Workshop on Seismic Wave Scattering and Noise Correlation Proceedings

February 16-17, 2009
Tohoku University, Sendai, Japan

Sponsors
Tohoku University Global COE Program
"Global Education and Research Center for Earth and Planetary Dynamics"
IASPEI task group on “Scattering and Heterogeneity”
Workshop on
“Seismic Wave Scattering and Noise Correlation”
Tohoku University, Sendai, Japan (Feb. 16 -17, 2009)

**Objectives:** For the study of heterogeneities in the earth medium, it is useful to analyze scattering phenomena of seismic waves. Coda wave envelope analysis is often used for quantifying statistically the distributed random heterogeneities. Envelope broadening of short-period seismic waves is also useful to detect the strength of medium heterogeneities along the seismic ray path. In addition to these methods focusing on the amplitude information, recently, there have been rapid developments in a method by using noise correlation to detect the medium heterogeneity and its temporal variation focusing on the phase information. This workshop will make a forum for exchanging scientific ideas in these scientific fields among geophysicists in Germany and Japan.

**Convener:** Haruo Sato (Tohoku University) and Michael Korn (Leipzig University)

**Period:** Feb. 16 -17, 2009

**Location:** Room #604, Physics Building A, Graduate School of Science, Tohoku University, Sendai, Japan

**Sponsors:** GCOE Earth Science, Tohoku University, IASPEI task group on “Scattering and Heterogeneity”

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**PROGRAM**

**Feb. 16 (Mon)**

15:00 **Opening Remark**  
H. Sato (Tohoku Univ.)

15:10 -15:40 **Oral Session**

Christoph Sens-Schöpfelder, Ludovic Margerin, Michel Campillo, Energy transfer simulations with laterally varying heterogeneity: An explanation for Lg-wave blockage by the western Pyrenees

16:00-17:30 **Poster Session**

Jens Przybylla and Michael Korn, Radiation transport of elastic waves in random media with multiple scales
Kaoru Sawazaki, Haruo Sato, and Takeshi Nishimura, Simulation of vector-wave envelopes in 3-D random elastic media for non-spherical radiation source based on the stochastic ray path method

Titi Anggono, Takeshi Nishimura, Haruo Sato, Hideki Ueda, and Motoo Ukawa, Temporal changes of seismic velocity of shallow structure associated with the 2000 Miyakejima volcano activity as Inferred from ambient seismic noise correlation analyses

Kentaro Emoto, Haruo Sato, and Takeshi Nishimura, Synthesis of seismic-wave envelopes on the free surface of a random medium by using angular spectrum

18:00  Ice Breaker (At restaurant “Shikisai”, Aoba-Kaikan, 15 min walk)

Feb. 17 (Tue)

9:00  Opening Remark  H. Sato

9:00 -12:15  Oral Session A  Moderator H. Nakahara

9:00 - 9:30  Michael Korn, Short period Greens function retrieval from ambient noise on the km to 10km scale: the NORSAR case

9:30 - 10:00  Christoph Sens-Schönfelder and Eric Larose, Studying a dynamic process in the lunar crust with passive image interferometry

10:00 - 10:30 Ulrich Wegler, Hisashi Nakahara, Christoph Sens-Schönfelder, Michael Korn, and Katsuhiko Shiomi, Temporal changes in the source region of the mid Niigata Prefecture earthquake of 2004

10:30 - 10:45 Break

10:45 - 11:15 Shiro Ohmi and Kazuro Hirahara, Temporal variations of crustal structure in the source region of the 2007 Noto peninsula earthquake, central Japan, using ambient seismic noises

11:15 - 11:45 Hisashi Nakahara, Ulrich Wegler, and Katsuhiko Shiomi, Monitoring seismic velocity changes using passive image interferometry: An application to the 2005 West off Fukuoka prefecture, Japan, earthquake (Mw 6.6)

11:45 - 12:15 Takuto Maeda, Yohei Yukutake, and Kazushige Obara, Recurrence of the
seismic velocity change associated with earthquake swarm activities in NE Kyushu, Japan, revealed by the seismic Interferometry

12:15 - 13:30 Lunch (Bento Lunch Box)

13:30 - 17:00 Oral Session B   Moderator U. Wegler

13:30 - 14:00 Kazuo Yoshimoto, Kenya Sakurai, Hisashi Nakahara, Shigeo Kinoshita, Hiroshi Sato, Seismic basement structure beneath the Kanto plain, Japan inferred from the seismic interferometry for strong motion records

14:00 - 14:30 Takashi Tonegawa, Kiwamu Nishida, Toshiki Watanabe, and Katsuhiro Shiomori, Seismic interferometry of teleseismic S-wave coda for detection of body waves-An application to the Philippine Sea slab underneath the Japanese Islands-

14:30 - 15:00 Kiwamu Nishida, Jean-Paul Montagner, and Hitoshi Kawakatsu, Global surface wave tomography using seismic hum

15:00 - 15:15 Break

15:15 - 15:45 Haruo Sato, Retrieval of the single scattering Green function from the cross-correlation function in a scattering medium illuminated by surrounding noise sources

15:45 - 16:15 Mare Yamamoto, Haruo Sato, and Takeshi Nishimura, Multiple scattering and mode conversion as revealed from active seismic experiments at active volcanoes

16:15 - 16:45 Eduard Carcole and Haruo Sato, Attenuation of short-period S-waves in Japan: high resolution maps of intrinsic absorption, scattering loss and coda decay

16:30 - 17:00 Discussion

17:00 Closing Remark   M. Korn

18:00 Reception (Some restaurant in downtown)

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ABSTRACT
Energy transfer simulations with laterally varying heterogeneity: An explanation for Lg-wave blockage by the western Pyrenees

Christoph Sens-Schönfelder¹, Ludovic Margerin², Michel Campillo³

1. Institut of Geophysics and Geology, University of Leipzig, Leipzig, Germany
   E-mail: sens-schoenfelder@uni-leipzig.de
2. Centre Européen de Recherche et d’Enseignement des Géosciences de l’Environnement (CEREGE), Aix-en-Proevnce, France, E-mail: margerin@cerege.fr
3. Laboratoire de Geophysique et Tectonophysics, Universite Joseph Fourier, Grenoble, France
   E-mail: campillo@ujf-grenoble.fr

Introduction

The phenomenon of Lg-blockage refers to an anomalous attenuation of seismic waves in the Earth’s crust. It is widely known from thick marine sediment basins that trap the crustal waves in the low velocity region. Some mountain ranges like the European Alpine range and the Pyrenees do also show such a strong attenuation of crustal seismic phases. In these cases the macroscopic velocity structure does not contain sufficient velocity contrast and attempts to model the Lg-blockage by the western Pyrenees on the basis of the large scale velocity structure failed (Chazalon, 1993). So it was speculated that scattering at small scale heterogeneity is an important factor for the attenuation of crustal waves. To test this hypothesis we analyzed new data of Spanish earthquakes that were recorded in France after crossing the Pyrenean mountain range and devised a Monte-Carlo algorithm for the simulation of these seismogram envelopes based on radiative transfer theory.

Observation and Modeling

The new data set supports the observation of Lg-blockage made in previous studies. Depending on the position where the waves crossed the Pyrenees the amplitude ratio between curstal and mantle phases changes dramatically. Whereas the waves that cross the eastern part of the Pyrenees show the typical shape of seismograms at regional distances that are dominated by the Pg and Sg (Lg) waves, these phases are almost absent after propagation through the western Pyrenees.

The algorithm that we implemented to model the seismogram envelopes simulates the elastic transfer of seismic energy in a model that consists of a layer with laterally variable heterogeneity above a half space with a vertical gradient of the mean velocity. The algorithm takes into account the conversion between P- and S-waves at the surface as well as at the interface between mantle and crust. To model the differences between the eastern and the western parts of the Pyrenees the model includes an additional body in the crust beneath the western part that differs in the scattering and attenuation properties from the surrounding material. Heterogeneity is modeled with fluctuations with an exponential auto-correlation function. With a genetic algorithm we estimated parameters of the random media in the mantle, the crust, and the supplementary body that best explain the observed envelopes for both, the propagation through the eastern and western parts of the mountain range.
Results
Our modeling indicates that the strong attenuation of crustal phases might indeed be caused by increased small scale heterogeneity. Intrinsic attenuation on the contrary does not suffice to explain the observation.
The best result is obtained with slightly increased attenuation and significantly increased heterogeneity in the body under the western Pyrenees. Figure 1 shows the observed and modeled seismogram envelopes for the eastern and western Pyrenees. Snapshots of the energy field are shown in figure 2.

Figure 1 Observed (black curves) and modeled (red curves) envelopes of seismograms for propagation through the undisturbed eastern (left figure) and western (right figure) part of the Pyrenees

Figure 2 Snapshots of the energy distribution at the surface of the crust 105s (left figure) and 180s (right figure) after an earthquake in Spain. The black box indicates the region of increased heterogeneity. Nicely seen is the gap in the Pg wave (left) and Lg wave (right) whereas the wave fronts of the mantle phases are continuous.
1. Introduction

Radiation transport theory (RTT) describes the propagation of wave energy in scattering media that means especially in media with small scale heterogeneities. For this we look at squared seismogram envelopes which are proportional to wave energy. RTT is one of the most powerful tools to picture the multiple scattering regime of waves and to obtain informations about small scale heterogeneities. Basic validity assumptions of RTT are: fluctuations of wave velocities are weak, waves are scattered incoherently and correlation length is of the same order of magnitude as the wavelength. One of the simplest models for small scale heterogeneities is a medium with random fluctuations around a constant background velocity, that are characterized by a autocorrelation function (ACF), a characteristic scale called the correlation length $\alpha$ and fluctuation strength $\varepsilon$. However, results from borehole velocity logs show, that there is a need for more than one scale to correctly characterise small scale heterogeneities of the earth medium.

2. Result

Here we present Monte Carlo simulations of RTT in random media with more than one scale. To obtain such a model we superpose two Gaussian ACF's, with different correlation lengths. The numerical simulations show, how wave energy can propagate through a random medium with multiple scales. We compare our results with Monte Carlo simulations in a single scale random medium. This comparison shows especially in P-coda clearly visible differences between a single scale and a multiple scale random medium (fig.1).

