



HOKKAIDO UNIVERSITY

Title	Effect of frequency-dependent site amplification and heterogeneous attenuation structure for the precise real-time prediction of ground motions
Author(s)	小木曾, 仁
Degree Grantor	北海道大学
Degree Name	博士(理学)
Dissertation Number	乙第7074号
Issue Date	2019-03-25
DOI	https://doi.org/10.14943/doctoral.r7074
Doc URL	https://hdl.handle.net/2115/75438
Type	doctoral thesis
File Information	Masashi_Ogiso.pdf



Effect of frequency-dependent site amplification and
heterogeneous attenuation structure
for the precise real-time prediction of ground motions
(周波数依存性を考慮したサイト特性と不均質減衰
構造が地震動即時予測に及ぼす効果について)

Masashi Ogiso

A thesis submitted to the Graduate School of Science, Hokkaido University
for the degree of Doctor of Philosophy

March 2019

Acknowledgment

First of all, I would like to express my sincere gratitude to Professor Kiyoshi Yomogida at Hokkaido University. He was my supervisor when I was a master course student at Hokkaido University. Even after my graduation, he has always been interested in my work and given me helpful comments. I have been inspired by discussions with him in many occasions. Also, he has suggested me to apply for a Ph. D. degree. This thesis would never have been completed without his continuous encouragement.

Secondly, I am grateful to Dr. Mitsuyuki Hoshiba at the Meteorological Research Institute (MRI). Since I started the current research on earthquake early warning systems at MRI, he has encouraged and helped me to conduct my research on site amplification and attenuation structure as well as seismic scattering. In addition, he has given me opportunities to attend scientific meetings, including several international ones. These experiences made me aware of the importance of broad perspectives in the scientific research in the world.

I appreciate the member of the examination committee of the doctoral thesis, Prof. Kazunori Yoshizawa, Prof. Masato Furuya, Prof. Youichiro Takada, and Prof. Hiroaki Takahashi at Hokkaido University, and Prof. Kazuo Yoshimoto at Yokohama City University for their careful reviews and constructive comments on this thesis and presentation.

I thank Dr. Marco Pilz at GFZ Potsdam for careful reviews and useful suggestions to improve the quality of this thesis.

I wish to thank all of my colleagues at the Seismology and Tsunami Research department and Volcano Research Department of MRI, especially Mr. Yuki Kodera and Ms. Emi Toshima, for their support in various occasions to conduct this study at MRI.

Chapter 2 is the outcome of collaborative researches with Mr. Shigeki Aoki at the Japan Meteorological Agency and Dr. Mitsuyuki Hoshiba, and a part of chapters 3 and 4 comes from the results obtained with Dr. Azusa Shito and Prof. Satoshi Matsumoto at Kyushu University,

and Dr. Mitsuyuki Hoshiba. I thank all of them. I also thank Ms. Sanae Tsuchiko for her efforts to produce the dataset of phase arrival times used in chapter 3.

In this study, I utilized seismic waveforms recorded by Hi-net, KiK-net and K-NET operated by the National Research Institute for Earth Science and Disaster Resilience. Seismic records observed by the seismic networks of Kyushu University, Kyoto University, and the Japan Meteorological Agency were also used. I thank the staffs of these institutions and organizations for their efforts to maintain seismic networks.

Some parts of this study were financially supported by the Joint Usage/Research program of the Earthquake Research Institute of the University of Tokyo and Grant-in-Aid for Scientific Research from the Japan Society for the Promotion of Science (JSPS KAKENHI) Grant Numbers JP25282114, JP17H02064, and JP18K13622.

Finally, I would like to thank my parents for their invaluable supports for the completion of this thesis.

