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ANISOTROPY AND SHEAR-VELOCITY HETEROGENEITIES IN THE UPPER MANTLE

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Abstract. Long-period surface waves are used to map lateral heterogeneities of velocity and anisotropy in the upper mantle. The dispersion curves are expanded in spherical harmonics up to degree 6 and inverted to find the depth structure. The data are corrected for the effect of surface layers and both Love and Rayleigh waves are used. Shear wave velocity and shear polarization anisotropy can be resolved down to a depth of about 450 km. The shear wave velocity distribution to 200 km depth correlates with surface tectonics, except in a few anomalous regions. Below that depth the correlation vanishes. Cold subducted material shows up weakly at 350 km as fast S-wave anomalies. In the transition region a large scale pattern appears with fast mantle in the South-Atlantic. S-anisotropy at 200 km can resolve uprising or downwelling currents under some ridges and subduction zones. The Pacific shows a NW-SE fabric.

Introduction

The present network of long-period digital seismographs has made possible the study of surface waves on a world-wide basis. We thus have an exciting new tool for the investigation of lateral heterogeneities within the earth. With the present station coverage, however, only a crude image can be retrieved. Our previous paper (Nakanishi & Anderson, 1983) presented spherical harmonic expansions of the phase and group velocities of Love and Rayleigh waves up to degree 6. The geodynamic implications of these data are important. We can develop a picture of density and shear-wave velocity heterogeneities in the upper mantle, with no need for a priori regionalization based on surface tectonics. Furthermore, since both Love and Rayleigh waves are used, we can resolve lateral variations in anisotropy. Both density and anisotropy are important geodynamic parameters. Density can be related to the geoid and to the dynamics of the mantle. Anisotropy contains information about the flow pattern, the kinetics of the mantle. We retrieve these parameters as a function of depth by inverting the dispersion data. This paper presents a spherical harmonic representation of shear-wave velocity and anisotropy in the upper mantle.

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Data and Inversion Method

The data we use is from Nakanishi & Anderson (1983). It consists of sets of 49 coefficients of degree 6 spherical harmonic expansions of phase and group slowness of Love and Rayleigh waves between periods of 100 and 250 seconds. The expansions were obtained from the single-station analysis of surface waves over about 200 paths for Love waves and 250 paths for Rayleigh waves. We invert each coefficient separately to find depth-profiles of the seismic parameters. We can then recombine the coefficients to construct maps of lateral heterogeneities at any given depth by:

$$\mathrm{Parameter}(z,\theta,\varphi) = \sum_{l}\sum_{m} \mathrm{Parameter}_l^m(z)\ Y_l^m(\theta,\varphi)$$

where $Y_l^m(\theta,\varphi)$ are fully normalized spherical harmonics, and $\mathrm{Parameter}_l^m(z)$ are the coefficients of the expansion at a given depth z for a given parameter (density, S-velocity, S-anisotropy,...). This linearized approach is appropriate since the slowness heterogeneities are less than 5%.

The inversion proceeds as follows :

1. phase and group slowness data are combined.
2. the coefficients are corrected for the influence of the uppermost layers of the earth.
3. smoothness and a priori correlation criteria between parameters are chosen.
4. partial derivatives are calculated, using PREM (Dziewonski & Anderson, 1981) as the reference model.
5. each spherical harmonic component is inverted separately; resolution kernels are calculated.
6. the coefficients are recombined to construct maps of the heterogeneities at different depths.

A detailed description of the method is presented elsewhere (Nataf et al.,in preparation).

Heterogeneities in the uppermost layers of the earth contribute significantly to the observed heterogeneities, even at long wavelength. This is illustrated in figure 1 where the power spectra of the phase slowness heterogeneity α of Love and Rayleigh waves and of the upper layers correction are compared at 200 seconds. The power is defined as:

$$\mathrm{Power}_l(\alpha) = \frac{1}{2l+1}\left(\sum_m (\alpha_l^m/\alpha_0^0)^2\right)^{1/2}$$

For l=2, the power in the correction is almost as large as the observed power. It is therefore necessary to correct for near surface effects as completely as possible. We have taken into account four factors: crustal thickness, P_n-S_n-velocities, water depth, and topography. For the first two factors, we used a recent compilation by Phoenix Corporation (Soller et al., 1981). The data were averaged on 15°x15° elements and completed, where data were lacking, by using a predictor based on tectonic setting. Then a spherical harmonic expansion was performed and correction coefficients calculated.