3. Discussion

Continuous random media with one correlation length and fluctuation strength are some of the simplest models for small scale heterogeneities of the earth. Probably these models are to simple for real earth heterogeneities. If the medium contains more than one typical scale, waves interact most intensively with scales that are of the same order of the wavelengths. The idea of the need of multiple scales is not a hypothetical one. Borehole analysis has shown this too (e.g. Goff und Holliger, JGR 1999). We have used a Gaussian medium with one and a Gaussian medium with two scales only (fig.1). In fig.1 we see clearly the influence of the smaller scale on the P coda. Certainly it is possible, that real earth structures need another kind of multiple scale media. The simulations in this poster show the behavior of waves propagating through media with multiple scales, in a qualitative way. The superposition of different random media is only an approximate step to describe the complicated structure in the small scale range of the earth medium.
Figure 1. Monte Carlo simulations of one- (black curves) and two-scale Gaussian random media (red curves), for different distances. The one scale medium is for $a k_s = 16$ and the two scale medium $a_1 k_s = 16$ and $a_2 k_s = 8$. Here is $a$ the correlation length of the single scale medium and $a_1$ and $a_2$ are the scales of the double scale medium. $k_s$ is the wave number of the S waves. In both random media fluctuation strength is $\varepsilon = 3\%$. 
Simulation of vector-wave envelopes in 3-D random elastic media for non-spherical radiation source based on the stochastic ray path method

Kaoru Sawazaki¹, Haruo Sato¹, and Takeshi Nishimura¹,

1. Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan,
   E-mail: sawa@zisin.geophys.tohoku.ac.jp

1. Introduction

In high frequency (>1Hz) seismograms of local earthquakes, seismic waves are clearly observed even in the direction of null-axis of the source radiation pattern, and also the excitation of the transverse (longitudinal) components is apparently observed for P(S)-waves. These phenomena are explained by scattering of seismic waves in random inhomogeneities in the lithosphere. Synthesis of vector-wave envelopes based on the Markov approximation has been precisely studied for an isotropic radiation source. However, those for a non-spherical radiation source have been studied only by Sato and Korn (2005) for 2-D case. In this study, we propose a method to synthesize vector-wave envelopes for a non-spherical radiation source in 3-D random media by using the stochastic ray path method.

2. Derivation of Angular Spectral Function

In the case that the wavelength is much shorter than the correlation distance of a random medium, forward scattering dominates and conversion scattering becomes negligible. In such a condition, the Markov approximation is very effective to describe wave envelopes near the direct wave arrival (Sato and Fehler, 1998). Based on this assumption, we describe the derivation of the angular spectral function (ASF).

We imagine an ensemble of random inhomogeneous media, which is characterized by a von-Karman type power spectral density function (PSDF)

\[ P(m) = \frac{8\pi^{3/2} a^3 \Gamma(\kappa + 3/2)}{\Gamma(\kappa)(1 + a^2 m^2)^{3/2}}, \]

where \( \epsilon, a, \) and \( m \) represents rms amplitude, correlation distance, and wavenumber of the velocity fluctuation, respectively. The parameter \( \kappa \) controls the roll-off of the PSDF at large wavenumbers. We first study the bending process of seismic rays radiated from a point source at the origin, where the medium is divided into many spherical layers with a thickness \( \Delta r \). We introduce the mutual coherence function (MCF) \( \Gamma \), which is an ensemble average of cross-correlation of the wavefield at different locations on the transverse plane which is orthogonal to the global ray direction. Neglecting backward scattering and using causality, we can derive the master equation for the MCF as

\[ \frac{\partial}{\partial r} \Gamma(r) + k_0^2 [A(0) - A(r_\perp, r)] \Gamma(r) = 0, \]

where \( A(r_\perp, r) \) is the longitudinal integral of auto-correlation function of the velocity fluctuation, and \( k_0 \) is the wavenumber of seismic wave. Solving eq. (2), we obtain the MCF at the distance \( r + \Delta r \) as

\[ \Gamma(r_\perp, r + \Delta r, k_0) = \Phi(r_\perp, \Delta r, k_0) \Gamma(r_\perp, r, k_0), \]

where
\[ \Phi(\mathbf{r}_{1d}, \Delta r, k_0) = e^{-k_0^2 [d(0) - d(r_{1d})]/\lambda}, \]  

which is the transfer function of the MCF for thickness \( \Delta r \). In the wavenumber domain, eq. (3) is converted into

\[ \tilde{\Gamma}_1(k_\perp, r + \Delta r, k_0) = \frac{1}{(2\pi)^3} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} d\mathbf{k}_\parallel \Phi(\mathbf{k}_\perp, \Delta r, k_0) \tilde{\Gamma}_1(\mathbf{k}_\perp, r, k_0), \]

where \( \tilde{\Gamma}_1 \) gives the ASF, which is equivalent to the distribution of ray angles.

3. Stochastic ray path method

For an increment of small distance \( \Delta r \) from a spherical layer to the next spherical layer, each seismic ray is bent due to the velocity inhomogeneity following a stochastic random process. The ASF at each layer boundary is repeatedly calculated by the Monte-Carlo method, where the probability density function of the scattering angle distribution is obtained from \( \Phi \). We shot energy particles from the source with the weight of the radiation pattern. The particles propagate through the medium with the mean velocity; \( V_P \) or \( V_S \). Taking the projection of the oscillation direction of the energy particle at the outermost layer to the unit base-vectors, we obtain the energy partition for the three components. Calculating the accumulated travel time from the source to the outermost layer for each particle, we obtain the histogram of travel times and obtain three-component MS envelopes. This method is called the stochastic ray path method originally proposed by Williamson (1972). We show the schematic illustration of the stochastic ray path method in Figure 1.

4. Results and conclusions

In Figure 2, we show the radiation pattern of P and S waves for a double-coupled source, and the definition of the spherical coordinate system. We show the three-component 10Hz envelopes for several directions at the 100 km distance in Figure 3 by assuming the following parameters: \( \varepsilon = 0.05, a = 5 \text{km}, \kappa = 0.5, V_P = 6.0 \text{km/s}, \) and \( V_S = 3.46 \text{km/s} \). We find an excitation of amplitude at a receiver in the direction of null-axis for both P- and S-waves, which is caused by bending of the particle by the inhomogeneity of the medium. The azimuthal dependence of the vector-wave envelopes reflects the source radiation pattern near the direct arrival; however, the azimuthal dependence gradually diminishes as the lapse time increases. In Figure 4, we show envelopes at the different hypocentral distances, which show the weakening of the azimuthal dependence with the increase of hypocentral distance. From Figure 5, we can see that envelopes are sharper in lower frequencies. These simulations give a positive insight into the observed fact that the wave envelopes are rather insensitive to the radiation pattern especially in high frequencies at long distances and for long lapse times.
Figure 1. Schematic illustration of the stochastic ray path method for a double coupled source.

Figure 2. Squared amplitude of the normalized radiation pattern for P- and S-waves for a double coupled source.

Figure 3. Azimuthal dependence of three component RMS envelopes (10Hz) for P- and S-waves at the 100km distance.
Figure 4. Square root of three-component sum of S-wave envelopes for 10Hz at different hypocentral distances.

Figure 5. Square root of three-component sum of P- and S-wave envelopes for 10Hz and 2Hz at the 100km distance.
Temporal Changes of Seismic Velocity of Shallow Structure Associated with the 2000 Miyakejima Volcano Activity as Inferred from Ambient Seismic Noise Correlation Analyses

Titi Anggono$^1$, Takeshi Nishimura$^1$, Haruo Sato$^1$, Hideki Ueda$^2$, and Motoo Ukawa$^2$

1. Graduate School of Science, Tohoku University, Aoba-Ku, Sendai, 980-8578, Japan, Email: titi@zisin.geophys.tohoku.ac.jp, nishi@zisin.geophys.tohoku.ac.jp, sato@zisin.geophys.tohoku.ac.jp
2. Research Institute for Earth Science and Disaster Prevention, Tsukuba, 305-0006, Japan, Email: ueda@bosai.go.jp, ukawa@bosai.go.jp

1. Introduction
Detection of the Earth internal structure is important issue for the earth science, and temporal changes of seismic velocity has become interesting issue in the prediction of earthquakes and volcanic eruptions. Coda wave interferometry (Snieder, 2006) tries to exploit the sensitivity of the coda waves, which are a later portion of a seismogram following the direct P or S-wave arrivals, to estimate slight changes in the medium from correlation of coda waves before and after the perturbation. Recently, small changes of seismic velocity have been detected by using the idea of coda wave interferometry associated with the earthquake (e.g. Peng and Beng-Zion, 2006) or volcanic eruption (e.g. Brenguier et al., 2008). In this study, we analyze the ambient seismic noise records at Miyakejima volcano, which is located at 170 km to the south of Tokyo, Japan, in order to study the behavior of volcanic structure associated with the 2000 Miyakejima eruption that started with the magma ascent and migration on June 26 – 27, and followed by caldera formation from July to August.

2. Data analyses and results
We analyze the ambient seismic noise recorded at three NIED seismic stations (MKK, MKT, and MKS) (Figure 1) at Miyakejima from July 1999 to December 2002. We apply cross correlation analyses to the seismic records of vertical component of short period seismometers (1 s). The data are sampled at a frequency of 100 Hz with an A/D resolution of 16-bit. We calculate cross correlation functions (CCFs) for time window of 60 s for each station pair. We stack the CCFs for each month and bandpass filter the stacked data at frequency band 0.4 – 0.8 and 0.8 – 1.6 Hz. The stacked CCFs, which may represent the Green function between two stations, at station pairs MKK – MKS (the distance is 1.8 km) and MKT – MKS (the distance is 3.9 km) show wave packets with large amplitudes at both sides (positive and negative time delays). The wave packets propagate at group velocities of about 0.8 – 1.0 km/s. The stacked CCFs for MKK – MKT (the distance is 3.1) is one sided (negative time delay). Such asymmetric might be due to the inhomogeneous distribution of propagation direction of ambient seismic noise, so we do not use the data for the following analyses.

Comparing the CCFs obtained for periods after the 2000 eruption with that of stacked from July 1999 to May 2002, we observed phase difference of the main wave packet. Our results show that for station pair MKK – MKS, whose path crosses the northern part of the island, velocity increased about 2.1 % and 2.0 % at frequency band 0.4 – 0.8 and 0.8 – 1.6 Hz after the 2000 volcanic activity. For MKT – MKS, whose path closely crosses the newly formed caldera, we estimate the velocity decrease of about 2.0 % at frequency 0.4 – 0.8 Hz (Figure 2).

