Results from Prior NSF Support, NSF OCE 02-21035; W. Menke, PI. Integrating Geophysical Data into New Axial Volcano Magma Chamber Model, 10/01/02-09/30/05; $163,658; 2 years with additional 1 year extension.

This study was directed at understanding the effects of tidal and tectonic loading on stresses within an undersea volcano, such as Axial Volcano (Juan de Fuca Ridge). Three-dimensional, time-dependent simulations of deformation were created using J. Deng's viscoelastic finite-element code, "FEVER". The simulation shown below, for example, examines the time-evolution of shear stress, which is taken as a proxy for seismicity, in the days following a dike intrusion. The westward (positive x) migration of the region of strong shear stress, is controlled by the interaction of magma chamber shape (an overhanging lid, in this case), and viscous relaxation of the magma. It explains the westward migration of shallow microseismicity observed after the 1998 eruption of Axial volcano. Other, comparable, simulations (not shown) explain the amplification of tidally-induced deformation observed by ocean bottom tiltmeters deployed above the Axial magma chamber.

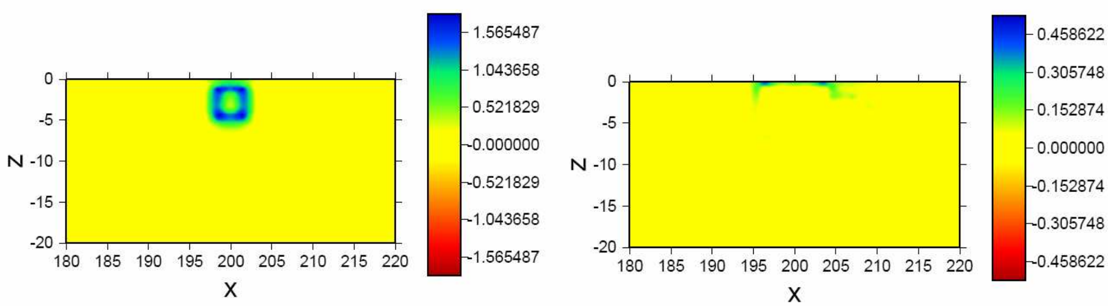

Figure. Cross-section through the earth showing shear stress, τ_{xz}, in MPa induced by a one meter opening of a vertical dike that crosses the center of the magma chamber. (Left) Initially, stress is concentrated in walls of magma chamber. (Right) One week after viscous relaxation of $\nu=10^{11}$ Pa/s magma, stress is concentrated in overhanging lid of magma chamber.

Publications arising from this Research

Menke, W., Stresses induced by tidal and tectonic loading at Axial volcano, Juan de Fuca Ridge, in preparation for submission to J. Geophys. Res, 2007.

Menke, W., R.C. Holmes and J. Xie, On the nonuniqueness of the coupled origin time - velocity tomography problem, in press in Bull. Seism. Soc. Am. 2006.

Menke, W. and D. Schaff, Absolute earthquake location with differential data, Bull. Seism. Soc. Am. 94, 2254-2264, 2004.

Menke, W., Case studies of seismic tomography and earthquake location in a regional context, in *Seismic Earth: Array Analysis of Broadband Seismograms*, Alan Levander and Guust Nolet, Eds., Geophysical Monograph Series 157. American Geophysical Union, 7-36, 2005.

West, M., W. Menke and M. Tolstoy, Focused magma supply at the intersection of the Cobb hotspot and the Juan de Fuca Ridge, Geophys. Res. Lett. 30, doi:10.1029/2003GL017103, 2003.

Note: E. Chesnokov has not served as PI's on an NSF proposal during the last 5 years.

Coupling of Surface Wave Anisotropy, Attenuation and Phase Velocity through Microscale Wavefield Processes occurring in the Mantle

Motivation: During the past five years, long-period surface wave studies have reached a level of fidelity and spatial resolution that has allowed them to contribute in striking new ways to a whole host of problems that are broadly connected with the structure, evolution and geodynamics of the lithosphere and asthenosphere. In many cases, the results are spectacular and the new insights surprising. Yet in some cases, reconciling the results with other research – and especially with results from shorter-period body wave studies – is problematical. In such cases, one asks, "Why is there a difference?" We highlight three general types of explanations: those that focus on *error*; those that focus on a *mismatch* of physical parameters; and those that focus on the *frequency-dependence* of earth properties. In many cases, all three might provide a plausible explanation for a discrepancy.

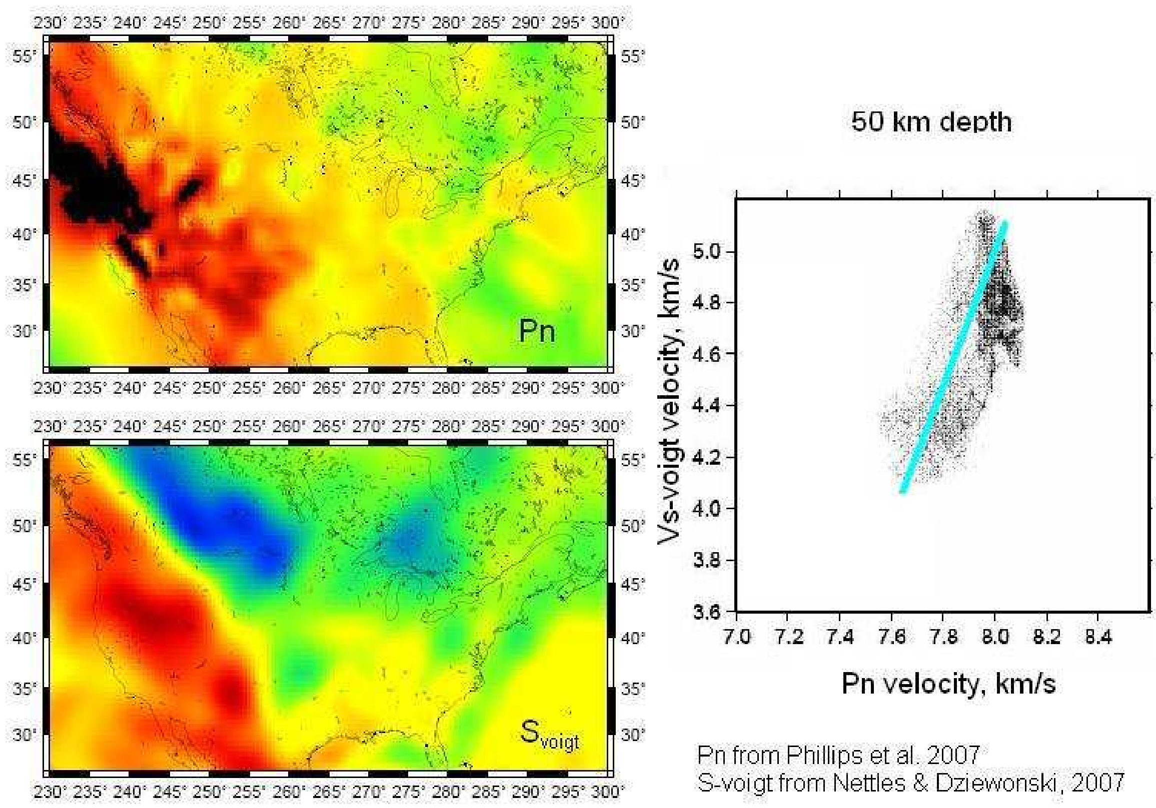

Figure 1. Maps of Pn velocity (top left, from Phillips et al. 2007) and Voigt average (isotropic) S velocity (bottom left, from Nettles 2005 and Nettles & Dziewonski, 2007), both at 50 km depth. A scattergram of these data (black dots) is shown at the right, with a line of slope=3 (blue) superimposed.

