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Attenuation Tomography of the Earth's Mantle: A Review of Current Status

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Abstract—Resolving the lateral variations of attenuation in the deep mantle by tomographic methods holds potential for constraining its thermal structure and dynamics. It is a challenging subject which has been addressed by only a few studies until now. We here review the main motivations behind pursuing this challenge, the difficult issues involved in separating effects of anelastic attenuation from scattering and focusing due to propagation in 3-D elastic structure and finally discuss the current status of global attenuation tomography.

Key words: Tomography, attenuation, earth's mantle.

Introduction

Global anelastic tomography is a difficult subject that can bring important constraints on the thermal structure of the mantle and therefore its dynamics, in complement to those provided by elastic tomography. Global elastic tomography has made great strides since the pioneering first studies of the last two decades (Dziewonski *et al.*, 1977; Woodhouse and Dziewonski, 1984; Nataf *et al.*, 1986). It is currently possible to resolve whole mantle structure at degree 12 with reasonable agreement between different studies (Su *et al.*, 1994; Li and Romanowicz, 1996; Masters *et al.*, 1996) and upper mantle structure with even greater detail, with lateral resolution reaching on the order of 500–1000 km (Montagner and Tanimoto, 1991; Ekström *et al.*, 1997; Laske and Masters, 1996; Trampert and Woodhouse, 1996). Such a resolution is also progressively attainable in the lower mantle, but only in the vicinity of subduction zones (Vasco *et al.*, 1995; Grand *et al.*, 1997; van der Hilst *et al.*, 1997).

In contrast, anelastic tomography has been lagging somewhat behind. One-dimensional profiles of the variations with depth of the quality factor Q have now reached a high level of consensus (e.g., Romanowicz, 1994a; Durek and

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EKSTRÖM, 1996; BHATTACHARYYA *et al.*, 1996) with generally only minor departures from the reference model PREM (DZIEWONSKI and ANDERSON, 1981). Degree 2 structure seems to be reasonably well constrained in the upper mantle from both normal mode and surface wave data (ROMANOWICZ, 1990; SUDA *et al.*, 1991; DUREK *et al.*, 1993), however very few studies have been completed that explore higher degrees. In fact only two groups to date have looked at both even and odd heterogeneity in the upper mantle at shorter wavelengths (ROMANOWICZ, 1994a, 1995, 1997; BHATTACHARYYA *et al.*, 1996), and only one published study addresses whole mantle 3-D global attenuation, using body-wave data (BHATTACHARYYA, 1996). In the upper mantle, agreement between body wave and surface wave results is variable. Emerging on-going studies may help resolve these disagreements (e.g., REID and WOODHOUSE, 1997).

In what follows, we briefly discuss the significance of global anelastic tomography, the technical reasons for which it is lagging behind elastic tomography and finally review its current status.

Anelastic Tomography: Goals and Issues

The existence of large lateral variations of Q in the crust (see MITCHELL, 1995 for a review) and the upper mantle, is well documented from various regional studies, either using surface waves (e.g., MITCHELL, 1975; CANAS and MITCHELL, 1978, 1981; NAKANISHI, 1979a; BUSSY *et al.*, 1993) or body waves, such as multiple *ScS* phases (NAKANISHI, 1979b; SIPKIN and JORDAN, 1980; LAY and WALLACE, 1983; CHAN and DER, 1988; SIPKIN and REVENAUGH, 1994) or multiple S and depth phases (SHEEHAN and SOLOMON, 1992; FLANAGAN and WIENS, 1990, 1994; DING and GRAND, 1993). These lateral variations can be an order of magnitude larger than observed lateral variations in velocity, commonly exceeding 50–100%.

There are two main reasons that make resolving 3-D anelastic structure in the deep mantle a worthwhile goal. First, the quality factor Q of the earth is considerably more sensitive to temperature than elastic velocity, as shown by laboratory and theoretical studies (e.g., MINSTER and ANDERSON, 1981; BERKHEMER *et al.*, 1982; GUEGUEN *et al.*, 1989; KARATO and SPETZLER, 1990; JACKSON *et al.*, 1992) and this sensitivity differs from that of elastic velocity. As argued by ROMANOWICZ (1994b), the nonlinear (Arrhenius law) sensitivity of Q to temperature implies that, in principle, attenuation tomography should be able to resolve hot regions (high attenuation) better than elastic tomography. In addition, elastic velocity can be affected to a large degree by compositional variations (e.g., YAN *et al.*, 1989; BINA and SILVER, 1997), so, ultimately, mapping regions with various degrees of agreement between velocity and attenuation distributions should help us constrain the distribution of chemical versus thermal heterogeneity in the mantle. At shallow depths, as discussed by MITCHELL (1995), and perhaps also near the core-mantle boundary, the effect of fluid inclusions and partial melt on attenuation also requires consideration.