3. Discussions
Nishimura et al. (2005) suggested that the changes of seismic velocity observed at Iwate volcano might be related to the dilatation due to the volcanic pressure source beneath the volcano. The dilatation in the crust due to the magma pressurization also might be responsible for the occurrence of velocity changes observed at Piton de la Fournaise (Brenguier et al., 2008). Wegler et al. (2006) proposed that the increase of shear wave velocity before the occurrence of 1996 Merapi volcano eruption assumed to be associated with the increasing pressure inside the volcano. Such dependency of rock velocities on stress was proven in laboratory experiments (e.g. Nur, 1971; Gret et al., 2006). That velocity increase and decrease at Miyakejima Island might also be caused by the stress increase or
decrease in the shallow structure due to volcanic pressure source. Volcanic gas permeation in the volcanic edifice might be the other candidate that causes changes in the seismic velocity.

4. Conclusions

We have observed changes in the seismic velocity from the stacked CCFs of ambient seismic noise associated with the 2000 Miyakejima activity. Velocity increase of about 2.0% and velocity decrease of about 2.0% are observed at frequency range of 0.4 – 1.6 Hz. Such velocity increase and decrease observed might be caused by the stress increase or decrease due to the 2000 activity although other candidates might also be considered.

Figure 1. (a) Map showing epicenter distribution during the 2000 activity and station locations in Miyakejima Island, (b) temporal distribution of earthquakes (Ueda et al., 2006).

Figure 2. Travel time difference for station pairs MKK – MKS and MKT – MKS for frequency band 0.4 – 0.8 and 0.8 – 1.6 Hz. Blue solid circles are time difference observed at the positive time delay and red solid circles are time difference observed at the negative time delay.
Synthesis of seismic-wave envelopes on the free surface of a random medium by using angular spectrum

Kentaro Emoto, Haruo Sato, and Takeshi Nishimura
Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan,
E-mail: emoto@zisin.geophys.tohoku.ac.jp

1. Introduction
Short-period seismograms of earthquakes are quite complex due to the high sensitivity to small-scale structures in the lithosphere. When the wavelength is smaller than the correlation distance of random medium, Markov approximation is a powerful stochastic method to synthesize seismic wave envelopes. Sato (2006) derived the analytical solutions of the vector-wave envelopes in random elastic medium. Kubanza et al. (2007) applied the solution to the analysis of teleseismic P-waves for the measurement of the lithospheric heterogeneity. In this study, we synthesize the vector-wave envelopes on the free surface of a random medium based on the Markov approximation to build a more realistic model.

2. Method
In the Markov approximation, mean square (MS) envelopes are calculated by using the Fourier transform of the two frequency mutual coherence function (TFMCF) with respect to the angular frequency. The TFMCF is statistically defined on the transverse plane which is perpendicular to the global ray direction. The Fourier transform of the TFMCF with respect to the transverse coordinates gives the angular spectrum, which describes the distribution of ray directions. By projecting the angular spectrum to each component and then integrating it in the wavenumber space, we obtain the MS envelopes in infinite media. In order to synthesize the each component of MS envelope on the free surface, we propose to multiply the amplification factor on the free surface instead of the simple projection. The amplification factor is the amplitude of each vector component on the free surface for the incidence of a plane P or S-wave with unit amplitude.

3. Result
For the calculations, we assume that the 3-D random media are characterized by a Gaussian autocorrelation function (ACF). We use the following typical stochastic parameters of the lithosphere: averaged P-wave velocity is 6 km/s, averaged S-wave velocity is 3.46 km/s, correlation distance is 5 km, root mean square fractional velocity fluctuation is 0.05, and the thickness is 100 km. For the incidence of an impulsive plane P-wavelet, both component MS envelopes are amplified by nearly a factor of 4 on the free surface (Figure 1). In precise examinations, however, amplification rate of the horizontal component varies with reduce time: 4.8 at the peak and gradually decreases. For the incidence of an impulsive plane S-wavelet, both components are also amplified by a factor of almost 4, but deformations of envelopes are stronger than those of the incident P-wavelet case. In the vertical component MS envelope on the free surface, the peak delay time becomes shorter than that in infinite media and the amplification rate is 4.6 at the peak. In the horizontal component MS envelope, the amplification rate is 3.2 at the peak.

4. Comparison of the Markov approximation with the finite difference simulation in 2-D
In order to confirm the validity of our method, we conduct finite difference (FD) simulations in 2-D Gaussian ACF type random media for the vertical incidence of a plane P-wavelet. By using the
same procedure as used in the 3-D case, we can obtain the MS envelopes on the free surface of a 2-D random medium based on the Markov approximation. As a result, we find a good coincidence between the MS envelopes derived by our method and those calculated by the FD simulations. (See Figure 2)

5. Conclusion

We have succeeded in the synthesis of MS envelopes on the free surface of random elastic media characterized by a Gaussian ACF for the vertical incidence of a plane wavelet on the basis of the Markov approximation. This study gives a solid mathematical base for the practical analysis of teleseismic waves for the spectral structure study of random velocity inhomogeneities in the lithosphere.

![Figure 1](image1.png)

**Figure 1.** (a) Comparison of MS envelopes between those on the free surface (solid) and those in the infinite media (dash) for the incidence of an impulsive plane P-wavelet. MS envelopes in infinite media multiplied by a factor of 4 are shown by broken curves with deep colors. (b) Zoom up of the horizontal component.

![Figure 2](image2.png)

**Figure 2.** (a) Comparison of MS envelopes calculated by our method with those calculated by FD simulation on the free surface of a 2-D random medium for the incidence of a Kupper wavelet with dominant frequency 2 Hz. Red and blue curves show the vertical and horizontal component Markov envelopes, respectively. Orange and green curves show the vertical and horizontal component FD envelopes, respectively. Light orange and light green shadows show the plus and minus 1 standard deviation of the vertical and horizontal components, respectively. (b) Zoom up of the horizontal component.
1. Introduction
It has been well established theoretically that the cross correlation of a diffuse wave field between two seismic stations is equivalent to the Green’s function, i.e. to the wave field generated by a point source at one of the receivers and recorded at the second one. Ambient seismic noise from distributed sources at the Earth’s surface around the receivers is often assumed to be sufficiently similar to such a diffuse wave field so that the Green’s function can be retrieved from continuous recordings of the noise. This opens a new way for structural imaging without the need of active sources, and it has a wealth of other possible applications like Passive Image Interferometry to monitor variations of structure or seismic velocities with time. It has been successfully used in detecting velocity changes associated with earthquakes, volcanic eruptions or changes of water table etc.

However, there is a number of issues that need further investigation before it should be used as a routine tool, e.g. to what extent the assumption of ambient noise as a diffuse wave field is valid in different frequency and distance ranges. Another point of interest is what amount of reliable and stable information except the fundamental mode surface waves is contained in the reconstructed Green’s functions. Short period waves are of special interest in this context as they are effectively scattered within the crust. and may be used in the context of passive image interferometry.

2. Data Analysis
In this study we use ambient noise data from the short period vertical sensors of the NORSAR array, southern Norway which consists of 7 subarrays with 6 receivers within each subarray. They are well suited as a test data set as they have high data quality and long time availability in a region where the crustal structure is well-known, and we do not expect any time variations. Continuous data sampled at 25 Hz has been used. Normalized cross-correlations have been computed for 24h traces and correlations stacked for up to 60 days. Inter-receiver distances vary between several km within one subarray, and some tens of km for the whole array.

3. Results
Stacked traces clearly show the Rg wave (Fig. 1), its velocity being consistent with what is expected from the crustal structure under NORSAR. Their amplitudes at positive and negative times strongly depend on the azimuth of observation (Fig.2). The amplitude pattern is consistent with a dominant noise source from a certain azimuth plus a significant amount of secondary scattered waves coming in from various directions. This points to the fact that oceanic microseism forms the most significant primary source of ambient noise, having a strong azimuth and time dependence and clearly deviating
from a diffuse wave field. Nevertheless the crustal scattering is strong enough to provide an azimuthal averaging that allows the retrieval of the Rg phase, but not of the amplitude information.

Above 2 Hz the Rg arrival breaks down for distances above 10 km, indicating that there are no noise sources at these frequencies that are strong enough to correlate over more than a few kilometers.

There is some indication that also body wave phases appear in the correlation traces at frequencies around 1 Hz, but they are not very clear. Several partly coherent phases can be identified within the coda after Rg. However, they are not stable with time and depend on the noise sources.

As a general result from this test we conclude that the success of Green’s function retrieval and its applications depends strongly on the properties and time stability of the ambient noise. Strategies for routine checking these properties should be developed.

**Figure 1.** Bandpass filtered correlation traces at various receiver distances. Left: 0.5-2 Hz, right: 1-4 Hz. Rg phases with 3.1 km/s group velocity travel from left to right at positive times and vice versa at negative times. Signal to noise ratio deteriorates for higher frequencies.

**Figure 2.** Amplitude ratio between waves at negative and positive times versus azimuth between receiver pairs. Maxima around 20° and minima around 160° indicate preferred propagation direction of ambient noise.
Studying a dynamic Process in the lunar crust with Passive Image Interferometry

Christoph Sens-Schönfelder1, Eric Larose2

1. Institut of Geophysics and Geology, University of Leipzig, Leipzig, Germay
   E-mail: sens-schoenfelder@uni-leipzig.de
2. Laboratoire de Geophysique et Tectonophysique, Universite Joseph Fourier, Grenoble, France
   E-mail: eric.larose@ujf-grenoble.fr

Introduction

Passive Image Interferometry (PII) as developed by Sens-Schönfelder and Wegler (2006) is a method for continuous monitoring of weak structural changes in the subsurface. The breakthrough of this technique is the easy applicability on existing data in seismology. The noise signal that is usually regarded as disturbing vibration becomes a valuable source of information in PII. This led to impressive applications of monitoring tectonic targets such as volcanoes or fault zones. In this paper we present another application that demonstrates the potential of PII.

Figure 1 Relative delay time variations of seismic waves in the lunar crust from August 1976 until May 1977. Gray background indicates lunar night at the Apollo 17 landing site. Black dots represent independent measurements and red curve their average.