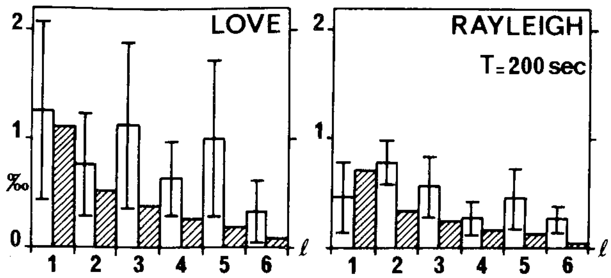

Fig. 1. Power spectrum in each degree ℓ of the spherical harmonic expansion of the phase slowness of 200 seconds Love and Rayleigh waves, with their 2σ error bar. The hatched rectangles give the power of the surface layers corrections.

Evidence for anisotropy in the earth's mantle is steadily increasing. Azimuthal anisotropy can reach 10% in the shallowest mantle, as is measured from P_n waves (Hess, 1964). Polarization anisotropy up to 5% is inferred from surface wave studies in order to fit Love waves (mostly sensitive to SH velocities) and Rayleigh waves (mostly sensitive to SV velocities) simultaneously (Forsyth, 1975; Dziewonski & Anderson, 1981; Cara et al., 1983). In the present study, azimuthal anisotropy has not been taken into account and is likely to have been averaged out in the spherical harmonic expansion. We are left with only polarization anisotropy, and use therefore a transversely isotropic parameterization (Anderson, 1966; Dziewonski & Anderson, 1981). This involves six inversion parameters: ρ, PH, SV, ξ, ϕ, η, where ρ is the density, PH is the horizontally polarized P-wave velocity, SV the vertically polarized S-wave velocity, ξ the anisotropy of S-waves, ϕ the anisotropy of P-waves, and η the fifth elastic parameter, as defined in Takeuchi & Saito (1972). We use a program written by Dziewonski to calculate eigenperiods and partial derivatives for a transversely isotropic earth.

Resolution kernels show that only SV and ξ can be resolved from the fundamental mode Love and Rayleigh waves. However, changes in ρ, PH, ϕ, and η affect these modes substantially. For example, a 5% P-anisotropy (ϕ) has the same effect on Rayleigh wave phase velocity as a 0.1km/s change in SV velocity. We must thus bring in further a priori information. We use the inversion method of Tarantola & Valette (1982), in which a priori information can be introduced naturally in the form of an a priori covariance matrix for the parameters. The a priori information we build in is based on physical considerations. If lateral variations in velocity are due to temperature variations, we can relate them to density changes using laboratory data. Similarly, if anisotropy is caused by the preferred orientation of olivine crystals, we can relate P-anisotropy to S-anisotropy (Christensen & Salisbury, 1979). We choose the following constraints:

$\Delta\rho/\Delta SV = 0.3 \times (1 \pm 50\%)$ $(g/cm^3)/(km/s)$
$\Delta PH/\Delta SV = 1.5 \times (1 \pm 40\%)$
$\Delta\phi/\Delta\xi = -0.5 \times (1 \pm 50\%)$

The a priori correlation length, which governs the degree of smoothness of the model we invert for, varies from 200 km at the base of the upper mantle to 100 km at the top of the model.

Results

Once the inversion has been performed, the coefficients are synthetized to produce maps of lateral heterogeneities at selected depths. Shear-velocity (SV) heterogeneities are shown in figure 2. With a few exceptions they exhibit a strong correlation with surface tectonics down to about 200km. Deeper in the mantle the correlation vanishes and some long wavelength anomalies appear. At 50km (top map) heterogeneities are closely related to surface tectonics. All major shields show up as fast regions (Canada, Africa, Antarctica, West-Australia, South-America). All

