

Results from Prior NSF Support: William Menke. Award Number OCE-11965 Amount \$304,958; Period 08/15/98-11/30/00, Title Active Seismic Imaging of Axial Volcano, PI's William Menke & Maya Tolstoy. The region of Axial Volcano, Juan de Fuca Ridge region provides an excellent opportunity to study the interplay between active "hot spot" and "mid-ocean ridge" magmatic systems. Important questions include how the two magma systems are fed; their magma and heat budgets; the degree of interconnectedness (or interaction) between them; their relationship to seismicity and geodetic strains; the role of each in plate-tectonic spreading and crustal formation; and their effect on the geochemistry (e.g. mixing, fractionation) of erupted basalts. Information on the physical layout of the magma systems is critical to the study of each of these issues. The purpose of this research was to investigate these questions through the tomographic imaging of the region using seismic data from an active seismic airgun-to-obs experiment. The experiment was remarkably successful, both in the sense that voluminous high-quality data were obtained, and in the sense that very clear signals associated with magma were detected in that data. The key elements of the new three-dimensional compressional velocity model of the Axial and Coaxial magma systems are: 1. A Very Large Axial Magma Chamber; A smaller Coaxial Magma Chamber, unconnected with the one at Axial; 3. Several other small low velocity zones are possibly outlier magma chambers from Axial; 4. Strong thickening of the crust beneath Axial volcano. The crust thickens from about 6 km far from Axial to 8 km near Axial to 11 km beneath the summit (West 2001). Publications:

Menke et al., Shallow crustal magma chamber beneath the axial high of the Coaxial Segment of Juan de Fuca Ridge at the "Source Site" of the 1993 eruption, Geology 30, 359-362, 2002.

West, The deep structure of Axial Volcano, Ph.D. Thesis, Columbia University, 2001.

West et al., Magma storage beneath Axial volcano on the Juan de Fuca mid-ocean ridge, Nature 25, 833-837, 2001.

Results from Prior NSF Support: J. Gaherty, EAR-9814565, An Analysis of Upper Mantle Heterogeneity and Anisotropy in Western North America Using Recordings from Broadband Permanent and Temporary (PASSCAL) Seismic Stations; duration 1/99-12/00 (no cost to 12/01), \$80,848. This grant funded several studies of the nature of seismic heterogeneity and anisotropy in the upper mantle. The primary effort has been a three-dimensional tomographic analysis of anisotropic heterogeneity beneath the California region [Gupta and Gaherty, 2000]. Models based on regional earthquakes show that radial anisotropy extends through the lithosphere with an average magnitude of about 2%, and the variability correlates with tectonics. Extension of this model to include azimuthal anisotropy required teleseismic data, which in turn required the development of a new array analysis [Freybourger et al., 2001]. We also investigated the lateral variation in upper-mantle anisotropy by characterizing radial anisotropy in distinct tectonic regimes. We found that models of tectonic and stable North America display over 6% variation in isotropic shear velocity in the upper 200 km, but have virtually identical radial anisotropy over path lengths of >2000 km [Hutko and Gaherty, 2000]. This implies that the significant contrast in thermal and mechanical properties does not correspond to a major difference in large-scale mantle fabric. Finally, we found that anisotropy associated with the northern EPR contrasts sharply with that along the Reykjanes Ridge, which most likely reflects hotspot-fueled buoyant upwelling beneath the ridge [Gaherty, 2001].

Freybourger, M., J.B. Gaherty, and T.H. Jordan, Upper-mantle structure of the Kaapvaal craton from surface-wave analysis, Geophys. Res. Lett., 28, 2489-2853, 2001.

Gaherty, J.B., Seismic evidence for hotspot-induced buoyant upwelling beneath the Reykjanes Ridge, Science, 293, 1645-1647, 2001.

Gupta, P. and J.B. Gaherty, Upper-mantle anisotropy beneath California and the surrounding region, EOS Trans, 81, fall meeting supplement, 2000. (manuscript in preparation).

Hutko, A. and J.B. Gaherty, Upper-mantle anisotropy beneath North America: Comparing active and stable tectonic regimes, EOS Trans, 81, spring meeting supplement, 2000. (manuscript in preparation).

Results from Prior NSF Support: Vadim Levin (with Jeffrey Park) Grant EAR-9805206 (\$248,117), Anisotropy and the Structural Geology of the Tectosphere (07/15/1998-07/01/2001). In this project we have used observations of shear-wave splitting and P-to-S mode conversions in teleseismic body waves to study the depth-dependent anisotropic structure of the lithosphere in a number of locations. We have inferred tectonic mechanisms that are consistent with the structures implied by data. Results obtained in stable continental regions (New England Appalachians and Arabian Shield) lead us to believe that continental lithospheric mantle preserves the rock texture from its last

large-scale tectonic event. In studies of tectonically active regions (Kamchatka and Cascadian subduction zones) we were able to identify likely anisotropic signatures of ongoing geodynamic processes. The grant supported in part the M.Sc. thesis work of H. Yuan, and the following publications (in part or in full):

Levin, V., and J. Park, 2000. *Shear zones in the Proterozoic lithosphere of the Arabian Shield and the nature of the Hales discontinuity*, *Tectonophysics*, v323, pp. 131-148.

Levin, V., W. Menke and J. Park, 2000a. *No regional anisotropic domains in the northeastern US Appalachians*, *JGR* v105, pp. 19029-19042.

Levin, V., J. Park, J. Lees, M. T. Brandon, V. Peyton, E. Gordeev, and A. Ozerov, 2002a. *Crust and upper mantle of Kamchatka from teleseismic receiver functions*, *Tectonophysics*, in press, 2002

Menke, W and V. Levin, *A Waveform-based method for interpreting SKS splitting observations, with application to one and two layer anisotropic Earth models*, submitted to *GJI*, 2002.

Park, J., and V. Levin, *Seismic anisotropy: Tracing plate dynamics in the mantle*, *Science*, 296, 485-489, 2002.

Peyton, V., V. Levin, J. Park, M. Brandon, J. Lees, E. Gordeev, A. Ozerov, *Mantle Flow at a Slab Edge: Seismic Anisotropy in the Kamchatka Region*, *GRL*, 28, pp 379-382, 2001.

Yuan, H., J. Park and V. Levin, *Skidmarks of trench-parallel terrane migration: Subduction-zone anisotropy structure under Corvallis, Oregon*, submitted to *Journal of Geophysical Research*, 2002.

Results from Prior Research- Spahr C. Webb and Wayne C. Crawford, OCE-9819159, Measuring Crustal and Moho Melt Beneath the EPR, 910 N, Using Seafloor Compliance, \$282088.

The compliance technique uses deformation of the seafloor under long period ocean waves to detect regions of low shear modulus associated with melt bodies and is much more sensitive to small regions of melt than conventional seismic techniques. A series of very broad band seafloor seismic instruments (OBS) has been developed for these measurements. In 1999 and 2000, compliance measurements were made at 31 sites on the EPR, including across-axis lines at 9°08'N (just north of the 9°N OSC), 9°33'N and 9°48'N, along-axis measurements between these sites and several sites within the OSC. Our analysis of the across- and along-axis data [Crawford and Webb, in press] reveal an asymmetric lower crustal partial melt zone (shear velocity less than 1.8 km/s) beneath the East Pacific Rise axis between 9° and 10°N, which we refer to as the Central Melt Zone (CMZ). The CMZ is comparable in width to the crustal magma chamber width inferred from seismic tomography but much narrower than the upper mantle melt zone. At 9°48'N, the CMZ is less than 8 km wide and is centered beneath the rise axis. The zone shifts west of the rise axis as the rise approaches the westward-stepping 9°N overlapping spreading center (OSC) discontinuity and is anomalously wide at the northern tip of the discontinuity. The ratio of the shear velocities to compressional velocities suggests that the melt is well-connected in high-aspect ratio cracks rather than in isolated sills. The shear and compressional velocities indicate less than 18% melt in the CMZ on average, but significantly more melt than in the AML. The compliance measurements reveal that the lower crustal melt seen 10 km off axis is isolated from the CMZ that there are Moho melt lenses at several locations both on and off-axis, but do not show a trend in the location of these bodies. These may be sites where the topography of the Moho buoyantly pools melt from over a larger area, as suggested by [Garmany, 1989; Garmany et al., 1994]. The results from this experiment suggests that the mantle melt supply controls the existence and location of crustal melt, but crustal processes provide the final control on the size and shape of the lower crustal melt zone. Our results also suggest that crustal melt may extend much farther off-axis than previously believed, both in the form of a central melt zone that is offset from the rise axis and in separate melt regions that begin several km off-axis. Compliance instrument data has also been used to investigate the mitigation of long period noise sources (infragravity waves and tilt noise) in seafloor seismic data.