Observation

Since investigations of the lunar environment are comparatively difficult its static properties were the primary interest of previous research. Only recently it was recognized that the Moon represents a dynamic system with notable changes. We reanalyze the almost historical data set from the Apollo 17 seismic experiment with PII. The data set contains the nine month of continuous seismic records from the four geophones of the Apollo 17 Lunar Seismic Profiling Experiment. The data show continuous excitation of ambient vibrations though there is no fluid atmosphere that could excite the vibrations like on Earth.
Analysis of the variable strength of the vibrations indicates that they are generated by acoustic emissions due to thermal cracking. Vibrations are strongest in the morning and at sunset with an almost exponential decay during the lunar night. Including auto-correlations the four geophones allow ten independent PII measurements that all show similar curves. Averaging the independent measurements we obtain a continuous time series of velocity variations in the lunar subsurface over nine lunations (figure 1). The velocity variations show a clear periodicity of 29.5 Earth day period. This period equals the synodic month and indicates that the velocity variations are linked to the position of the sun. During lunar day the seismic phases get increasingly delayed indicating decreasing velocities. During lunar night the velocity increases again.

**Modeling**

![Figure 2](image)

Oberved and modeled relative delay times (RDT) of one lunation. Red curve: observed RDT averaged over the nine lunations with gray background indicating one standard deviation of this average. Black curve: modeled RDT. Blue curve: surface temperature as predicted by our model as check of consistency.

We test the hypothesis that the delay of the seismic phases is caused by the sun by a simple modeling of the thermal processes during a lunation. We assume the following causal relation for the influence of the sun on the relative delay of the seismic waves:

- The energy balance of the lunar surface is determined by the influx from the sun depending on the incidence angle and the thermal radiation form the surface into free space.

- Diffusion of heat from the surface into the crust follows the 1D diffusion equation with an additional term that takes into account the heat transport by radiation in the uppermost centimeters of the lunar soil.

- Relative variation of seismic velocity is proportional to the relative temperature changes.

- The wave field is equally sensitive to velocity changes in any depth

Especially the last item is a strong approximation and might be far from reality. But it is justified here since the thickness of the layer that is influenced by temperature changes is small compared to any length scale of the wave field.

We solve the 1D heat diffusion equation subject to the variable boundary condition at the surface with material constants known from the Apollo heat flow experiments. From this model of the depth dependent temperature distribution we deduce the expected delay of seismic phases. The model correctly reproduces the observed changes in the wave field (figure 2). This make us confident that we observe a dynamic process in the lunar environment based on the seismic noise. A detailed inversion of the relative delay times will allow to study the heat conduction processes in the lunar crust with new means.
Temporal Changes in the Source Region of the 2004 Mid Niigata Prefecture Earthquake of 2004

Ulrich Wegler¹, Hisashi Nakahara², Christoph Sens-Schönfelder³, Michael Korn³, and Katsuhiko Shiomi⁴,

1. Federal Institute for Geosciences and Natural Resources (BGR), Hannover D-30655, Germany,
   E-mail: ulrich.wegler@bgr.de
2. Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan,
   E-mail: naka@zisin.geophys.tohoku.ac.jp
3. Institute of Geophysics and Geology, University of Leipzig, Germany,
   E-mail: mikorn@uni-leipzig.de, sens-schoenfelder@uni-leipzig.de
4. National Research Institute for Earth Science and Disaster Prevention, Ibaraki 305-0006, Japan,
   E-mail: shiomi@bosai.go.jp

1. Abstract

Passive Image Interferometry (PII) uses ambient seismic noise to monitor temporal changes of mean shear wave velocity in the earth crust. In a first step, the elastic Green's tensor between two seismometers is computed by cross-correlating seismic noise recorded during a certain time period. In a second step, the constructed seismograms of different time periods are treated as earthquake multiplets and small time shifts in their coda are used to invert a relative change in mean shear wave velocity.

We applied this technique to the source region of the Mid Niigata Prefecture earthquake of 2004, Japan, which had a centroid-depth of 5 km and a moment Magnitude of 6.6. We used noise recorded at 5 seismometers of Hi-net, the Japanese High-Sensitivity seismograph network, and one station of F-net, the Japanese Broadband Seismograph Network. All stations are located at a distance of less than 25 km from the epicenter. We construct seismograms using noise in the two different frequency bands of 0.1 - 0.5 Hz and of 2 - 8 Hz. Using high frequency noise (2 - 8 Hz) one day of data is generally sufficient to estimate the source-receiver co-located Green's function, which leads to a temporal resolution of one day. Using lower frequencies (0.1 - 0.5 Hz), on the other hand, much longer noise time series in the order of weeks are required to compute the Green's function. The advantage of using lower frequencies is that Green's functions for larger station distances can be computed, whereas for high frequencies due to the lack of coherence in many cases only source-receiver co-located Green's functions can be constructed from the auto-correlation of noise at a single station. Applying the technique to the source region of the Mid-Niigata earthquake we revealed a rapid co-seismic drop in relative seismic velocity of some tenths of percent, that spatially roughly coincides with the earthquake source area. The fact that the velocity decrease measured in the 2 – 8 Hz frequency band has a similar amplitude as the velocity decrease measured in the 0.1 - 0.5 Hz frequency band is some indication that the change is not restricted to the shallow subsurface.

The physical mechanism causing the co-seismic velocity drop could not be completely clarified. A non-linear site response in the shallow subsurface layer due to strong ground motion and structural weakening due to the creation of new fractures in the fault zone are consistent with our data. Static stress changes, on the contrary, cannot explain the fact that only decreases in velocity are observed, whereas regions of increasing velocity are not observed in our study.
**Figure 1.** Source region of the Mid Niigata prefecture earthquake of Oct. 23., 2004: Locations of Hi-net sensors KWNH, MUIH, NGOH, STDH, and YNTH as well as F-net station KZK (triangles). The beachball indicates the hypocenter of the mainshock according to JMA and its focal mechanism determined by F-net. Small gray dots indicate hypocenters of aftershocks during Oct. 23-31 according to the JMA catalogue.

**Figure 2.** Temporal evolution of the source-receiver co-located Green’s function constructed from the auto-correlation of seismic noise at station KZK during a period of four months. The Green’s function is averaged over one day and shown as a function of the day of the year. The black arrow indicates the occurrence of the Mid-Niigata earthquake. Red and blue wiggles correspond to positive and negative amplitudes of the Green’s function, respectively. White space near the day of the Mid-Niigata earthquake is caused by the lack of data for four days. Right: Same as left, but for an enlarged time window from 7 to 9 s. Note the time-shift in the Green’s function after the earthquake.

**References**


Temporal Variations of Crustal Structure in the Source Region of the 2007 Noto Peninsula Earthquake, Central Japan, using Ambient Seismic Noises

Shiro OHMI¹ and Kazuro HIRAHARA²

1. Disaster Prevention Research Institute, Kyoto University, Kyoto, 611-0011, Japan, E-mail: ohmi@rcep.dpri.kyoto-u.ac.jp
2. Graduate School of Science, Kyoto University, Sakyo-ku, Kyoto, 606-8502, Japan E-mail: hirahara@kugi.kyoto-u.ac.jp

1. Introduction

The passive image interferometry technique (Sens-Schoenfelder and Wegler, 2006) is applied to the continuous seismic waveform data obtained around the source region of the 2007 Noto Peninsula Earthquake (Mw6.6, occurred on March 25, 2007, Noto EQ), central Japan, to detect the temporal variation of the subsurface structure around the source region. We computed the autocorrelation function (ACF) of band-pass filtered seismic noise portion recorded with vertical component short-period seismometer at several seismic stations for each one day. Figure 1 shows the location of the epicenter together with several seismic stations used in this study.

2. Results

Around the source region of the Noto EQ, station N.TGIH (epicentral distance 4km), DP.NNJ (36 km), and DP.HRJ (45 km) exhibit the change of ACFs. In these stations, changes of lag time of the particular phases in ACF are observed. They are attributed to the change of seismic wave velocity in the volume considered. In some stations, temporal evolution of ACFs preceding the mainshock is also detected. Figure 2(a) shows the temporal variations of the ACF at N.TGIH during 6 months. Clear increase of the lag time is observed after the mainshock, which would be attributed to the decrease of the subsurface seismic velocity. It is also seen that the time shift is smaller on phases with shorter lag time, and larger time shifts are observed on the phases with larger lag time.

We also preliminarily investigated the temporal evolution of the decay factor of the ACFs. It is also indicated that decay of ACF is equivalent to that of coda waves (e.g. Sens-Schoenfelder and Wegler, 2006, Wegler and Sens-Schoenfelder, 2007). Thus we assume the envelope of ACF should obey the typical relation of the coda Q theory. The obtained Q values during a period of one year including the mainshock at some stations exhibit temporal variations (Figure 2(b)). At station N.TGIH, Q gradually decreases since September 2006 and kept lowermost values from mid November 2006 to mid March 2007, and then gradually increased after the mainshock.

3. Discussion and Conclusion

Temporal variation of the autocorrelation function (ACF) of ambient seismic noise around the source region of the 2007 Noto Hanto Earthquake is analyzed to detect possible change in the subsurface structure associated with the earthquake. Sudden change in lag time of the ACF associated with the
occurrence of the mainshock is detected in some stations.

The decay rate of ACFs also exhibit temporal change in some stations. Many previous studies reported the temporal change of coda Q values associated with seismic activity including precursory change. Although the Q values in our analysis is not identical to the coda Q, the decay rate of ACFs would be also a powerful tool for monitoring the stress state of the crust if we could imply the correlation between the coda-Q and the ACF-Q.

Figure 1. Map around the epicenter of the 2007 Noto Peninsula earthquake. Solid star denotes the epicenter of the mainshock, while solid squares represent seismic stations used in this study. Fault plane solution together with the surface projection of the fault plane obtained by Horikawa (2008) are also shown.

Figure 2. Temporal variation of the ACFs in the source region of the Noto EQ. (a) Temporal evolutions in lag time of ACFs at station N.TGIH (left), and (b) temporal change of the decay rate of the ACFs (Q) at stations N.TGIH and N.SHKH (right). Dots represent daily Q values while lines show moving average of 10 days.
Monitoring seismic velocity changes using Passive Image Interferometry: An application to the 2005 West Off Fukuoka Prefecture, Japan, Earthquake (Mw 6.6)

Hisashi Nakahara¹, Ulrich Wegler², and Katsuhiko Shiomi³,

1. Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan, E-mail: naka@zisin.geophys.tohoku.ac.jp
2. Federal Institute for Geosciences and Natural Resources (BGR), Hannover D-30655, Germany, E-mail: wegler@szgrf.bgr.de
3. National Research Institute for Earth Science and Disaster Prevention, Ibaraki 305-0006, Japan, E-mail: shiomi@bosai.go.jp

1. Introduction
 Monitoring seismic velocity changes in fault areas and volcanic regions is interesting in terms of predictions of earthquakes and volcanic eruptions. Passive Image Interferometry (PII) was developed by Sens-Schoenfelder and Wegler (2006, GRL) as a monitoring tool. So far, the method has been widely applied to detect changes in seismic velocity associated with earthquakes (e.g. Wegler and Sens-Schoenfelder, 2007, GJI) and volcanic eruptions (e.g. Brenguier et al., 2008, Nature Geoscience). In this study, we present another application of the method to the 2005 West Off Fukuoka Prefecture, Japan, Earthquake (Mw 6.6; the Fukuoka event), which is a strike-slip one which took place on March 20, 2005 off the northern part of Kyushu Island.

