Results from Prior NSF Support

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 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 signials 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 (West, 2001):

1. **A Very Large Axial Magma Chamber.** At a depth of 2.25 to 3.5 km beneath Axial caldera lies an 8 by 12 km magma chamber containing 10−20% melt (West 2001). At depths of 4−5 km beneath the sea floor there is evidence of additional melt, in lower concentrations (a few percent) but spread over a larger area. Residence times of a few hundred to a few thousand years are implied (West et al. 2001).
2. **A smaller Coaxial Magma Chamber, unconnected with the one at Axial.** The magma chamber is loacted at the "Source Site" of the 1993 eruption (Menke et al., 2001). It is at least 6 cubic km in volume and contains at least 0.6 cubic km of melt, enough to supply at least several eruptions of size equal to the one in 1993.
3. **Several other small low velocity zones are possibly outlier magma chambers from Axial.** Two other low−velocity zones occur in the shallow crust near Axial volcano, one about 10 km north of the caldera on the North Rift, and the other about 10 km south of the caldera but displaced to the west of the South Rift (West 2001). They appear unconnected to the main Axial magma chamber and might possibly represent small accumulations of melt left over from past lateral diking events.
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:

1. Menke−W, 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, submitted to Geology, 2001.
2. West−M, The deep structure of Axial Volcano, Ph.D. Thesis, Columbia University, 2001.
3. West−M; Menke−W; M−Tolstoy; S−Webb; R Sohn, Magma reservior beneath Axial volvano, Juan de Fuca Ridge is far larger than eruption size; submitted to Nature, 2001.

Data and other products This project collected new data, which is freely available on−line at http://www.ldeo.columbia.edu/user/menke/AX/. Some software, including the tomograhy code, that was written for the project is available at http://www.ldeo.columbia.edu/user/menke/software/.

Results from Prior NSF Support

Title: Collaborative Research: Complex Upper Mantle Structure Beneath Northeastern US Investigated Through Shear Wave Tomography (collaborative project between LDEO and Yale University)

Period 08/15/97-07/31/99
William Menke(LDEO), **Award Number:** EAR-9706195, **Amount:** $63993.00
Jeffrey Park & Vadim Levin (Yale), **Award Number:** EAR-9707189, **Amount** $53193.00

The purpose of this research was to test the proposition that the mantle beneath northeastern North America is divided into several "anisotropic domains" that are the seismic expression of the plate tectonic process of "terrain accretion". Were such the case, we would expect different directions of the shear wave fast direction and different mean shear wave velocities in each of the several terrains (whose existence has been established geologically).
Thus we assembled shear wave traveltime and splitting databases for all the broadband seismic stations that were operated - even only temporarily - in northeastern North America for the past 5 years. We analyzed the splitting data by comparing it to synthetic measurements drawn from synthetic seismograms computed for anisotropic models. We tomographically inverted the traveltime data. Much of the data analysis and modeling code was custom-written (by us) for this project.

The results are quite suprising, and show:

- The pattern of shear wave fast directions across northeastern North America is very homogeneous. No anisotropic domains occur.
- At a given station, the pattern of shear wave fast directions varies rapidly with the backazimuthal angle to the earthquake epicenter. This pattern has a strong "four-theta" component that can be explained in a most excellent manner by postulating two layers of mantle anisotropy. .
- These layers are laterally homogeneous across northeastern North America.
- The top layer has a shear wave fast direction oriented toward/away from the center of the craton. We believe it to be unrelated to the dynamics of the Precambrian craton, and instead to be related to a period of intense strain experienced by all the terrains during a lithospheric delamination, likely to have occurred during the during the Appalachian orogeny.
- The bottom layer has a shear wave fast direction oriented parallel to the edge of the craton. We believe it to be related to asthenospheric flow.
- Shear wave velocities at 100 km depth are quite heterogeneous, with the western Adirondacks being particularly slow. We postulate that this is a chemical heterogeneity that is unrelated to the strain-induced anisotropy.

We have written these papers describing the results:
Levin, V., W. Menke and J. Park, 1999. Shear wave splitting in the Appalachians and the Urals: A case for multilayered anisotropy **J. Geophys. Res.** Vol. 104, No. B8, p. 17,975-17,987.
Levin, V., J. Park, M. Brandon and W. Menke, Thinning of the upper mantle during the late Paleozoic Appalachian orogenesis, **Geology** 28, 239-242, 2000.
Levin, V., W. Menke and J. Park, No Regional Anisotropic Domains in Northeastern US Appalachians, **J. Geophys. Res.**, Vol. 105 , No. B8 , p. 19,029, 2000.

Data and other products This project collected no new data. Some software that was written for the project is available at http://www.ldeo.columbia.edu/user/menke/software/

INTRODUCTION

Importance of lithospheric erosion as a geodynamic process. The process of thermo−mechanical erosion has a significant impact on the recycling of material from the lithosphere to the asthenosphere. Such erosion occurs in a variety of tectonic settings, including subduction (where the uppermost of the two plates is abraded) (e.g. Karig 1974; von−Huene−Roland & Lallemand 1990); collisional orogenies (where thickened lithosphere can delaminate) (e.g. Seber et al. 1996); and rifting and plumes (where the hot upwelling asthenosphere can warm and shear away the bottom of the lithosphere) (e.g. Campbell & Griffiths 1992; Kerr 1994; Furman & Graham 1999). The effect of such erosion is manifest in changes in freeboard (i.e. uplift and subsidence), the chemistry of erupted magmas (both through changes in depth of melting and in lithospheric contamination), surface heat flow and in changes in the the lithosphere's mechanical strength (ie. flexural response, faulting and seismicity).