Crawford, W.C., S.C. Webb and J.A. Hildebrand, , *Lower crustal melt variability beneath the East Pacific Rise 9°N segment*, *EOS Trans. Am Geophys. Union*, 1999.

Crawford, W.C., and S.C. Webb, *Removing tilt noise from low frequency (<0.1 Hz) seafloor vertical seismic data*, *Bull. Seismo. Soc. Amer.*, 90(4), 952-963, (2000).

Crawford, W.C. and S.C. Webb, *Variations in the distribution of magma in the lower crust and at the Moho beneath the East Pacific Rise at 9-10N*, *Earth Planet Sci. Lett.*, in press, (2002).

Webb, S.C. and W.C. Crawford, *Long period seafloor seismology and deformation under ocean waves*, *Bull. Seismo. Soc. Amer.*, 89(6), 1535-1542, (1999).

Rationale This is a proposal to collect new data on the structure of the lithosphere and underlying

asthenosphere across the passive margin of eastern North America, and to use these data to gain new insights into the processes of continental rifting, initiation of ocean-floor spreading, the subsequent evolution of the continent-ocean transition region. The target is primarily the upper mantle (as contrasted to the crust). Our motivation is that the structure of, and the deformation within the upper mantle are key to understanding both the formative process and the present-day dynamics of passive margins, but that good structural data bearing on that deformation is currently lacking, especially at the characteristic length scales of 250 km and less. Key parts of the new data collection effort are offshore observations made with long-deployment ocean bottom sensors (OBS's). Our contention is that data from both the landward and the seaward side of the margin are needed to constrain the details of its structure. We assert that the significance of a given structural feature can only be properly interpreted when one knows how it behaves across the margin.

Introduction When a continent breaks apart and a new ocean basin forms, the structure of the lithospheric plate in the region of the initial rifting event is extensively modified. The crust is thinned dramatically, much of the lithosphere is removed, and volcanism associated with the rifting alters what is left. This bounding region remains distinct from the new oceanic lithosphere, however, in terms of both geometry and composition of the crust, and the mode of formation of the mantle lithosphere. Despite this distinction, the evolution of the lithosphere at the newly formed passive margin largely follows the path of the oceanic lithosphere to which it is now attached. As the spreading center moves away, the passive margin subsides due to both cooling of the lithosphere and the sedimentary load. If no other tectonic event takes place, the structure of the lithosphere at the passive margin will preserve a record of the continental breakup on the landward side, a record of the oceanic lithosphere evolution on the seaward side, and a distinct structure shaped by the processes of the breakup at the continent-ocean transition. Also, the passive margin will affect the pattern of mantle flow in the underlying asthenosphere through its evolving new geometry.

The shallow subsurface structure of passive continental margins has been explored by numerous active-source seismic experiments, yielding much information about the thinning of the continental crust, the role of volcanism in continental break up, and the evolution of the nascent oceanic basin. Our knowledge of the deeper structure, meaning the mantle lithosphere and the asthenosphere, is poorer. Seismological techniques that provided a wealth of new information about the fine structure of the upper mantle beneath "dry" regions have, to date, been applied to only a handful of locations covered by water. As a result, many first-order questions about the passive margins' structure and evolution remain open, for example:

- 1) How is the ocean-continent boundary expressed in the mantle, and what is the relationship between structures present there and the tectonic events that have occurred?**
- 2) What is the nature of 3D lithospheric and asthenospheric velocity heterogeneity along the margin, and how are these structures associated with tectonics**
- 3) What is the pattern of mantle flow (as inferred from seismic anisotropy) beneath the passive margin? Is it any different beneath the oceanic side of the margin relative to the continental side?**

Regional Background The northeastern edge of North America is a "textbook-case" passive margin (Grow and Sheridan, 1988; Sheridan et al., 1995). It was formed during a Mesozoic (0.2 Ga) rifting event that created the Atlantic Ocean, which was associated with voluminous mafic magmatism (see review in Mahoney and Coffin, 1997). It has experienced very little subsequent tectonic activity, the most significant event being the passage of the hot spot that formed the New England sea mounts in the Cretaceous at 0.1 Ga (Sleep, 1990).

Crustal structure A general thinning of the continental crust from 45-50 km at the center of the craton to ~35 km at the coast, and to as low as 15 km within the continent-ocean transition zone is evident from seismic refraction data (Hughes and Luegert, 1992; Hennes et al., 1991; Keen and Barrett, 1981, Holbrook et al., 1992). The subareal thinning mostly reflects the different provenance of the several terranes, with the younger accreted terranes generally having thinner crust. The submarine thinning reflects extension during the Mesozoic rifting event that can exceed 100% (Speed, 1994). Fifty km east of the continental margin, the Atlantic oceanic crust has normal (7 km) thickness (Morris et al.,

1993).

In a deviation from the general trend, the continent-ocean transition (COT) region offshore northeastern North America is characterized by an increase in crustal thickness [Kelemen & Holbrook 1995]. Seismic structure of the thickened crust is distinct from both the thinned continental crust landward and the oceanic crust seaward of the COT. Seaward-dipping reflectors identified within the crust of the COT are believed to be intrusive bodies emplaced during the final breakup of the continent. They are the likely cause of the prominent East-Coast Magnetic anomaly that delineates the ocean-continent boundary.

Upper-mantle structure. A general thinning of the lithosphere from the interior of the craton towards the coast may be inferred from a variety of seismic observations, although constraints are much weaker, and consensus is far from complete. For example, upper mantle shear velocity models, based on long-period waveform inversion, show a general decrease in lithospheric velocity (e.g. the velocity at 100 km depth) along a traverse from the craton's center into the western Atlantic (van der Lee and Nolet, 1997; van der Lee, 2002), see Figure 1. However, the changes are far from being progressive and smooth. Rather, a number of regions with distinct lithospheric thickness may be inferred: a Proterozoic region where lithosphere exceeds 200 km, a region beneath Appalachians that lacks apparent lithosphere, and a coastal region where lithosphere appears to be less than 100 km thick (van der Lee, 2002). A strong "one-theta" azimuthal variation of Rayleigh wave phase velocity, detected using a small-aperture array in the New York - southern New England region (Menke and Levin, 2002), suggests that localized lithospheric thickness variations occur at some of these boundaries (Figure 2).

Interestingly, in the coastal northeastern US, seismic body wave techniques provide evidence for some distinction between the upper 250 km of the mantle and the region below. P-wave tomography identified a significant drop in the degree of heterogeneity deeper than 250-300 km (Levin et al., 1995), which the authors interpreted to denote the depth extent of the continental lithosphere. On the other hand, prominent P-S converted phases are seen from a feature at the depth of ~250 km, interpreted by (Li et al., 2002) as a lower bound of the low-velocity asthenospheric zone beneath a region with thin lithosphere.