To illustrate the problem, consider the maps of North American upper mantle elastic wave velocities (depth of 50 km) shown in Figure 1. The top-left map shows the compressional velocity, V_p, derived from short-period Pn traveltimes (Phillips et al. 2007). The bottom-left shows the isotropic ("Voigt average") part of the transversely anisotropic shear velocity, V_s, derived long-period Rayleigh and Love waves (Nettles 2005; Nettles & Dziewonski 2007). The maps are similar in the sense that they both show the eastern, cratonic part of the continent to be fast (green to blue) and the western, tectonically-active part to be slow (orange to red). Yet the maps have major differences, among the most striking of which is a 300-500 km disagreement on the location of the boundary between these two regions. We would like to know why this difference and other differences occur.

One possibility is that one or both of the images have significant error, perhaps caused by poor ray geometry in the underlying tomography or noisy data. This possibility could be examined by a thorough inspection of the details of the tomographic inversions. Resolution tests, for instance, might shed light on the ability of the tomography to locate the boundary between the tectonically-active west and cratonic east.

A second kind of explanation for the discrepancies would involve pointing out that compressional velocity is not the same a shear velocity. They are independent physical parameters that need not correspond at all. Yet while true, this assertion must be tempered by the fact that we know quite a bit about the way in which plausible variations in earth conditions impact the V_p/V_s ratio. And in fact the observed correlation of compressional and shear velocity (Figure 1, right) is roughly what is expected if temperature, T, were a controlling factor: $[dV_s/dT]/[dV_p/dT] \approx 3$ (e.g., see compilation by Goes et al. 2000, their Table 1). The rather large scatter of about ±0.2 km/s about the trend for both V_p and V_s would, or course, require further explanation.

A final kind of explanation would focus on the extreme disparity in the wavelengths of the short-period body waves (~5 km) that were used to infer V_p and the wavelengths of the long-period surface waves (~500 km) that were used to infer V_s, and the disparity between both these length scales and plausible scale lengths of heterogeneity of the earth (which go right down to the scale of individual mineral grains). In this interpretation, the maps in Figure 1 would represent the 'effective' properties of the earth in specific and different period bands, and might well be expected to be different because different period ranges average earth properties in different ways.

While long-wavelengths in some sense "average out" the small-scale properties of the earth, the physical mechanism that causes this averaging is the rather complicated process of wave field scattering. Consequently, effective properties are not simple averages of the earth's fine-scale structure but rather can have non-intuitive behavior. Finely layered isotropic media, for instance, behaves homogenous but transversely anisotropic at extremely long periods (Backus 1962). And at intermediate periods it is attenuating (O'Doherty & Anstey 1970), and has a slower V_p but a faster V_s than its component layers (Menke 1983). Whether such an explanation of the differences in the North American V_p and V_s maps would work is problematical. On the one hand, we have strong reason to believe that the mantle is heterogeneous at all scales (e.g. Anderson 2006; Helffrich 2006). However, we do not currently have a way to calculate the effect of small scale heterogeneity on surface wave propagation. Developing such ability is a part of this proposal that we will discuss below. But in a rough way, it seems plausible to us that some of the regional difference between the V_p and V_s maps might be due, say, to systematic regional differences in near-Moho structure that influences surface wave propagation but which has no effect on Pn propagation, because of its shorter wavelength.

More Examples. We briefly examine several other interesting surface wave results that raise challenging issues whose explanation might possibly involve the *frequency-dependence* of earth properties.

1. Why is there such variation in the lithospheric shear velocity gradient? Nettles (2005), Nettles & Dziewonski (2007) and Kustowski (2007) provide high-resolution shear wave profiles in a variety of continental regions. These transversely-anisotropic models exhibit rather large variation of the isotropic (Voigt average) part of the shear velocity, with both strong positive (+0.003 s^{-1}) and strong negative (–0.003 s^{-1}) gradients in the upper 100 km (Figure 2). These results can be compared with continental-scale traveltime inversions. Nettles & Dziewonski's (2005) shear gradient of –0.0037 s^{-1} for cratonic North America is a different sign than Iyer et al.'s (1967) compressional gradient of +0.001 to +0.002 s^{-1} (based on traveltimes from Early Rise chemical explosions in Lake Superior). Kustowski's (2007) shear gradients of +0.015 to +0.003 s^{-1} in northern Eurasia are the same sign, but significantly lower than Pavlenkova & Pavlenkova's (2006) compressional gradients of +0.004 to +0.006 s^{-1}•(based on traveltimes from Peaceful Nuclear Explosions). Once again, we are comparing apples to oranges (that is, compressional to shear velocity gradients), yet the correspondence unclear and the variability not easy to interpret.

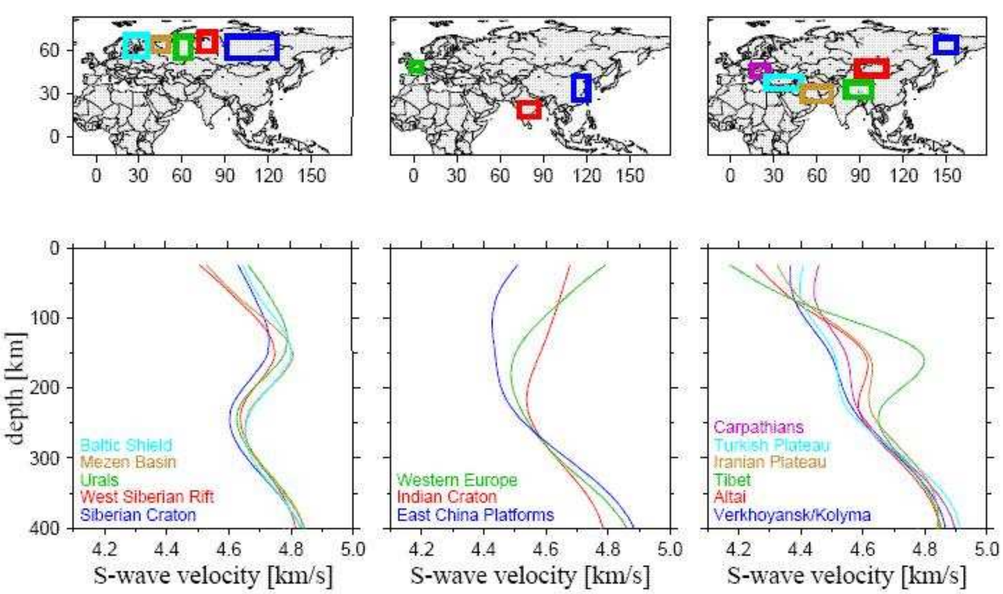

Figure 2. Upper mantle shear velocity (Voigt average isotropic) for various patches of Eurasia, from Kustowski (2007). Note the large variation of the velocity gradient in the upper 100 km, with some areas having a large positive gradient (e.g. Mezen Basin, dVs/dz=+0.003 s^{-1}) and others having a large negative gradient (e.g. Western Europe, dVs/dz=–0.003 s^{-1}).