As the lithosphere is defined in terms of its mechanical strength, its development is controlled both by its cooling history and by any physical or chemical differences (e.g. degree of depletion, grain size) between it and the surrounding asthenosphere that effect its rheology. Whether the erosion of the lithosphere is long−lasting will depend upon circumstances. Shallow, stagnant asthenosphere left after an erosive event will tend to cool conductively, adding to the lithosphere. If continued asthenospheric flow advects heat to shallow depths, this thickening can be impeded. Even should the lithosphere return to its original thickness and rigidity, a "scar" left by the erosive event might still be detectable, through the chemical and textural differences between the old and new material. A scar could be detected in a fossil mantle fabric that leads to seismic anisotropy [Levin et al, 2000ab]. Since undepleted mantle contains a larger proportion of iron, relative to magnesium, in its olivine and pyroxene components, its seismic velocity can be several percent slower than surrounding lithosphere, even after thermal equilibration [Anderson and Isaak, 1995; Collins and Brown, 1998; Chai et al, 1997].

Lithospheric Erosion by Plumes. Plume erosion mechanisms have been advanced to explain a variety of observations. Bonneville et al. [1997], for instance, used heat flow anomaly data from Reunion hot spot that were carefully referenced against the background heat flow of the surrounding plate to conclude that a persistent thermal anomaly beneath the lithosphere does cause significant thinning after 10 Myr. More generally, lithospheric thinning by continental plumes have been used to explain episodes of regional metamorphism, anatexis, crustal extension, flood basalts development, and rifting (Hill et al. 1992, Davies 1994). In the oceans, lithospheric erosion has been postulated to influence the local geoid and the chemistry of the plume magmas (e.g. Rhiannon et al. 1995). Thus there seems to be ample data that supports an association between the passage of a plume and lithospheric modification and erosion.

The fluid dynamics of the erosion processes has been investigated by a variety of modeling techniques (e.g. Sleep 1987; Olsen et al. 1988; Davies 1994). The rate of thinning is very dependent on the lateral scale of the upwelling, with narrower upwelling associated with greater thinning potential, and on the viscosity difference between the lithosphere and asthenosphere (Davies 1994). Whether or not plume can cause significant erosion is still very controversial. Many of the early papers on this subject (e.g. Davies 1994) concluded that erosion was likely to occur, at least in a narrow zone around the "plume tail". However, the most recent geodynamical modeling studies suggest that the plume thermal flux is too weak to erode the lithosphere directly [Ribe and Christensen, 1994; Cserepes, et al. 2000]. Our reading of this controversy is that there is considerable evidence that plumes do erode the lithosphere, but that the mechanism by which they do so is not understood. Other processes in addition to simple reheating and viscous transport likely play an important role. Therefore, an association between the past passage of the New England hot−spot track and eroded lithosphere near the North American margin is a viable hypothesis that can be tested in such a way as to significantly contribute to resolving this issue.

Research at the Yellowstone and Hawaiian Plumes. The Yellowstone plume, which interacts with continental lithosphere in western United States, and the Hawaii plume, which interacts with old oceanic lithosphere in the central Pacific, have both been region of vigorous study. Seismic models have been interpreted to conflict with the geodynamic models of plume−generated erosion, and motivate further study of the process in new locales.

The Yellowstone plume is associated with a parabola−shaped region of tectonic activity to its west. Models of this parabola have suggested that it is due to plume material that had ponded at the base of the lithosphere, and then has been viscously dragged horizontally and spread out. However, recent tomographic imaging of compressional velocity (Humphreys et al. 1999; Saltzer and Humphreys, 2001) on a line that cuts across the parabola, identifies features that are inconsistent with this model. A 200 km wide zone of low velocities extends from 50−300 km depth. They have interpreted this as indicating that the plume material has both eroded further upward into the lithosphere, and extends deeper into the asthenosphere, than has been anticipated by the modeling. The low velocity region is bounded by high velocity edges, interpreted as the descending limbs of convective rolls. Flow in the plume is thus more complicated than anticipated, perhaps because of the presence of melt−buoyancy as an added driving force. Lithospheric anisotropic fabric in the interior of the parabola, determined by SKS splitting studies, is indicative of lithospheric strain parallel to the direction of absolute North American plate motion. This suggests that the region of thinned lithosphere experiences more deformation than its surroundings, where the mantle anisotropy pattern is more complicated.

Many authors have investigated the role of the lithospheric thinning associated with the Hawaii plume, and especially whether it is confined to a narrow corridor along the island chain or does include the much broader region of the swell. McNutt & Shure (1986) argue that the geoid data require a shallow depth of compensation, indicative of thinning. Subsequent modeling (Robinson et al. 1987) has indicated that such shallow compensation is not required, if low viscosity layers are present in the asthenosphere. Ribe and Christensen [1994] argued that plume−generated density anomalies beneath the lithosphere, not lithospheric erosion, are responsible for its rapid uplift (~2−3 Myr) and geoid signature. Estimates of shear velocity structure from the surface waves are inconsistent with widespread thinning beneath the swell (Wood & Okal 1996), although they do detect slower seismic velocities (Laske et al. 1999) and a major change in the anisotropic fabric (Levin & Park 1998) close to the plume itself. Thus the seismic data at Hawaii contradict the notion of widespread erosion under the swell, but may be consistent with more narrowly confined erosion by the plume−tail.

Part of the problem in understanding lithospheric erosion at Hawaii and Yellowstone is that the plumes themselves are a major source of "geophysical signal" that complicates any attempt to quantify erosion. For instance, the geoid signal from the plume material flowing horizontally along the base of the lithosphere mimics, to some degree, that of the lithospheric thinning. Hence the question of lithospheric erosion may profitably be addressed with seismic data along an old hot spot track. We propose to study the offshore track of the the New England hot spot (or "Monteregian" hot spot Sleep [1990]). This plume is now far gone, but may have left a prominent shear wave velocity anomaly in the oceanic lithosphere. This anomaly could be thermal and/or compositional (i.e. a furrow of undepleted mantle that has replaced the depleted peridotite of the oceanic plate). Assuming that the seismic anomaly of Van der Lee & Nolet [1997] extends off the continental shelf, our proposition is that by studying an old plume track, one in which hot plume material is no longer present, we will be better able to detect the permanent changes that the plume has had on the lithosphere. The point is not so much that the NEA is unique as a geophysical phenomenon − we expect that there are many similar features worldwide − but rather that it is the first to have been identified. It is the only example known to us where a well−documented and no longer active plume track is in a region well−resolved by seismic tomography. Furthermore its position in oceanic lithosphere makes interpretation of its significance easier than a continental setting, since the process of initial formation of oceanic lithosphere is so much simpler and better understood than continental lithosphere.