Beneath the Western Atlantic, lithospheric thickness may only be estimated from large-scale seismic studies. Thicknesses of about 100 km may be inferred from models by Grand and Helmberger [1984], van der Lee [1997, 2002] (see Figure 1) and by the University of Colorado group. (see <http://ciei.colorado.edu/~nshapiro/MODEL/index.html>). In the tectonically similar Brazil basin, a deep seismic sounding profile yielded estimates of lithospheric thickness between 60 and 70 km (Pavlenkova et al., 1994, Pavlenkova 1996).

While changes in upper-mantle properties across the Atlantic passive margin of North America are significant, equally dramatic variation in properties is seen laterally along the margin (van der Lee and Nolet, 1997; van der Lee, 2001) (Figure 1). A number of smaller-scale (<400 km) but large amplitude (+/- 6%) heterogeneities are present, some of which are elongate, and trend sub-perpendicular to the margin. Regional P wave, S wave and surface wave tomographic studies using land stations have confirmed this strong heterogeneity (Levin et al, 1995; Levin et al. 2000; Li, 2000), and also suggest that fine scale (< 100 km) structure unresolved by the continental-scale studies exists as well. One of these slow offshore anomalies has been confirmed by timing arrivals of surface waves from Mid-Atlantic ridge earthquakes to coastal stations (Menke and Levin, 2002).

None of the studies performed to date has the ability to address the issue of the lithospheric structure in the vicinity of the continent-ocean transition (which is located well east of the coast). Being fairly narrow, the region will not be resolved by the continent-scale imaging studies, and data is lacking for more detailed regional-scale efforts. Active source studies (Holbrook et al. 1994) indicate presence of dense, high-velocity bodies in the lowermost crust at the ocean-continent transition. However these studies do not address the properties of the mantle underneath these anomalies.

Mantle fabric. Investigations of seismic anisotropy in the upper mantle beneath northeastern North America (Fouch et al., 2000; Levin et al., 1999;2000; Li, 2000) differ in details, but agree in two fundamental aspects: 1) systematic fabric exists within the upper mantle and 2) there is vertical stratification of the fabric, with distinct contributions likely from both the lithosphere and the asthenosphere. Specifically, (Levin et al., 1999, 2000) show that there are two distinct layers of mantle "fabric" that are more-or-less uniform across much of northeastern North America. The upper layer has an anisotropic fast-axis (N115E) locally perpendicular to the Appalachians, oblique to the cratonic edge.

The lower layer has an anisotropic fast axis of S53W that is sub-parallel to both the continental margin and to the S65W absolute motion of the North American plate. Associating upper-mantle anisotropy-inducing fabric with the alignment of the olivine crystals, the upper layer has been interpreted as being in the mantle lithosphere, and the lower layer as being in the mantle asthenosphere. The upper layer is thus reflective of the past tectonics, while the lower layer should reflect the motion of the North American plate relative to the mantle. Fouch et al. (2000) document, through computational modeling, the influence a lithospheric depth profile has upon the mantle flow beneath the plate and, consequently, upon the pattern of observed anisotropic indicators.

No direct measurements of seismic anisotropy have yet been made in the submerged part of the proposed study area (except by low-resolution, global tomography). The stretched lithosphere of the continental margin is likely to retain the marks of that process. The lithosphere of the Atlantic ocean should retain the olivine fabric formed during its formation, however the orientation of this anisotropy will depend on the nature of the upwelling regime at the spreading center. Gaherty (2001) shows that upper 100 km of the mantle beneath the Atlantic east of Reykjanes Ridge are characterized by anisotropy ~5%, with a vertical symmetry axis. Such orientation probably reflects the buoyancy-driven upwelling at the ridge close to a hotspot. Lithosphere formed at a ridge where the upwelling regime is "passive" is more likely to have a ridge-normal alignment of olivine crystals. If the upwelling regime of the Mid-Atlantic ridge has changed over its lifetime, we will find an age progression of anisotropic properties preserved within the lithosphere. Finally, at the continent-ocean transition we may find the signature of the along-axis (i.e., coastline-parallel) mantle flow hypothesized by Lizzaralde and Holbrook (1997). Mantle fabric beneath the lithospheric plate should reflect plate motion relative to the asthenosphere. If the asthenosphere is stationary, this direction should be close to the S65W vector of absolute plate motion. Independent flow within the asthenosphere, e. g. due to small-scale convection beneath the plate, will change this value.

Questions to be answered and Hypotheses to be Tested

1) How is the ocean-continent boundary expressed in the mantle, and what is the relationship between structures present there and the tectonic events that have occurred?

Is there any significant difference between the lithosphere and asthenosphere on different sides of the continent-ocean margin? Van der Lee's continental-scale surface wave images show a large change in lithospheric thickness at the edge of the North American craton (several hundred kilometers inland from the coast) but no significant change across the COT margin. These images, however, are based on land stations only, and thus may lack the necessary resolution to detect change across the margin.

The Mesozoic rifting event that created the Atlantic ocean was associated with voluminous mafic magmatism (often referred to as the Atlantic Large Igneous Province, or LIP). This magmatism is expressed as numerous dikes, sills and flows preserved within the rift basins along the coast, as seaward dipping reflectors observed offshore, and as extremely thick (20 km) oceanic crust just east of the continent-ocean transition. The apparent synchronicity (Paul Olsen, personal communication) of this magmatism and its chemical homogeneity all along the coast has been used to argue for a single localized source. We will examine the mantle lithosphere just below the margin, looking for anomalously fast regions that would be indicative of loci of high-degree melting and thus of depleted, seismically fast peridotite. We will also look for a narrow zone of margin-parallel anisotropy that might be associated with the mantle flow during the final stage of rifting, as proposed by Lizzaralde and Holbrook (1997).

The ocean-continent transition should also be delineated in the upper-mantle discontinuities. Oceanic upper mantle tends to be characterized by a sharp decrease in seismic velocity with depth associated with the base of the lithosphere (the Gutenberg, or G, discontinuity; Revenaugh and Jordan, 1991; Gaherty et al., 1999). Continental regions often have sharp increases with velocity near ~80-km (Hales, or H) and ~250-km (Lehmann, or L) depth. The interpretation of these features includes phase changes (e.g Revenaugh and Jordan, 1991), transitions in anisotropic fabric (Gaherty and Jordan, 1995; Bostock, 1997, 1998; Levin and Park, 1998), and the base of the low-velocity zone (e.g. Li et al., 2002). We will attempt to image these boundaries across the ocean-continent transition, thereby illuminating the processes underlying these features. In particular, we will focus on the nature of the L discontinuity, which Li et al. (2002) detected ~300 km deep beneath the craton and which shallows to ~250 km deep at the coast. They hypothesize that this reflector represents the bottom of the

asthenosphere, but we speculate that it might have an anisotropic origin. Figure 5 illustrates hypothetical receiver functions for an ocean site containing these discontinuities. We will utilize the complementary constraints of receiver-function and surface-wave analyses to specifically evaluate the nature of this boundary (Gaherty and Jordan, 1995).

2) What is the nature of 3D lithospheric and asthenospheric velocity heterogeneity along the margin, and how are these structures associated with tectonics?

Do the margin-cutting heterogeneities cross the continent-ocean boundary? Where are they the most intense? Does their depth extent indicate that they are mostly lithospheric or mostly asthenospheric? Do they have anomalously high seismic attenuation, as would be consistent with a thermal origin? Do they correlate with crustal or tectonic structure? What mechanisms are consistent with the details of their structure?

3) What is the pattern of mantle flow (as inferred from seismic anisotropy) on the oceanic side of the margin?

Is that pattern consistent with the idea that the shape of the lithosphere is controlling the pattern of flow? The flow models of Fouch et al. (2000) suggest that anisotropy should diminish away from the coast, over the ~1000 km length of the ocean part of the proposed array. Or does it continue unabated well seaward of the continental margin? In this latter case it would likely be controlled by a large-scale coherent shear between the plate and return flow in the underlying asthenosphere. What does the fabric retained in the oceanic lithosphere tell us about continental breakup and the initial evolution of the Atlantic? Does the fabric reflect buoyancy-driven flow during or following breakup? Does it retain a signature of margin-parallel transport (Lizarralde and Holbrook, 1997)?