An important question is whether variations in simple earth properties, such as temperature and chemistry (e.g. depletion) can explain such variability. Certainly temperature alone would seem unable to explain the different *sign* of the shear gradient between cratonic North America and Northern Eurasia. We speculate that regional variations in small scale structure, averaged in different ways by the surface and body waves, may be responsible for some of the variability.

2. Is the Low Velocity Zone (LVZ) caused by partial melting? Thybo and Perchuc (1997) and Thybo (2006) argue that the global ubiquity of a P wave shadow zone that starts at about a source-receiver range of about 10-12° is evidence of the mantle being at near-solidus temperatures at depths in the 100-200 km range, even in stable continental regions. Surface wave tomography also provides evidence for LVZ's. For example, almost all of Kustowski's (2007) profiles shown in Figure 2 have very large LVZ's (and some, such as Western Europe, have extremely large ones). Ray tracing through these 3D models corroborates the existence of large body wave shadow zones (though agreement in detail with body wave S data has not been addressed) (Figure 3).

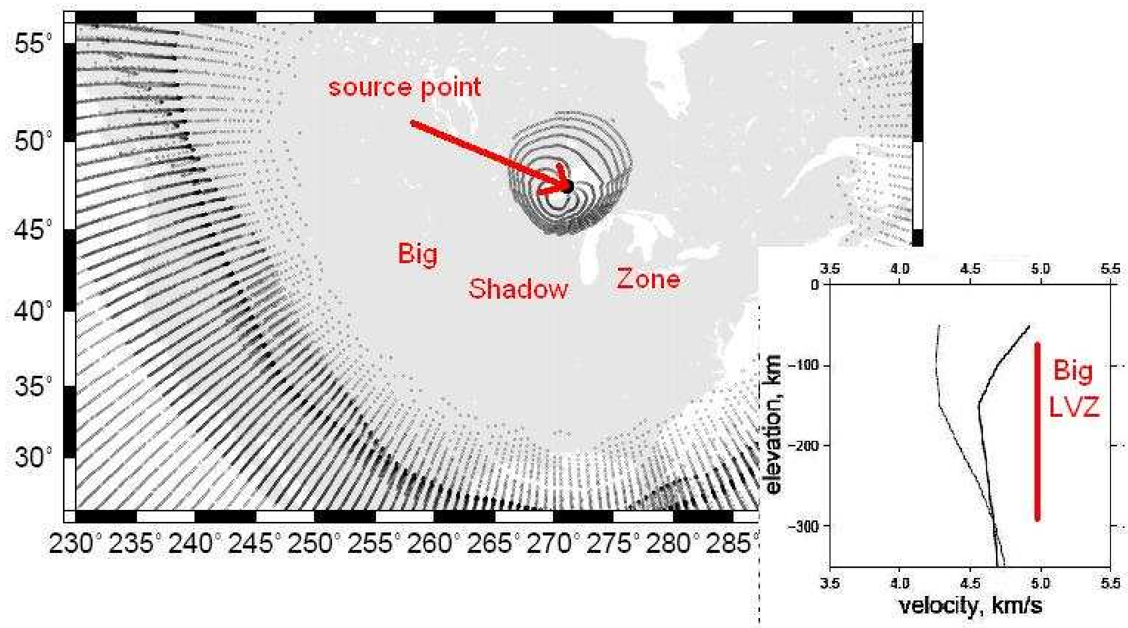

Figure 3. Ray exit points (dots) for shear waves from a source point of the Early Rise experiment, for a 3D earth model based on Nettles & Dziewonski's (2007) isotropic (Voigt average) shear wave velocities. Note large shadow zone. (Inset) Shear velocity profiles through this model typical of central (bold) and western (solid) North America. Note large Low Velocity Zone. Menke's (2005) raytracing code was used to compute raypaths.

On the other hand, Dalton and Ekstrom's (2006) measurements of surface wave attenuation beneath cratonic North America and northern Eurasia show no difference in attenuation between periods of 75 and 150 seconds (Figure 4). These periods have peak sensitivities in the ~100 km and ~200 km depth range, respectively. Taken at face value,

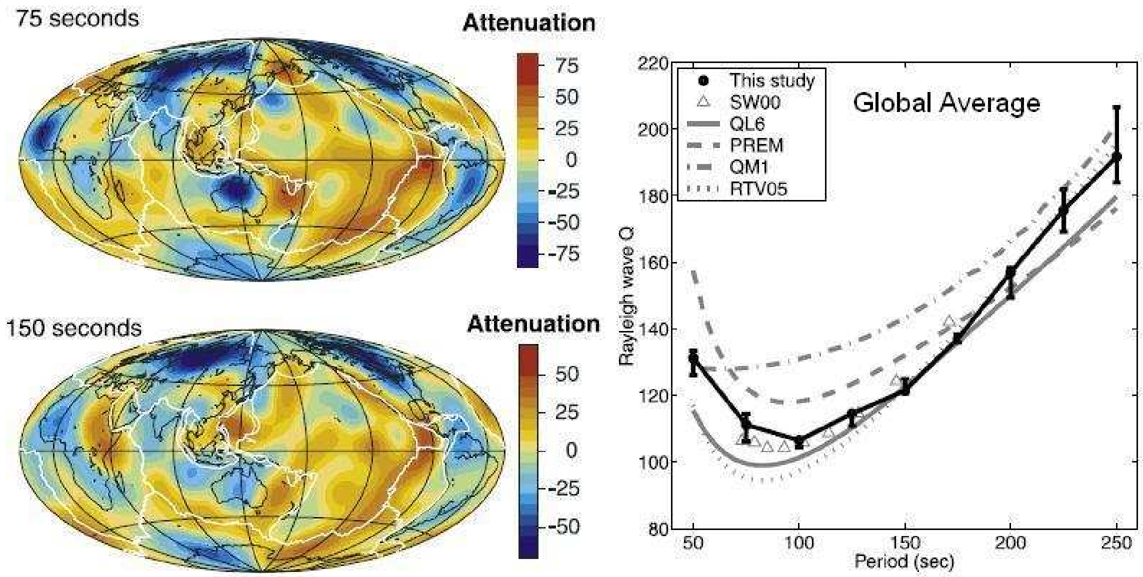

Figure 4. (Left) Rayleigh wave attenuation (the quantity $\delta Q^{-1}/Q^{-1}$) at periods of 75 and 150 seconds. The attenuation beneath the Great Lakes region of North America is $\delta Q^{-1}/Q^{-1}=-55$ and beneath western Siberia is about –60 at both periods. (Right) Clear variation with period is observed in the global average. From Dalton and Ekstrom (2006).

this result is difficult to reconcile with the hypothesis that the LVZ beneath the regions is due to near-solidus temperatures, since we expect extremely high attenuation in that case (Kampfmann & Berckhemer 1985). We are left wondering what kinds of attenuation mechanisms (if any) might be able to explain this discrepancy, and speculate that a mechanism that has a strong frequency-dependence would be needed.