2. Data Analysis and Results
 SBR station of the F-net broad-band network is located about 30km away from the epicenter as shown in Figure 1. Auto-correlation function (ACF)s of ambient noises at the station are calculated day by day during a period including the Fukuoka event. Frequency bands between 1-8 Hz are focused. A one-day record of ambient noises, which is sampled at a rate of 20Hz, is divided into segments with a length of 51.2s. Only the segments whose maximum amplitude is less than 10 times the root mean squared amplitude of ambient noises in a quiet day are used for the calculation of ACFs. For individual segment satisfying the criterion, ACFs are stacked. The number of segments per day is 1687 at maximum. In Figure 2, temporal evolutions in the stacked ACFs for UD component in 2-4Hz band during the period including the mainshock are shown. The phases between 4s and 7s in lag time are found to be delayed by about 0.1s after the mainshock in comparison to before the mainshock . Investigating data in much longer time period, we find that the phases are very stable for about 1100 days before the mainshock and the delay still remains as of two years after the mainshock.

3. Discussions
 A problem is wave types of the delayed phases in the stacked ACFs. To get a clue for it, decay rates in envelopes of the stacked ACFs are analyzed. Based on the analysis, we prefer to interpret the delayed phases as fundamental Rayleigh wave. Concerning the origins of the phase delay, we calculate induced static strain (stress) changes at the station due to a fault model of the Fukuoka event. But we find that the change in static strain (stress) due to faulting expects phase advances contradicting our observation. So far,
the damage in soils at very shallow depths due to strong ground motion seems one possible cause for the phase delay.

4. Conclusions
We have detected changes in the stacked ACFs of ambient noises between 2-4Hz at station SBR associated with the 2005 West Off Fukuoka Prefecture earthquake. The phases between 4s and 7s in lag time have been found to be delayed by about 0.1s after the mainshock. The phases are very stable for about 1100 days before the mainshock, and the delay still remains as of two years after the mainshock. So far we speculate that the damage beneath the station caused by strong ground motion is a possible cause for the delay.

![Figure 1](image1.png)

**Figure 1.** The F-net station SBR is shown by a solid triangle. A fault model and the epicenter of the 2005 West Off Fukuoka Prefecture Earthquake are shown by a rectangle and a solid star, respectively. Epicentral distances of 10km, 20km, and 30km are shown by 3 concentric circles.

![Figure 2](image2.png)

**Figure 2.** Temporal evolutions of Stacked ACFs of ambient noises in UD component (2-4Hz) during two months including the mainshock are shown. The arrow on the right indicates the mainshock.
Recurrence of the Seismic Velocity Change Associated with Earthquake Swarm activities in NE Kyushu, Japan, revealed by the Seismic Interferometry

Takuto Maeda1, Yohei Yukutake2, and Kazushige Obara1

1. National Research Institute for Earth Science and Disaster Prevention, Ibaraki 305-0006, Japan
   E-mail: maeda@bosai.go.jp
2. Hot Springs Research Institute, Kanagawa Prefectural Government, 586 Iriuda, Odawara, Kanagawa, 250-0031, Japan

1. Introduction
Auto-correlation functions (ACFs) of ambient noise, which can be interpreted as a seismic wavefield or Green's function for the collocated source and receiver, is a powerful tool for searching temporal change of crustal structure. After the pioneering works by Sens-Schönfelder and Wegler (2006) and Wegler and Sens-Schönfelder (2007), there are many reports of the detection of the seismic velocity change [e.g., Brenguier et al., 2008]. In this study, we report recurrence of quite large phase delay of ACFs, which corresponds to the velocity reduction, associated with the earthquake swarm activities.

2. Earthquake swarm at northeastern Kyushu, Japan, in 2007
Earthquake swarm activity had started from 6 June 2007 at mid Oita prefecture, in NE Kyushu, Japan (Figure 1). High-activity lasts about 6 days. Earthquake sources are located Beppu Bay to northern portion of Yufuin fault system, having depth of about 10km. Fine-scale relocation by using precise measurement of relative traveltime by waveform cross correlation technique shows that the swarm earthquake activity migrates from NNE to WSW direction with increasing time. Four months later, small-size earthquakes are re-activated on the shallower portion along the extended line of the swarm migration at 30 October.

3. Seismic Interferometry for detecting velocity change
We calculated ACFs of noise at OITA2 station that is located very close to the epicenters of earthquake swarm, operated by Japan Meteorological Agency. At first, one-day continuous trace of vertical component is divided into 24 of one-hour segments. Then, filtered trace at the frequency band of 1-3 Hz is used to calculate autocorrelation by one-bit correlation technique [Campillo and Paul, 2003]. By taking ensemble average of ACFs among 24 hours, the one-day ACF is estimated. Following to the above method, one-day ACFs are calculated from January 2006 to March 2008. From the ACFs in 2006 in which there are no swarms, we confirmed the stability of the ACFs in this frequency range. We found that there is prominent phase delay for the lag times of 7 s and 9.5-12 s from the mid of June 2007, when earthquake swarm had started (Figure 2). We note that the delay is not uniformly increases with increasing lag time. For example, there is no delay around 8s in lag time, whereas there are clear delays at lag times before and after that time. The estimated phase delay was up to 0.2 s, and this phase delay remains about four months after the termination of the earthquake swarm activities with decreasing the amount of phase delay. However, in 30 October, sudden phase delay is observed again, accompanying the second earthquake swarm activity.

4. Discussion
We found the delay and recovery of phase in ACFs with the long-time duration. This observation suggests that there is a velocity change and its recovery associated with the earthquake swarm. ACF could also be changed by seasonal and/or rainfall changes. However, phase delay observed in this case does not match rainy season. Also, change for both of the two earthquakes swarm activity support that this change is due to the swarm. We note that the peak ground amplitude by these earthquakes are generally small; The maximum magnitude of the activity is 5.0. Therefore, we cannot expect that there is a non-linear effect of elasticity caused by the strong oscillation. The non-uniform distribution of the
phase delay with increasing lag time suggests that the region of velocity change is localized. One possible scenario to explain such characteristics is that the water flow induced in the shallow crust by the earthquake swarm activity cause the seismic velocity change. Long duration of recovery of the phase may reflect the diffusion process of the water around the source region.

Acknowledgement
We have used continuous seismic velocity trace at the station operated by Japan Meteorological Agency.

Figure 1. (a) The station by Hi-net (triangles) and JMA (inverse triangles), known active faults (gray lines), and the epicenters of the swarm. (b) The epicenter distribution in the rectangle in the map (a). Colors indicate the sequential number of the earthquake measured from June, 2007. (c) M-T diagram of the swarm activity. (d) Depth-section of the hypocenter along the line A-B in the map (b).

Figure 2. Day-by-day ACFs of vertical component seismic trace for the period from January 1 2007 to March 31, 2008 at OITA2 station for the lag time between 7 and 12 sec. The day corresponds to the start of the earthquake swarms are indicated by the arrows in the right hand side of the panel.
Seismic basement structure beneath the Kanto plain, Japan inferred from the seismic interferometry for strong motion records

Kazuo Yoshimoto¹, Kenya Sakurai¹, Hisashi Nakahara², Shigeo Kinoshita¹, Hiroshi Sato³

1. International Graduate School of Arts and Sciences, Yokohama City University, Yokohama, Kanagawa 236-0027, Japan
   E-mail: yoshi@yokohama-cu.ac.jp (K.Y.), i050280g@yokohama-cu.ac.jp (K.S.),
   kkk001@yokohama-cu.ac.jp (S.K.)
2. Graduate School of Science, Tohoku University, Sendai, Miyagi 980-8578, Japan
   E-mail: naka@zisin.geophys.tohoku.ac.jp
3. Earthquake Research Institute, University of Tokyo, Bunkyo, Tokyo 113-0032, Japan
   E-mail: satow@eri.u-tokyo.ac.jp

1. Introduction

   Seismic basement (pre-Neogene rock basement) structure beneath the Kanto plain has been investigated by using many geophysical approaches (e.g. seismic reflection survey, microtremor array analysis). However, mainly because of the insufficient investigation points in the target area, there is still ambiguity in the local variation of seismic basement structure. In this study, we show the effectiveness of the seismic interferometry (e.g. Nakahara, 2006, GJI) for the investigation of seismic basement structure from strong motion records of local earthquakes.

2. Data and Analysis

   We analyzed the seismic waveforms of 59 local earthquakes recorded at 503 stations of SK-net and K-net. In order to estimate the reflection response of S-waves for shallow underground structure, we adopted a seismic interferometry technique proposed by Nakahara (2006, GJI). Firstly, acceleration waveforms of each event were high-pass-filtered to remove long-period microtremors, and then were integrated to obtain displacement waveforms. The reflection response at each station was evaluated by stacking the autocorrelation functions of the S-waves (SH component) from all available events.

3. Results

   On the most of reflection responses, we observed a clear seismic basement phase with negative polarity. Figure 1 shows the local variation of the appearance time of this phase. Since this time corresponds to the two-way travel time of S-waves between the free surface and the seismic basement, it can be used as a measure of seismic basement depth and its local variation (Figure 2). For example, in the central Tokyo metropolitan area, the two-way travel time of S-waves between the free surface and the seismic basement reaches a maximum value of about 6-7 s in Nerima Ward, implying a local subsidence of the seismic basement with a maximum depth exceeding 3 km. This result is consistent with that reported by Tokyo Metropolitan Government (2004) from the seismic reflection survey. Figure 2 indicates that the depth of seismic basement varies very irregularly beneath the Kanto plain.

