RESULTS FROM PRIOR AND CURRENT NSF SUPPORT

B. Romanowicz. (P.I.),

- 1-*Studies of the CMB and inner core using body waves and free oscillations* EAR-9417862 (3/95-2/98) $181,472
- 2-*High resolution modeling of the core-mantle boundary region and the inner core* EAR-9902777 (8/99-7/03) $220,000.
- 3-*Global attenuation model of the whole mantle using waveform tomography* EAR-9706380 (09/01/97-12/31/99), $140,000.
- 4-*3D Variations in temperature and composition in the upper mantle* EAR-0001955, (8/00-8/01), $19,391 to UMN, EAR-0001965 $27,703 to UCB, collaborative with S.Karato.
- 5-*Collaborative Research: Toward mapping 3D variations in temperature and chemical composition in the upper mantle through a mineral physics-based inversion of seismic waveforms* EAR-0112409, (08/15/01-07/31/03), $104,999
- 6-*Numerical Modeling of Strong Heterogeneity in the Deep Mantle* EAR-0106000, (07/15/01-06/30/04), $238,525.

These grants have supported work on several problems of deep earth structure: (1) and (2) - anisotropy in the inner core using PKP travel time and attenuation data, as well as free oscillations splitting measurements; structure of D" from observations of S-diffracted waves at long ranges as well as PKP travel times. (3) - Global mantle tomography, in particular attenuation tomography, introducing a waveform modeling approach. (4) and (5) are CSEDI grants that have supported a pilot study to prepare the ground for mapping out three-dimensional variations of temperature and water content in the earth's upper mantle directly from seismic waveform data. (6) Forward modeling of D" structure using a coupled spectral element/normal mode approach. This is the only grant that will still be active after the beginning of the period covered by the present proposal. Some of the recent results are described in more detail in later sections of this proposal.

Publications supported by these grants:

These publications are listed in full in the reference section of the proposal and indicated by (**): Souriau and Romanowicz (1996; GRL); Souriau and Romanowicz (1997; PEPI); Romanowicz, Li and Durek (1996, Science); Mégnin et al. (1997, GRL); Souriau and Romanowicz (1997, PEPI); Bréger and Romanowicz (1998, Science); Bréger, Romanowicz and Vinnik (1998, GRL); Romanowicz (1998, Pageoph.); Mégnin, C. and B. Romanowicz (1998, GJI); Vinnik, Bréger and Romanowicz (1998, Nature); Vinnik, L., L. Bréger and B. Romanowicz (1998, GRL); Bunge et al. (1998, Science); Durek and Romanowicz (1999, GJI); Bréger, Romanowicz and Tkalčić (1999, GRL); Mégnin, C. and B. Romanowicz (2000, Geophys. Monogr.); Mégnin, C. and B. Romanowicz (2000, GJI); Bréger, Tkalčić and Romanowicz (2000, EPSL); Bréger, Romanowicz and Rousset (2000, GRL); Romanowicz and Bréger (2000, JGR); Romanowicz and Durek (2000, Geophys. Monogr. 117); Clévédé et al. (2000, PEPI); Igel et al. (2000, PEPI); Bréger, Romanowicz and Ng (2001, GRL); Romanowicz (2001, GRL); Tkalčić, Romanowicz and Houy (2002, GJI); Tkalčić and Romanowicz (2002, EPSL); Romanowicz and Gung (2002, Science); Kuo and Romanowicz (2002, GJI); Romanowicz (2002, CRAS); Romanowicz (2002, AGP); Romanowicz, Bréger and Tkalčić (2002, Geoph. Monogr); Capdeville, Romanowicz and Toh (2003, in press GJI); Romanowicz (2003, Annu. Rev. Geoph.); Gung, Panning and Romanowicz (2003, Nature); Gung and Romanowicz (2003, Geophys. J., submitted);

Graduate students and post-doc's supported at least partially by these grants: X. D. Li; E. Clévédé; J. Durek; L. Bréger; C. Mégnin; C. Kuo; H. Tkalčić; Y. C. Gung; M. Panning; A. Toh.
Undergraduate students involved in these projects: C. Ng; S. Rousset; N. Houy.

PROJECT PLAN

Background

With its precambrian center surrounded by progressively younger geological provinces (Figure 1), the North American continent is, in many ways, an ideal region to investigate timely geophysical questions such as the relation to geological age of the variations in thickness of continental lithosphere , the nature of the lithosphere/asthenosphere boundary, the relation of upper-mantle anisotropy to present-day flow and/or past tectonic events. The backbone permanent network component of USArray, complemented by temporary BigFoot deployments, will provide an unprecedented density of recordings to address these questions. In anticipation of this unique dataset, it is time to develop theoretical and data analysis methodologies that will allow an effective exploitation of the broadband waveforms that will be collected. A central issue is that of measurement and interpretation of seismic anisotropy in the upper-mantle beneath the continent. Indeed, upper mantle anisotropy is usually attributed to lattice preferred orientation (LPO) of olivine and pyroxene (e.g. Estey and Douglas, 1986) and is an indicator of either paleo or recent deformation (e.g. Park and Levin, 2002; Zhang and Karato, 1995), which may be complicated by other factors such as water content (e.g. Jung and Karato, 2001). We focus this proposal on investigation of 3D upper mantle anisotropic structure at the scale of the north American continent.

Different kinds of seismic anisotropy are observed in the upper mantle and have been documented over the last 40 years. Discrepancies in the dispersion measurements of Love and Rayleigh waves have been attributed to polarization or "radial" anisotropy (e.g. Anderson, 1961), which can be described by 5 anisotropic parameters (A, C, F, L, N, "Love parameters"), and is characterized by a *vertical* symmetry axis. On the other hand, azimuthal anisotropy, originally observed from the analysis of Pn velocities in the oceanic mantle (Hess, 1964), has also been found in surface waves (e.g. Forsyth, 1975). Both kinds of anisotropy observed in surface wave data can be simultaneously modelled by introducing 13 anisotropic parameters (Montagner and Nataf, 1986). Another manifestation of upper mantle anisotropy comes from the observation of splitting of SKS phases (e.g. Vinnik et al., 1984). The latter is generally interpreted in terms of anisotropy with a *horizontal* symmetry axis. Under the assumption of weak anisotropy and a horizontal symmetry axis, it is possible to relate in a simple way, the SKS splitting to surface wave azimuthal anisotropy (Montagner et al., 2000). Surface wave measurements have the advantage of providing depth resolution, but have limited lateral resolution, whereas shear wave splitting measurements have practically no depth resolution, but provide spatial resolution. Combining the advantages of both types of measurements is therefore desireable, and the relatively dense distribution of the proposed USArray backbone (and eventually BigFoot) deployments should lend itself particularly well for this purpose.

Until now, in north America, seismic studies involving the depth range appropriate for the investigation of anisotropy relevant to lithosphere/asthenosphere processes at the continental scale have been primarily of two kinds: 1) global or regional scale tomography and 2) the analysis of shear wave splitting.