New England Seismic Anomaly. The New England plume caused continental volcanism in New York, New England and southern Canada and the linear New England seamount chain (e.g. Kelvin seamount) in the western Atlantic. The volcanism is time−progressive to the east (with an azimuth of about N120E), with the continental activity at 140−120 Ma, the westernmost seamounts at 100 Ma and the easternmost at 80 Ma (Duncan 1984). This track correlates well with a prominent shear wave low velocity anomaly that appears in images (figure 1A) based on long−period seismic waveform inversions (e.g., Van der Lee & Nolet 1997). We shall henceforth call this feature the New England Anomaly (NEA). It is about 5−6% slow, and extends from the general vicinity of the New York Adirondack mountains, eastward across New England out to a distance of at least 1000 km into the western Atlantic. In the tomographic model the NEA appears to be quite wide, 300−500 km (perhaps an upper limit in width, since these images are of relatively low resolution). It extends from about 60 to 150 km depth beneath the Atlantic and rises to even shallower depths (about 40 km) beneath New York. These depths are much shallower than the expected depth of the lithosphere/asthenosphere boundary in a plate of Jurassic age (~120 km) and suggest that the lithosphere has been strongly eroded. (The "apparent age" of lithosphere of this thickness is only 30 Ma).

Because the NEA appears on the edge of the Van der Lee & Nolet (1997) tomographic image − in a position less well−constrained by the data − we first need to establish that this anomaly is indeed real, and not an artifact arising from some minor instability in the inversion. We have examined two completely independent data sets that support the existence of the NEA, at least in some form. First, the continental part of the NEA shows up in the shear wave inversion (figure 1B) of New England by Levin et al. (2000a), which is based on traveltimes of teleseismic S waves. Both the position and strike of the westernmost part of the NEA are similar in the Vs−anomaly patterns of Van der Lee & Nolet (1997) and Levin et al. (2000a). Second, traveltimes of 70 s Rayleigh waves from mid−Atlantic ridge earthquakes arrive systematically late on coastal New England seismic stations, compared to stations in New York (Menke & Levin, 2001). The delay (~4 s) is consistent with the reported position and intensity of the oceanic part of the NEA, and seems too large to be caused by a modest low−velocity anomaly that is limited to the continental margin (figure 1C).

The genetic association of the NEA with the New England plume track is based mainly on their geographic juxtaposition, and is thus circumstantial. Levin et al [2000b] and Tucker et al. [2001] suggest that lithospheric delamination occurred beneath New England at 400 Ma, as part of the continental collision that marked the final closure of the Iapetus Ocean. A compositional seismic wavespeed depression, due to the replacement of slab or continental lithosphere by resurgent Fe−rich asthenosphere, could perhaps explain the NEA on land. If the seismic anomaly indeed cross−cuts the Jurassic−age continental margin (as the surface wave data indicate), a connection with the New England plume is more plausible, as no other significant post−Jurassic tectonic events are known in its vicinity. It is somewhat surprising (at least to us) that such a significant slow shear anomaly would be present today, roughly 100 Ma after the passage of the plume, at least if the low shear velocity is equated with high mantle temperatures. Three competing explanations are: that the anomaly is 1) **anisotropic** or 2) **compositional** in nature; and 3) that a thermal anomaly is being **actively sustained** by some present−day asthenospheric flow field.

Since upper mantle anisotropy can cause variations in seismic velocity of up to +/− 6%, a resetting of the usual spreading−parallel fabric of the oceanic lithosphere to a different direction could plausibly cause a long−lived anomaly. As the normal spreading−parallel anisotropy of the Atlantic has a fast axis oriented sub−parallel to many of the ray−paths used in the Van der Lee & Nolet (1997) inversion, any disruption of it would appear seismically slow. There is indeed some evidence that the continental−scale pattern of North American mantle fabric is disturbed in northeastern North America (an area that includes the NEA), possibly because of the flow around an irregularity in the edge of the craton (Fouch et al. 2000). On the other hand, the continental part of the NEA has no small−scale anisotropic signal (Levin et al. 2000ab). Whether the oceanic part of the NEA has an anisotropic signal is not known, and will be a target of investigation in the proposed research.

Compositional heterogeneity has long been thought to be able to cause only very subdued (less than 2−3%)

seismic perturbations in the mantle lithosphere. However, the major−element compositional differences that distinguish "undepleted" from "depleted" mantle peridotite could explain a large portion of the NEA velocity contrast (Anderson and Isaak, 1995; Chai et al. 1997; Collins and Brown, 1998), and must be considered carefully, especially as the amplitudes of seismic velocity images can scale up and down with the choice of damping in tomographic inversion computations. In this scenario, the low velocities could indicate the replacement of the depleted lithospheric mantle with material from the undepleted asthenosphere.

If anisotropic and compositional effects can be ruled out, we are left with elevated temperature as the cause of low velocity anomaly. If the NEA were indeed thermal in origin, how would it be maintained? The absolute plate motion of North America (about N245E) (Gripp & Gordon 1990) is oblique to the trend of the New England seamounts (about N300E), making an angle of about 55 deg. It is perhaps possible that the flow of the asthenosphere past a preexisting groove in the lithosphere (as might have been caused originally by the plume) might set up more or less permanent perturbation in flow that advects enough heat upward into the groove to sustain it.