   Our result shows that the seismic interferometry for strong ground motion data is quite effective for investigating the local variation of seismic basement depth even in the densely populated area with high ground noise.
Acknowledgments
Data provided by SK-net and K-net are gratefully acknowledged. We thank Tokyo metropolitan government, Chiba, Gunma, Ibaraki, Kanagawa, Saitama, and Tochigi prefecture, and Yokohama city. We also thank Earthquake Research Institute, University of Tokyo and National Research Institute for Earth Science and Disaster Prevention.

Figure 1. Map showing local variation of the appearance time of seismic basement phase. Shaded areas indicate the outcrops of Pre-Neogene rocks, Neogene sedimentary rocks, and volcanic rocks (Geological Survey of Japan, 2003). Thin lines are prefecture borders.

Figure 2. Map showing local variation of seismic basement depth (preliminary result). S-wave velocity model at Iwatsuki from VSP measurement (Yamamizu, 1996) were used in the depth conversion.
Seismic Interferometry of teleseismic S-wave coda for detection of body waves
-An application to the Philippine Sea slab underneath the Japanese Islands-

Takashi Tonegawa¹, *, Kiwamu Nishida¹, Toshiki Watanabe², and Katsuhiko Shiomi³

1. Earthquake Research Institute, Univ. of Tokyo, Tokyo, 113-0032, Japan
   E-mail: tonegawa@eri.u-tokyo.ac.jp, knishida@eri.u-tokyo.ac.jp
2. Graduate School of Environmental Studies, Nagoya University, Nagoya, 464-8602, Japan
   E-mail: watanabe@seis.nagoya-u.ac.jp
3. National Research Institute for Earth Science and Disaster Prevention, Ibaraki, 305-0006, Japan
   E-mail: shiomi@bosai.go.jp

1. Introduction
The reconstruction of surface waves from spatial correlation of random wavefield has recently been extensively inspected from theoretical and experimental approaches. In addition to the surface wave, the retrievals of body waves have recently been reported by several papers (e.g., Roux et al. 2005: Miyazawa et al. 2008). The subsurface reflectivity could also be imaged by taking cross correlations. (e.g., Yu and Schuster 2006: Draganov et al. 2007: Abe et al. 2007). However, whether all direct P, S, and reflected waves can simultaneously be retrieved by cross-correlating a wavefield observed at two receivers is still under debate in the application approaches. In this study, we present a method to extract the propagations of body waves, that is, direct P and S waves, and the reflected wave from the Philippine Sea slab underneath the Japanese Islands.

2. Data and Processing
We used the spatial correlation of the wavefield generated by teleseismic S waves observed by the Hi-net tiltmeters with a passband of 0.07-0.5 Hz. The number of teleseismic events used in this study is 193, which occurred between April 2003 and Dec 2007 with 5.5 ≤ Mw ≤ 7.5. The epicentral distances of the teleseismic events are between 30° and 85°, which correspond to incident angles of the direct S wave between 15° and 30°.

The waveforms after the direct S arrival presumably contain something other than random signals, such as surface waves, microseisms, and the source-time function. In this study, we attempted to eliminate the above factors with preserving the contributions of S coda. Practically, we depressed the source-time function and the later arrivals of deterministic phases, sS, ScS, SS, and surface wave. The cross correlation functions are computed using the processed S coda with a time length of 500 sec. Then, we stacked the cross correlations of different earthquakes.

3. Results
To verify whether the CCFs describe direct and reflected waves, we aligned the CCFs as a function of separation distance between two stations. Figure 1 successfully shows the propagations of direct P and S
waves with approximately 5-7 km/s and 3 km/s, and more importantly, a S-to-S (SS) reflection, indicating that this technique is applicable to extract body waves. The amplitude difference and the slow seismic velocities (5-7 km/s and 3 km/s) lead us to conclude that the extracted P and S waves propagate at relatively shallower depths, and impinge with shallow incident angles to the stations.

In order to enhance the reflected waves, we search for the reflected points by assuming that the later phases in the CCFs are SS reflections and mapping the amplitudes onto depth sections. As a result, the negative phases dipping to the north can be traced right below the hypocenter distribution, probably corresponding to the oceanic Moho within the Philippine Sea slab. We also show that the oceanic Moho can also be traced by assuming the PP reflections.

4. Conclusion

We presented that the spatial correlation of the teleseismic S coda is applicable to detect the body waves including the direct and reflected waves. To effectively enhance the contribution of the S coda, we depressed the source-time function and the deterministic phases, and used the recordings with good S/N. In the resultant seismic images, the reflectivity of the Philippine Sea slab can be imaged with the stacked CCFs, assuming that the CCF contains PP and SS reflections. These results indicate that the CCFs plausibly contain the information of both P and S waves between the two receivers, and are capable of detecting reflected phases in addition to the direct waves.

Fig. 1 The cross correlation functions (CCFs) and phases as a function of station distance. (a) The CCFs with the processed S coda as a function of station distance. The amplitudes are normalized by the maximum value, after multiplied by square of the station distance, (b) Same as Fig. 6(a), but for the phase. (c) Interpretation of Fig. 6(b). Bold black lines indicate the direct P and S waves and the reflected phase, and broken line indicate a ghost phase associated with the deterministic phases, such as the direct S, sS, ScS, and SS, in the S coda. Thin black lines indicate velocity gradients of 7, 5, 3 km/s for comparison with the observed travel time curves.
Global surface wave tomography using seismic hum

Kiwamu Nishida¹, Jean-Paul Montagner², and Hitoshi Kawakatsu¹,

¹. Earthquake Research Institute, The University of Tokyo
   E-mail: knishida@eri.u-tokyo.ac.jp
2. Institut de Physique du Globe de Paris

1. Introduction

Recently a technique of ambient noise tomography has been developed. For random wave field, a cross-correlation function between a pair of stations exhibits Green function like signals [Snieder, 2004]. In principle this technique is spatial-time domain representation of spatial autocorrelation method, which has been developed since an early work by Aki [1957]. Shapiro et al. [2005] obtained group velocity map in California by the cross-correlation analysis of background noise between many pairs of stations at around 0.1 Hz. These studies resulted in group-speed maps at short periods (7.5-15 s) that display a striking correlation with the principal geological units in California with low-speed anomalies corresponding to the major sedimentary basins and high-speed anomalies corresponding to the igneous core of the main mountain regions. Since then group-velocity maps have also been obtained at larger scales and longer periods across much of Europe [Yang et al., 2007], in South Korea at very short periods [Cho et al., 2007], and in Tibet at long periods [Yao et al., 2006]. However there is no global upper mantle model using this method.

In the frequency range of microseisms (0.05-0.2 Hz), the background Love and Rayleigh waves are attenuated and dissipated in global scale. For the tomography of global upper mantle structure, we must observed global propagation of background surface waves in low frequency range below 20 mHz. Nishida et al. [2002] shows clear propagation of background Rayleigh waves by cross-correlation analysis as shown in Fig. 1. The waves are closely related to background free oscillations know as seismic hums [Suda et. al. 1998, Kobayashi and Nishida, 1998]. These results suggest possibility of global surface wave tomography using seismic hums.

An excitation theory of these waves with an assumption of stochastic stationary excitation has been developed [Fukao et. al, 2002; Nishida and Fukao, 2007]. The can explain most part of observed cross-correlation functions as shown in Fig. 1. This figure promises measurements of their phase difference. We inverted them for global 3-D upper mantle structure in this study.

2. Data Analysis and Results

We analyze 10-second continuous sampling records in a time period from 1988 to 2000 through the very--long--period high--gain (VH) channel from the vertical STS-1 seismometers at 54 FSDN stations at the lowest ground noise levels of slightly less than 3x10⁻¹⁸ m² s⁻³ [Nishida and Fukao, 2007]. The records are provided by the Incorporated Research Institutions for Seismology Data Management Center [IRIS DMC: IRIS, 1994]. For each station, we remove glitches and divide the whole record into about 5.6 hour segments with an overlap of 1 hour. Each of the segments is Fourier-transformed to obtain the power--spectrum. The spectrum might have been disturbed by transient phenomena such as earthquakes and local nonstationary ground or instrumental noise. We discard all the seismically disturbed segments, which are defined in terms of the mean power spectral densities (PSDs) greater than 3x10⁻¹⁸ m² s⁻³ in a frequency range 2.5-7.5 mHz. We also discard noisy segments if their mean PSDs over the four frequency ranges, 7.5-12.5, 12.5-17.5 and 17.5-22.5 mHz are greater than 3x10⁻¹⁸ m² s⁻³ [Nishida and Kobayashi, 1999]. We calculate the cross-correlation function and cross-spectrum between every pair of different stations for their common record segments. Such calculations are made for the thirteen years. We then obtain the cross-correlation functions and cross-spectra of the records between every pair of two stations as shown in Fig. 1. We measured phase difference between the observed cross-correlation functions and synthetic ones [Nishida and Fukao, 2007] for 906 R1 paths and 777 R2 paths. We inverted the observed phase velocity anomalies for phase velocity maps from 3 to 10 mHz. Then, we inverted obtaining phase velocity maps for 3-D S-wave velocity structure in the upper mantle [Montagner, 1986]. At the
depths shallower than 200 km, our results show low velocity anomalies associated with mid-ocean ridges and back arcs, and fast velocity anomalies in continental shield and platform area as shown in Fig. 2.

Figure 1. Observed cross-correlation functions (black lines) and synthetic ones (green lines).

Figure 2. Resultant S-wave velocity structure at depths from 140 km to 600 km.
Retrieval of the single scattering Green function from the cross-correlation function in a scattering medium illuminated by surrounding noise sources

Haruo Sato
Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan, E-mail: sato@zisin.geophys.tohoku.ac.jp

1. Introduction
Since seismic interferometry was proposed, there have been many measurements on the Green function retrieval by using the cross-correlation function (CCF) of seismic waves (e.g. Campillo and Paul, 2003). Most of them are measurements of the average velocity from the peak lag time of the CCF, and there were theoretical models for the CCF of waves in a homogenous medium illuminated by noise sources spatially distributed (e.g. Snieder, 2004; Roux et al., 2005). Recently, there have been attempts to measure precisely the temporal change in the coda portion of the CCF (or ACF) which might reflect the medium change in the crust (e.g. Wegler and Sens-Schonfelder, 2007).

Here, we propose a theoretical derivation of the Green function having coda from CCF in a simple scattering medium illuminated by surrounding noise sources.