Elastic tomographic models, whether constructed from global (e.g. Montagner and Tanimoto, 1991; Masters et al., 1996; Ritsema et al., 1999 ; Mégnin and Romanowicz, 2000a; Ekström et al., 1997; Grand, 2002) or regional datasets (e.g. Romanowicz, 1979; Grand, 1994; Alsina et al., 1996; van der Lee and Nolet, 1997) consistently exhibit faster than average upper-mantle velocities under the cratonic and stable platform part of the continent to depths of at least 200 km and document the presence of relatively sharp lateral transitions both at the western (rocky mountain front, e.g. Romanowicz, 1979; Grand, 1994)) and eastern (e.g. Li et al., 2003a; van der Lee, 2002) edges of the precambrian shield. At the global scale, Tanimoto and Anderson (1984) and Laske and Masters (1998) have considered azimuthal anisotropy (and the latter have illustrated how surface wave polarization data can be used as constraints), whereas Montagner and Tanimoto (1991) have taken into into account both radial and azimuthal anisotropy. Most of the regional studies are based on the analysis of Rayleigh wave dispersion in the framework of elastic isotropy, with some recent efforts at also mapping azimuthal anisotropy (e.g. Li et al, 2003ab).

Figure 1. Geological Provinces of North America. Red: Precambrian; Pink: Platform cover; brown: Paleozoic; blue: Mesozoic-tertiary; yellow: passive Margins; green: tertiary/quaternary volcanic rocks *From Bally et al., 1989).*

The primary tool used so far, in north America, to study upper mantle anisotropy has been the analysis of shear wave splitting. Many SKS splitting measurements have been performed (e.g. Vinnik et al., 1989; Silver and Chan, 1988; Silver, 1996; Savage et al., 1996; Barruol et al., 1997; Levin et al., 1999; Savage and Sheehan, 2000; Fouch et al., 2000). These and other such measurements around continental areas in the world have long fueled a lively debate concerning the depth range of origin of the detected splitting: is it related to "frozen anisotropy" in the continental lithosphere, marking the deformation associated with the last tectonic event (e.g. Silver, 1996; Rondenay et al., 2000), or is it located deeper, and marks the shear imposed by present day plate-motion on the asthenosphere and/or the base of the continental lithosphere (e.g. Vinnik et al., 1992; Bormann et al., 1996; Savage et al., 1996; Savage and Sheehan, 2000)?. In north America, resolving this issue using only shear wave splitting is made more difficult by the fact that in central and eastern US, the fast direction of anisotropy deduced from shear wave splitting is very similar under both hypotheses. On the other hand, rapid variations of the fast axis direction have been documented in the western U.S. (e.g. Savage et al., 1996).

Several recent studies indicate that both interpretations may, in fact, be right: SKS splitting measurements may contain signal from both depth ranges and may result from the combination of relatively shallow frozen anisotropy in the continental lithosphere, as well as deeper anisotropy related to asthenospheric flow (e.g Levin et al., 1999; Fouch et al., 2000). In particular, Fouch et al. (2000) performed quantitative modelling of the flow around the keel of the north American continent and predict fast directions of anisotropy in agreement with the observed SKS splitting in the central/Eastern US (Figure 2). Measurements of surface wave azimuthal anisotropy may also reflect a combination of causes (e.g. Simons et al., 2001).

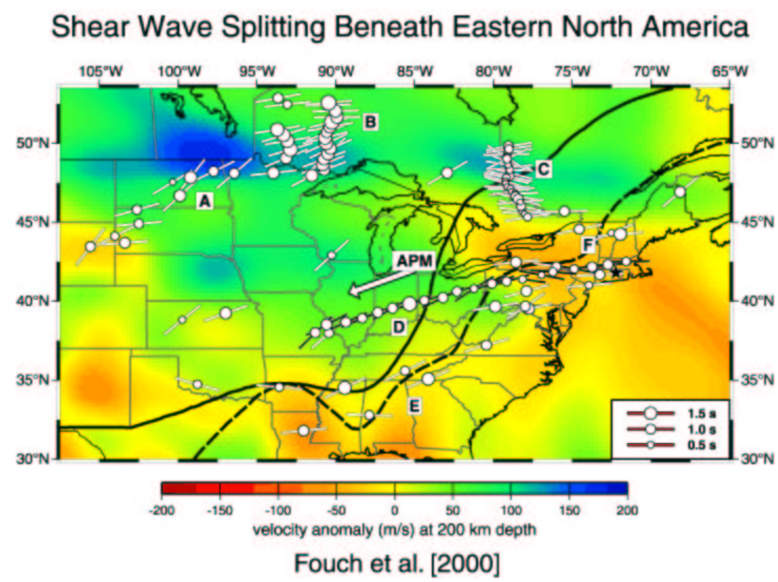

Figure 2. Shear wave splitting parameters for permanent and temporary stations in eastern North America. APM gives the direction of absolute plate motion (*from Fouch et al., 2000*).

In a recent study of radial anisotropy in the upper mantle at the global scale, using a waveform tomographic approach, we have shown evidence for the presence of significant radial anisotropy, with $V_{sh} > V_{sv}$ in the depth range 200-400 km beneath continental cratons (Gung et al., 2003), and in particular beneath the Canadian shield (Figure 3). The style of this anisotropy is similar to that observed at shallower depth under the ocean basins (Montagner and Tanimoto, 1991; Ekström and Dziewonski, 1998), and we suggested a similar interpretation under oceans and continents in terms of shear in the sub-lithospheric asthenosphere: the mapping of radial anisotropy would thus allow us to map the depth variations of the lithosphere/asthenosphere boundary around the world (Figure 4). A similar interpretation was also proposed by Plomerova et al. (2002).

We also suggested that the Lehmann discontinuity may represent the boundary between a weakly anisotropic lower lithosphere and a more strongly anisotropic asthenosphere, in agreement with the interpretation of Leven et al. (1981), but opposite to that of Gaherty and Jordan (1992) and Karato (1992), who suggested that Lehmann marked the bottom of the anisotropic upper mantle region, possibly marking a transition from dislocation to diffusion creep. On the other hand, our interpretation also provides an explanation as to why the Lehmann discontinuity is preferentially detected under stable continental areas (e.g. Revenaugh and Jordan, 1991; Gu et al., 2001) but not systematically detected globally (e.g. Shearer, 1990): under ocean basins, the lithosphere/asthenosphere boundary is located at much shallower depths ($\sim 80-100km$) and is marked by the Gutenberg discontinuity (e.g. Revenaugh and Jordan, 1991).