The other main reason pertains to the dispersion effect of Q on elastic velocities (FUTTERMAN, 1962). For example, LIU *et al.* (1976) and KANAMORI and ANDERSON (1977) demonstrated that the baseline shift between velocity models of the mantle obtained from low frequency free oscillations and surface waves on the one hand, and short-period body waves on the other, could be accounted for by introducing the dispersive effect of Q, assuming an absorption band model with high frequency cut-off in the vicinity of 1 sec, as implied by body-wave observations (e.g., SIPKIN and JORDAN, 1979). When lateral variations of Q are present, the dispersion corrections also vary laterally. This has implications not only for the comparison of elastic models obtained in different seismic frequency bands, but also for geodynamic studies that jointly utilize seismic tomographic models and geoid data to infer the viscosity structure of the mantle (e.g., HAGER *et al.*, 1985; FORTE *et al.*, 1996). Indeed, the conversion factor between velocity and density is a crucial parameter in these studies, and, in the presence of 3-D Q structure, it should vary both with position and with the dominant frequency of the elastic models considered. KARATO (1993) illustrated the importance of anelasticity correction for the radial profile of dv/dr (the density derivative of velocity) and ROMANOWICZ (1990) showed that the misalignment in phase between the degree two pattern of fundamental mode-free oscillation frequency shifts, and that of the geoid, could be explained by the effects of lateral variations in anelastic dispersion.

Seismic Measurements of Attenuation in the Mantle

To measure attenuation in the deep mantle, one can use either low frequency surface wave and free-oscillation data, or deep turning body-wave data. Low frequency techniques typically employ either a travelling wave or a standing wave, normal mode formalism. The travelling wave formalism is well adapted to fundamental mode surface waves to periods of about 250 s, that are well isolated on the seismograms and sample the first 400 km of the upper mantle. Typically, in this approach, the amplitude spectrum $A_i(\omega)$ of each wave packet i is computed at frequency ω after appropriate windowing and tapering and an attenuation coefficient η is defined such that:

$$A_i(\omega) = A_o(\omega)\exp(-\eta_i(\omega)X_i) \tag{1}$$

where X_i is the epicentral distance in (km), and $A_o(\omega)$ represents the amplitude at the source.

The standing wave approach allows us to sample deeper into the mantle and to measure Q not only along the fundamental mode branch. In this approach, two types of methods are commonly used. One relies on the fitting of resonance peaks to spectra of individually observed free oscillations. The fitting is done either on the complex spectra (e.g., MASTERS and GILBERT, 1983) or on the amplitude spectra (e.g., DUREK and EKSTRÖM, 1997) of long time windows (at least 12 hours or more, depending on the Q of the mode). DUREK and EKSTRÖM (1997) have shown that the differences between the two approaches are not significant. The other type of method relies on measuring the amplitude decay with time of individual normal modes, by computing the spectrum successively shifted in time and measuring the slope of the resulting amplitude/time curve (e.g., ROULT, 1975; SAILOR and DZIEWONSKI, 1978). When applied carefully, both modal approaches yield similar results. However, significant discrepancies (on the order of 15%) exist between fundamental mode measurements of Q using the travelling wave and the standing wave approach, in their common frequency domain of application, that are still currently the subject of controversy (e.g., DUREK and EKSTRÖM, 1997).

The measurement of attenuation of body wave is generally a differential measurement and involves two "related" phases (such as S and SS or successive multiple ScS phases) to help minimize effects of the source and near-source and near-receiver structure. These measurements can also be done either in the frequency domain, by looking at the spectral ratios of two related phases, and measuring the slope of the spectrum as a function of frequency, or in the time domain, by computing a transfer function from the wave form of the first phase to the second one (e.g., BHATTACHARYYA, 1996).