2. Green function in a scattering medium embedded in a homogenous medium
Let us imagine a distribution of $N$ velocity anomalies in a volume of dimension $L^3$ in a homogenous medium with the background velocity $V_0$. The wave velocity is written as $V(x) = V_0 + V_0 \sum_{j=1}^{N} \epsilon_i L^3 \delta(x-y_j)$ by using a delta function in space. This is a mathematical representation of the case that the $j$-th velocity anomaly at location $y_j$ is small, $|\epsilon_j|<<1$, and its spatial scale is also small compared with the wavelength. We call the region $D$ having velocity anomalies the scattering medium (see Figure 1a). The wave equation is written as

$$\Delta \phi(x,t) - \frac{1}{V_0^2} \partial_t^2 \phi(x,t) + 2 \sum_{j=1}^{N} \epsilon_j \frac{L^3}{V_0^2} \delta(x-y_j) \partial_t^2 \phi(x,t) = f(x,t). \quad (1)$$

For the delta function source term $f = \delta(x) \delta(t)$, solving (1) we obtain the retarded Green function on the basis of the first order Born approximation. For a receiver at $x_A = (0,0,h_0)$ and a source $x_B = (0,0,-h_0)$ on the z-axis with a separation of $h_{AB} = 2h_0$, the Green function in the angular frequency domain, which satisfies the radiation condition, is written as

$$\hat{G}(x_A, x_B, \omega) = -\frac{1}{4\pi h_{AB}} e^{i\omega h_{AB}} + 2L^3 k_0^2 \sum_{j=1}^{N} \epsilon_j e^{i\omega h_{Aj}} e^{i\omega h_{Bj}} \frac{4\pi h_{Aj}}{4\pi h_{Bj}}, \quad (2)$$

where $\omega = V_0 k_0$ and a caret means the Fourier transform with respect to time. The second term shows that the scattering is isotropic and no phase shift for the delta function velocity anomaly. The retarded Green function in the time domain is given by

$$G(x_A, x_B, t) = -\frac{1}{4\pi h_{AB} V_0} \delta\left(t - \frac{h_{AB}}{V_0}\right) H(t) - 2 \frac{L^3}{V_0^2} \sum_{j=1}^{N} \epsilon_j \frac{1}{4\pi h_{Aj} V_0} \frac{1}{4\pi h_{Bj} V_0} \delta\left(t - \frac{h_{Aj}}{V_0} - \frac{h_{Bj}}{V_0}\right) H(t), \quad (3)$$

where $h_{Aj}$ is the distance between the receiver $x_A$ and the scatterer $y_j$ and $h_{Bj}$ between the source $x_B$ and he scatterer $y_j$. This solution explicitly shows the source-receiver reciprocity.
second term shows coda waves composed of single isotropic scattering.

3. CCF of waves in the scattering medium illuminated by surrounding noise sources

We imagine the scattering medium \( D \) is embedded in an infinite homogeneous medium with velocity \( V_0 \) as illustrated in Figure 1b. Noise sources with the spectrum \( \hat{N}(x, \omega) \) are distributed on a spherical boundary \( \partial D \) of radius \( r \), which is much larger than \( L \), the source-receiver distance \( h_{AB} = 2h_j \), and the wavelength. Waves at two locations \( x_A \) and \( x_B \) excited by those noise sources are written as

\[
\hat{\phi}(x, \omega) = \int_{\partial D} \hat{G}(x, \omega) \hat{N}(x, \omega) df(x),
\]

where \( df(x) \) is the infinitesimal surface element. We imagine an ensemble of noise sources \( \{ N \} \) on the spherical surface. When the spatial distribution of noise sources is random on the spherical surface and each noise source has the same spectrum, we may write the ensemble average of noise spectra on the spherical boundary as

\[
\lim_{T \to \infty} \frac{1}{T} \left\langle \hat{N}(x, \omega) \hat{N}(x', \omega) \right\rangle = \delta_2(\mathbf{x} - \mathbf{x}') \hat{S}(\omega),
\]

where \( \hat{S}(\omega) \) is the power spectral density function.

We evaluate this integral in the following. In the spherical coordinate system, we have

\[
\mathbf{x}_A = (h_j , 0 , 0), \quad \mathbf{x}_B = (h_j , \pi , 0), \quad \mathbf{y}_j = (h_{ij} , \theta_j , \phi_j),
\]

where \( h_{ij} = |\mathbf{y}_j - \mathbf{x}_B| \), \( r_A = |\mathbf{x} - \mathbf{x}_A|, r_B = |\mathbf{x} - \mathbf{x}_B| \) and \( r_j = |\mathbf{x} - \mathbf{y}_j| \). The surface integral in (6) is written as

\[
\int_{\partial D} df \hat{G}(x, \omega) \hat{\phi}(x, \omega)^* = \int_{\partial D} d\Omega(\theta, \phi) \left[ e^{-ik_0 r_A} e^{ik_0 r_B} \sum_{j=1}^{N} e^{rac{-ik_0 h_j}{4\pi h_{ij}}} e \frac{e^{-ik_0 h_j}}{8\pi^2} \right],
\]

where \( d\Omega(\theta, \phi) = \sin \theta d\theta d\phi \) and the approximation \( r_A = r_B = r_j = r \) is used in geometrical factors. Distances on the exponent are approximated as \( r_A \approx r - h_j \cos \theta, r_B \approx r + h_j \cos \theta \), and \( r_j \approx r - h_{ij} \cos \psi_j \), where \( \theta \) is the angle between \( \mathbf{x} \) and the \( z \) axis at the origin, \( \psi_j \) the angle between \( \mathbf{y}_j \) and \( \mathbf{x} \) at the origin. Using the following formulas,

\[
e^{ikr \cos \zeta} = \sum_{l=0}^{\infty} i^l (2l + 1) J_l(kr) P_l(\cos \zeta) = 4\pi \sum_{l=0}^{\infty} i^l j_l(kr) \sum_{m=-l}^{l} Y_{lm}(\theta, \phi) Y_{lm}^*(\theta', \phi'),
\]

where \( \cos \zeta = \cos \theta \cos \phi' + \sin \theta \sin \phi' \cos (\phi - \phi') \), and
\[
\sin k\sqrt{r_1^2 + r_2^2 - 2r_1r_2 \cos \zeta} \sqrt{r_1^2 + r_2^2 - 2r_1r_2 \cos \zeta} = \sum_{j=0}^{\infty} j_j (kr_j) j_j (kr_j) (2l+1) P_l (\cos \zeta),
\]

we can write the surface integral (7) as

\[
\int_{\partial D} d\vec{f} \hat{G}(\vec{x}, \vec{x}, \omega) \cdot \hat{G}(\vec{x}, \vec{x}, \omega) = -\frac{1}{k_0} \text{Im} \left[ \frac{-1}{4\pi} \frac{e^{ik_0 h}}{h_{AB}} + 2L' k_0^2 \sum_{j=1}^{\infty} \epsilon_j e^{ijkr_j + \epsilon jkr_j} \right] = -\frac{\text{Im} \hat{G}(\vec{x}, \vec{x}, \omega)}{k_0}.
\]

Taking the derivative of the CCF with respect to lag time, we have

\[
\frac{1}{V_0} \lim_{T \to \infty} \frac{1}{T} \int_{-T/2}^{T/2} \langle \phi (\vec{x}_A, t - \tau) \phi (\vec{x}_B, t) \rangle dt = \int \left[ G(\vec{x}_A, \vec{x}_B, \tau - \tau') - G(\vec{x}_A, \vec{x}_B, -\tau - \tau') \right] S(\tau') d\tau'.
\]

If the noise spectrum is white, \( \hat{S}(\omega) = \hat{S}_0 \), the noise ACF is a delta function of time \( \hat{S}_0 \delta(\tau) \) and the RHS becomes an antisymmetric Green function with respect to lag time. We note that Wapenaar and Fokkema \[2006\] derived the relation (9) for a general case by using the reciprocal theorem of the correlation type. Sanchez-Sesma et al. \[2006\] derived a similar result for a cylindrical inclusion in 2D.

4. Single isotropic scattering model for coda envelope

For a distribution of delta function-type anomalies, the retarded Green function has coda waves of which the envelope decays with lapse time increasing according to the first order Born approximation. If the distribution is random and the interference of scattered waves is negligible, the mean square envelope of the scattering part of the Green function (3) is well approximated by the single isotropic scattering model (Sato, 1977) as

\[
|G_1(\vec{x}_A, \vec{x}_B, t)|^2 \propto \frac{1}{(4\pi)^2} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \delta \left( |\vec{z} - \vec{x}_A| + |\vec{z} - \vec{x}_B| - V_0 t \right) \left( \frac{V_0 t}{h_{AB}^2} \right) d\vec{z} = \frac{1}{4\pi h_{AB}^2} K(\nu) H \left( t - \frac{h_{AB}}{V_0} \right),
\]

where

\[
K(\nu) = \frac{1}{\nu} \ln \frac{\nu + 1}{\nu - 1} \quad \text{for} \quad \nu > 1
\]

We show an example of the time derivative of CCF showing an antisymmetric Green function with coda in a scattering medium and the square root of (11) in Figure 2.

5. Conclusion

We have theoretically shown that the Green function with coda in a scattering medium can be retrieved from the time derivative of the CCF when the finite scattering medium is illuminated by surrounding noise sources on the basis of the first order Born approximation. This model gives a solid base for the practical analysis of the coda portion of CCF or ACF. It will be necessary for us to examine the following cases: there are scattering terms of higher orders, noise sources are distributed uniformly in a scattering medium, and velocity anomalies of different types.
Fig. 1 (a) Two receivers A and B in a distribution of point scatterers. (b) Geometry of the scattering medium and noise sources distributed on a spherical shell.

Fig. 2 Example of the time derivative of CCF showing an anti-symmetric Green function with coda (solid curve) for a scattering medium with the envelope according to the single isotropic scattering model (broken curve).
Multiple scattering and mode conversion
as revealed from active seismic experiments at active volcanoes

Mare Yamamoto, Haruo Sato, and Takeshi Nishimura

Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan,
E-mail: [mare, sato, nishi]@zisin.geophys.tohoku.ac.jp

1. Introduction

Volcano is one of the most heterogeneous fields in the Earth’s crust, and the understanding of such inhomogeneities in volcanoes may provide us important information on the shallow small-scale structure which eventually controls various volcanic processes. In the previous studies on the wave propagation at active volcanoes, the diffusion model, which reflects strong multiple scattering of the seismic energy, has been widely used to model the energy transportation in heterogeneous media. Almost all of these studies, however, assume the diffusion of a single mode (usually S-wave), and the contribution of P and S waves and their mode conversion have not been well recognized partly due to lack of dense seismic observations capturing spatio-temporal pattern of energy propagation. In this presentation, we present the characteristics of the wave propagation in the shallow heterogeneous structure of Asama volcano, Japan revealed from an active seismic experiment with dense seismic network, and discuss about the multiple scattering and mode conversion of P and S waves.