Of course there is a fourth possibility, namely that the the NEA represents an episode of lithospheric erosion that has nothing to do with the New England seamounts, but rather is being caused by some very recent (i.e. Tertiary) or even present−day process. We do not think this scenario very likely, since there is no recent volcanism or seismicity along the strike of the NEA. On the other hand, it might represent a hithertofore unrecognized type of lithosphere−asthenosphere interaction and therefore be quite exciting.

PROPOSED RESEARCH

General outline of experiment. We are proposing to image a swath across the NEA in the oldest oceanic lithosphere just east of the North American continental margin, centered on the position of Kelvin seamount. New seismic data (teleseismic body and surface wave) will be used to determine the width of the low−velocity anomaly, the depth to its top and bottom, its velocity contrast, whether it is bounded by fast edges, whether there is any disruption in anisotropic fabric, and its seismic attenuation.

In particular, we seek to answer the following specific questions, which will enable us to construct and interpret the nature of the lithospheric thinning:

1. **How much of the lithosphere was eroded?** We are operating under the assumption that the region above the slow velocities are the remains of the original, Jurassic−age lithosphere, and that all the material below it is either new lithosphere added after the Cretaceous plume event or is asthenosphere. This working hypothesis will of course have to be re−evaluated in light of the new measurements, but it serves to order the discussion here. The most important question we seek to answer is how much of the Atlantic lithosphere was eroded by the plume event. We want to establish the shape and cross sectional area of the erosion. This will be accomplished through teleseismic P, S and Rayleigh wave imaging, which when taken together, will be be able to determine both the seismic structure of the normal lithosphere at the edges of the study area, the structure in the central, eroded region, and the position of the transition. Of particular interest is whether the width is really the ~300 km estimated from the Van der Lee & Nolet (1997) image, or whether it is actually narrower.

2. **Is the remaining original lithosphere normal, or has it been altered?** Plume models such as those of Davies (1994) model erosion as a viscous flow problem, and do not model the effect of melt and volatile transport through the lithosphere. Nevertheless, such fluids can potentially have an important effect on the chemistry of the lithosphere, at least in regions close the channels in which transport occurs. We seek here to measure what fraction of the uppermost lithosphere shows evidence of being affected by such processes. The seismic structure of the shallow (<50 km depth) lithosphere will be determined by the dispersion of shorter−period Rayleigh waves.

3. **What is the cause for the slow velocities? Is this region lithosphere or asthenosphere?** As noted in the discussion above, the slow velocities could represent a chemical or fabric difference between the old Jurassic lithosphere and newer post−plume lithosphere, or it could represent hot asthenosphere at unusually shallow depths. We will seek to discriminate between these alternatives using measurements of seismic anisotropy (which is sensitive to fabric) and seismic attenuation (which is sensitive to temperature). We will use both SKS splitting and quasi−Love phases to quantify the anisotropy. SKS splitting provides an estimate of the vertically−integrated amount of anisotropy, and thus might be particularly good at quantifying the degree to which the "background" spreading−parallel fabric of the original Jurassic−age lithosphere has been disturbed. Quasi−Love phases are particularly useful in detecting the lateral boundary between two fabrics, and thus may be helpful in mapping out the extent of any disturbed region. SKS splitting has successfully been employed in an oceanic setting by Wolfe et al. (1998) and Smith et al. (2001). As noted above, quasi−Love waves (observed mainly on the vertical component of motion) have been successfully employed at Hawaii (Levin & Park 1998). P and S wave attenuation measurements are sensitive to the presence of near−solidus material, and thus can detect changes in the depth to the asthenosphere. Thus we would expect much higher attenuation in the central, plume−disturbed part of the study region, than at the normal lithosphere at its edges, assuming that the slow velocities represent a thermal anomaly. Thus images of differential P and S wave attenuation, based on teleseismic body wave phases, will be employed. Measurements of attenuation have been successfully made using OBS data by Smith et al. (2001).

4. **Is there any deep−rooted perturbation in the asthenosphere?** If the slow velocities are thermal in nature, then we would expect that some kind of perturbation in the mantle flow might be sustaining them. The anisotropy measurements might give some sense of the direction of this flow. Furthermore, we would expect some perturbation in the temperature field of the deeper part of the asthenosphere (>120 km depth) that correlates with the pattern of upwelling and downwelling. The teleseismic tomography, which will extend to 300−400 km depth, will provide a means of addressing this question. We would, for instance, be able to determine whether any localized regions of downwelling (such as apparently occur at Yellowstone) (Saltzer and Humphreys, 2001) are present.

EXPERIMENT DESIGN

OBS Array Design We are proposing to deploy an 18−OBS array centered roughly on Kelvin seamount, which is located on oceanic crust approximately 600 km offshore Massachusetts (figure 3A). The array is 650 km by 550 km, somewhat wider in the cross−plume−track direction than the along−plume−track direction. The pattern of OBS's is chosen to include several lines of OBS's oriented NNE−SSW, pointing towards the seismically active Caribbean, northern South America and northern mid−Atlantic ridge regions. These lines improve the resolution in the cross−plume−track direction. Each OBS will be equipped with a Differential Pressure Transducer (DPG), for sensing P and Rayleigh waves. At least 10 OBS's will be equipped with 3−component seismometers capable of recording S waves. The array deployment is planned to last 1 year, during which time we expect to record numerous teleseisms (figures 3B and 3C).

Anticipated Seafloor Noise Levels Some reviewers will be familiar with the difficulties of using seafloor seismic data for the types of analyses described here. We have two large OBS experiments for a basis of comparison: MELT on the southern EPR and LABATTS in the Lau back arc region (Zhao et al., 1997). Signal−to−noise levels at LABATTS were much superior to those at MELT, allowing a much broader seismic band to be recorded at the former, permitting traveltimes, shear wave splitting and attenuation to be measured with greater precision (Smith et al. 2001). Our proposed western Atlantic data should be comparable or better to the LABATTs data because the microseism peak in the western Atlantic has much less low frequency energy than seen in the Lau basin. Some sample SKS and P wave data, observed on a station in southwestern Long Island, NY, are shown in figure 4. While not from OBS's, they nevertheless give an indication of data quality along the Atlantic margin.