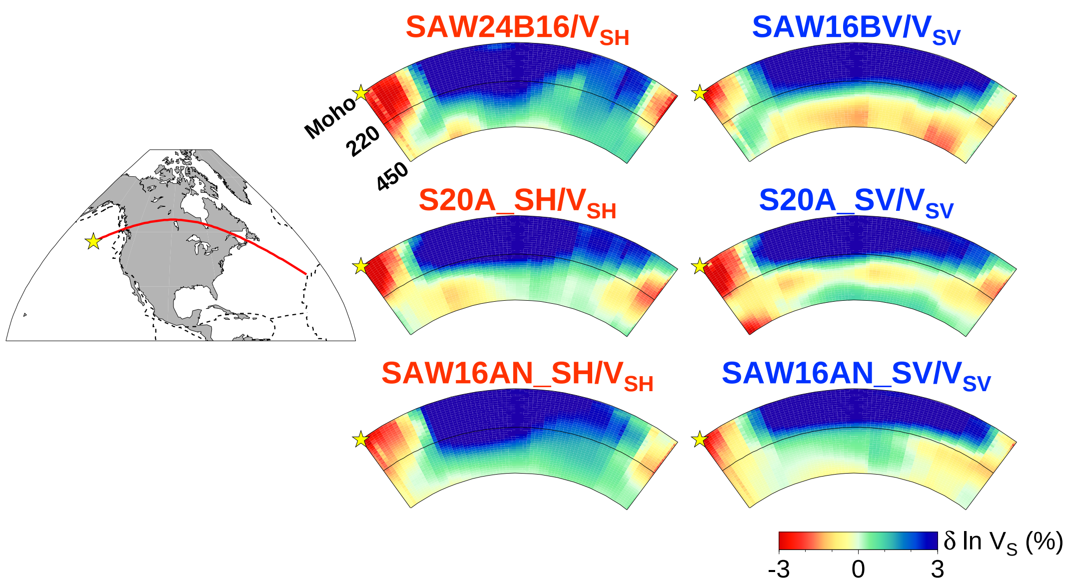

Figure 3. Depth cross-sections across the Canadian shield showing the SH (left) and SV(right) parts of several global S velocity models: SAW24B16 (Mégnin and Romanowicz, 2000a), S20RTS (Ritsema et al., 1999), S20A (Ekström and Dziewonski, 1998), SAW16AN (Gung et al., 2003). The SH sections generally indicate fast velocities extending to depths in excess of 220 km, whereas the SV sections do not.((*adapted from Gung et al., 2003*).

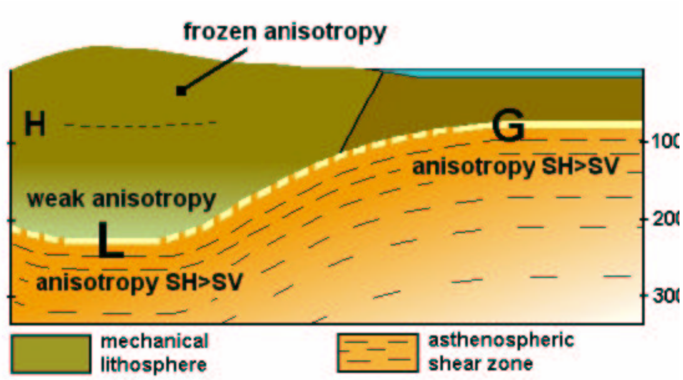

Figure 4. Sketch illustrating our interpretation of the observed radial anisotropy in relation to lithospheric thickness, and its relationship to the Lehmann (L) and Gutenberg (G) discontinuities. The Hales discontinuity (H) is generally observed as a positive impedance embedded within the continental lithosphere in the depth range 60-80 km (Levin and Park, 2000) and may bear no relation to G(*from Gung et al., 2003*).

The presence of anisotropy under stable continents in the depth range 200-400 km explains discrepancies, in this depth range, between upper mantle tomographic models based primarily on transverse component data, dominated by Love waves (e.g. Li and Romanowicz, 1996; Mégnin and Romanowicz, 2000a) and those dominated by Rayleigh wave data (e.g. Ritsema et al., 1999). In particular, continental lithospheric thickness is at most on the order of 200-250 km, in agreement with other geophysical data (Rudnick et al., 1998; Jaupart et al., 1998; Hirth et al., 2000). This view is also consistent with regional observations under the Australian shield (Debayle and Kennett, 2000). There is then no need to invoke the existence of a "tectosphere" that would translate coherently with continental keels, a controversial notion that has been the subject of debate in the last 25 years (e.g. Jordan, 1975; Anderson, 1979).

The notion that the continental lithosphere is not strongly coupled to the thermal structure of the underlying mantle has also received support recently from the absence of observation of significant topography of the 400 km discontinuity under the north American continent (e.g. Bostock, 1996; Li et al., 1998). There is also some evidence, at least locally, for the presence of a low velocity zone in shear waves under the north American craton (Rodgers and Bhattacharyya, 2001; van der Lee, 2002).

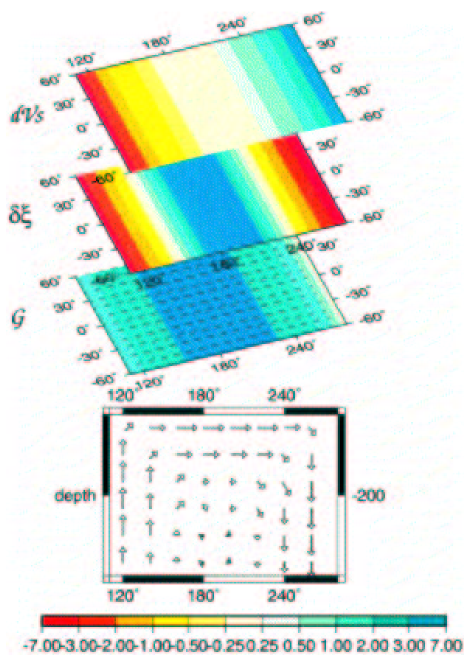

Figure 5. Seismic parameters V_s, ξ, G, Ψ_G as considered by Montagner (2002) to describe the effects of a simple convecting cell in the upper mantle, assuming LPO of anisotropic minerals such as olivine. A vertical flow is characterized by a negative ξ radial anisotropy and a small azimuthal anisotropy (G), and a $\delta V_s > 0$ or $\delta V_s < 0$, respectively, for an upwelling or downwelling. A predominant large-scale horizontal flow corresponds to a significant amplitude of the azimuthal anisotropy G and $\delta\xi > 0$, and its orientation will reflect the direction of flow (*from Montagner, 2002*).

Modelling of radial anisotropy as we have done, so far relies on several important assumptions: it ignores the contribution of azimuthal anisotropy and assumes a vertical axis of symmetry for the anisotropic structure, and thus does not allow a comprehensive analysis of the relation of anisotropic pattern to flow. Essentially, all that can be said is that the detection of significant

$dln\xi > 0$ (where $\xi = (V_{sh}/V_{sv})^2$) is indicative of the presence of significant horizontal shear, such as might be expected under the "flat" portion of continental keels (Figure 5). Moreover, contamination by azimuthal effects cannot be ruled out, unless sufficient azimuthal coverage is available, which is only possible at very long wavelengths and on the global scale. To gain a more accurate view of anisotropic structure, one must analyze both radial and azimuthal anisotropy simultaneously, using three component data.