While it is easily recognized that accurately resolving 3-D anelastic structure of the mantle could be very useful to further our understanding of mantle dynamics, progress has to date been slow because of the inherent difficulty of measuring attenuation and especially its lateral variations. Indeed, the amplitude of seismic waves travelling through the earth is affected not only by anelastic attenuation but also by focusing and scattering effects due to propagation in a 3-D elastic medium. The latter can be as large or larger than anelastic effects and depend strongly on the short wavelength details of the elastic structure, which are at present not very well constrained. Indeed, as shown by WOODHOUSE and WONG (1986) in the framework of ray theory and by ROMANOWICZ (1987) and PARK (1987) in the framework of asymptotic normal mode theory, to first order, the focusing terms due to elastic structure depend on the transverse gradients of velocity along the propagation path.

Asymptotically, if A_l is the amplitude of a normal mode of angular order l, then the perturbation due to focusing takes the form:

$$\delta A_l = (1 + \delta F_l) \tag{2}$$

where δF_l, which represents the focusing/scattering term, has the form (ROMANOWICZ, 1987):

$$\delta F_l = \frac{-a\Delta}{2Uk}\tilde{D} \tag{3}$$

here a is the earth's radius, Δ epicentral distance, U group velocity and $k = l + 0.5$. $\tilde{D}$ is the minor arc average of the transverse derivative term D, which, in turn, can be expressed in terms of local coordinates (θ, ϕ) on the surface of the sphere as:

$$D = \frac{\sin(\phi - \Delta)}{\sin(\Delta)} [\partial_\theta^2 \delta\omega_k^0(\theta, \phi) \sin \phi - \partial_\phi \delta\omega_k^0(\theta, \phi) \cos \phi]. \tag{4}$$

In equation (4) $\delta\omega_k^0$ is the "local frequency" (JORDAN, 1978), which, to zeroth order, represents the integrated effect of structure beneath the local point (θ, ϕ) on the surface of the earth.

For a velocity model described by an expansion in spherical harmonics with coefficients (Cst, Sst), transverse gradients depend on terms of the form s^2Cst, s^2Sst, and are therefore sensitive to large values of s, in other words small wavelengths.

If the elastic structure of the earth were perfectly known, one could first correct for its effects using either linear theory (Born approximation), as described above, or, preferably, a more complete formalism including multiple scattering effects (e.g., LOGNONNÉ and ROMANOWICZ, 1990; FRIEDERICH, 1997; GELLER and HARA, 1993). We expect that in the near future, global 3-D models will become reliable enough at short wavelengths so that, combined with the increase in computer power, such corrections will become feasible and accurate.

Until now however, indirect methods of dealing with focusing and scattering have generally been used. ROMANOWICZ (1990) and DUREK *et al.* (1993) exploited the fact that, for surface waves and in the case of linear theory, focusing and anelastic effects could be separated by combining measurements over several consecutive wavetrains for a single recording, because focusing terms change sign with the direction of propagation whereas attenuation terms are always additive. This is for example how elastic focusing can be visually detected in actual recordings, when successive wavetrains present alternating high and low amplitudes (e.g., LAY and KANAMORI, 1985). Figure 1 shows an example of vertical component recordings for the M 7.5 Chile earthquake of 03/03/1985, observed at Geoscope stations and compared with predictions calculated for the PREM model (DZIEWONSKI and ANDERSON, 1981). Several records (stations SSB, KIP, WFM) exhibit anomalously high amplitudes for later arriving trains. In order to remove focusing effects, at least 4 consecutive surface wavetrains are needed. As pointed out by ROMANOWICZ (1994a), the drawback of this approach is that, the longer the travel path, the more the waves are affected by 3-D elastic structure, and the harder it is to account for that in an approximate, linear fashion. Moreover, such a technique is only applicable to surface waves. An alternative approach, favored by ROMANOWICZ (1995) for surface waves and BHATTACHARYYA *et al.* (1996) for body waves, is to reject data that are strongly affected by focusing, after visual inspection. In the case of body waves, BHATTACHARYYA *et al.* (1996) used a technique in which the attenuation operator t^* is inferred from the slope of the variation with frequency of *SS* to *S* amplitude ratios. Data for which a smooth variation of this ratio with frequency cannot be obtained are rejected. In the case of surface waves, ROMANOWICZ (1994a) devised a technique that allows to 1) keep only data for first arriving trains that have travelled the shortest paths (e.g., R1 and R2 for Rayleigh waves), and 2) among those data, reject those that do not present a smooth variation of attenuation coefficient with period. This approach appears to be successful, provided very strict rejection criteria are applied, which limits the coverage of the earth that can be achieved and therefore the spatial resolution of three-dimensional structure.