3. How can transverse and azimuthal anisotropy be reconciled? Gaherty and Jordan (1995), using long-period Love and Rayleigh surface wave data, showed that the upper 200 km of the North American craton has strong transverse anisotropy, with horizontal ($S^{\bullet}$) polarizations being systematically faster than vertical (S^{V}) polarizations (Figure 5, right). Fouch et al. (2000), using the splitting of intermediate period SKS body waves observed along the MOMA Array showed that this region has a consistent pattern of azimuthal anisotropy, with a northeast/southwest fast direction (Figure 5, bottom left). Superficially, these two studies would seem to be looking at complete different aspects of anisotropy. However, Gaherty (2004) has pointed out that the Love/Rayleigh transverse anisotropy is observed even along northeast/southwest azimuths that are parallel to the splitting fast direction, and has argued that no single layer of any plausible anisotropic material can have this behavior. We are thus left with the sense that either the anisotropic structure of this region is very complicated, with several distinct layers in different depth intervals, or that the anisotropy is strongly frequency-dependent, so that the intermediate period SKS waves and longer period surface waves are being influenced by two completely different anisotropies. This would likely be the case if Backus-type (1962) layering were contributing to the anisotropy, in addition to lattice-preferred orientation (LPO) of olivine crystals in the mantle.

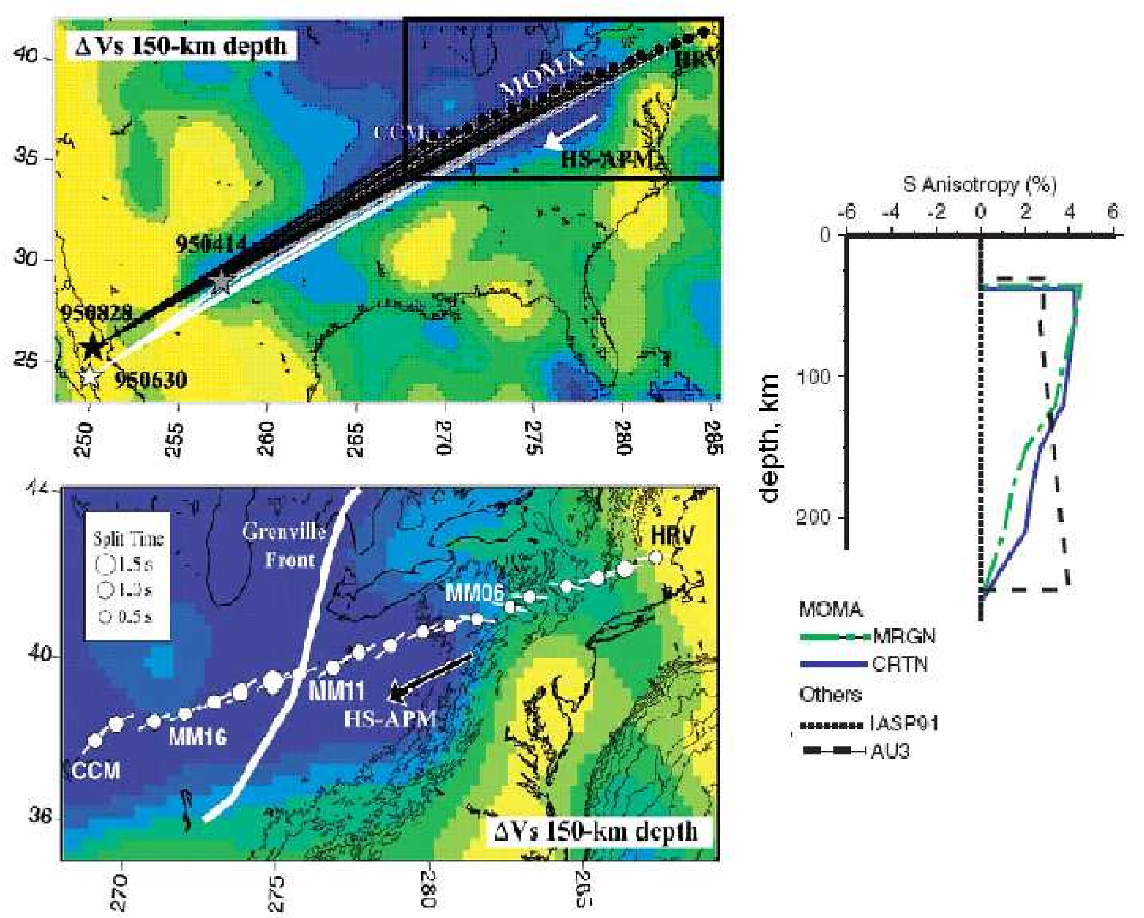

Figure 5. (Top-left) Propagation paths for several earthquakes across eastern North America is parallel to the MOMA array. (Bottom-left) Azimuthal anisotropy along MOMA array, as determined by shear wave splitting (adapted from Fouch et al. 2000). (Right) Transverse anisotropy along these paths deduced from Love and Rayleigh waves (adapted from Gaherty 2004)

Are multiple layers of anisotropy present in the mantle? Several authors have used SKS splitting observations to argue for multiple layers of anisotropy beneath North America (e.g. Silver & Savage, 1994; Levin et al. 1999; Menke and Levin 2003). One of the limitations is that SKS splitting data are not sensitive to the depth of the anisotropy (except that it is probably above the transition zone). Correlations of SKS-derived anisotropic layers with geodynamical units such as lithosphere or asthenosphere are thus speculative. Li et al. (2003) attempt to resolve this ambiguity for northeastern North America by correlating Rayleigh wave azimuthal anisotropy with the SKS-derived models. However, they do not detect significant Rayleigh wave anisotropy, and thus conclude that the source of the SKS signal must be deeper than the limit set by the penetration depth of the longest periods in their study (120 s, corresponding to depth of about 200 km). Marone and Ramonowicz (2007) perform a joint inversion of surface wave and SKS data, to produce a three-layer model of anisotropy. Their lowest (300 km deep) layer has a fast direction that most closely corresponds to observed northeast/southwest SKS fast directions (red bars in Figure 6), with the top (100 km deep) layer that is highly oblique to it (black bars in Figure 6). This inversion