To date, published work on surface wave derived other-than-radial anisotropy on the north American continent, only addresses azimuthal anisotropy using Rayleigh waves (Li et al., 2003ab). These authors have detected weak azimuthal anisotropy in the eastern U.S.. The most complete study remains the global anisotropic model of Montagner and Tanimoto (1991), who used a technique called "vectorial tomography" developed originally by Montagner and Nataf (1988) and first applied to the Indian Ocean (Montagner and Jobert, 1988), which we will discuss further below. The dataset considered by these authors only included fundamental mode Rayleigh and Love wave dispersion data, so that their depth resolution was poor below about 250-300km. Nevertheless, under a simplifying hypothesis of radial anisotropy with a symmetry axis of arbitrary orientation (corresponding to 7 anisotropic parameters: the 5 Love parameters plus two angles to define the orientation of the symmetry axis), and ignoring non-linear terms, they were able to map significant lateral variations on the global scale of both radial anisotropy (parameter ξ) and of the orientation of the fast axis down to 250 km depth. Their model showed the systematic alignement of the fast axis perpendicular to ridges in young oceans, and a correlation of the style of anisotropy with geologic province in the continental lithosphere (Babuśka et al., 1998), as well as the predominance of $dln\xi > 0$ in ocean basins both in the Pacific and Indian Ocean.

A better understanding of the depth distribution and orientation of anisotropy in the lithosphere/asthenosphere depth range under continents (i.e. down to at least 400 km) is important, to address such questions as the nature and strength of lithosphere/asthenosphere coupling and the driving mechanisms of plate motions (e.g. Bokelmann, 2002ab, Bokelmann and Silver, 2002). In order to improve the resolution of the distribution of anisotropy at the continental scale, both laterally and vertically, we propose to develop a methodology to analyze three component waveform data, that would not only include fundamental mode surface waves but also exploit the information contained in overtones and body waves, and would take into account both radial and azimuthal anisotropy. In what follows, we describe the proposed approach and discuss the simplifying assumptions that will be necessary to make the problem tractable.

Theoretical considerations

Waveform tomography

In order to take advantage of the complementary sampling and depth sensitivity ,to structure in the mantle, of surface waves and body waves, it is desireable to combine them and work with waveforms. Here, we propose such an approach to study anisotropic structure in the upper mantle beneath North America.

In the past 10 years, we have introduced a waveform tomographic approach based on asymptotic normal mode coupling theory (Li and Tanimoto, 1993) which allows us to invert entire long period seismograms (including fundamental mode, overtone and body wave portions of the record) for global three dimensional elastic structure. Traditional approaches in global tomography combine surface wave data (waveforms or dispersion measurements) treated under the "path-average" approximation (PAVA, e.g. Woodhouse and Dziewonski, 1984) with body wave travel time data, treated using ray theory. The PAVA is a zeroth order asymptotic approximation which is equivalent to including coupling of modes along a single dispersion branch (Park, 1987; Romanowicz, 1987). It assumes that the surface waves are sensitive to the average 1D structure along the source-receiver great circle path. Ray theory assumes that the sensitivity of body waves is restricted to the infinitesimal ray path and uniformely distributed along the ray. The advantage of Non-linear Asymptotic Coupling theory (NACT, Li and Romanowicz, 1995) over traditional PAVA on the one hand, and infinite frequency "ray theory", on the other, is that we consider broadband kernels that more accurately reproduce the sensitivity of body waveforms to structure along and around the ray-path. It allows the incorporation of diffracted phases, and it also allows to consider wavepackets that contain several phases arriving quasi-simultaneously, but with different sampling of the mantle (For example SSS and ScS_2), thus providing a way to include more information in the inversion than contained in isolated body wave phases for which travel times can be measured. In fact, this body-wave character of the kernels is also important to accurately model the waveforms of surface wave overtones (e.g. Figure 6).

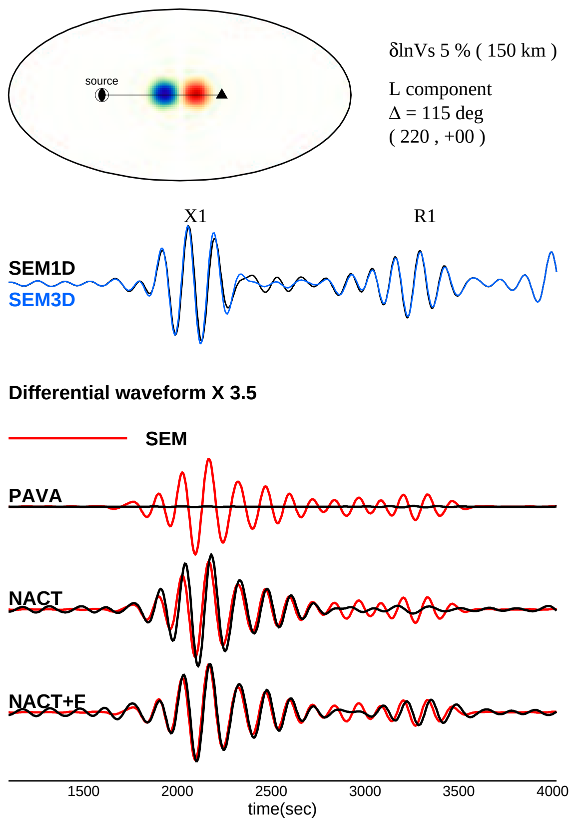

Figure 6. Comparison of synthetic seismograms for the heterogeneous structure shown in the top panel. The structure is centered at 150km depth and each S velocity anomaly is approximately 2000 km diameter horizontally and 200 km vertically, with strength ±5%. In these examples, the seismograms are computed down to a period of 100 sec, at a distance of 115^o, for a thrust event source, on the longitudinal component. The seismograms at the top compare the SEM (spectral element method) calculation for the 3D structure to the reference 1D seismogram (plotted with a small vertical offset). The other seismograms are magnified differential seismograms, comparing the 3D contribution in SEM (red) with that obtained under 3 successive approximations, from top to bottom: PAVA, NACT and NACT+focusing. This figure illustrates how adding across-branch coupling (NACT) improves the agreement with SEM for the higher mode wavepacket (X1), whereas for this particular geometry, inclusion of focusing improves the fit for the fundamental mode Rayleigh wave.

In the asymptotic formalism, the interactions of modes with structure appear in integrals over the source-receiver great circle, which is arrived at using the stationary phase approximation at several levels of approximation (Romanowicz and Roult, 1986). This approach is valid theoretically for teleseismic data and smoothly varying structures. It provides an improvement with respect to standard zeroth-order approaches, while keeping the numerical computations much less time consuming than more exact numerical computations such as can now be obtained with the Spectral Element Method (Komatitch and Vilotte, 1998; Komatitch and Tromp, 1999).