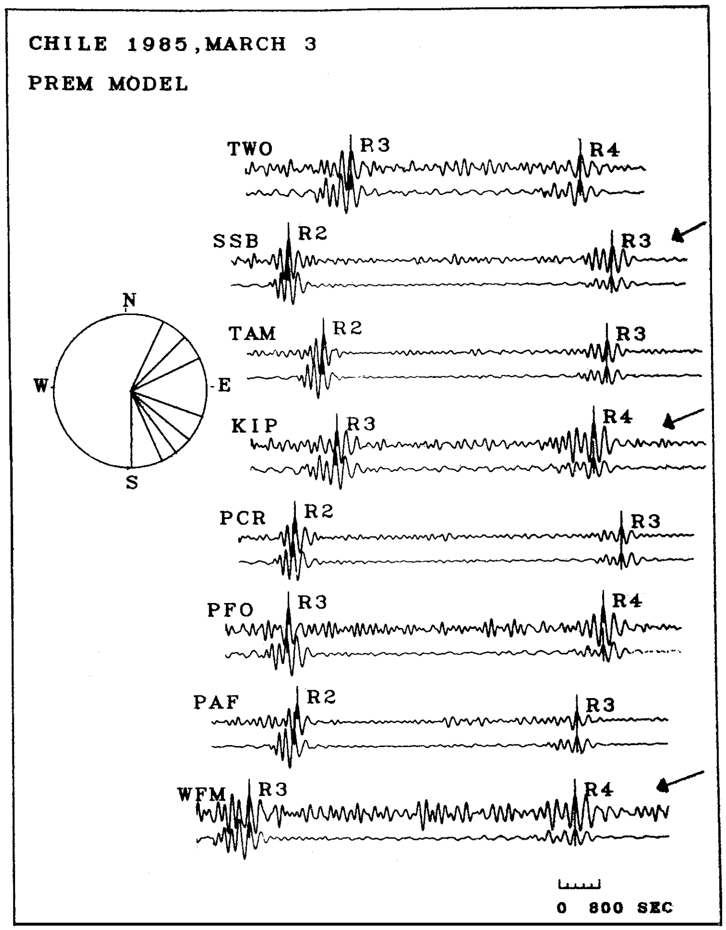

Figure 1
Example of vertical component records for the Chile earthquake of 03/03/85 at Geoscope stations (top traces) compared to PREM synthetics (bottom traces). Stations indicated by an arrow exhibit anomalously large amplitudes of R3 or R4 trains compared to preceding trains, indicating the presence of focusing effects.

An additional concern is that potentially major uncertainties in the amplitude at the source (scalar moment) must be allowed for. In the case of body waves, this is dealt with by considering spectral ratios between phases for which source take-off angles are not very different. In the case of surface waves, the method designed by ROMANOWICZ (1994a) involves computing the scalar moment bias by comparing attenuation coefficients measured on first arriving trains with those measured using three consecutive trains. In the latter case, the source effect is cancelled out, however the attenuation measurement is generally less accurate due to the increased influence of elastic structure over the longer R3 (or G3) path. To determine the source correction factor, an interactive graphic procedure was designed, which involves the superposition of the attenuation curves obtained both ways.

The comparison of maps of lateral variations of attenuation coefficient of Rayleigh waves at different periods obtained using the former approach, in which four consecutive wavetrains are required (ROMANOWICZ, 1990) and the latter approach, in which only first arriving trains (R1 and R2 for Rayleigh waves) are used (ROMANOWICZ, 1994a), indicates good agreement at long wavelengths.

The published models to date generally rely on amplitude measurements in the frequency domain. This is suitable for isolated phases, such as fundamental mode surface waves or specific body-wave phases such as *S*, *SS* and *ScS*. More recently, we have started to explore the possibility of using time-domain wave-form information to invert for anelastic structure (ROMANOWICZ *et al.*, 1996; ROMANOWICZ, 1997), which should allow us to extend attenuation measurements to a larger portion of the seismogram, and therefore increase sampling of the deep mantle. In the case of surface waves, a time-domain method based on the comparison of observed and synthetic wave forms would also help resolve the problem of contamination of fundamental modes by higher mode energy, inherent in frequency domain methods, and, more generally, the problem of length of time window considered to compute the spectrum, and the dependence of amplitudes on chosen tapers. Such an approach requires the ability to accurately account for 3-D elastic structure. We will discuss it briefly in a later section.