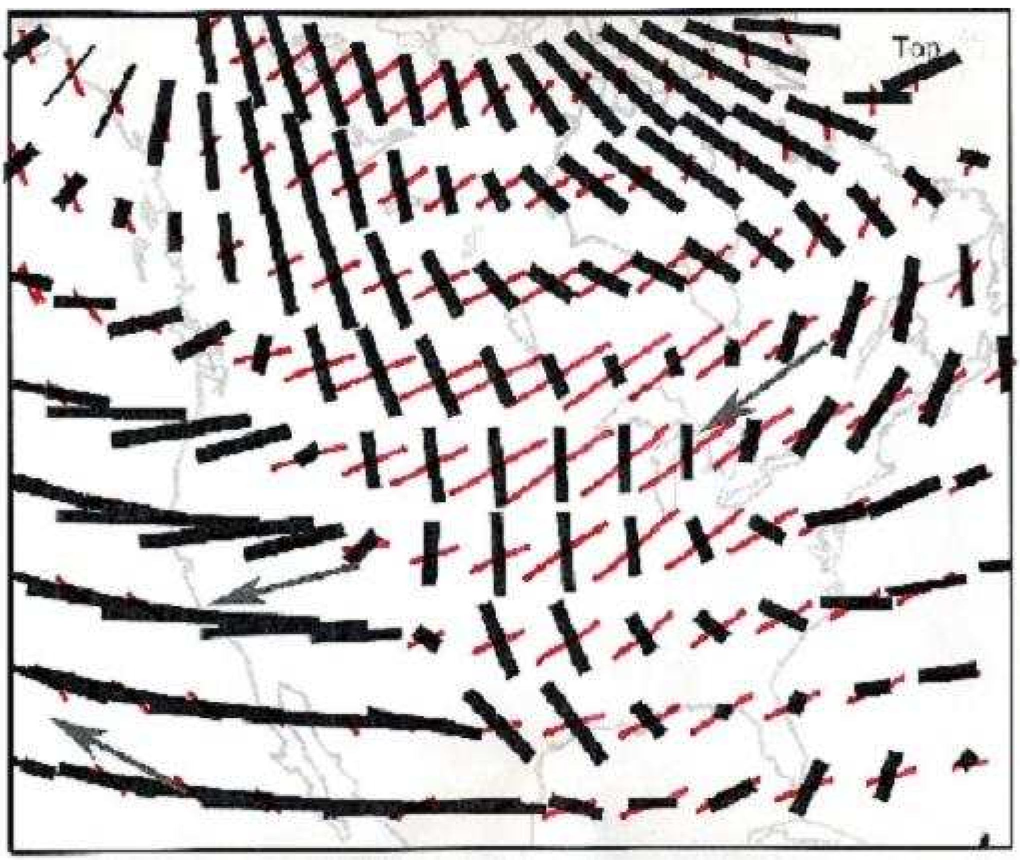

Figure 6. Azimuthal anisotropy at depths of 100 km (black) and 300 km (red), adapted from Marone & Ramonowicz (2007). Bar direction is parallel to the fast direction of anisotropy and bar length is proportional to the amplitude of anisotropy.

is successful in the sense of reducing the variance of both the surface wave and SKS splitting data. Marone and Ramonowicz (2007) argue that the lower layer is associated with absolute plate motion, an explanation that has also been used to explain the SKS data (e.g. Levin at al. 1999). However, the azimuth of the nearly-perpendicular top (100 km) layer fast axes defies easy explanation, since it is sub-parallel to no known tectonic trend. One is left wondering whether the SKS and surface waves really are sampling the same anisotropic earth, or whether (following our previous speculations) the anisotropy of the earth really looks radically different at these two period ranges.

Effective Properties of the Earth. Effective medium theory is collection of approximate methods for determining how seismic waves of a given period average earth properties. The goal is to derive an effective elasticity tensor and an effective density for patches of the earth that contain an ensemble of small heterogeneities (described statistically), and then to use the effective elasticity and density in wave propagation simulations. As in all wave propagation problems, one must start with a wave equation that embodies the essential physics of wave propagation in the earth. Some choices must be made:

1. Whether to model the earth is fully anisotropic, described by an elasticity tensor C_{ijpq} *(Definitely yes, since anisotropy is a key seismic observable).*
2. Whether to model intrinsic attenuation, by allowing the elasticity tensor to have an imaginary part *(Probably yes, since attenuation is also a key observable);*
3. Whether to allow for fluid inclusions *(Probably yes, since inclusions of partial melt may be important in the asthenosphere)* and whether to allow the wave field to excite Biot-style (1956) motions of those fluids *(Possibly yes, since fluid motions will cause wavefield attenuation);* and
4. Whether to model self-gravitation *(Probably no, since this is important only for the gravest of earth motions).*

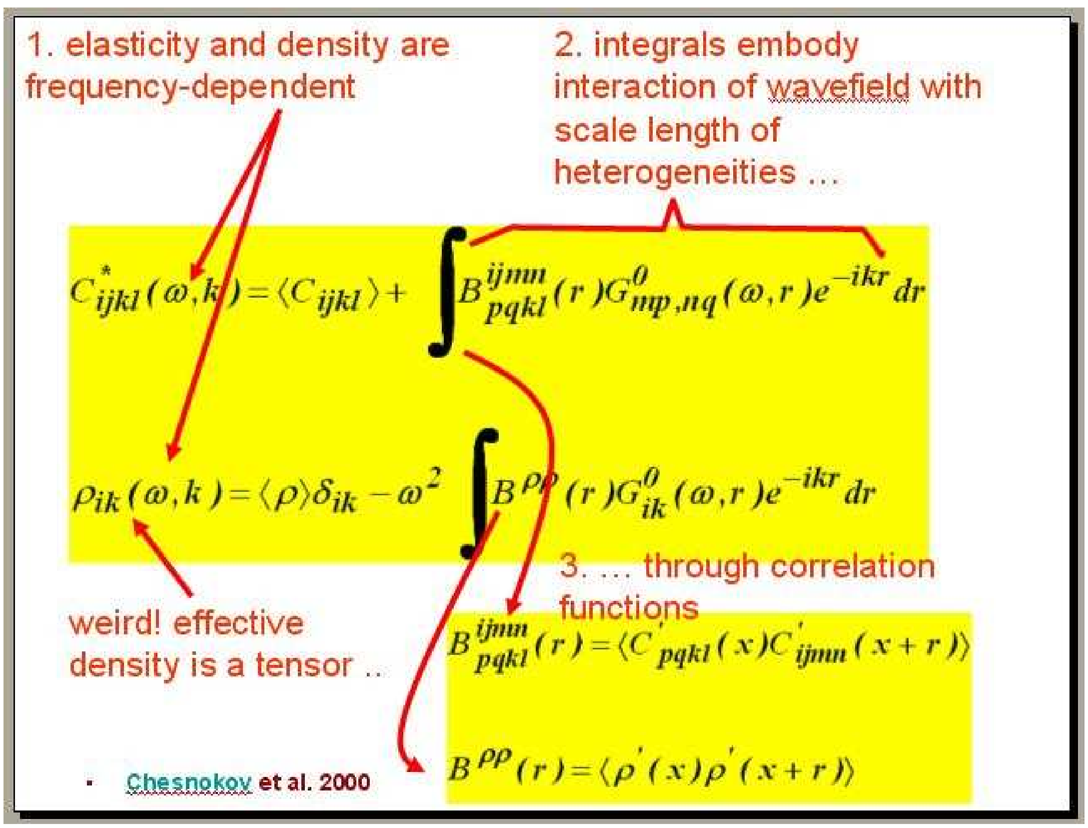

Figure 7. Some important elements the effective media calculations. See text for further discussion.