The MELT experiment was at a poor site for tomographic studies because high attenuation associated with the rise axis limited detection of short period arrivals to the band between 3 and 7s − right in the middle of a very energetic microseism noise peak. In contrast, attenuation under New England is low and high frequency teleseismic P waves (several Hz) are often seen. Noise levels at a few hertz in the oceans can be nearly as low as the best continental sites. The microseism peak in the Atlantic is invariably very subdued compared to the mid−Pacific (Webb, 1998) so that signal to noise ratios for short period arrivals could be as much as 20 dB better in the western Atlantic than for MELT.

Rayleigh waves were routinely detected with good signal to noise ratios during MELT in the period band from 20 to 60s on both the vertical component and the DPG (Forsyth et al., 1998). We expect to be able to extend the useful range to considerably longer period for this experiment because of improvements in the seismometer amplifiers (Webb et al., 2001), and because of a much subdued low frequency ocean wave climate which limits DPG observations in the Pacific to periods shorter than about 60s (Webb, 1998).

Long period horizontal component noise on freely deployed OBS's is primarily due to tilt noise from rocking of the instrument in ocean currents (eg. Crawford et al, 2000). This noise is much higher at longer period. The subdued Atlantic microseism peak coupled with low attenuation for shear waves should allow for much better observations of shear wave splitting because it will be possible to make these measurements at considerably shorter period than the 15−25s period used in MELT.

No seismic attenuation measurements were obtained from the MELT experiment because of the limited useful band of observation for the body waves, although high attenuation was evident from the low frequency content of the body waves. The useful bandwidth during MELT was limited by 1) a very energetic microseism peak and 2) high attenuation. The much lower attenuation coupled with a much less energetic microseism peak between 6 and 10 s period should permit good observations of attenuation (observations comparable to LABATTs; Roth et al., 1999).

OBS Deployment and recovery The OBS deployment of 18 instruments from the new OBSIP pool will be scheduled for the second year of the project. All of these instruments will be equipped with broadband differential pressure transducers, and at least ten with 3−component seismometers. Two cruises on an Oceanus−class research vessel, each 11 days long, will be necessary to deploy and recover the OBS's (assuming nominal Woods Hole MA to Woods Hole MA legs):

station time: 18 x 3.1 hrs per OBS = 56 hours
steaming time: 1800 nmi / 10 kt = 180 hours
contingency (e.g bad weather) = 24 hours
Total = 260 hrs = 11 days

Land Data We will also retrieve coeval data from all publicly−available permanent observatories in northeastern North America (Figure 3A, triangles) and Bermuda (assuming the GSN deployment schedule for that not−yet−installed station remains on track). These data will augment the ocean−based data and will enable us to include the part of the plume track that is near the continental margin in the interpretation. Data from earlier PASSCAL deployments in the northeast US (e.g., MOMA and NOMAD) will also be examined where data set integration is profitable.

METHODS OF DATA ANALYSIS

Traveltime Tomography. Traveltimes of telesesmic P and S body waves, determined using cross−correlation methods, will be used in first−order tomographic imaging. We have prior experience using teleseismic tomography imaging methods from the New York and southern New England region (Levin et al. 1995, 2000a), and are confident that we can achieve a resolution of about 50−75 km resolution across the NEA. One of us (W. Menke) has recently finished a new tomographic imaging code

ftp://lamont.ldeo.columbia.edu/pub/menke/raytrace3d.tar.Z

that uses a three−dimensional tetrahedral representation of velocity, and can image seismic velocity and attenuation for a variety of source and receiver geometries.

Local Estimates of Surface Wave Phase Velocity. Estimates of shear wave velocity variation with depth can be obtained in a standard way by inverting the regionally−localized, frequency−dependent phase velocity of fundamental−mode Rayleigh waves. Differential Pressure Gauges (DPG's) on the OBS's will provide the primary data. Both the 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 to determining lithospheric structure. Phase velocity localization will be performed using an estimation technique similar to the one popularized by Forsyth, in which the phase velocity is perturbed so as to match the evolution of the Rayleigh waveform across the array. This method provides a natural way of including the effect of both anisotropy and multipathing (which is sometimes present at the shorter periods).

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, the DSP's on the ocean−bottom instruments will be able to record them.

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, 2000a) the PIs measured hundreds of such phases on both temporary and permanent stations in nearby New York and New England. Horizontal component measurements made on OBS's are typically more noisy than those made on the vertical component. Nevertheless, SKS measurements were successfully made in the MELT and Southwest Pacific experiments (Wolfe et al. 1998; Wiens et al. 1995). 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; and the generally lower asthenospheric attenuation in this region, compared to that of a ridge or back−arc spreading setting. The PI's 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 (ie. at several depths), should any be present.

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.

MANAGEMENT PLAN

All of the PI's will broadly participate in all phases of the project. 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, will head the OBS array deployment. Once the data is collected, Menke will head the traveltime tomography analysis, Park will head the surface wave related data analysis, and Levin the splitting analysis. All participants will be involved in the interpretation and write−up of the results.

USE OF FACILITIES

The proposal requests use of two NSF−funded facilities: 1) We will request 18 long−deployment OBS's from the OBS pool for 12 months; and 1) We will request two cruises on an intermediate−class research vessel. Requests to use these facilities are have been filed with the relevant parties.

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.

1. Year 1: Cruise to deploy OBS array. Collect and analyze all available nearby land data for teleseisms relevant to the study.
2. Year 2: Cruise to recover OBS array; Initial data processing of OBS data; Begin tomography, dispersion and shear wave analyses. Present preliminary results at a scientific meeting.
3. Year 3: Final data processing; Interpretation of results; Writing up papers. Presenting initial results at a scientific meeting.

DISSEMINATION OF RESULTS

We will submit the OBS data to the OBSIP facility within two years of its collection, and to any other public archives that are required by NSF and the other institutions 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 make a best−faith effort to publish them rapidly in a peer−reviewed journal.