New Data Collection Effort with Onshore/Offshore Array. We propose to perform an onshore-offshore passive seismic experiment that will use observations of distant and regional seismicity to characterize lithospheric structure.

The array (Figure 3) is designed to probe the lithosphere across a region extending from Lake Superior in the interior of the Canadian craton to the North Atlantic abyssal plane near Bermuda, from as far south as Delaware to as far north as northern New England. We will combine data from several distinct arrays (Figure 2), which total to about 60 seismic stations:

1. An array of 20 OBS's deployed for 12 months on the sea floor in a broad region between Bermuda and the continental shelf.
2. A temporary deployment of 4 PASSCAL-type broadband observatories along the Atlantic coast.
3. The existing network of about 35 broadband seismic observatories in the Eastern US, Canada and Bermuda. The stations include the US National Seismic Network, Canadian Seismic Network and Global Seismic Network, and the newly expanded Lamont Cooperative Seismic Network in New York and New England area. The Bermuda station, which began operation in 2002, is of particular importance in this study since it will provide high quality data for this ocean site over a longer period of time than the OBS's. The new Canadian POLARIS array provides dense coverage in the cratonic region of southeastern Canada (<http://www.polarisnet.ca/>). LDEO has an excellent working relationship with Dave Eaton, a PI of the Polaris project, and he assures us both that the data will be freely available and that he would be willing to collaborate with us in assembling it (see attached letter).

As a result, station spacing within the array will be a uniform 100-200 km throughout the study region. An important aspect of this array will be that both land and sea stations will operate contemporaneously, thus permitting observations of the same set of earthquakes. This "synoptic" dataset will allow precise discrimination of structure effects from source effects.

PASSCAL-type Stations The land part of the array, 4 portable broadband seismic observatories owned by LDEO will be deployed during the first year of the project, and will continuously acquire data for 1 year. These sites will be chosen as close to the coast as practical, so as to fill in the gap between existing land observatories and the ocean.

Noise levels are, of course, higher for coastal sites than for inland sites. In order to test the

premise that we will nevertheless obtain a reasonable amount of good data, we operated a portable seismic observatory (with REFTEK data logger and a CMG40T broadband seismometer) for one month each at two sites on Long Island, NY (one in the southwest at East Rockaway, NY, the other in the southeast, at East Hampton, NY). In both cases the sensors were placed on a concrete platform overlying sand. Although noise levels were somewhat higher than on hard rock, excellent data was nonetheless recorded (Figure 4). The East Rockaway, NY station recorded several core phases from western Pacific events that could be used to measure shear wave splitting directions. The splitting direction matches the previously-observed pattern for the NY/New England region, although the amplitude is somewhat smaller. The smaller amplitude, should it be observed for more data, would indicate a thinning of the LPO layers towards the continental margin.

OBS Deployment We plan a 12 month deployment of 20 four-component (three component seismometer plus DPG) instruments from the OBSIP pool, during the first year of the project. These instruments have successfully recorded both regional and teleseismic data in many different settings, including the East Pacific Rise and the Lau Basin. The work will require two 12-day cruises on an intermediate-class research vessel (assuming the ports are Woods Hole, MA and Bermuda).

The proposed deployment is longer in duration than the MELT experiment, which yielded adequate shear wave splitting data and excellent Rayleigh wave observations. Recording conditions for this experiment should be greatly superior to MELT: the distribution of events is superior, attenuation should be much lower here than under the MELT EPR sites, and microseisms (seafloor noise levels) are expected to be 10 to 20 dB lower in amplitude, and of higher frequency (thus providing a wider useful bandwidth for detection of long period arrivals) (Webb 1998). The instruments are sited off the shelf, in water deep enough to avoid strong ocean bottom currents.

Data Sources Figure 3 illustrates the location of the proposed onshore/offshore array with respect to global seismicity exceeding $M_b=6.0$ in a two-year interval. We anticipate excellent data coverage from the northwest, west and southwest, more modest coverage from the northeast and east. In addition, we expect to record a number of smaller events from Mid-Atlantic Ridge and Caribbean Basin, which promises to fill the coverage gap. Local seismicity within and close to the array will be utilized as well.

Methods

Local Estimates of Shear Velocity from Surface Waves. Estimates of shear wave velocity variation with depth can be obtained by inverting the regionally-localized, frequency-dependent phase behavior of fundamental- and higher-mode surface waves. On the OBS data, this analysis will predominantly focus on Rayleigh waves, as bottom currents often produce prohibitive noise levels on the horizontal-component data at frequencies below ~ 0.04 Hz (Collins et al., 2001a). Both the Differential Pressure Gauges (DPG's) and seismometers have proved a very reliable source of Rayleigh wave data in previous OBS-based experiments, such as MELT (Forsyth et al. 1998) and SWELL (Laske et al. 1999), especially in the the 20-70s period range important for determining lithospheric structure. Phase velocity localization will be performed using an estimation technique similar to Forsyth et al. (1998) in which the phase velocity is perturbed so as to match the evolution of the Rayleigh waveform across the array, as well as the array-based GSDF technique of Freybourger et al. (2001). Both methods provide a means to map propagation velocities as a function of azimuth, providing a natural way of estimating seismic anisotropy and associated mantle fabric. The Forsyth algorithm also accounts for the effect of multipathing, which is sometimes present at the shorter periods. The GSDF analysis can incorporate higher modes, improving our depth resolution, and it also improves the anisotropy estimates by modeling Love as well as Rayleigh velocities.

Love waves cannot be routinely expected from the OBS deployment, but we are hopeful that they will be observed for at least a handful of large events, as in the MELT and OSN pilot experiments [Webb and Forsyth, 1998; J.A. Collins, personal communication]. If so, the localized phase delays can be inverted for both radially and azimuthally anisotropic structure [Gupta and Gaherty, 2000]. The azimuthal anisotropy estimates are improved relative to those determined by Rayleigh alone because they allow the decoupling of isotropic and anisotropic heterogeneity. Radial anisotropy provides additional first-order estimates of critical anisotropy characteristics such as the depth dependence (lithospheric vs. asthenospheric affinity) and dominance of horizontal vs. vertical fabric. The variations in lithospheric anisotropy across the ocean-continent transition will provide clues to the

dynamic processes associated with continental collision, rifting, and ocean basin formation, while the asthenospheric variations will illuminate fabric associated with ongoing mantle flow. This analysis will be performed using the on-shore stations even if the OBS fail to deliver adequate transverse data.

Surface Wave Mode conversion. Surface waves observed by the array will be analyzed for the presence of mode-converted phases, e.g. quasi-Love waves (Park & Yu 1993; Yu et al., 1995). Observations of quasi-Love provide strong spatial constraints on the location of regions where abrupt changes in anisotropic properties occur (Levin and Park, 1998), while their spectrum contains information on the depth provenance of the anisotropic features. Since the diagnostic quasi-Love phase has P-SV motion, both vertical and DPG components on the ocean-bottom instruments will be able to record them.

Body Wave Tomography Travel times of P and S body waves will be used in first-order tomographic imaging. We expect large variation of seismic velocity field along the array as it goes from the craton across the passive rifted margin and into the ocean. We hope to provide significant refinement to the currently available images of upper-mantle velocity field, in particular where the width of transition regions is concerned. The lithospheric North American shear-velocity model reported by van der Lee and Nolet (1997) and van der Lee (2001) has significant lateral heterogeneity in the region sampled by our array. Other studies hint at smaller-scale variability, but are limited in their lateral extent (Levin et al., 1995, 2000; Li, 2000). We hope to improve the resolution of this heterogeneity, both laterally and in depth, to relate it to the findings of the other techniques (i.e. not tomographic) and to gain better understanding of its nature.