Existing Global Models and Stable Features

As mentioned, only a few models of attenuation have been published, because of the difficulties involved in obtaining reliable measurements of attenuation. The status of attenuation tomography today is comparable to that of elastic tomography in the early 1980s, with stable long wavelength features emerging, although still little quantitative information at scales beyond degrees 5–6 of spherical harmonic expansion of lateral heterogeneity. Moreover, most studies as yet assume that lateral variations in Q are confined to the upper mantle.

Even though the degree 2 pattern in Q in the upper mantle is less prominent than that of elastic velocities in studies of free oscillations and surface waves (SMITH and MASTERS, 1989), it appears to be reasonably well constrained (ROMANOWICZ *et al.*, 1987; ROULT *et al.*, 1990; ROMANOWICZ, 1990; SUDA *et al.*, 1991; DUREK *et al.*, 1993), with significant correlation between the results of surface-wave and body-wave studies (BHATTACHARYYA *et al.*, 1996). This structure is also well correlated with degree 2 in elastic velocities, and in particular manifests a similar shift in phase towards the west as one goes from the uppermost mantle (first 300 km) into the transition zone (400–600 km), which is indicative of the predominantly thermal nature of heterogeneity at these long wavelengths, (e.g., MONTAGNER and ROMANOWICZ, 1992). Even at such long wavelengths, there is however some disagreement concerning the depth distribution of attenuation. DUREK *et al.* (1993) argued that the lateral variations in Q are primarily confined to the depth range corresponding to the seismic low velocity zone (80–220 km), whereas more recent studies indicate significant heterogeneity persisting into the transition zone (ROMANOWICZ, 1995; BHATTACHARYYA *et al.*, 1996).

At shorter wavelenghts there are currently only two published global models that consider both even and odd terms of lateral heterogeneity in Q: one using surface waves (ROMANOWICZ, 1995) and the other body waves (BHATTACHARYYA *et al.*, 1996). Comparison of these models evidences agreement in some regions of the world, with low attenuation in Eurasia, western Australia and the Himalayas (BHATTACHARYYA *et al.*, 1996; Fig. 2) and high attenuation in the mid-Pacific, eastern Australia and China. In general, models obtained to date at the global scale concur with regional scale models on the existence of a correlation of lateral variations in Q with tectonic province, in the first 250 km of the upper mantle, with high Q under shields and low Q under oceans, particularly so under young oceans. Surface wave derived models (ROMANOWICZ, 1990, 1995; DUREK *et al.*, 1993) tend to indicate high attenuation under young oceans (as confirmed by regional studies, e.g., DING and GRAND, 1993). ROMANOWICZ (1994b) demonstrated the correlation of Q with the age of the sea floor in the first 250 km of the upper mantle, both in the Pacific and in the Atlantic Oceans (Fig. 3). In this depth range, the correlation of Q structure with heat flow is also observed. At greater depths in the South Pacific the shift to the west of the high Q maximum observed at degree 2 also persists at shorter wavelengths, perturbing the correlation with age of the sea floor. In this depth range, the correlation of Q structure with hotspot distribution appears to be significant (ROMANOWICZ, 1994b, 1995). The disagreement between the predictions of model QR19 (ROMANOWICZ, 1995) and the cap-averaged t^* measurements of BHATTACHARYYA *et al.* (1996) under the young ocean in the South Pacific may come in part from the different way in which *SS* waves average structure over depth beneath their bounce point as compared to surface waves, which have inherently greater depth resolution.