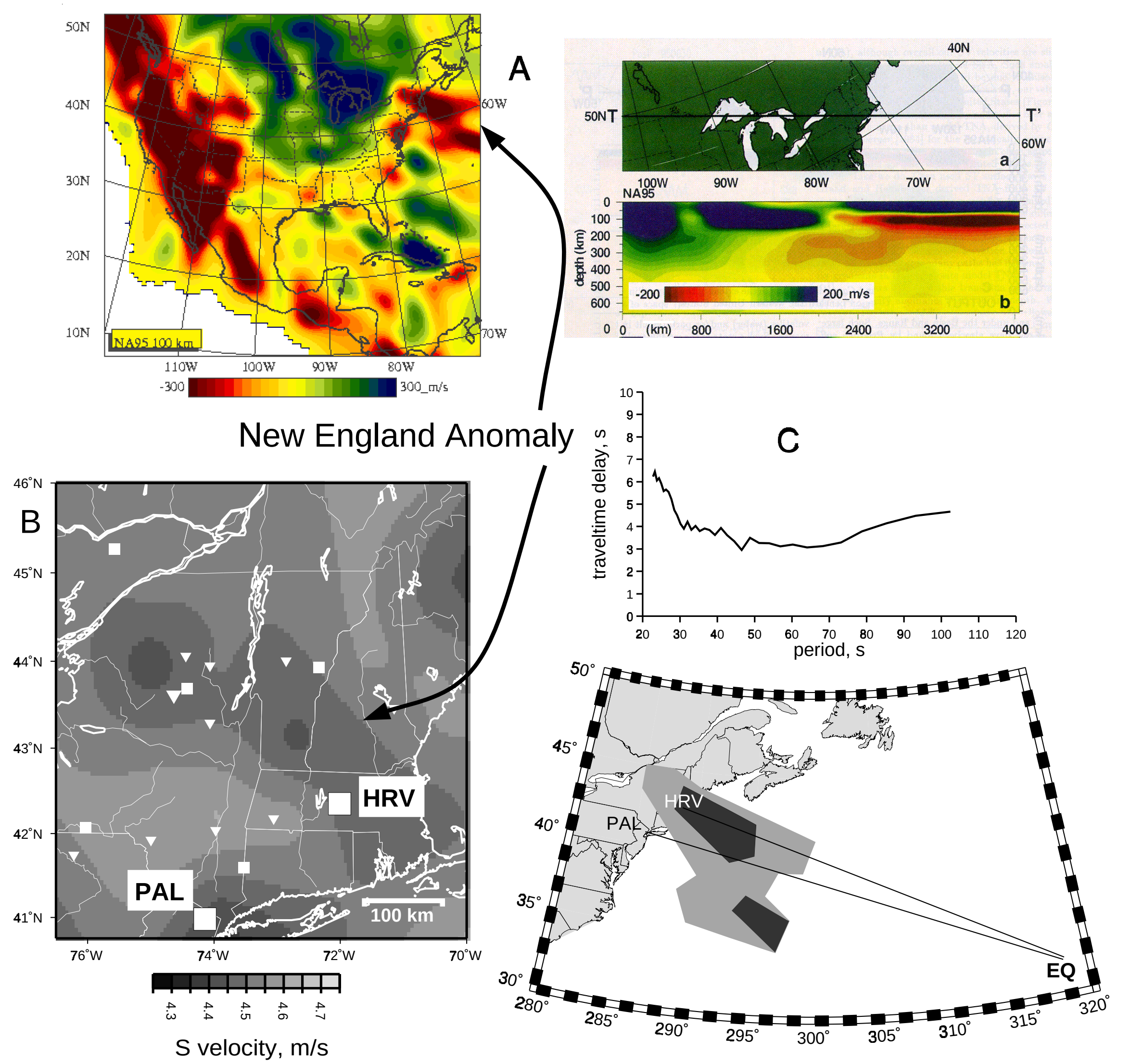


Figure 1. (A)Tomographic image of shear wave velocities under the North America and the adjuscent western Atlantic (from van der Lee and Nolet, 1997 clearly shows a major slow velocity anomaly). (B) S wave velocity distribution 100-200 km under New England and New York (from teleseismic S wave tomography, Levin et al., 2000) suggests an anomaly with a similar strike, but not as wide. (C) Differential traveltime (upper graph) between HRV and PAL for a Rayleigh wave from the Oct. 5, 2000 Mid-Atlantic Ridge earthquake (lower graph), after a correction that adjusts the two stations to the same source-receiver range. The 4s delay at HRV refects propagation along a path that is about 3% slower than the path to PAL, an amount consistent with path to HRV being through the center of the New England Anomaly, and the path to PAL being along its southern edge. Outline of the New England Anomaly is sketched in dark and intermadieate shading.