A second issue is how to describe the heterogeneity in the earth. At the largest scales, we view the earth as having a deterministic structure that is named parts (crust, lithosphere, LVZ, etc.) with a known, or at least roughly known, structure that can be specified in terms of layers or voxels or some similar deterministic method. It is, of course, infeasible to think about small scale heterogeneity in this way. One must instead resort to a statistical description, but one that is rich enough to capture our notion that the earth's heterogeneity has both a variety of scale lengths and some intrinsic fabric (meaning alignment). Here effective medium theory itself is helpful in suggesting an appropriate statistical description: two-point correlation functions (e.g. 3D autocorrelation and cross-correlation of elastic parameters) arise quite naturally in their development and are also able to capture a great deal about the statistical "texture" of the earth's heterogeneity.

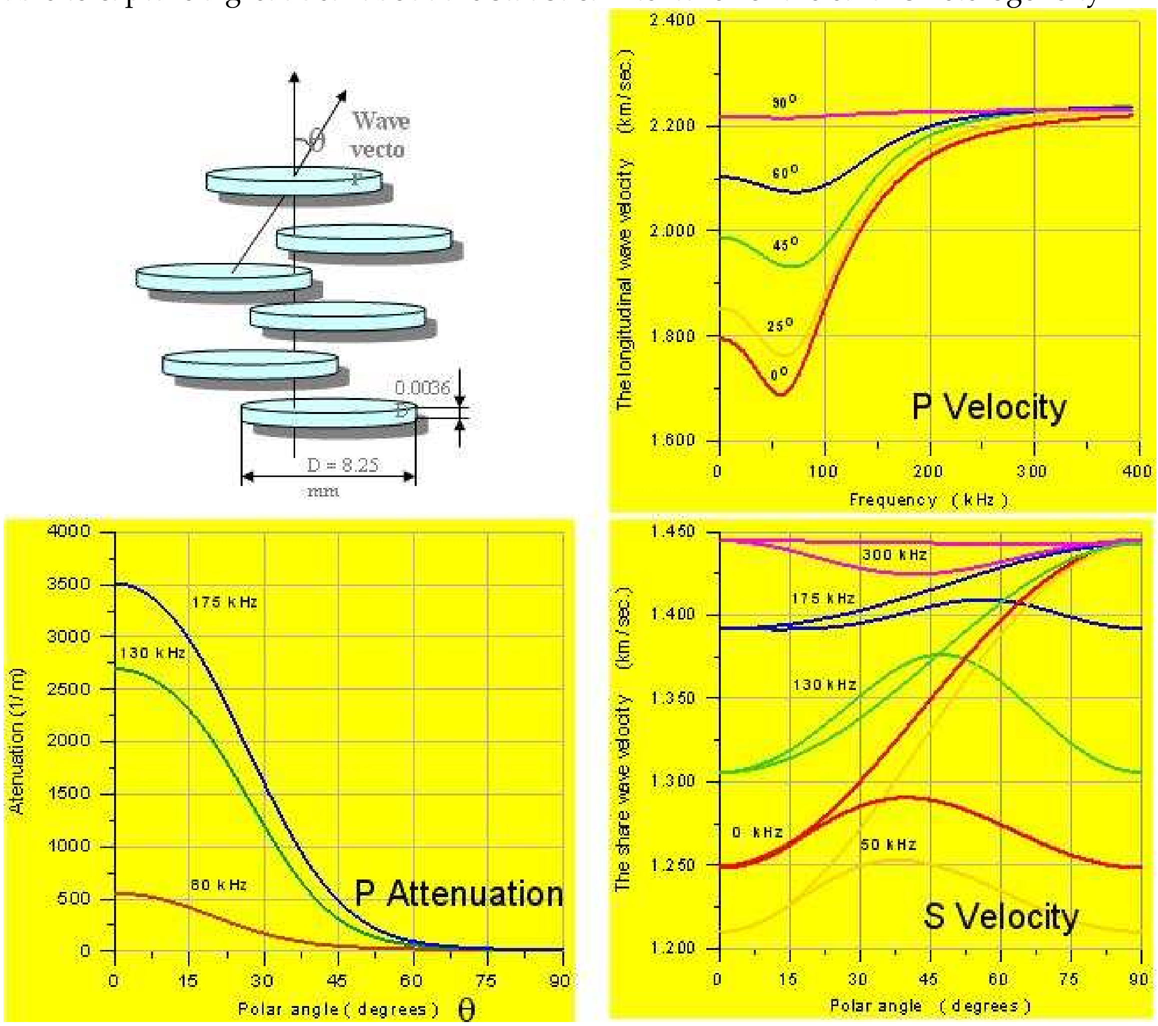

Figure 8. Example of effective P and S velocities and P attenuation, for model with aligned gas-saturated inclusions, calculated using effective medium theory. Note strong angular dependence (anisotropy).

The particular effective medium theory that we use is developed by Chesnokov's group (Chesnokov et al. 1995, 1998; Chesnokov & Kukharenko 1996, Bayuk & Chesnokov

1998). We review it only briefly here (Figure 7). The key idea is that the effective elastic constants C^*_{ijpq} can be written as the sum of two terms: a frequency-independent part, $\langle C_{ijpq} \rangle$, that is just a simple spatial average of the heterogeneous constant; and a frequency-dependent part that depends upon both the statistics of the heterogeneities (through correlation functions, B) and the properties of the wavefield (through a Green function, G^0_{mp}). Most of the effort of calculation is in the construction of the Green function, because it embodies the effect of scattering of the wavefield from small-scale heterogeneities. The Green function satisfies a Dyson-type integral equation (not shown here) that must be solved numerically (and in Chesnokov's formulation, using the so-called general singular approximation) (Bayuk and Chesnokov 1998). Sample results are shown in Figure 8.