We have prior experience using teleseismic tomography imaging methods on smaller datasets from the New York and southern New England region, and are confident that the improved dataset can achieve a similar resolution across the entire proposed array. One of us (W. Menke) has recently finished a new tomographic imaging code, "raytrace3d" (available at <ftp://lamont.ldeo.columbia.edu/pub/menke/raytrace3d.tar.Z>) that uses a three-dimensional tetrahedral representation of velocity.

Receiver Function Analysis . Isolation and analysis of P-S mode-converted phases within the coda of teleseismic body waves - the so-called receiver function (RF) method - will be the primary tool to probe for the presence of sub-horizontal interfaces of the lithosphere. In the region of transition from oceanic to continental lithosphere we expect dramatic changes in lithospheric thickness, nature of the crust-mantle transition, and layering and fabric within the subcrustal lithosphere.

In addition to the usual Radial-component receiver functions (which detect interfaces where P-SV conversion occurs), we will also employ Transverse-component functions (which detect interfaces with changes in anisotropy) (Levin & Park, 1997, 1998). These functions have proved extremely useful in constraining mantle fabric in other parts of the world (e.g., Bostock, 1997, 1998; Levin & Park 1997, 2000a, Levin et al, 2002).

We will use a set of analysis and forward-modeling algorithms developed by Jeffrey Park and Vadim Levin (software distributed via the "Seismic Software Library" maintained by IRIS DMC). In particular, the multitaper spectral coherence (MTC) tool for receiver function estimation has an exceptional noise resilience, and is able to combine contributions from sources at virtually any distance (Park and Levin, 2000; 2001). To date, Levin successfully used this tool to analyze data from a variety of environments, including coastal regions of an active subduction zone (Kamchatka) and the ice-covered interior of Antarctica. Given likely noise spectrum on the seafloor, the ability of the MTC tool to analyze high frequency energy in both core phases (abundant in the study region) and regional seismic waves will be extremely useful, allowing to work in the spectral window above the microseismic peak. Also, resolution of vertical scales for interfaces like the Moho, the Hales and the Lehmann discontinuities will be greatly improved by the analysis of receiver functions at relatively high (1-3 Hz) frequency (see Levin and Park, 1998). Low attenuation beneath the site will provide much better short data than obtained from the MELT experiment. Tests with synthetic seismograms show that sediment-induced reverberations will complicate analysis of the high-frequency receiver functions on the seafloor where sedimentary thickness exceeds 1 km. The eastern part of the proposed OBS array will be sited over regions with modest sedimentary cover (as per sedimentary cover model of Gabi Laske), and we expect to avoid this problem at least on some of our sites.

The location of the proposed array will allow us to construct epicentral-distance receiver-

function gathers for the northwestern and southern directions on the basis of North- and Western Pacific, and South American seismicity, respectively. Epicentral distance controls the timing between the parent (P) and daughter (Ps) phases in the receiver function, offering additional means for signal enhancement, e.g. via stacking with appropriate moveout corrections or migration (Bostock, 1998; Dueker and Sheehan, 1997; Kosarev et al, 1999). We are particularly interested in mapping the transition of the G discontinuity from the oceanic lithosphere (where it should be well developed, e.g. Gaherty et al., 1999; Grand and Helmberger, 1984) into the continental lithosphere, as well as the character of the L discontinuity, which has been previously observed beneath eastern North America (Li et al, 2002). We will specifically test whether L represents an anisotropic boundary (Gaherty and Jordan, 1995; Bostock, 1997, 1998) by utilizing RF constraints on the depth of L in the surface wave analysis (e.g. Gaherty and Jordan, 1995).

Seismic Attenuation. Differential attenuation of P and S wave across the array will be determined using standard spectral techniques (e.g. Menke et al. 1995). Source areas along the mid-Atlantic ridge and in the Caribbean are expected to provide adequate numbers of relatively high-frequency seismograms. Path-averaged measurements of attenuation, such as can be made on individual P and S waves will be inverted for a 3D image of the attenuation using the "raytrace3D" code described above. Low microseism noise levels in the Atlantic broaden the useful bandwidth for attenuation measurements compared to Pacific seafloor sites.

Shear Wave Splitting. Measurements of shear wave splitting require observations of teleseismic SKS waves made on the horizontal components of the OBS's. The array is located favorably with respect to sources of SKS waves in the western Pacific. In several previous studies (Levin et al. 1999, 2000) the PIs measured hundreds of such phases on both temporary and permanent stations in nearby New York and New England.

As noted above, horizontal component measurements made on OBS's are typically more noisy than those made on the vertical component. Nevertheless, SKS splitting measurements were successfully made in the MELT, Southwest Pacific, and OSNP experiments (Wolfe et al. 1998; Wiens et al. 1995; Collins et al., 2001b). Furthermore, two effects will likely improve our ability to measure shear waves over these past experiments: the generally lower microseism level of the Atlantic over the Pacific (Webb 1998), and the probable lower asthenospheric attenuation in this region, compared to that of a ridge, back-arc spreading, or hotspot setting. The PIs have developed both a data analysis technique capable of making accurate splitting measurements, ftp://ftp.ldeo.columbia.edu/pub/menke/ah_splitest2.tar.Z and a modeling method capable of determining the anisotropic parameters from the observations, ftp://ftp.ldeo.columbia.edu/pub/menke/SPLITTING_MODELER.tar.Z. This methodology is also capable of detecting and modeling the effect of several distinct layers of anisotropy (i.e. at several depths), should several be present.

Methodological Synergy. Techniques discussed separately above will be applied to the same dataset by the same group of PIs and students. We hope to avoid the common pitfall of compartmentalization, by cross-validating findings of mutually complimentary techniques. For example, surface-wave studies and receiver function studies compliment each other in sensitivity to velocity structure and interface structure, respectively. Similarly, observations of seismic anisotropy via shear-wave splitting, surface-wave dispersion and mode-conversion, and body-wave mode conversion provide complimentary constraints on intensity and location of anisotropy-inducing fabric.

Management Plan. All of the PIs will broadly participate in all phases of the project. Levin, with significant experience in previous PASSCAL-type array efforts, will lead the land array deployment. We will make a special effort to involve undergraduates in the fieldwork, a goal that is facilitated its being conducted within driving distance of LDEO and Rutgers. Webb, who has long-term experience with the manufacture and use of OBS's and Menke, who has used OBS's in a tomographic imaging experiment on the Juan de Fuca ridge (see Menke's Prior), will head the OBS array deployment. Once the data is collected, Gaherty will head the surface wave related data analysis, Levin the receiver function and splitting analysis, Menke and Webb the body wave tomography. The PIs will be assisted in this data analysis by LDEO and Rutgers graduate students.

Timetable The main constraints on timing are on the deployment/recovery of the OBS's, which are best done in May to September of the year, and the deployment of the LDEO broad band instruments, which is best done when snow does not interfere with driving.

Year 1: Initial deployment of four land stations; Service and collect data from these stations at 3 month intervals; Begin initial assembly and data processing of data from all (i.e. temporary and permanent) land stations; Cruise to deploy OBS array.

Year 2: Continue to service and collect data fromm land array at 3 month intervals; Continue initial assembly and data processing of all land data; Cruise to recover OBS array; Recover four temporary land stations. Initial data processing of OBS data; Merge OBS and land data into synoptic dataset and perform preliminary analysis and interpretation. Present initial results at scientific meeting.

Year 3: Final data processing; Interpretation of results; Writing up papers. Presenting final results at scientific meetings.

Dissemination of Results . We will submit the LDEO broad band instrument data and OBS data to the IRIS DMC within two years of its collection, and to any other public archives that are required by the Facilities involved. We will maintain archives of data and preliminary results on our institutional web sites (as we now do for previous studies, see for example <http://www.ldeo.columbia.edu/user/menke>). We will present results at scientific national meetings, such as the Fall AGU and IUG, and make a best-faith effort to publish them rapidly in a peer-reviewed journal.