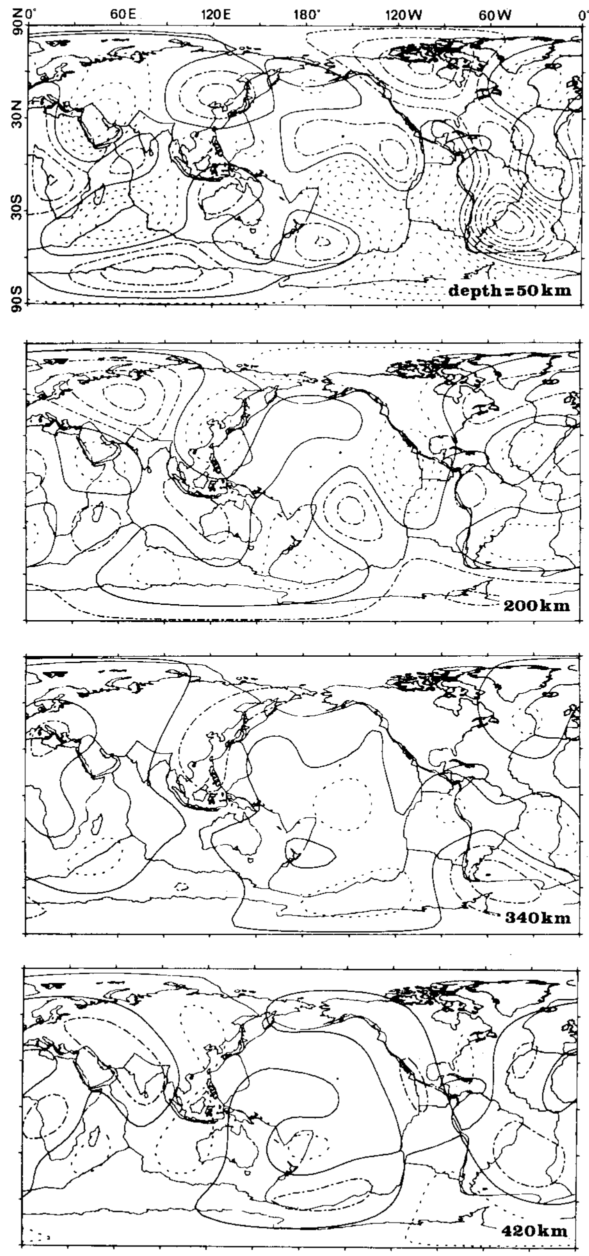

Fig. 2. Shear velocity (SV) distribution maps at selected depths, synthetized from degree 6 spherical harmonic expansion. The countour interval is 0.1 km/s. The solid, chain, and broken lines indicate spherical average (PREM), higher than average, and lower than average velocity.

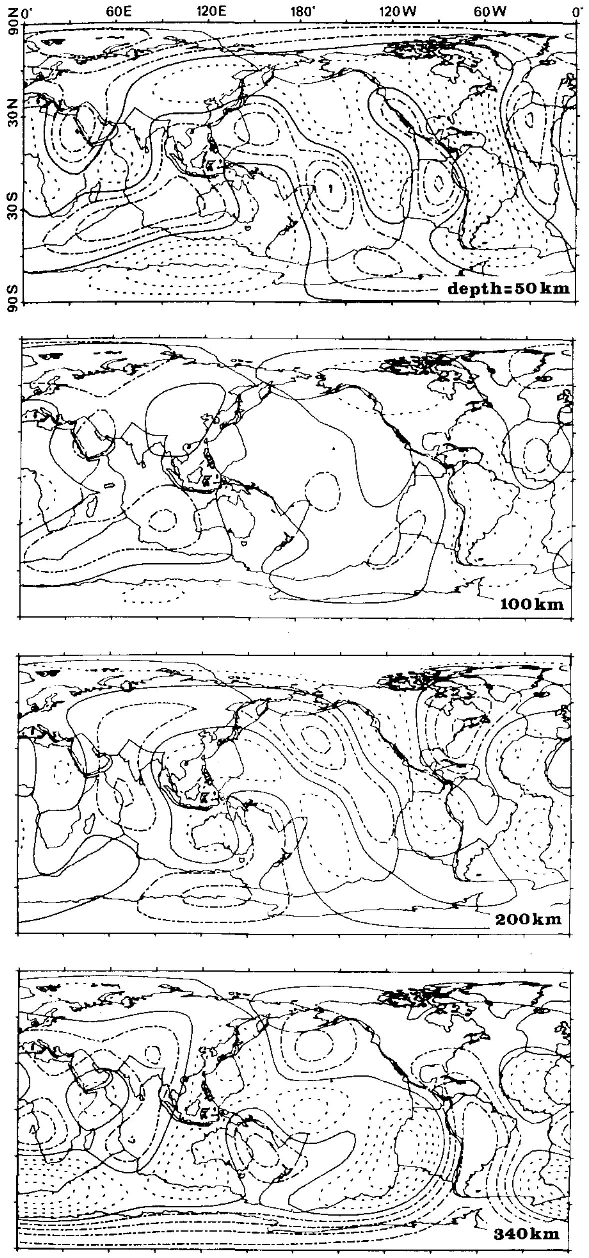

Fig. 3. Anisotropy (SH-SV)/SV distribution at selected depths, synthetized from degree 6 expansion. The contour interval is 2.5%. Symbol conventions as in figure 2.