Abstract

The 2011 Off the Pacific coast of Tohoku earthquake (hereafter called the Tohoku earthquake) and its following seismicity around Japan revealed the limitation of any earthquake early warning (EEW) systems based on the estimation of source parameters in their first stage. In general, the ground motion at a given site is represented by the convolution of source, path, site amplification, and instrumental response terms. Because the observed waveforms should reflect the complex effects of source and path terms up to the arrival time, a ground motion prediction directly from observations, that is, a wavefield monitoring approach for EEW is one of the promising schemes to overcome the issues arisen from the Tohoku earthquake. To improve the precision of ground motion predictions in the wavefield monitoring approach, we focus on the effects of site amplification and path terms.

Chapter 2 discusses about site amplification factors. Site amplification factors usually have frequency dependency. However, site amplification has been measured by scalar values in the ground motion prediction equation adopted in most of the current operational EEW methodologies. To consider the frequency dependency of site amplification characteristics for the purpose of EEW, we estimate frequency-dependent site amplification factors and corresponding time-domain minimum-phase filters which approximate frequency characteristics for strong motion stations in Japan. These time-domain filters can be applied to the observed waveforms in a real-time manner so that they are utilized for a real-time ground motion prediction. The effect of real-time correction of frequency-dependent site amplification has been investigated through experiments of seismic intensity prediction by the correction of the site amplification factors to the observation at an adjacent station. The frequency-dependent site amplification correction has yielded results with smaller residuals than the correction by scalar values. The shorter the interstation distance become, the more the frequency-dependent correction improves. In other words, some path terms are required for accurate prediction

with a long interstation distance (i.e., long lead time).

In chapter 3, we focus on attenuation properties as a part of the above mentioned path effects. Seismic attenuation for high-frequency (≥ 1 Hz) seismic waves is classified into two mechanisms: intrinsic and scattering attenuations. Since seismic energy is conserved during scattering, scattering attenuation affects the duration of the ground motion as well as the amplitude. Separation and precise estimation of intrinsic and scattering attenuation structures would lead to the better prediction of the ground motion, not only the maximum amplitude but also the duration of ground motion. We propose a new method to estimate intrinsic and scattering attenuations separately in 3D space. The proposed method consists of two parts, the estimation of attenuation parameters by the envelope fitting method followed by mapping these parameters into 3D space with sensitivity kernels of the attenuation parameters. A depth-dependent velocity model has been considered to include the effect of a realistic seismic velocity structure on the observed shape of seismogram envelopes. We have adopted a non-isotropic scattering model to treat an envelope broadening effect which is important for the duration of ground motions. The newly estimated 3D heterogeneous intrinsic and scattering attenuation distributions in southwestern Japan has been compared with the 2D structure estimated by the multiple lapse time window analysis. Strong intrinsic and scattering attenuations in the Kyushu region, whose tectonics is more active than the other regions, has been detected by the two analyses. In addition, we have found depth dependency of both intrinsic and scattering attenuations, especially in the Chugoku region. Furthermore, a low intrinsic attenuation zone was found at the depth of 10-50 km in the central Shikoku region. Such features could not be retrieved by the conventional 2D analysis of attenuation structure.

In chapter 4, the effect of heterogeneous attenuation structure in the Numerical Shake Prediction scheme, which is based on the wavefield monitoring approach, is investigated by the simulations of ground motion prediction for the 2016 Kumamoto earthquake. We have extended the calculation code of the Numerical Shake Prediction scheme to incorporate the

heterogeneous attenuation structure estimated in chapter 3. First, the 2D heterogeneous attenuation structure derived from the multiple lapse time window analysis based on the isotropic scattering model was used in the simulation. Comparing the prediction residuals of maximum seismic intensity, the introduction of 2D heterogeneous attenuation structure yielded 12% lower for 10 s ahead and 15% lower for 20 s ahead predictions. Next, we conducted the same experiment but based on the non-isotropic scattering model. The prediction accuracy of maximum seismic intensity was improved about 10% for predictions 10 s ahead and 9% for predictions ahead 20 s, with a slightly different tendency to those of the isotropic scattering model, probably originated from the difference in the dimension of the estimated attenuation structure and the ground motion prediction simulation. Comparing the four prediction models, i.e., isotropic or non-isotropic scattering model and homogeneous or heterogeneous structure, the heterogeneous structure with the non-isotropic scattering model resulted in the smallest root-mean-square value in residuals of the whole predicted envelopes. These results indicate that the non-isotropic scattering model has an ability to predict the whole envelope with the Numerical Shake Prediction scheme and that the heterogeneous attenuation structure enables to improve the prediction precision of the scheme.