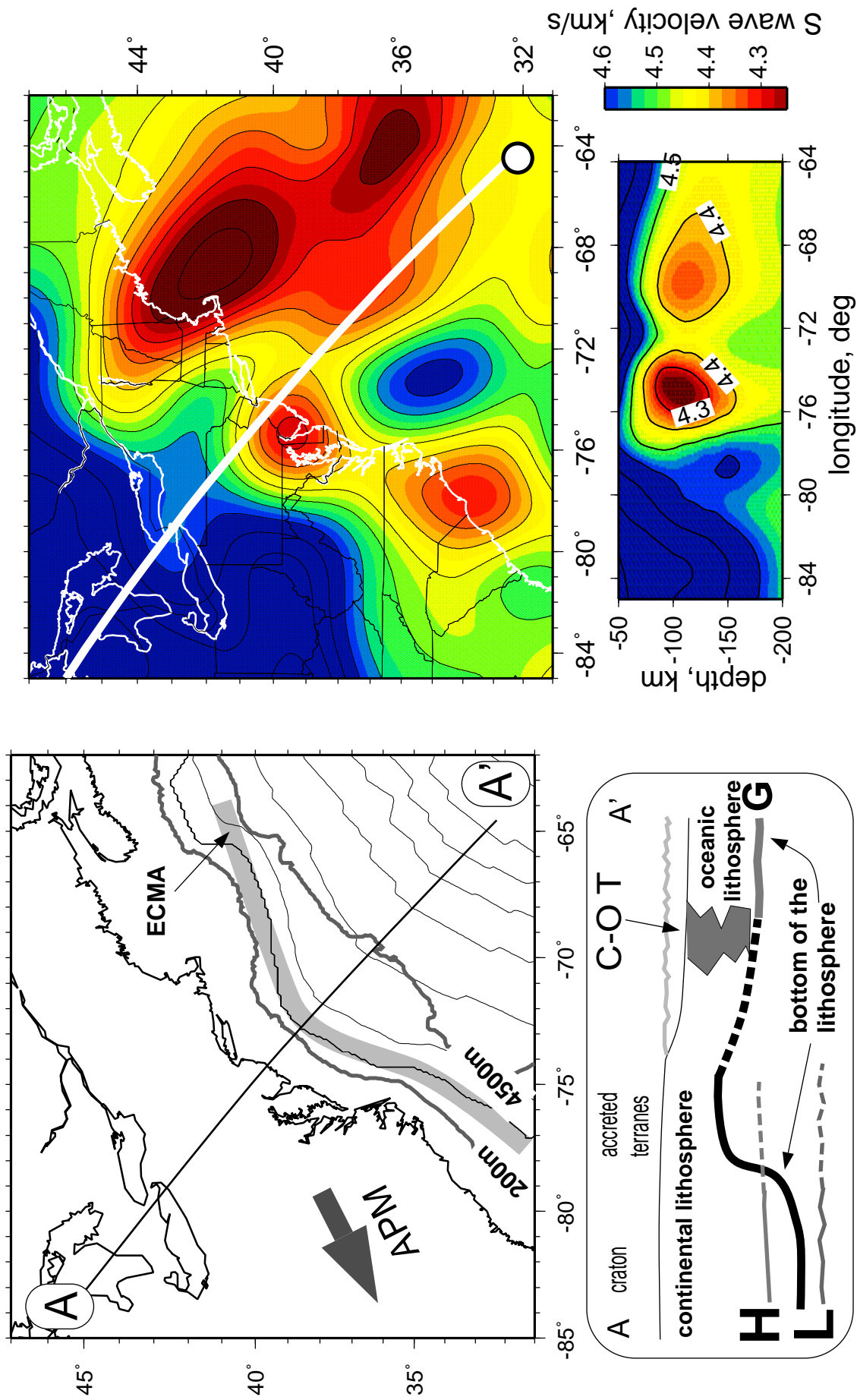


Figure 1. (LEFT) Map of study region showing the outline of the bathymetry (shaded lines), the age of the oceanic crust (thin solid lines) and the approximate location of the continent-ocean transition, as marked by the East Coast Magnetic Anomaly (ECMA). Large arrow shows absolute plate motion. The schematic cross section AA' illustrates hypothesized locations of Hales (H), and Lehmann (L) seismic discontinuities beneath the continent, the Gutenberg (G) seismic discontinuity beneath the ocean, the tentative profile of the lithospheric thickness (inferred from Li et al., 2002; van der Lee, 2002) and the location of the continent-ocean transition. (RIGHT) Seismic velocity of shear waves at depth of 130 km (map) and a cross-section along the cross-section from Bermuda to the Great Lakes. Velocity field extracted from the model NA00 by Susan van der Lee (<http://www.sg.geophys.ethz.ch/geodynamics/suzan/na/na00.html>).