We have developed several generations of global S velocity models using NACT 2D sensitivity kernels (e.g. Li and Romanowicz, 1996; Mégnin and Romanowicz, 2000A; Gung and Romanowicz, 2003). More recently, we have also included effects of off-plane propagation (focusing effects), which are particularly important for the study of attenuation (e.g. Romanowicz and Gung, 2002) and involve the expansion of asymptotic mode expressions to higher order (order $1/l$ where l is the angular order of a normal mode, Romanowicz, 1987; Park, 1987).

We have also extended the elastic isotropic NACT formalism to include radial anisotropy with a vertical axis of symmetry and constructed a global upper mantle model of lateral variations of isotropic S velocity Vs, and the anisotropic parameter ξ, as discussed earlier (Gung et al., 2003). We restricted the number of physical parameters in the inversion to those two which our data are most sensitive to (isotropic S velocity V_{siso} and $\xi = (V_{sh}/V_{sv})^2$, by considering relations between the different anisotropic parameters as appropriate for upper mantle minerals (Montagner and Anderson, 1989).

In order to extend our approach to a more general case involving both radial and azimuthal anisotropy, we need to consider the theoretical expressions giving the perturbations to the seismogram for the general case of anisotropy. The various elements needed in the framework of normal mode theory are available in the literature from the work of Tanimoto (1986), Mochizuki (1986) and Park (1997). The mode coupling terms were also derived in a form suitable for application to NACT in Romanowicz and Snieder (1988).

In the Born approximation, the3D perturbation $\delta u(t)$ to the seismogram calculated in a spherically symmetric earth can be written:

$$\delta u(t) = Re\left(\Sigma_{k,k'} \frac{exp(i\omega_k t) - exp(i\omega_{k'} t)}{\omega_k^2 - \omega_{k'}^2} \times \Sigma_{mm'} R_{k'}^{m'} Z_{kk'}^{mm'} S_k^m\right) \tag{1}$$

where k, k' are mode indeces, with azimuthal orders m, m' and ω_k, $\omega_{k'}$ their respective spherical eigenfrequencies, $R_{k'}^{m'}$ and S_k^m are receiver and source terms, and $Z_{kk'}^{mm'}$ is the interaction matrix which contains the structure perturbation terms.

In the isotropic case, the interaction of the modes with the structure can be described through the introduction of a "local frequency" $\delta\omega_{k,k'}$ (e.g. Jordan, 1978; Woodhouse and Girnius, 1982), which includes the perturbation to the 3 elastic parameters (ρ, Vs, Vp) (as well as interface perturbations which we will not explicitely write here). If we define:

$$A_{kk'} = \Sigma_{mm'} R_{k'}^{m'} Z_{kk'}^{mm'} S_k^m \tag{2}$$

we obtain an expression of the form:

$$A_{kk'} = \Sigma_{N,M} R_{k'N} S_{kM} \int\int \delta\omega_{k,k'}(\lambda, \mu) Y_l^{\prime N,0}(\lambda) Y_l^{M,0}(\lambda') d\Omega \tag{3}$$

where k and k' are mode indeces, $R_{k'N}$ and S_{kM} are receiver and source terms as defined in Woodhouse and Girnius (1982), and $Y_l^{M,0}$ are generalized spherical harmonics (Phinney and Burridge, 1972). The integration is over the unit sphere, and :

$$\delta\omega_{k,k'}(\lambda,\mu) = \int_0^a \left(\frac{\delta V_s}{V_s} M_s^{k,k'}(r) + \frac{\delta V_p}{V_p} M_p^{k,k'}(r) + \frac{\delta\rho}{\rho} M_r^{k,k'}(r) \right) r^2 dr \tag{4}$$

The radius of the earth is a, $M_x^{k,k'}(r)$ are depth dependent coupling kernels, and the angles $\lambda, \lambda' and \mu$ are defined in Figure 7.

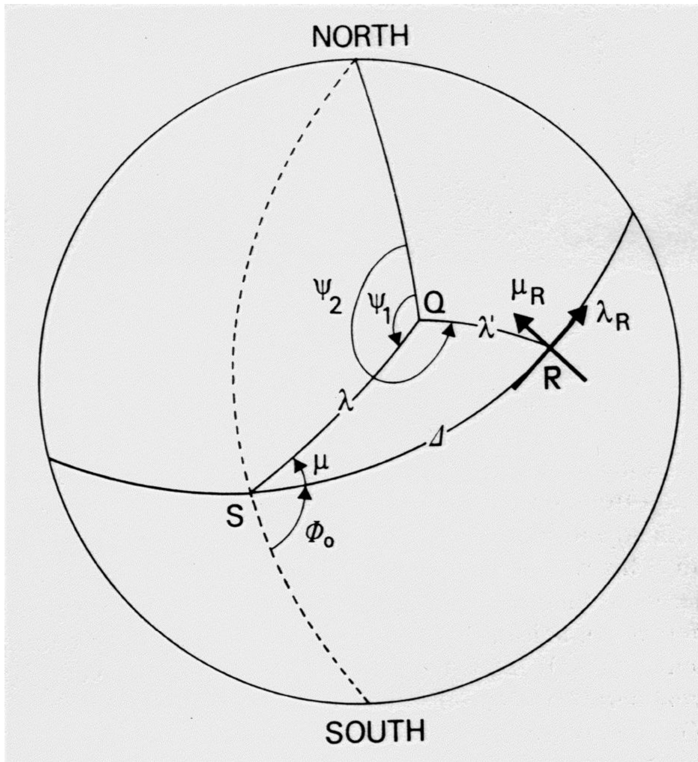

Figure 7. Geometry considered. S is source location, R receiver location. Q is the running integration point of epicentral coordinates (λ, μ).

This can be generalized to the more general case of anisotropy, as described by 21 independent elements of the elastic tensor (e.g. Romanowicz and Snieder, 1988). The perturbation to the seismogram then involves the introduction of additional local frequencies $\omega^j_{k,k'}(\lambda,\mu)$, which contribute to the modified expression of $A_{k,k'}$ as follows:

$$A_{k,k'} = \Sigma_j \left(\Sigma_{N,M} R_{k'N} S_{kM} \int\int \delta\omega^j_{k,k'}(\lambda,\mu) Y_l'^{N,p}(\lambda) Y_l^{M,q}(\lambda') d\Omega \right) \tag{5}$$

where integer indeces p, q take the values $\pm 0, \pm 1, \pm 2$. The local frequencies now depend on the cosine and sine of linear combinations of angles Ψ_1 and Ψ_2 as defined in Figure 7. As shown in Romanowicz and Snieder (1988), in the zeroth order asymptotic case (case of coupling along a mode branch), the angle dependence reduces to a dependence on cosine and sine of even multiples (0, 2, 4) of the scattering angle $\Psi = \Psi_2 - \Psi_1$ and lead to expressions equivalent to the surface wave expressions of Smith and Dahlen (1973), which give the dependence of phase velocity (directly related to local frequency of the corresponding mode) as a function of the azimuth along the ray path (along the great circle Ψ becomes the azimuth, Figure 7):