2. Data and Analysis

Asama volcano is one of the most active volcanoes in Japan. The volcano is an andesitic composite stratovolcano, and located at the junction of the Izu-Marianas and NE Japan volcanic arcs. It last erupted in Nov. 2004, and the magma transport associated with the eruption has been successfully delineated by seismic and geodetic observations. To gain more insights into the magma pathway and the dynamics of magma transport beneath the volcano, an active seismic experiment was conducted in Oct, 2006 as a part of the national project for the prediction of volcanic eruptions. One of the most impressive characteristics of the seismograms observed by about 450 seismometers deployed every 50-150 m is their envelopes; the seismograms from artificial shots, emitting mainly P-energy around 10 Hz, are characterized by spindle-like envelopes having small P-onsets and long codas. Furthermore, the spatial distributions of the propagating energy at fixed time shows a clear pattern exhibiting two slopes which are indicative of propagation of two modes having different scattering coefficients (Fig. 1). To explain the observed energy distribution, we consider the multiple isotropic scattering taking into account the mode conversions between P and S modes, and model the energy propagation in 3-D scattering media using the radiative transfer theory to estimate the scattering parameters.

3. Results and Discussions

From the comparison between the observed energy distributions and those obtained from the radiative transfer theory, we found that the energy propagation would be well explained by the multiple isotropic scattering of P energy and S energy converted from the P energy radiated from the active source. From the energy distribution at the early stage of the wave propagation (t = 1.0-1.5 s), the total scattering coefficients for P-S and S-S scattering are found to be 2.4 and 3.0 times higher than that of P-P scattering, respectively, and the mean free path of S-wave is estimated as about 1 km at 4-16 Hz band. The estimated mean free path is about one order shorter that the typical one in the crust and it seems to reflect the strong heterogeneity in the complex edifice of the active volcano. In contrast,
the amount of intrinsic absorption and energy leakage (window effect) can be estimated using the
energy distribution at long lapse time, and found to be about $Q=100$. The estimated scattering and
absorption parameters are consistent with the decay rate of observed direct P-wave amplitude with
distance (Fig. 2), and the result suggests that the direct wave amplitude in heterogeneous volcanic
environment is mainly controlled by the strong scattering attenuation.

4. Conclusions

We find an observational evidence of mode conversions and multiple scattering at Asama volcano
using an active seismic experiment with dense network. Observed spatial distributions of propagating
seismic energy suggest the dominant mode conversion at small lapse time and following multiple
scattering of P and S waves. The mean free path for S wave estimated using the radiative transfer
theory is as short as about 1km for 8-16Hz band. These results suggest that the mode conversion from
P to S and significant multiple scattering have an indispensable effect in the modeling and analysis of
seismic wave propagation in heterogeneous volcanic environments.

Another observational evidence of mode conversion may be obtained from the mode decomposition
of the wave field. In the last of this presentation, we will report some preliminary results from our
3-component array observation at Sakurajima volcano.

Figure 1. Comparison of observed (red) and
modeled (blue) spatial energy distribution.
The blue and green arrows indicate the wave
front of P and S waves, respectively.

Figure 2. Comparison of observed (top) and
modeled (bottom) dependency of direct
P-wave amplitude with distance.

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Attenuation of short-period S-waves in Japan: high resolution maps of intrinsic absorption, scattering loss and coda decay

Eduard Carcolé, Haruo Sato,
Graduate School of Science, Tohoku University, Aoba-ku, Sendai 980-8578, Japan,
E-mail: eduard@zisin.geophys.tohoku.ac.jp, sato@zisin.geophys.tohoku.ac.jp

1. Introduction
Detailed information on the attenuation structure provides insight into the nature of complexities in the earth structure and composition. We compute scattering loss $Q_s^{-1}$, intrinsic absorption $Q_i^{-1}$ and the seismic albedo, which is the ratio of scattering loss to the total S-wave attenuation, by using the Multiple Lapse Time Window Analysis (MLTWA) (Hoshiba, 1991; Fehler et al. 1992, Sato and Fehler, 1998) in Japan. Also coda attenuation $Q_c^{-1}$ is computed for the same data set and results are compared with those from other attenuation parameters. This work may be considered a reappraisal or extension of the pioneer work done by Hoshiba (1993). In his paper, he showed strong regional dependence of attenuation parameters in Japan. However density of the stations was very low (he used only 16 stations). In our case, we take benefit of the high quality Hi-net data in order to elaborate high resolution maps.

2. Data Analysis and Mapping Process
For this study, we have collected Hi-net data from more than 135,000 events with magnitudes higher than 1.5 and maximum 3.5 within the period June 2002-December 2007. Hi-net stations uniformly cover the Japanese Islands with a spacing of 20–30 km, allowing us to plot high resolution maps. From each station we obtained data for the three components and the records provide signals sampled at a 100Hz rate. For every station, we collected events within a radius of 100km and a maximum depth of 40km. At least 20 events are collected at every station to perform each calculation. If a smaller amount of data is available calculations are not performed. Then our results will show lateral variations of the average properties of the crust around each station. We consider the square sum of the three component amplitudes of the incoming S-waves and the following coda as a measure of the energy arriving to the station. Then, a single energy envelope, representing the total energy arriving to each station will be considered for each event. From this envelope, attenuation parameters will be evaluated.

In our implementation of the MLTWA we consider a single station for each calculation. For each event located at a certain hypocentral distance, the addition of the mean squares of amplitudes of each component, within a lapse time of 45s, measured from the onset of S-waves, is computed. We only consider time traces with signal to noise ratio higher than 3 for the three windows. Then, we use the same amount of data for each window. We believe this might be convenient since lateral variations in Japan are sometimes strong and it is important to make sure that the same amount of information is contained in each window. We were able to obtain more than 190,000 useful time traces. To filter the envelopes, we use bandpass Butterworth filters of order 3, back and forth, to avoid phase delays. We
consider the following frequency bands: 1-2Hz, 2-4Hz, 4-8Hz, 8-16Hz and 16-32Hz. After the filtering process we keep a 100 Hz sampled signal. In advance to filtering, we assigned zeroes to all the part of the seismogram that corresponds to times smaller than the arrival of the S-wave in order to avoid further contributions from P waves. This way, for the first time window, the integration process may start -1s before the arrival of the S-wave in order to take into account correctly the energy contained in the first window (Hoshiba, 1993). We consider three 15 seconds time windows in total. We normalize each integral by dividing the signal with the integration window of +/-5s seconds around tref = 65s. In order to determine the parameters, it is necessary to compare the observed results with the ones provided from a certain theoretical model by means of a non-linear fitting process. We use Paasschens' solution (1997) in this study for modeling coda waves’ envelopes. To perform the non-linear fitting we used the Levenberg-Marquard algorithm. The Levenberg-Marquard algorithm is a wise combination of the steepest descent and the Gauss-Newton algorithms (Press et al., 2007) and it is much faster and exact than a grid search. Moreover, it provides a linear estimate of the errors of the parameters from the values of the variance-covariant matrix.

Since coda decay depends on the lapse time, and we wish to compare the coda decay with the corresponding values of intrinsic absorption and scattering loss, we have to consider a similar portion of the time trace to compute the coda decay. Then, we compute it from 2r/v up to r/v + 45s from the origin time. This procedure ensures that at least 20 seconds are considered. It also ensures the consistency between the measurement of \( Q^{-1} \) and the measurements of the other attenuation parameters.

The results of every measure for each attenuation parameter are assigned to the coordinates of each station. This is an important decision since we wish to show the dependency of the results with the location of the station used to perform each measurement. The result in each station should represent an average of the attenuation parameters around the station. Taking into account that the maximum hypocentral distance under consideration is 100 km and the maximum depth is 40 km, this is quite a large volume of about 100 km depth (in the single scattering approximation).

The maps have been plotted by means of the GMT routines. A grid with a spacing of 0.04º (about 4.5 km) from the original data was created and we used the function “surface” to generate a continuous map with a tension factor of 1.0, which is the maximum value. In this way we are able to show an informative continuous distribution and at the same time we keep the maximum amount of original information about local variations of the parameters being measured.

3. Results

In the intrinsic absorption maps of Figure 1, we can see the following: i) high levels of absorption in Hokkaido, ii) higher absorption in the west side than in the east side of Tohoku area, iii) high absorption in the volcanic arc in Tohoku for the lower frequencies, iv) high absorption in Central Japan, v) high absorption in the “Kinki spot” in Kii peninsula and iv) higher absorption in Kyushu than in Chugoku and Shikoku regions. In Figure 2, \( Q^{-1} \) maps show similar characteristics to those corresponding to intrinsic absorption as well as for the frequency dependence.
Figure 1. Maps of intrinsic absorption for several frequency bands in the range 1-32Hz.

Figure 2. Maps of $Q_c^{-1}$ for several frequency bands in the range 1-32Hz.
In the scattering loss maps of Figure 3, we notice the following: i) scattering loss is high in the volcanic arc in Hokkaido and Tohoku areas for the 1-2Hz band, ii) high values in some regions of Central Japan, iii) high values in Chugoku region for 4-8Hz, 8-16Hz and 16-32Hz as well as for the “Kinki” spot, iv) low values in Shikoku for all the frequency bands and higher values in Kyushu. It is possible to correlate these results with results found by means of velocity tomography. A good example of this correlation corresponds to the Chugoku region and to the Kinki spot. In both regions we can see strong frequency dependence for the scattering loss. In these regions, velocity tomography studies have been performed. These studies show the existence of very important velocity anomalies that have been interpreted as upwelling of fluids from the mantle (Nakajima and Hasegawa, 2007). Another good example of this is the behavior of the northern region of Hokkaido, where a large velocity anomaly has been detected (Nakamura et al. 2008).

4. Conclusions

The maps show strong regional variation of the attenuation parameters. These variations are related mainly with the geotectonic setting of each region and its volcanism mechanisms. For the lower frequency bands, it is possible to identify volcanic areas of Japan because they show strong scattering loss and/or strong intrinsic absorption. For higher frequency bands there are also high scattering and/or absorption regions that correlate with the existence of low velocity anomalies, detected by means of velocity tomography, that are interpreted as regions containing fluids.