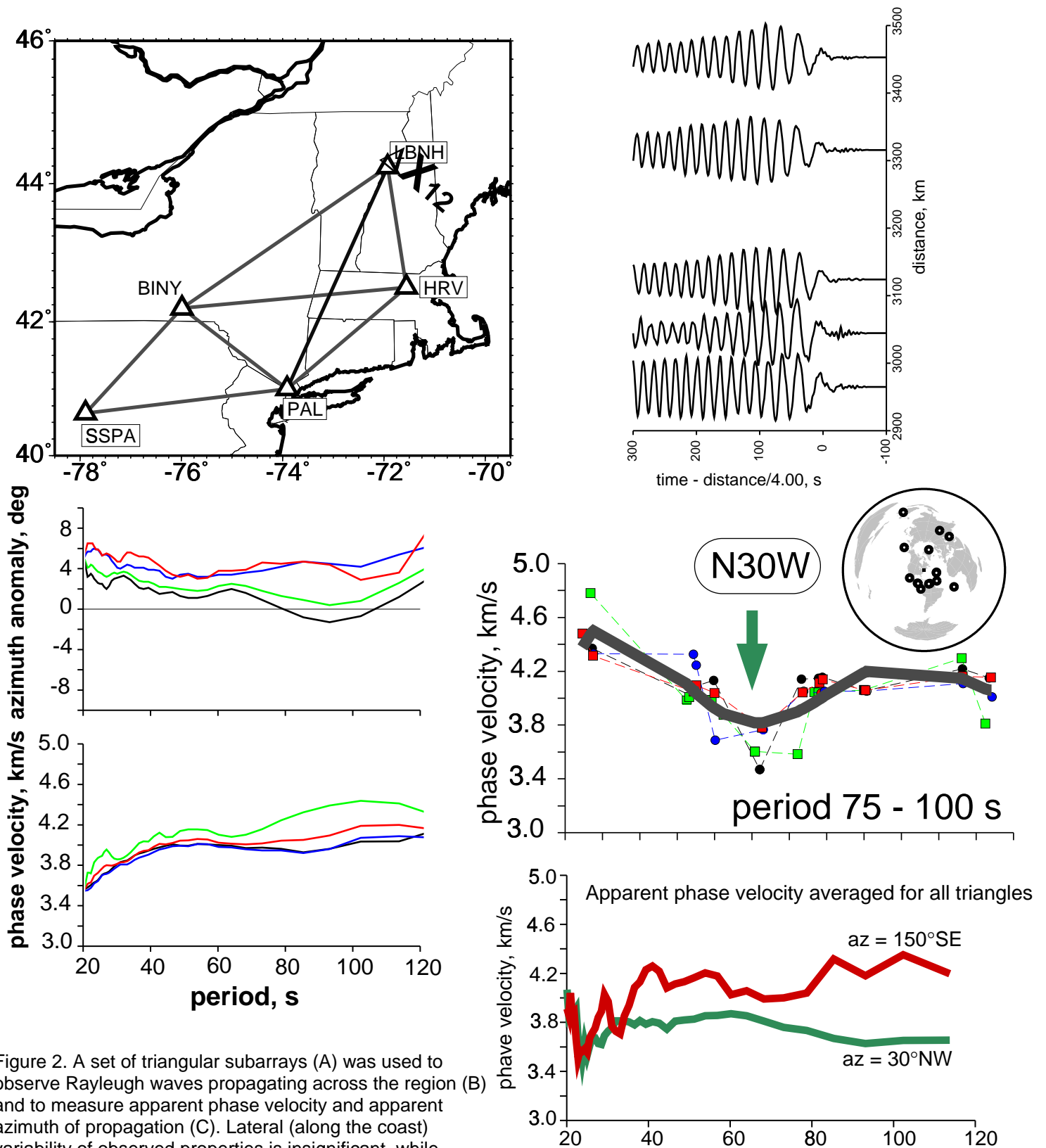
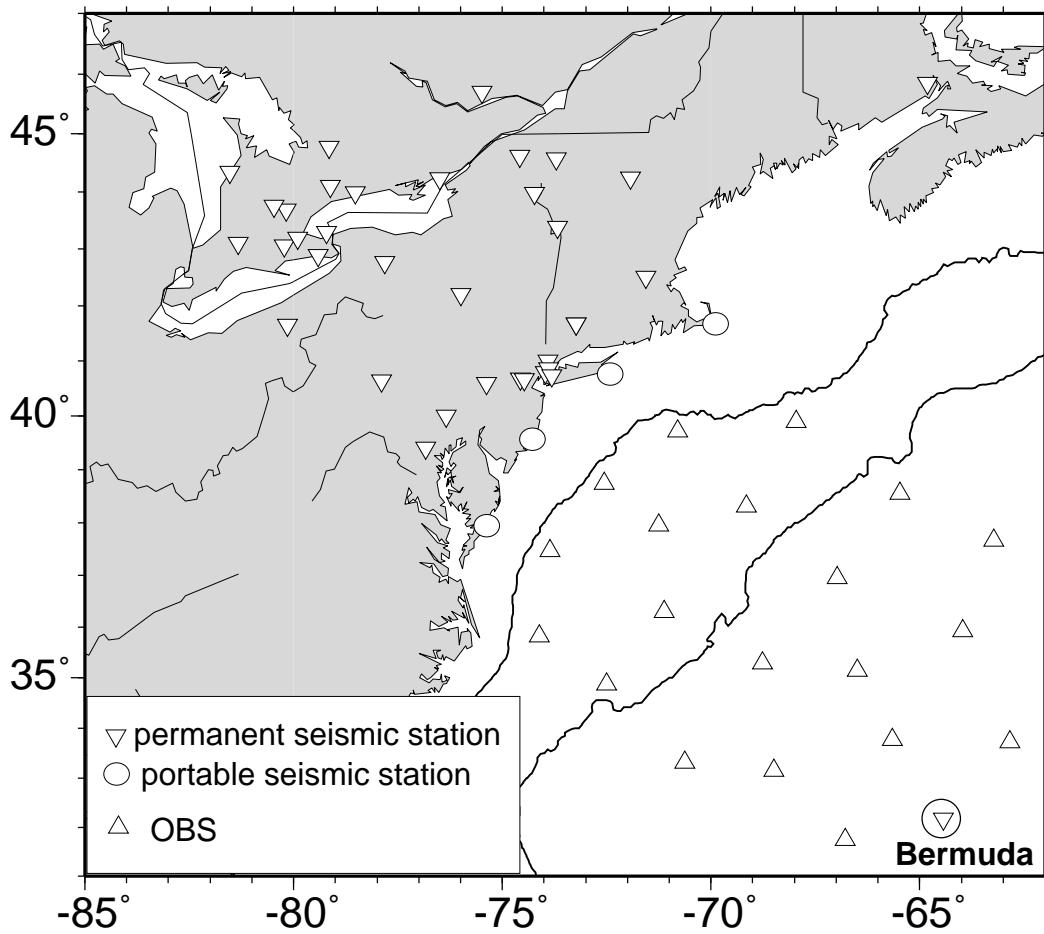
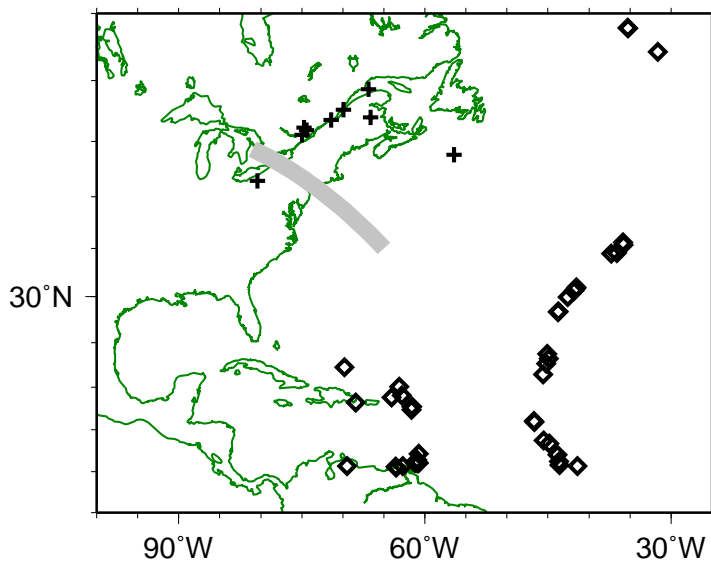


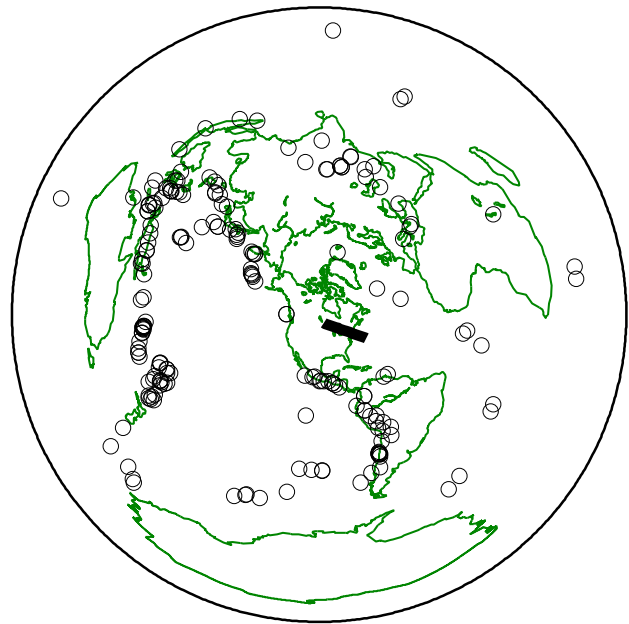
Figure 2. A set of triangular subarrays (A) was used to observe Rayleigh waves propagating across the region (B) and to measure apparent phase velocity and apparent azimuth of propagation (C). Lateral (along the coast) variability of observed properties is insignificant, while all subarrays display pronounced directional dependence of phase velocity (D and E). Waveforms presented in B are vertical-component Rayleigh wave records from a Mid-Atlantic Ridge earthquake of October 5th, 2000. An inset in D shows all sources used. The direction of propagation along azimuth N30W (fromocean to the continent) is the slowest. This behavior is consistent with a southeasterly (N150E) dip of the lithosphere-asthenosphere boundary beneath New England. This dip is probably associated with the shoaling of the low velocity anomaly (see Figure 1) from the ocean towards the interior of the continent. Detailed description of this study is presented in Menke and Levin (2002)



Teleseismic and regional seismicity in 1997-98



cross - $M_b > 4$ "local"
diamond - $6 > M_b > 5$ "regional"



$M_b > 6.0$

Figure 3. (TOP) A map of existing permanent seismic observatories in northeastern North America and on Bermuda (inverted triangles), and a tentative schematic of the proposed deployment (circles and triangles). (BOTTOM) Seismic activity in a 2-year period (1997-98) illustrates the likely distribution of sources the proposed array will observe.