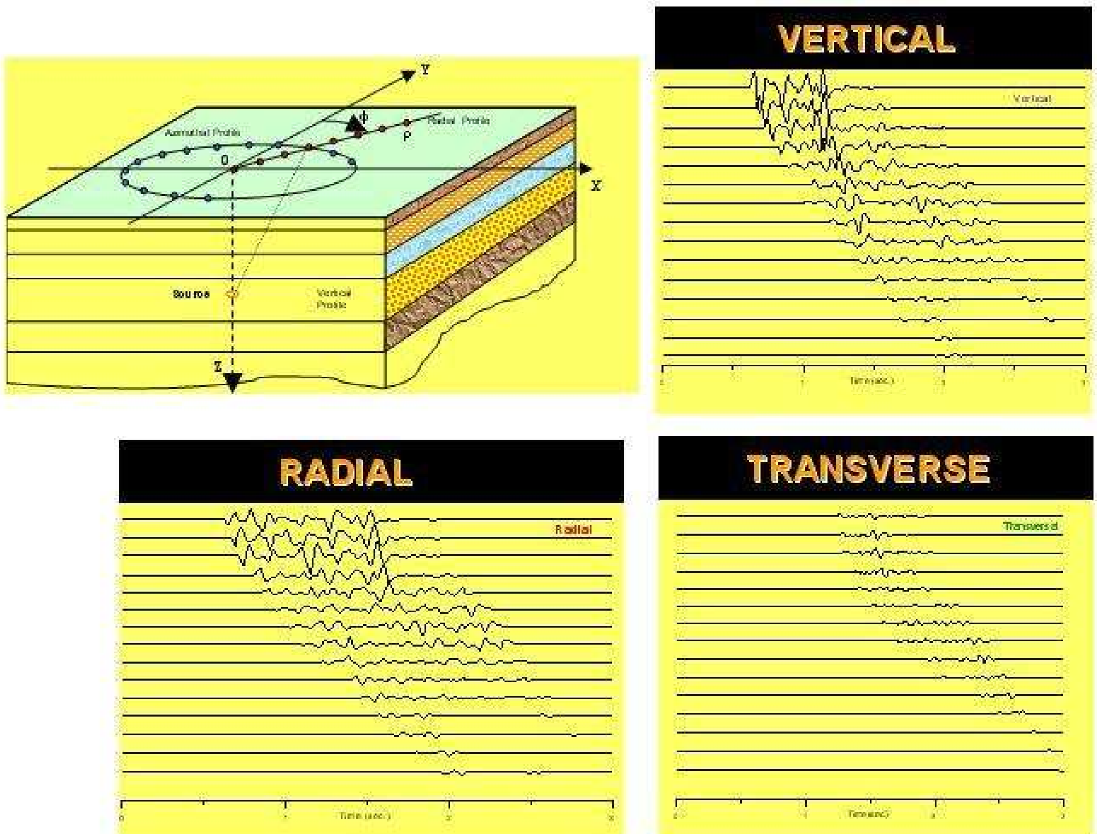

Figure 9. Sample seismogram calculated using extended Reflectivity Method. See text for further discussion.

A key advance in effective medium theory is its introduction into synthetic seismogram calculations through an extension of the reflectivity method (also called the *frequency-wavenumber method* and the *propagator matrix method*) (Figure 9). Over the years the reflectivity method has become a standard technique for computing synthetic seismogram in layered earth models, because it permits a complete and "realistic looking" seismogram, including both body and surface wave forms, to be calculated. It was first developed for isotropic, lossless earth models (Kind 1979) and was later extended includes layers with a frequency-dependent attenuation (O'Neill and Hill 1979) and anisotropy (Booth, 1980; Chesnokov & Abaseev 1986, 1987, Tsvankin and Chesnokov, 1990a,b). Chesnokov's approach is to allow each layer to contain small-scale heterogeneities, cracks and inclusions, as described by correlation functions. The effective medium problem is then solved in a self-consistent way, so that each layer acquires a frequency-wavenumber dependent anisotropic elastic tensor. The result of this calculation can be quite different than the case where anisotropy is merely prescribed *a priori*, since in that case its frequency-dependence (if any) would not necessarily be consistent with the wavefield.

Anisotropy of Attenuation as a Possible New Discriminator of Earth Properties. Seismic attenuation can arise from a variety of sources, including atomic-scale deformation within crystals or along grain boundaries, shear within fluid-filled cracks and inclusions, and scattering from small heterogeneities. If the crystals, cracks, inclusions or heterogeneities have a preferred alignment, then the attenuation will be

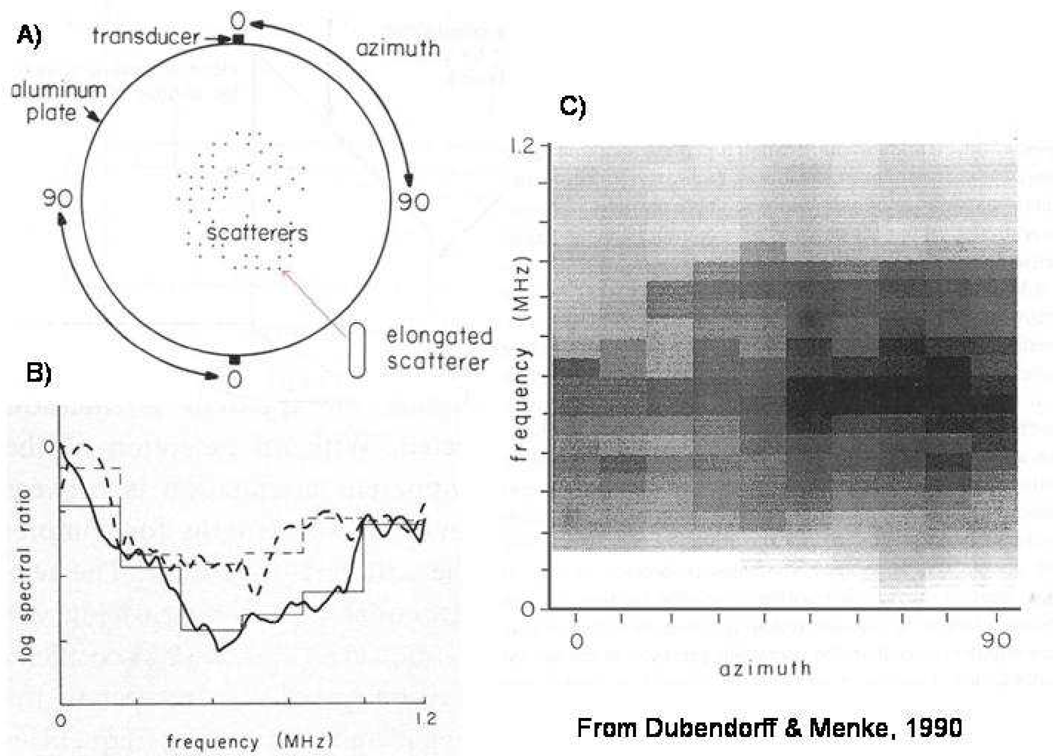

Figure 10. Anisotropy of attenuation in metal sheet with aligned elliptical scatterers. A) Experimental geometry. B) Compressional wave attenuation, as quantified by log spectral ratio, for propagation paths parallel to (dashed) and perpendicular to (solid) the long axis of the scatterers. Note attenuation is largest for perpendicular propagation. C) Attenuation depicted as grey-scale plot, with darker shades implying more attenuation, as a function of frequency and propagation direction.

anisotropic, that is, the amount of attenuation will depend upon the direction of propagation. Effective medium theory provides a way to calculate this attenuation anisotropy, given a statistical description of the heterogeneities and inclusions (Figure 8, lower left).

Anisotropy of attenuation was investigated using physical modeling experiments by Dubendorff and Menke (1990), who examined P wave propagation in metal plates containing aligned elliptical inclusions (Figure 10). They detected strong anisotropy of attenuation in both P and S waves, with the least attenuation experienced along propagation directions parallel to the long axes of the inclusions.

Laboratory measurements have also been used to quantify anisotropy of attenuation in rock samples. Kern et al. (1997) measures both velocity and attenuation in samples of serpentinite and amphibolite, rocks with a strong metamorphic fabric. Both velocity anisotropy and attenuation anisotropy are detected. Interestingly, they conclude that "the nature of anisotropic behavior of velocities seems to be different than that of attenuation". The attenuation anisotropy of the rock seems more controlled by more by inclusions than by LPO, while the reverse seems to be true for velocity anisotropy.