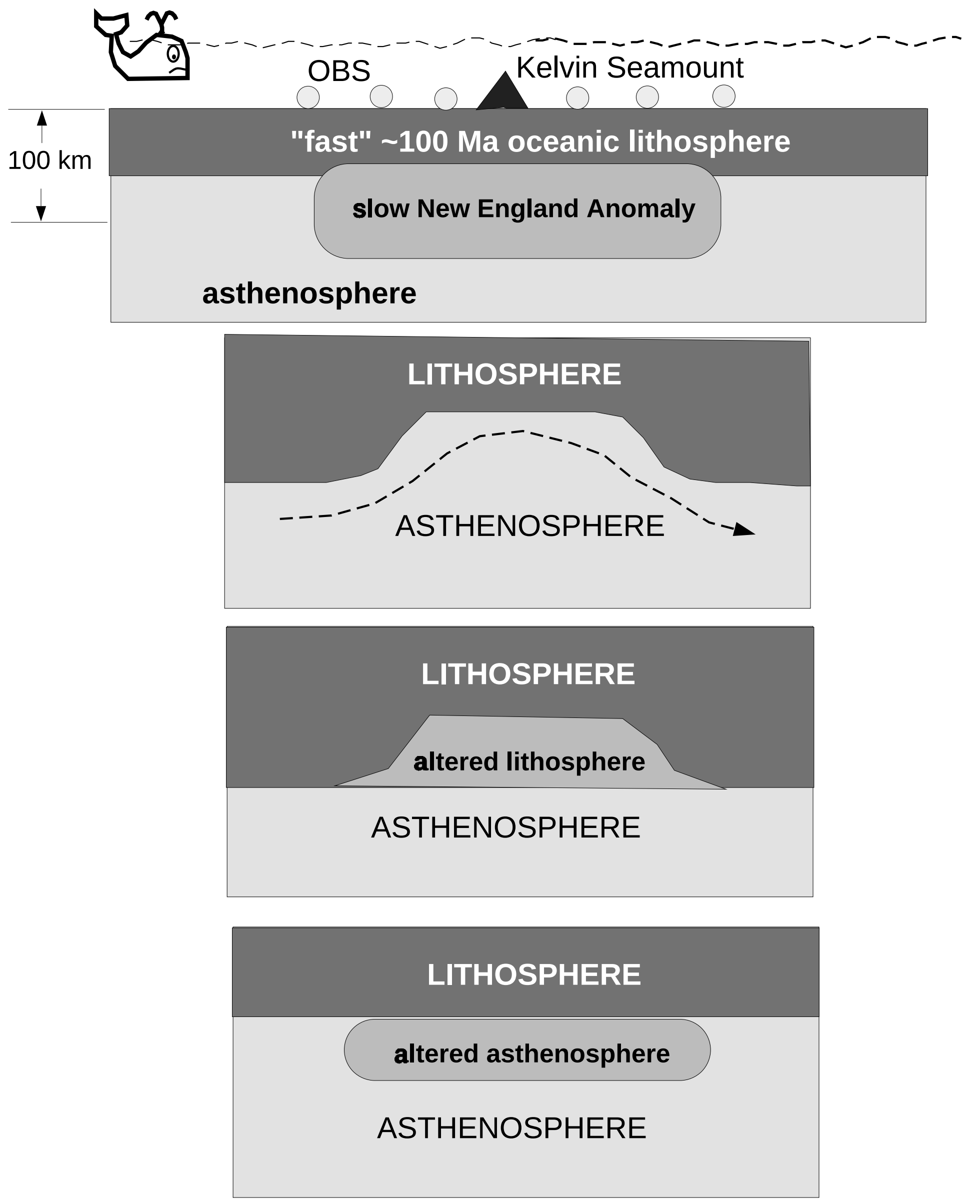
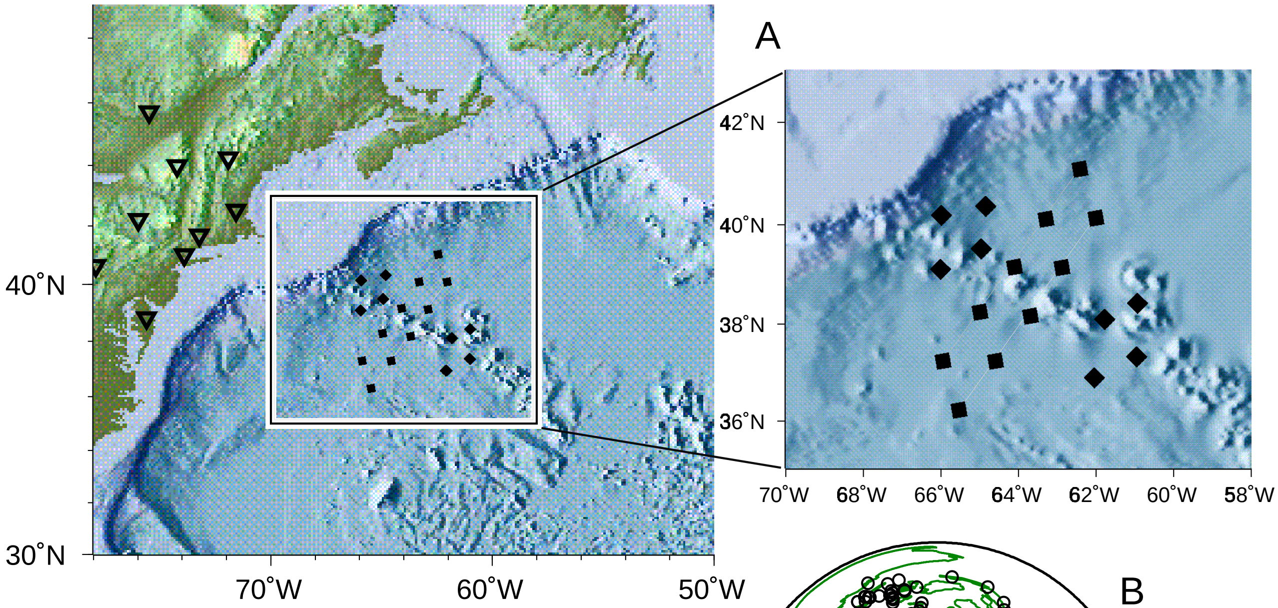
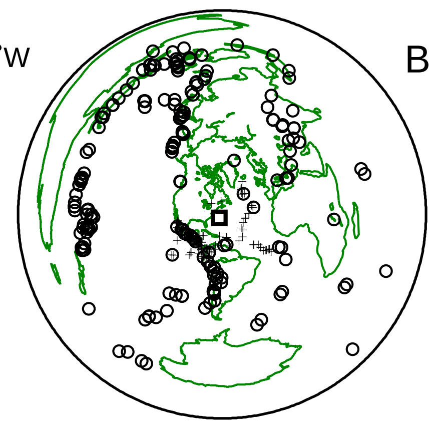


Figure 2. Schematic rendering of the New England Anomaly seismic structure, and possible scenarios that might expalin it.