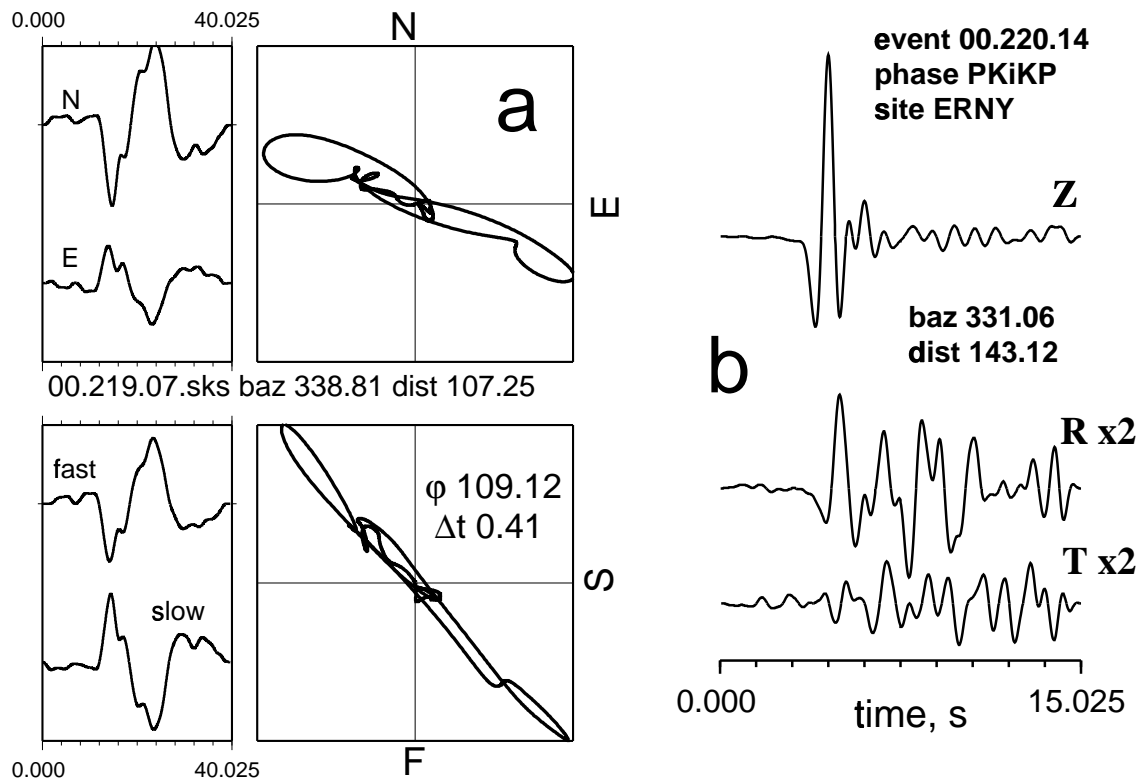


Figure 4. Sample observations from EARNY (East Rockaway, NY). Installation consisted of a REF TEK data logger and a CMG40T sensor placed on the surface in a small fiberglass enclosure.

- a)** Shear wave splitting measurement on an SKS phase. Waveforms are filtered in 0.025-0.3 Hz passband. Upper panels show observed data, lower panels show waveforms corrected for the effect of anisotropy. Splitting parameters are shown in the lower right panel: fast direction $\phi = 109.12^\circ$, delay is 0.41 s. Note reduced ellipticity of particle motion. This observation agrees well with measurements further inland (Levin et al, 2000), suggesting continuity of structures. **b)** PKiKP phase from an earthquake 143° away is particularly bright due to a caustic in the travel time curve. Waveforms are filtered in 0.025-1.5 Hz passband. Motion on horizontal components is a result of near-surface reverberations that form data for receiver function analysis. Of particular interest is SH-polarized motion (T component) indicative of lateral heterogeneity and/or anisotropy. **c)** Location of the station on the oceanward shore of Long Island.

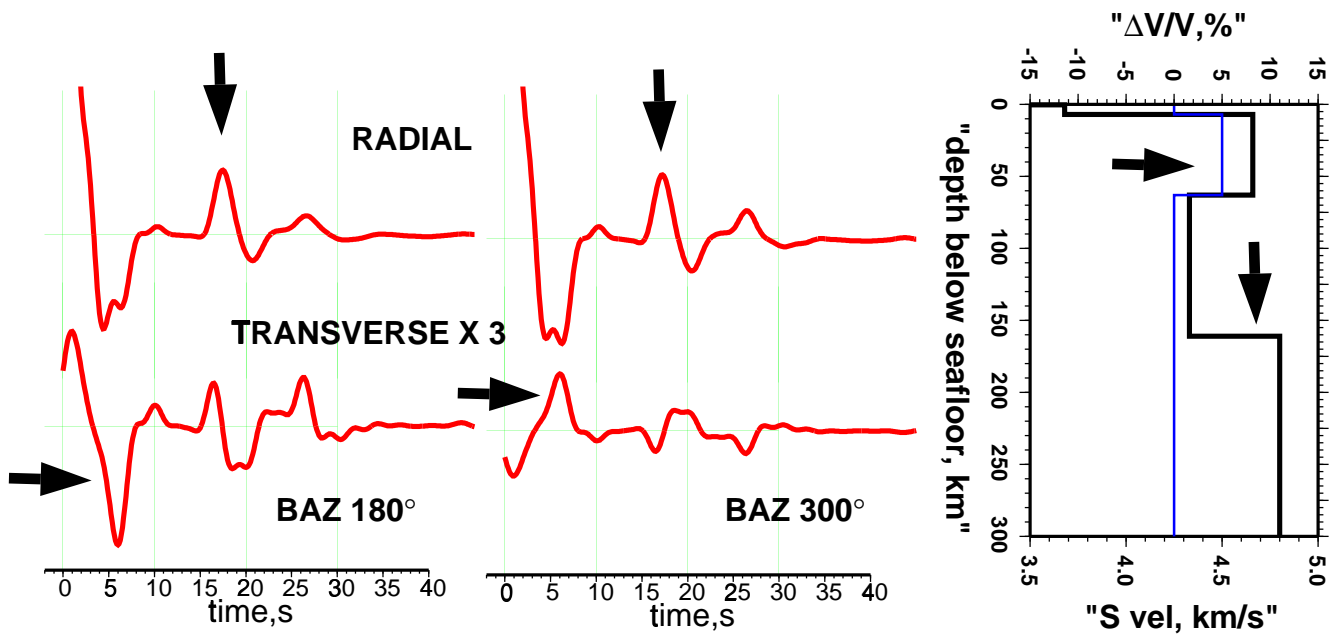


Figure 5. Synthetic receiver functions computed in a hypothetical velocity structure to illustrate features we may expect in the data from the OBS deployment. Vertical profile of SV velocity (right panel, bold line) is modified from the Pacific lithosphere model of Gaherty et al., (1996) by an introduction of a sharp velocity increase at 166 km. This feature is meant to be similar to the bottom of the low-velocity zone identified by Li et al. (2002) beneath the coastal region of the North American continent. The converted wave from this feature (marked by a vertical arrow) will arrive 18-20 sec behind the parent P wave. Another feature of the hypothetical model is the 5% anisotropy in the lid (right panel, thin line), with a horizontal axis of symmetry aligned at N135W (normal to the lines of equal age on the seafloor, see Figure 1). A prominent transversely-polarized phase (marked by a horizontal arrow) will be associated with the anisotropic lid. This phase will change its polarity depending on the direction of the incoming P wave. Computations are shown for backazimuths 180 and 300, which are illuminated well by earthquakes in South America and the Pacific, respectively. Spectral content of the synthetics is limited to below 0.25Hz.