a model obtained using NACT theory and a wavepacket weighting scheme with several other models obtained: using NACT but no wavepacket weighting (solid line, 1), versus using PAVA theory and wavepacket weighting (dashed line, 2) or no wavepacket weighting (dotted line, 3). From (Mégnin and Romanowicz, 1998).

Figure 100. Top: map view of S-velocity heterogeneity in the Pacific at a depth of 2800 km for a modified mantle model that explains S-SKS and S-ScS travel time residuals along

paths from Fiji-Tonga to North America. A sample path is indicated in green. Middle and bottom: Depth cross-section along the path shown in top plot for the original tomographic model (SKS12, Liu et al., 1994) and the modified model (bottom). Sample ray paths of S,ScS, Sdiff and SKS are shown in green (from Bréger et al., 2001).