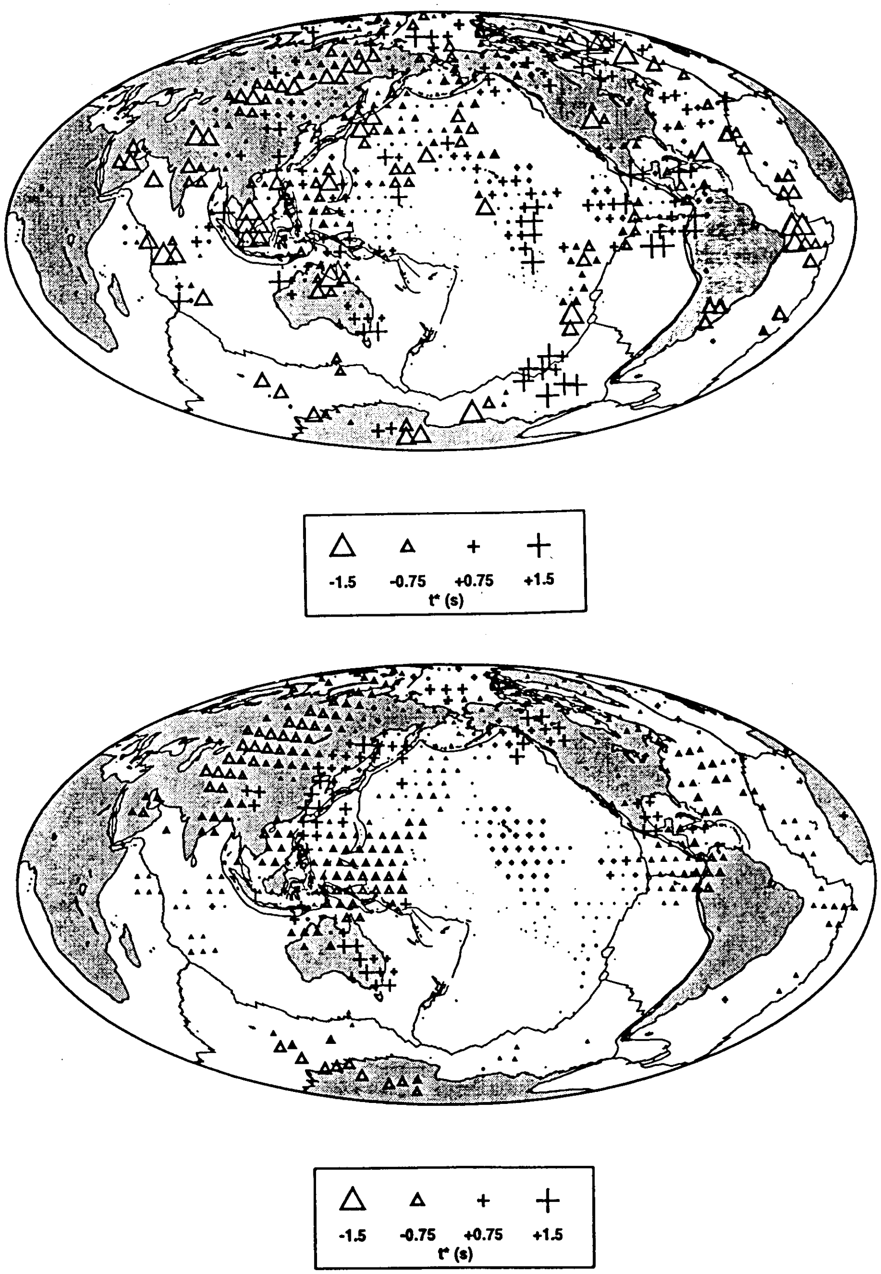

Figure 2

Top: Cap averaged residual t^* values obtained from SS/S spectral amplitude ratios by BHATTACHARYYA *et al.* (1996). Bottom: For comparison, average t^* values predicted by model QR19 (ROMANOWICZ, 1995) plotted at 5° caps. In both cases, average values have been subtracted before plotting. From BHATTACHARYYA *et al.* (1996).

In an on-going study, we are investigating the retrieval of global 3-D mantle Q structure using surface and body wave-form data (ROMANOWICZ *et al.*, 1996; ROMANOWICZ, 1997). This approach follows the general framework of global wave-form inversion for elastic structure developed by LI and ROMANOWICZ (1995, 1997) and proceeds in an iterative manner: in the first step, a spherically symmetric attenuation model is assumed, and wave-form data are inverted for 3-D elastic structure. In the second step, the derived 3-D elastic structure is used as the starting