major ridges show up as slow regions (East-Pacific, triple junctions in the Indian Ocean and in the Atlantic, East-African rift). Old oceans are fast but not as fast as shields. A few interesting regions seem to be anomalous: a slow region around French Polynesia in average age ocean; a fast region centered southeast of South-America. At 100 km, the picture is not drastically changed. Overall variations are smaller (±6%). At 200 km depth, the correlation with surface tectonics starts to break down. Shields are fast, in general, but ridges do not show up systematically. The Afar region is still slow. The south-central Pacific is faster than most shields. An interesting feature appears: a belt of slow mantle at the Pacific subduction zones. This may be a manifestation of the volcanism and marginal sea formation induced by the sinking oceanic slab. At 340 km, the same belt shows up as fast mantle: we are now seeing the presence of the cold slab. However, we cannot exclude the possibility that part of this effect comes from oscillations of our model due to a lack of resolution with depth. At 420 km, the pattern of heterogeneities shows no obvious correlation with surface tectonics. The South-Atlantic is fast, as is the Himalayan-Alpine region. This might indicate the presence of cold subducted material. Many ridge segments are now fast. At larger depths the resolution becomes poor but these trends seem to persist.

Figure 3 shows contour maps of S-anisotropy. The correlation with surface tectonics is more tenuous. At intermediate depths, regions of uprising (ridges) or downwelling (subduction zones) have an SV>SH anisotropy, in agreement with olivine crystals aligned in a vertical flow. At shallow depth (50km), our results show very large anisotropy variations (±10%). However our resolution kernels indicate a rather strong trade-off between SV-velocity and S-anisotropy at this depth. A clear anti-correlation does appear when comparing the two maps at 50km. At 100km the amplitude of the variations is much smaller (±5%) but the pattern is similar. At 200 km, the trade-off with SV-velocity is minimum. The mid-Atlantic ridge has SV>SH whereas the other ridges show no clear-cut trend. Under the Pacific there appears to be some parallel bands trending NW-SE with a dominant SH>SV anomaly. At 340 km, most ridges now have SV>SH (i.e. vertical flow). Antarctica and South-America have a strong SH>SV anomaly (i.e. horizontal flow). North-America and Siberia are almost isotropic at this depth.

Discussion and Conclusions

The correlation between S-velocities and surface tectonics, and in particular, with the age of the ocean floor, has been thoroughly documented (e.g.: Forsyth, 1975). Our analysis confirms the correlation but only down to a depth of about 200 km. Even then, some anomalous regions appear (French Polynesia, eastern South-America, southwestern Indian Ridge). From a geodynamic point of view, these anomalous regions may be of greater interest than 'normal' regions. They might be the seat of large scale thermal or chemical anomalies. Below 200 km, SV heterogeneities are not directly linked to surface tectonics. Subduction zones, however, are still evident. Recently, Woodhouse & Dziewonski (1983) have obtained a spherical harmonic expansion up to degree 8 of the S-structure of the upper mantle. Their results are very similar to ours at shallow depth. In the transition zone, our results show some differences but the major trends remain: in particular the large region of fast mantle in the South-Atlantic. That region is also fast in the even order expansion of Masters et al. (1982). The differences between our maps and those of Woodhouse and Dziewonski (1983) can be attributed to their neglect of lateral variations in anisotropy and their simplified treatment of the crust. On the other hand they have a denser path coverage.

Our results for S-anisotropy have no worldwide counterpart to compare them to. At intermediate depths, S-anisotropy seems to trace the uprisings and downwellings inferred from plate tectonics. To resolve the shallow S-anisotropy, it is necessary to use overtone data (Cara et al., 1983).

Seismology offers a unique way to investigate the structure of the mantle under the lithospheric plates. This ability need not be constrained by using regionalizations based on surface tectonics. The development of digital networks (IDA,GDSN) and the improved understanding of source mechanisms make it possible to undertake studies of lateral heterogeneities with no a priori biases. Our results only offer a crude picture of these heterogeneities. The image will become sharper in the near future as analyses using waveform fitting techniques develop (Woodhouse & Dziewonski, 1983) and use is made of overtones (Cara et al., 1983).

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