$$c(\theta,\phi,\Psi) = c_0(\theta,\phi) + c_1(\theta,\phi)cos2\Psi + s_1(\theta,\phi)sin2\Psi + c_2(\theta,\phi)cos4\Psi + s_2(\theta,\phi)sin4\Psi \tag{6}$$

where c_0 is the azimuth independent part of phase velocity and c_i, s_i are expressed in well-known combinations of elastic parameters.
In the most general case of anisotropy, each element of the elastic tensor $C^{\alpha\beta\gamma\delta}$ (described in contravariant components as appropriate to the use of generalized spherical harmonics) contributes to a local frequency, for a total of 21 independent terms. In the asymptotic limit, we obtain expressions which depend on combinations of the elastic tensor multiplied by $cos(n\Psi) and sin(n\Psi)$, where Ψ is the scattering angle, and the integer n can take the values 0,1,2,3,4. This also includes coupling between spheroidal and toroidal modes which can give rise to the observation of quasi-Love waves on the vertical component and quasi-Rayleigh waves on the transverse component (e.g. Yu and Park, 1994). Asymptotically, we obtain expressions that are equivalent to those derived using ray theory for surface waves (Woodhouse and Wong, 1986; Larson et al., 1998) and can implicitely account for polarization anomalies induced by anisotropy (e.g. Laske and Masters, 1998).

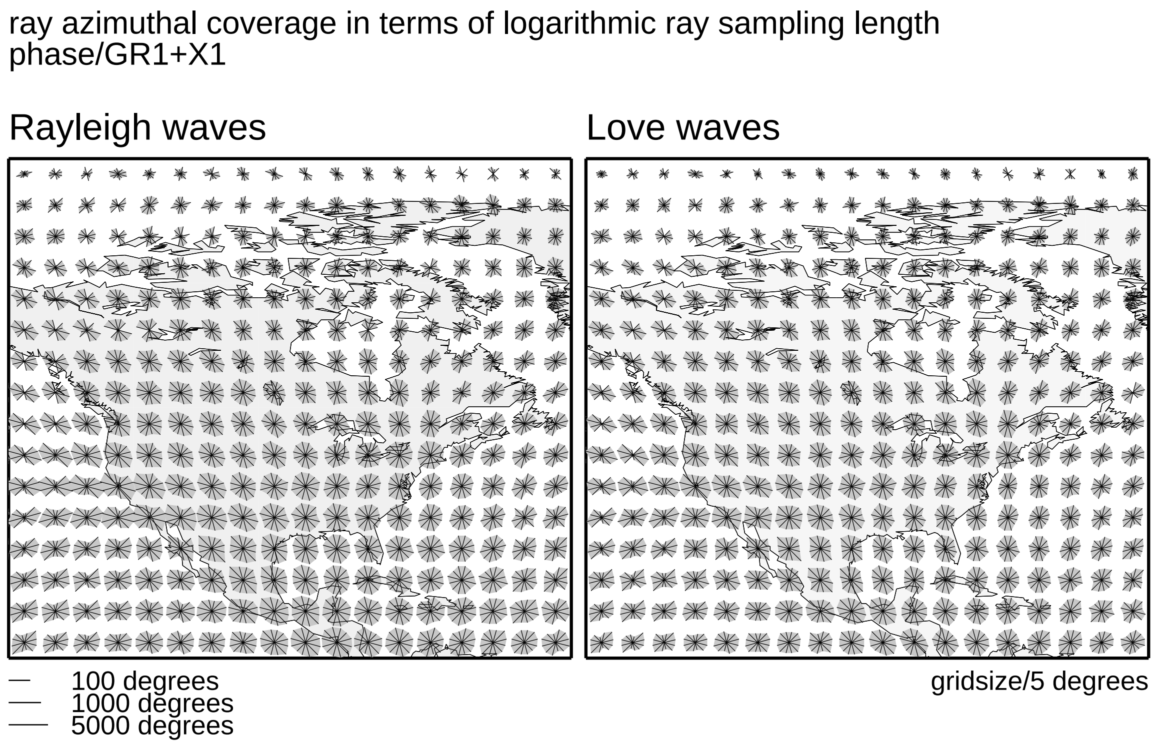

Figure 8. Azimuthal coverage available on minor arc source-station paths crossing north America in our global dataset fundamental mode and overtone waveforms. The region is divided into 5×5^o elements, and in each of these, each segment length is proportional to the sampling in the corresponding azimuthal section (each section is 30^o wide). The scale is in log_{10}(sampling ray length) where "samping ray length" refers to the cumulative path length (in degrees) in the corresponding block and azimuth range.

Clearly, resolving all 21 elements of the elastic tensor and their spatial and depth variations is beyond our present or near-future capabilities. A manageable case, which is also physically realistic given what we know about the properties of mantle minerals, is to consider a model of anisotropy consisting of radial anisotropy with an axis of symmetry of arbitrary orientation (Montagner and Nataf, 1988; Park, 1997). Such a model can be most generally described by the 5 Love parameters (plus density) and two angles, describing the orientation of the axis of symmetry (Θ, Φ).

Park (1997) gives the relations between these 7 parameters and the elastic tensor elements $C^{\alpha\beta\gamma\delta}$ in the contravariant notation which we use to define $\omega^j_{k,k'}$ above. The corresponding expressions for $\omega^j_{k,k'}$ depend non-linearly on $x = cos(2\Theta)$ (through dependence on x^2). Montagner and Nataf (1988) provided arguments for ignoring the non-linear terms in the case of the earth's upper-mantle. We propose to examine the importance of the non-linear terms in the framework of our application to long period waveforms and NACT formalism.

Just as in our global waveform tomography, we will first consider frequencies down to 32 sec and later extend to 16 sec. If we find that we cannot neglect the non-linear terms, we will include them accordingly in a non-linear iterative inversion scheme. We will also assume realistic relations between the anisotropic parameters (and also density), and in particular explore those proposed by Montagner and Nataf (1988) and Montagner and Anderson (1989), which will lead us to reduce the number of parameters in the inversion. Even though this is at least twice as many parameters to solve for than we have ever considered so far at the global scale, if we restrict this parametrization to the well resolved part of north America, we anticipate that the problem should be tractable, assuming a good azimuthal distribution of data.

To illustrate this assertion, we show in Figure 8 the azimuthal coverage available over north America for our current global data set (as used in our most recent global models). This coverage comprises all the source-station direct paths crossing north America, for which we have fundamental mode (Rayleigh or Love) surface wave data as well as X_1 overtone wavepackets.