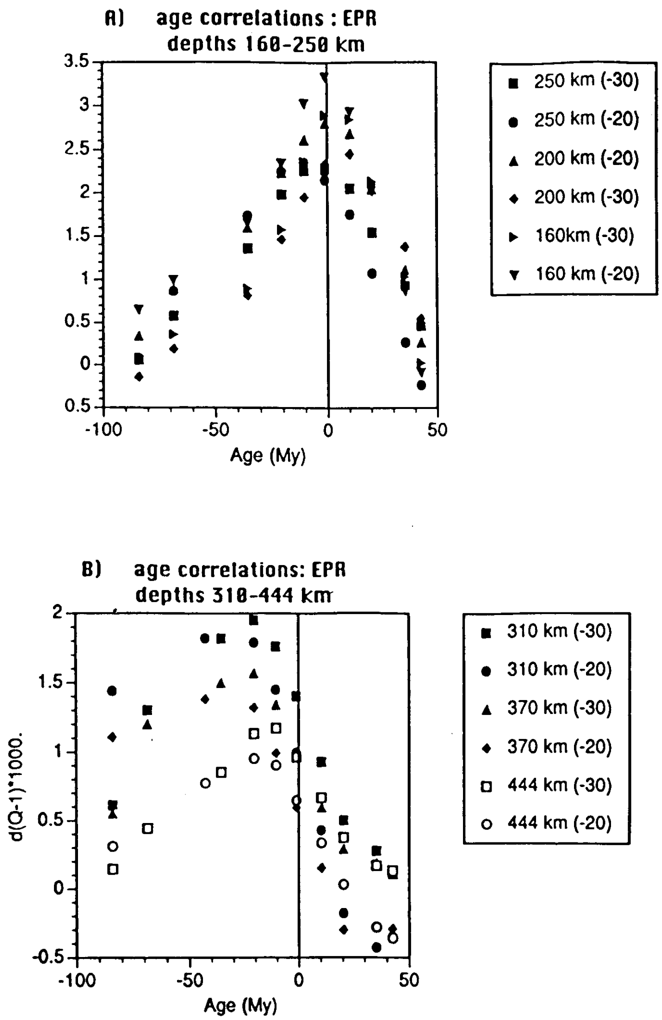

Figure 3

Dependence on age of the sea floor of the southern Pacific Ocean of the anomaly in Q in the upper mantle in different depth ranges, according to model QR19 (ROMANOWICZ, 1995). Top: 0–25 km; Bottom: 250–450 km. The symmetry around the ridge is broken in the deeper range, where lowest Q is found in the central Pacific. From ROMANOWICZ (1994b).

model for an inversion for 3-D anelastic structure. In this second step, focusing terms due to the elastic structure can be incorporated in the forward computation of seismograms. Preliminary results indicate that this approach is promising and results in models that are at least qualitatively compatible with those obtained earlier using a different dataset and a spectral approach. (Fig. 4).

Whether any significant lateral variations in Q exist in the lower mantle is currently an open question. Regional studies based on multiple *ScS* data cannot discriminate between lateral variations spread over the entire mantle or concentrated in the upper mantle. It may be that lateral variations in temperature in the lower mantle are mostly confined to narrow upwellings or plumes and their detection must await a significant increase in our ability to resolve small-scale lateral variations in anelastic structure. Conversely, it is expected that stronger lateral variations in Q exist in D'', because of the boundary layer nature of this portion of the mantle (e.g., LOPER and LAY, 1995). Our preliminary whole mantle wave-form inversion results for *SH* waves (ROMANOWICZ, 1997) indicate a stable pattern of degree two in the lowermost mantle, correlated with that which is well constrained in elastic tomographic models (Fig. 5), with, as in the upper mantle, high attenuation corresponding to low velocity. This would confirm the thermal origin of the velocity lows observed in the central Pacific and under Africa, that have often been interpreted as associated with rising thermal plumes (e.g, STACEY and LOPER, 1983). The existence of melt inclusions (e.g., WILLIAMS and GARNERO, 1996) could also contribute to correlated patterns of velocity and attenuation.

The preliminary pattern obtained up to degree 4 is highly correlated with a global map of P velocities in D'' obtained by WYSESSION (1996) from the study of travel times of P-diffracted waves. High attenuation in D'' in the Pacific Ocean and beneath Africa has also been suggested by BHATTACHARYYA (1996) in a whole mantle inversion in which the bulk of the lower mantle is assumed to have laterally homogeneous Q.