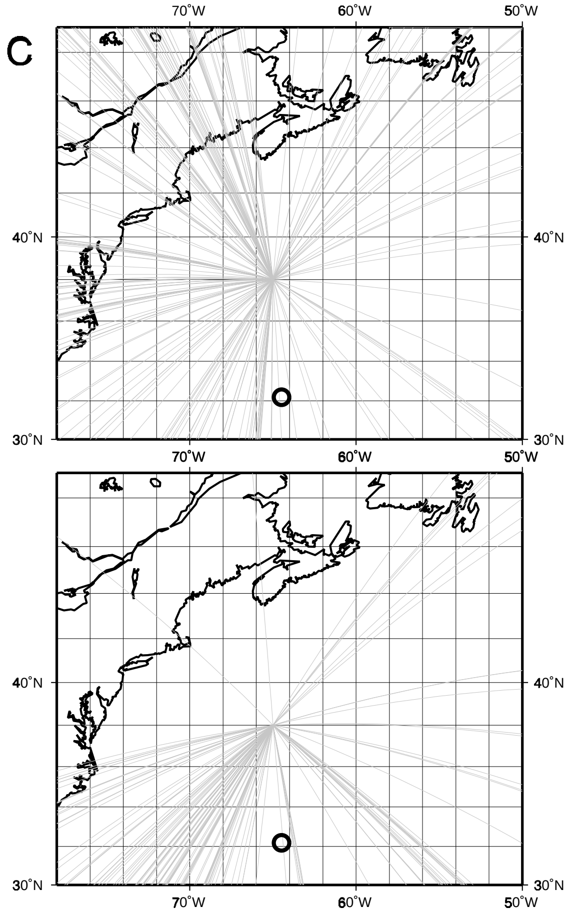
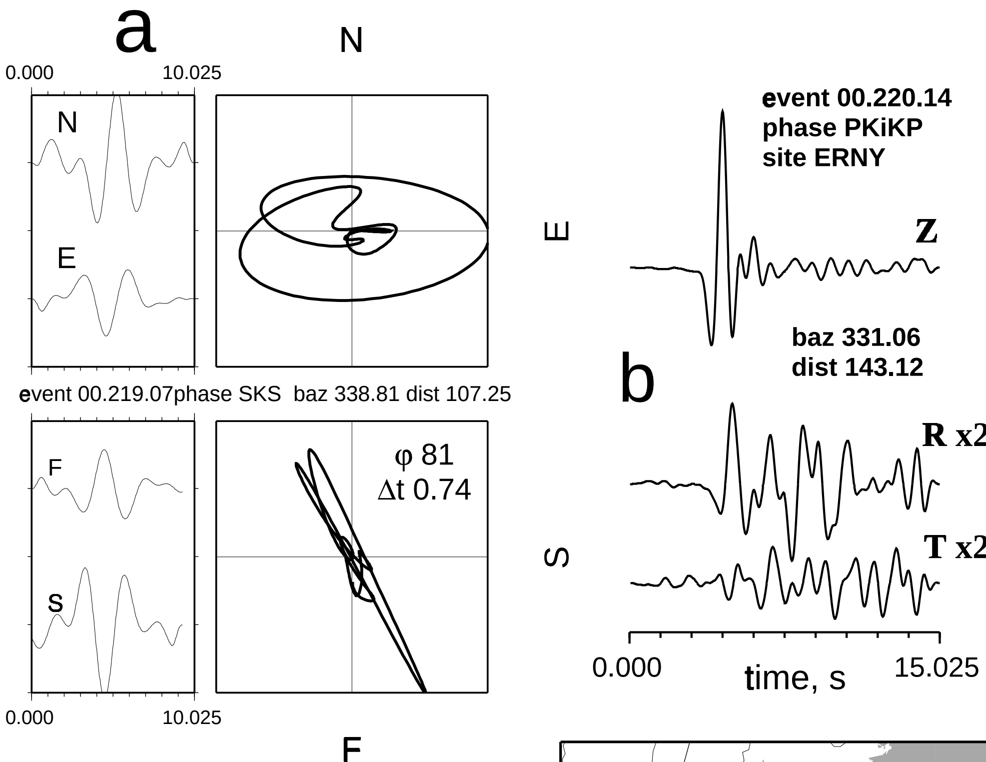


Figure 3. (A) Geometry of the proposed 18-OBS array. Solid diamonds show provisional OBS locations, inverted triangles show existing permanent seismic observatories on land. (B) Typical distribution of global seismicity relative to the proposed array (open box). Circles show earthquakes with M>6 during 1997-98. Ray distribution for these earthquakes is shown in C (top panel). Crosses show earthquakes with M>5 within 40° from the center of the proposed array. Ray distribution for these earthquakes is shown in C (lower panel). An open circle on ray diagrams denotes Bermuda.



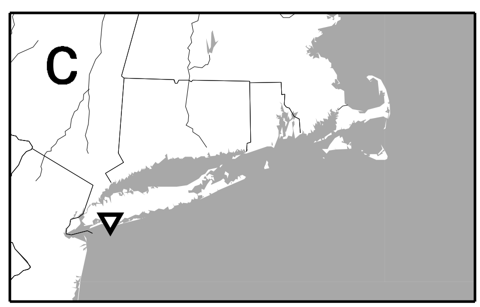


Figure 4. Sample observations from ERNY (East Rockaway, NY). Installation consisted of a REF TEK data logger and a CMG40T sensor placed on the surface in a small fiberglass enclosure.
a) Shear wave splitting measurement on an SKS phase. Waveforms are filtered in 0.2-0.5 Hz passband. Upper panels show observed data, lower panels show waveforms corrected for the effect of anisotropy. Splitting parameters are shown in the lower right panel: fast direction $\varphi = 81°$, delay is 0.74 s. Note reduced ellipticity of particle motion.
b) PKiKP phase from an earthquake 143° away is particularly bright due to a caustic in the travel time curve. Waveforms are filtered in 0.025-1.5Hz passband.
c)Location of the station on the oceanward shore of Long Island.