This project derives from work we have been involved in independently in global tomography. We plan to start from our current global elastic radially isotropic models, for which we have two versions: one with anisotropy restricted to the upper mantle (Gung et al., 2003) and another, with radial anisotropy in the whole mantle, the development of which is under construction (Panning and Romanowicz, 2003), supported by an active NSF grant. Parametrization within north America will be expanded to allow for anisotropy with an axis of symmetry with an arbitrary direction. This will necessitate reparametrizing our models in terms of local basis functions, such as spherical splines (they are currently parametrized in terms of spherical harmonics up to degree 16). We are in the process of developing such a procedure for our global models independently of this proposal, and it will be operational by the time of the start of this project. The advantage of the regional focus, not addressed in our other work centered on global modelling, is that the spatial extent limited to the continent combined with the density of data to be collected in the framework of Earthscope (specifically, for the time of duration of the proposed work, the USArray/NSN backbone network, later to be extended to BigFoot data) allows us to investigate a more complete anisotropic parametrization.

The upper mantle model under north America will be progressively updated, as additional data to north American stations become available. To give a sense of the potential dataset, we show, in Figure 9, the distribution of events of $M > 6$ surrounding the north American continent over the last 12 years (since 1990), and the distribution of seismicity as a function of time over the last 6 years, indicating that western and southern azimuthal coverage can be achieved over a period of 2 years, but eastern azimuths are less well sampled and require longer times. Figure 10 shows the present distribution of broadband stations in north America. Not all these stations presently provide high quality seismograms, however, we expect that this will soon be corrected with a better dataflow from the NSN to the IRIS-DMC, as well as through progressive upgrade of existing NSN stations to USArray/ backbone standards. Figure 11 gives an example of 3 component recording of the quality that we would expect to utilize. Finally, Figure 12 shows two examples of copmparisons of observed and synthetic waveforms, filtered with a corner frequency of 80 sec, as we currently use for fundamental modes and overtones in our global modelling effort), showing

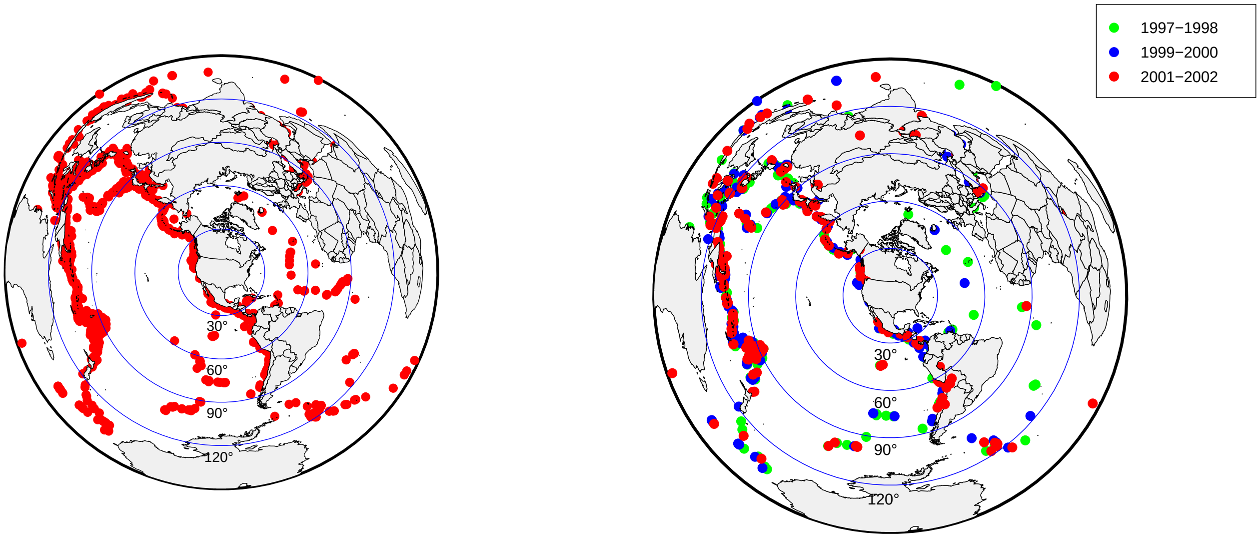

Figure 9. Map centered on north America, showing the distribution of M¿6 teleseismic events. Left: during the last 12 years; Right: in intervals of 2 consecutive years, for the last 6 years. The Pacific rim is well sampled in any given time period, and the red dots corresponding to 2001-02 cover some of the seismicity from previous years.

the improvement of fit in the 3D global models.

The examples shown emphasize the surface wave (fundamental mode and overtone) part of the dataset, for which we have been working in the band-pass (80-250 sec) in our global modelling (e.g. figure 12). The first stage of our proposed data analysis will consist in applying our extended NACT anisotropic formalism to this type of data. We will, in particular extend the bandpass down to 60 sec. Below 60 sec, effects of multipathing start to be important and we cannot accommodate those. Our formalism allows for inclusion of body wavepackets, which we filter with a low-pass corner frequency of 32 sec in our current global models. Increasing the band-pass to shorter periods represents significant additional computation time in the global case, however, for a limited region like north America, we can accommodate a corner frequency of 16 sec without problem and we will consider even shorter periods (i.e. 10 sec). We are particularly interested in including SKS waveforms (in the distance range 80 to 120 deg.), because the splitting of SKS waves can be readily attributed to the upper mantle beneath the receivers. We will also consider other body waveforms containing multiple S arrivals, however their analysis will require careful consideration of anisotropic effects outside of the region of study.

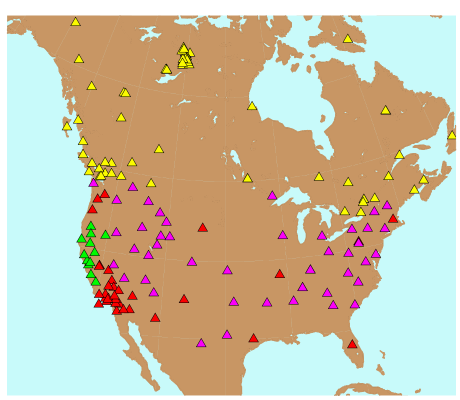

Figure 10. Current distribution of broadband stations in north America. Canadian broadband array (yellow); northern (green) and southern (red) California broadband networks; National Seismic Network (purple). GSN stations are also shown in red.

Project summary timeline

During the first year of the project, we will focus on finalizing the theoretical expressions for the perturbation to the seismogram in the NACT framework in a anisotropic model with a symmetry axis of arbitrary orientation, as well as explore realistic assumptions on relations among elastic parameters that could further simplify the parametrization, and in particular examine the non-linearities involved. We will adapt our waveform modelling code to this case and test it on our existing dataset of surface waves and overtones on paths covering north America. We will also start collecting additional data at north American stations, and in particular extending the bandwidth to 60 sec.

In the second year, we will extend our data collection as the backbone network upgrade proceeds and focus on including body waveforms (with extension to shorter periods) in the inversion and exploring how to include wavepackets other than SKS while minimizing contamination from regions outside of the north American upper mantle. We will also address issues of the accuracy of crustal corrections (for which there are numerous available models, e.g. Mooney et al., 1989; Mooney et al., 1998; Li et al., 2003a;) as well as uncertainties in source parameters.