In summary, the time-domain filters introduced in chapter 2 enables us to correct site amplification factors for the observed waveform in real time. In addition, the heterogeneous attenuation structures estimated in chapter 3 results in a more precise prediction of ground motions in the Numerical Shake Prediction scheme, as shown in chapter 4. These results will lead to more precise real-time prediction of ground motions based on the wavefield monitoring approach.

概要

平成 23(2011) 年東北地方太平洋沖地震とそれに引き続く地震活動によって、震源要素の早期推定に基づく緊急地震速報の手法としての限界が明らかになった。一般的に、地震動は周波数領域において震源項・伝搬項・サイト特性項の積で表現される。緊急地震速報ではある時点までの実際の観測地震動を予測に用いることが一般的な地震動予測と異なる点である。観測地震動はその時刻までの複雑な震源項や伝播項の影響を含んだものであるから、緊急地震速報においては波動場のリアルタイムモニタリングに基づいた地震動即時予測というアプローチが可能である。本研究では波動場のモニタリングに基づいた地震動即時予測の精度向上を目的として、サイト特性及び伝播項の効果に着目する。

一般的にサイト特性は周波数依存性を持つが、現在運用されている緊急地震速報システムでは単一のスカラー量として補正される。第 2 章では、地震動即時予測においてサイト特性の周波数依存性を考慮するため、日本全国の強震観測点における周波数依存性をもつサイト特性を推定するとともに、そのサイト特性を近似する時系列フィルタを推定した。このフィルタは最小位相特性を満たしており、観測波形に対して時系列上でリアルタイムに適用することが可能である。周波数依存性の考慮がどの程度有効か調査するため、サイト特性を補正することによって隣接観測点の震度を予測するという実験を行った。2 点間の平均震度差を補正量とするスカラー補正法に対して、本研究で推定した時系列フィルタを用いて観測波形のサイト特性を補正した後に予測震度を計算する手法のほうがよい予測成績を示した。観測点間距離が短くなるほど周波数に依存した補正の優位性が大きかった。これは、観測点間距離が長い、すなわち、猶予時間が長い予測のためにはサイト特性以外の影響の考慮が必要であることを示唆している。

地震動の振幅はその伝播経路にも依存する。第 3 章では伝播経路特性のうち、地震波減衰構造について考察する。おおよそ 1Hz より高い周波数帯においては、地震動の振幅は媒質の内部減衰及び散乱減衰というメカニズムの異なる 2 種類の減衰の影響を受ける。内部減衰は地震波エネルギーの減少による振幅減衰をもたらし、散乱減衰は地震動の振幅減

衰に加えてその継続時間にも影響を及ぼす。したがって、正確な内部減衰と散乱減衰の分離推定は地震動の振幅のみならず継続時間をも含めた高度な地震動予測につながると期待される。本研究では、地震波エンベロープに着目して、新たな内部減衰と散乱減衰の3次元空間分布推定手法を提案する。提案手法は2つの部分からなり、最初にエンベロープフィッティングを用いて各震源-観測点ペアごとに内部減衰と散乱減衰のパラメータを推定する。次に、得られたパラメータをセンシティブリティカーネルを用いて3次元空間に投影していく。どちらの過程においても、深さ依存性を持つ地震波速度構造の影響も考慮した。エンベロープのモデリングにあたっては、主要動付近の形状を的確に表現するために前方散乱モデルを採用した。新たに提案した手法を用いて推定した西南日本地域の3次元内部・散乱減衰構造と従来の分離推定手法を用いて推定した2次元の内部・散乱減衰構造を比較したところ、地殻活動が活発な九州地方は他の地域より内部減衰・散乱減衰がともに大きいという共通した特徴が見られた。また、新たに推定した3次元構造では、特に中国地方において下部地殻の内部・散乱減衰がともに上部地殻のそれより小さいという特徴が明瞭に見られた。さらに、深さ10-50kmにおいては四国中央部に低内部減衰域が推定された。これらの特徴は従来の2次元解析では発見できないものである。