Anisotropy of attenuation has been detected in the earth's inner core (Yu and Wen 2004). In this case, the direction of highest attenuation is parallel to the P wave fast direction.

These results suggest that attenuation anisotropy, if it is detectable within the asthenosphere, might provide information that would help discriminate between different attenuation mechanisms that are occurring there. In particular, it might be extremely useful in determining whether melt is present, since we would expect melt to be distributed in inclusions that are aligned by the stress field (especially in extensional regimes such a mid-ocean ridges) and by buoyancy. To our knowledge, no attempts have been made to identify anisotropy of attenuation in surface wave data. It is a subject that we wish to explore further (and will discuss below).

Proposed Research and Issues We Plan to Explore

1. Assessing the effect of small-scale heterogeneities on surface wave propagation.

A. Model building (years 1 and 2). There is ample evidence that the upper mantle is heterogeneous at scales that are much smaller than the 100-1000 km wavelength of the Love and Rayleigh waves discussed in this proposal (e.g Anderson 2006; Thybo 2006). It includes not only microscale heterogeneity, but also kilometer-scale features such as diapers (Nicholas et al. 1988), melt conduits (Speigelman et al. 2001), frozen magma lenses (Nedimovic 2005), since scale layering (Thybo 2006), shear zones and layered reflectors (e.g. McBride et al. 1995; Thybo and Perchuc 1997). However, we see as a major challenge the process of going from these anecdotal and fairly qualitative descriptions of heterogeneity to a well-documented parametric description that can be used in numerical modeling. We plan to construct a suite of test models for each of several major tectonic units (cratonic, rifted continent, old oceanic,

oceanic ridge), with the models consisting of a deterministically specified large-scale stratification with a statistically-specified small-scale heterogeneities superimposed.

B. Synthetics Seismograms and surface wave dispersion estimates (years 2 and 3). We will then compute synthetic seismograms and dispersion curves surface waves propagating in these models, using the effective-medium adapted Reflectivity code, with the goal of understanding the impact of heterogeneity on critical surface wave properties. One critical question is the amount of "signal" (e.g. in the variation of phase velocity with frequency, and in the amount of anisotropy) that can be caused by the small-scale heterogeneities. Here comparisons between fully-deterministic modes and models that include statistical fluctuations will be very helpful.

C. Impact on Inversion (years 2 and 3). The final element of this analysis is understanding the impact that ignoring statistical heterogeneity has on structures determined though standard, deterministic surface wave inversion. Recall that in a deterministic structure the frequency dependence of the dispersion is being caused by the interaction of the frequency-dependent surface wave penetration depth and the vertically-stratified elastic structure of the earth. But effective medium theory adds a new frequency-dependent element to the problem, the effective elasticity itself. Thus it is reasonable to presume that omitting this frequency-dependence will cause error in recovered deterministic part of the velocity structure. Such error, we speculate, might account for some of the variability of upper mantle structures exhibited in Figure 2.

2. Assessing whether effective media approach can reconcile surface wave and body waves anisotropy. The forward modeling in part 1 also allows us to examine short-period body wave propagation in the same suite of model as used for the surface wave analysis. Thus we can quantitatively evaluate two extremely important issues that were discussed in the Examples section, above:

A. Does the presence of statistical heterogeneity help to reconcile SKS and surface wave anisotropy measurements (years 1 and 2)? The working hypothesis that we have is that while a single tectonic process might be responsible for the creation of a given region of the upper mantle, the creation process will induce both LPO of mantle minerals and small-scale structural heterogeneities. Since the anisotropy associated with the heterogeneity is strongly frequency dependent, the earth might look quite different at periods typical of SKS and surface waves, even though the fabric of the heterogeneities might have some simple geometrical relationship to the mineral alignment. This we will experiment, for instance, with models in an attempt to reproduce behaviors typified by the Marone and Ramonowicz (2007) results depicted in Figure 6.

B. Does the presence of statistical heterogeneity help to reconcile surface wave velocities with those inferred from body waves (and especially from P waves) (years 2 and 3)? Following the same general logic as in A, above, we will examine whether small-scale heterogeneity might be able to make lithospheric velocity gradients seem more intense when determined by surface waves than when determined by body waves.

3. Understanding anisotropy of attenuation in long-period surface wave data.

A. Can anisotropy of attenuation be detected? (year 1) By a fortuitous coincidence, Colleen Dalton, who is the lead author on the surface wave attenuation research cited here, will be joining the Lamont Seismology Group as a fully-funded Postdoctoral Fellow, and has agreed to work with us on this problem. Her already-assembled database of ~50,000 path-averages of surface wave attenuation is an ideal dataset for exploring this question. Our approach will be to start with the world's mid-ocean ridges, where we might expect highly-attenuating melt to be present and to occur in conduits and inclusions that have a simple orientation with respect to the known spreading direction.

B. What does it tell us about the earth? (year 2) Should such anisotropy of attenuation be detected, we will use the use new forward modeling and inversion capabilities to infer mantle properties (e.g. statistical distribution of melt-filled cracks).

4. Algorithmic Improvements. We will also address two "algorithmic" issues that are important in applying effective medium theory to upper mantle problems.

A. Improving effective medium codes to better model mantle melt (year 2 and 3). Effective medium theory allows us to take into account both the elastic symmetry of peridotite matrix in which melting is occurring and the statistics of the inclusions containing melt (i.e. their aspect ratio, orientation and the viscoelastic properties of the melt). Some code improvements, however, will be needed to fully account for the frequency dependence associated with the viscosity of the melt. Once extended in this fashion, the methodology which will allow us to distinguish between structures with various mechanisms of attenuation (scattering or intrinsic viscosity of inclusions).

B. Is "effective medium inversion" practical (year 3). A natural question to ask is whether it is possible to perform waveform inversions that recover both a deterministic part and a statistical part of earth structure, and if so, what the relevant practical issues are in implementing such an inversion. Our experience with a related "up-scaling" problem in exploration geophysics gives us some confidence that such an inversion might be possible. In up-scaling, sonic log data is used to infer the statistical description (correlation functions) of small-scale heterogeneity in the earth, and then this heterogeneity is used to predict the seismogram that would be observed in a Vertical Seismic Profiling (VSP) experiment. Up-scaling is not quite inversion, but it embodies some of the underling algorithms that inversion would rely upon.

Project Management. The PI's Evgeni Chesnokov and William Menke will work together on all aspects of the project. Chesnokov, assisted by Dr. Vikhorev, will be responsible for the design, implementation and testing of new codes, with all effective medium calculations being performed on OU computers. Menke, assisted by a GRA, will be responsible for designing numerical experiments that test particular hypotheses about the heterogeneity of the earth, and for inversion methods, and (in conjunction with Dalton) for studies of surface wave anisotropy of attenuation.

Dissimination of Results. Menke will maintain an archive of preliminary results on his web site, www.ldeo.columbia.edu/users/menke, which is already quite extensive. We will present results at national scientific meetings, such as the Fall AGU, and make as best faith effort to publish results as rapidly as possible in peer-reviewed journals.