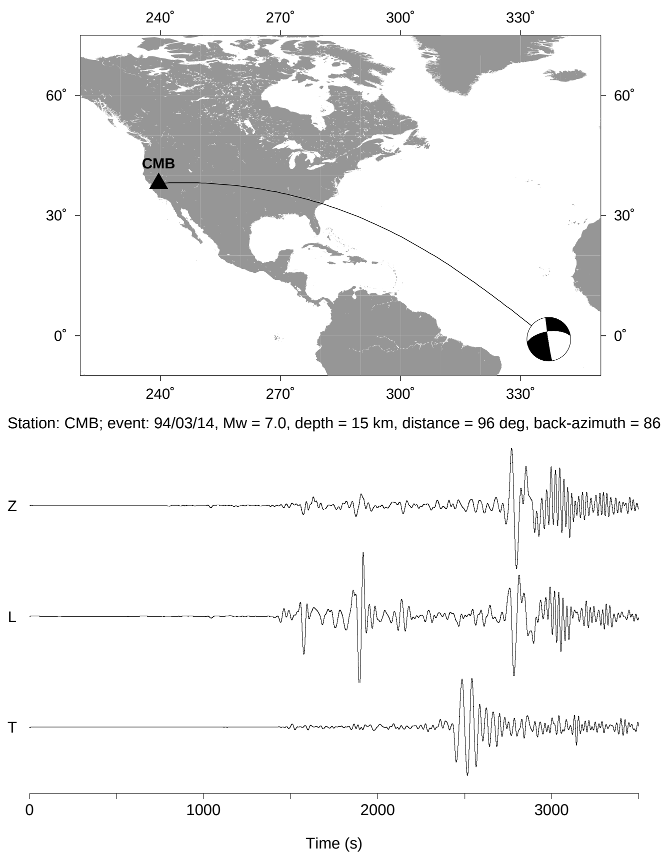

Figure 11. example of 3 component data for an event on the mid-Atlantic ridge (94/03/14, Mw = 7.0) observed at station CMB in California

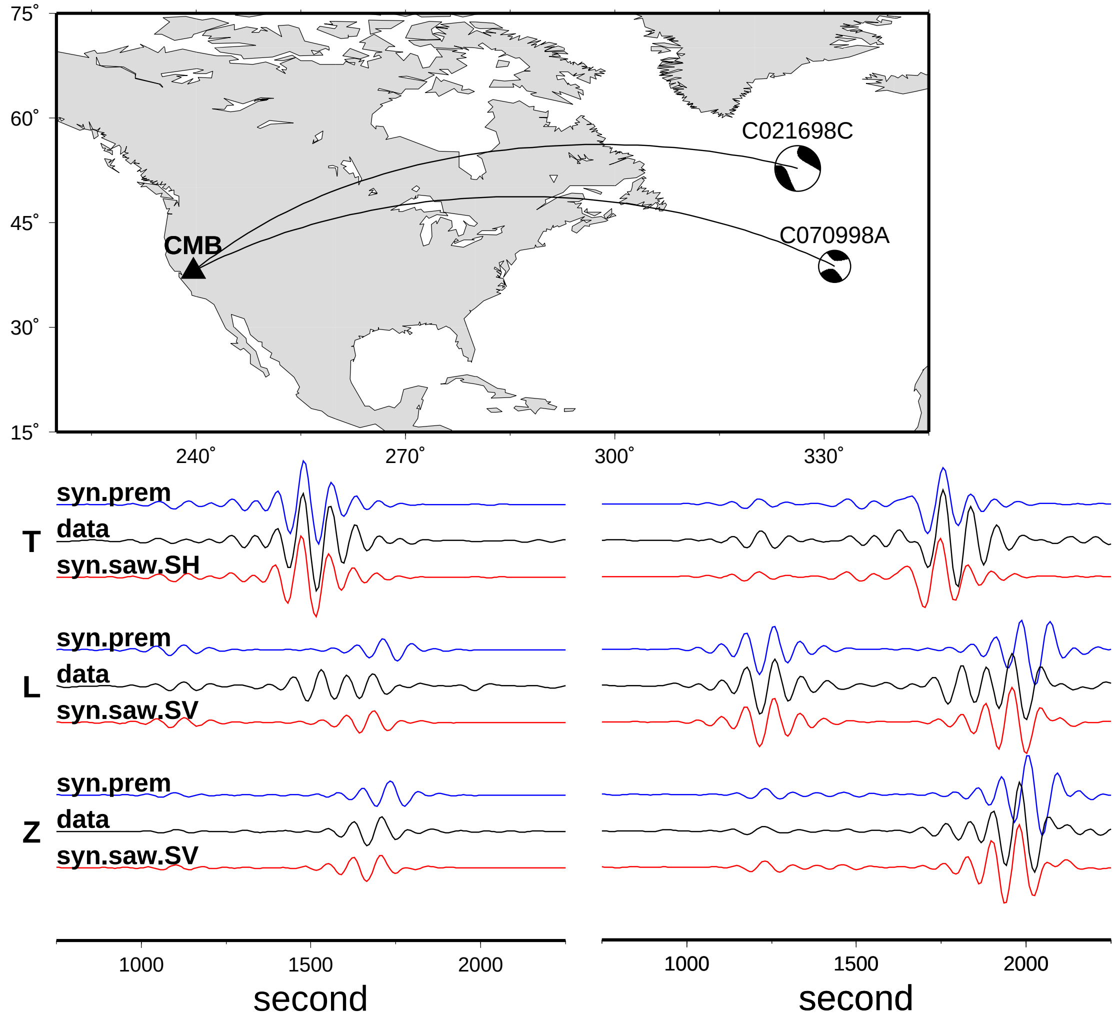

Figure 12. Examples of three component waveform data (black) and corresponding synthetics for PREM (Dziewonski and Anderson, 1981: blue) and our 3D global elastic models (red: SAW24B16 for SH, M‘’egnin and Romanowicz, 2000; SAW16BV for SV, Gung et al., 2003). The data shown are for two events that occurred in the mid Atlantic in 1998 (Ms = 6.6 and 6.0 respectively) observed at station CMB, and have been bandpass-filtered with corner frequencies of 80 and 250 sec.

Broader Impact Statement

The proposed investigation of anisotropy in the upper mantle of north America, which will exploit data collected under the Earthscope program, will benefit a broader community of non-seismologists in geophysics (geodynamicists and mineral physicists), in that it aims at providing seismological constraints on geodynamical models of the earth's upper mantle and may shed light on the relations of strain and seismic anisotropy at the microscopic level. This project will partially support the work of 1 graduate student and one post-doc. Through regularly scheduled discussions within the geophysics group at U.C. Berkeley, this research will contribute to the education of 10 graduate students in seismology and geophysics.