Conclusions

The study of lateral variations of Q in the deep mantle is still in its infancy. Quantitative models are few and their reliability is difficult to assess, since many discrepancies exist between different studies. Some qualitative features of the models can nevertheless be considered as well established. There is general agreement that lateral variations of Q are strongest in the crust and the uppermost mantle, where they are correlated with tectonic features and elastic velocities, and that some heterogeneity persists into the transition zone, at least at the longest wavelengths, indicating a strong thermal component to the low degree elastic structure in this depth range. In this depth range, the long wavelength distribution of Q manifests a correlation with that of hot spots. Lateral variations of Q in the

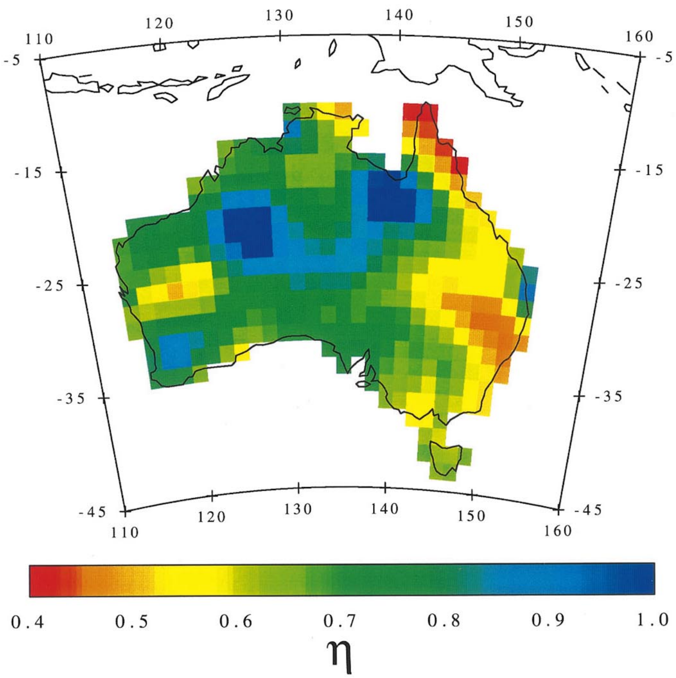
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Fig. 4

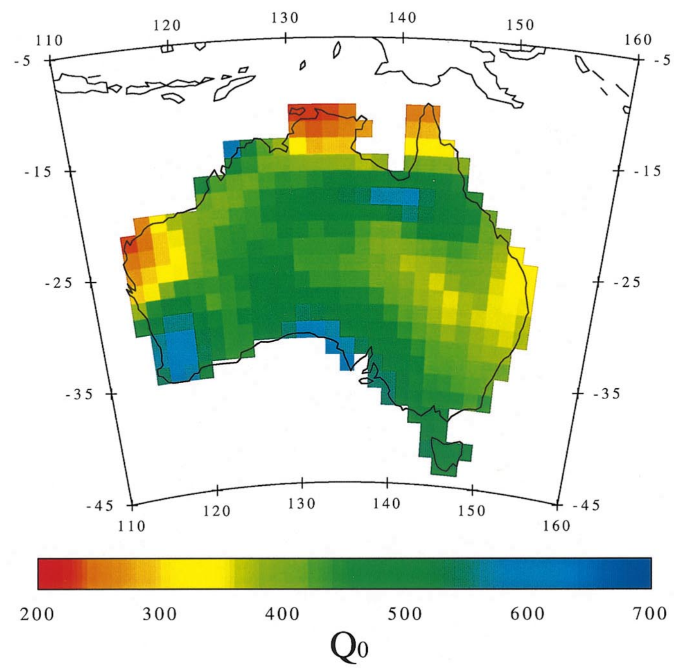
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Fig. 3

lower mantle are much less well constrained, although there seem to be indications of correlation of anelastic and elastic structures in the lowermost mantle, at least at very long wavelengths.

Our ability to further constrain Q models quantitatively relies strongly on how successful we will be, in the near future, in modeling the effects of elastic structure on seismic wave amplitudes (both for low frequency surface waves and shorter period body waves). This is contingent on the construction of global elastic 3-D models with well constrained small-scale features, likely beyond degree 30, which are beginning to be available in subduction zone regions (VAN DER HILST *et al.*, 1997; GRAND *et al.*, 1997) but also on the incorporation of more exact theoretical formalism in the inversion (e.g., ROMANOWICZ, 1987; LOGNONNÉ and CLEVEDE, 1996).

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Figure 4

Comparison of maps of lateral variations in Q in the upper mantle obtained for (top) a preliminary model based on wave-form data inversion (ROMANOWICZ, 1997) and (bottom) model QR19 (ROMANOWICZ, 1995).

Figure 5

Comparison of degree 2 maps in D'' for S velocity (top) as in model SAW12D (LI and ROMANOWICZ, 1996) and for Q (bottom), from a preliminary inversion of wave-form data (ROMANOWICZ, 1997).

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