第4章では、不均質減衰構造が地震動即時予測に及ぼす効果を調査するため、波動場のモニタリングに基づく地震動即時予測手法である「揺れの数値予報」において不均質減衰構造が考慮できるよう拡張し、2016年熊本地震を対象としてこの手法に基づく地震動即時予測実験を行った。まず、等方散乱モデルに基づく、従来の分離推定手法で推定した2次元内部・散乱減衰構造を用いた場合と均質な減衰構造を用いた場合を比較した。最大震度の予測残差のRMSを比較すると、不均質減衰構造を考慮した場合、均質構造の場合と比較して10秒後予測の場合は12%、20秒後予測の場合は15%の改善が見られた。次に、前方散乱モデルを用いて同様の実験を行った。その際、不均質減衰構造としては3章で推定した3次元減衰構造のうち、第2層目の構造を用いた。その結果、最大震度の予測残差のRMSは10秒後予測で10%、20秒後予測で9%の改善がみられた。等方散乱モデルの場合と比較すると改善傾向がやや異なるが、その原因として減衰構造と波動伝播シミュ

レーションにおける次元の違いが影響していることが考えられる。4種の予測モデルにおいてエンベロープ全体の予測残差の RMS を計算すると、前方散乱モデルのもとで不均質構造を考慮した場合が最も良い成績であった。これらの結果から、前方散乱モデルは「揺れの数値予報」において地震動のエンベロープ全体の正確な予測を可能と示うこと、また、不均質減衰構造の導入により「揺れの数値予報」の震度予測精度が向上することが明らかとなった。

本研究で得られた時系列フィルタはリアルタイムで波形レベルでのサイト特性補正を可能とし、また、本研究で推定された不均質減衰構造は「揺れの数値予報」において地震動の推定精度を向上させることが明らかとなった。これらの成果は、波動場のモニタリングに基づく地震動即時予測において、その予測精度の向上に重要な役割を果たすであろう。

Contents

1	Motivation of this study	1
1.1	A brief introduction to earthquake early warning	1
1.2	The 2011 off the Pacific coast of Tohoku earthquake (Mw 9.0) and EEW	5
1.3	Improvement of the EEW methodologies after the Tohoku earthquake	6
1.4	Outline of this study	7
2	Frequency-dependent site amplification for real-time ground motion prediction	9
2.1	Introduction	9
2.2	Experimental methods of seismic intensity prediction	10
2.3	Estimation of frequency-dependent site amplification factors	11
2.4	Estimation of digital filters	14
2.5	Data and procedures for the frequency-dependent site amplification	15
2.6	Waveform prediction using site-correct filters	17
2.7	Comparison of two intensity prediction methods	19
2.8	Discussion	24
2.9	Conclusion	27
3	Estimation of heterogeneous attenuation structure	32
3.1	Introduction	32
3.2	2D heterogeneous attenuation structure of southwestern Japan estimated by the Multiple Lapse Time Window Analysis	35
3.3	Estimating 3D distribution of intrinsic and scattering attenuations in southwest- ern Japan	38
3.4	Discussion	59

3.5	Conclusion	65
4	Numerical shake prediction with heterogeneous attenuation structure	68
4.1	Introduction	68
4.2	Ground motion prediction scheme: Numerical Shake Prediction	69
4.3	Target earthquake, data and preprocessing for ground motion prediction experiments	74
4.4	Results of the Numerical Shake Prediction with a heterogeneous attenuation structure	76
4.5	Discussion and conclusion	91
5	Conclusion and future perspectives	101
5.1	Summary of this study	101
5.2	Perspectives for future studies	103

1 Motivation of this study

1.1 A brief introduction to earthquake early warning

History of the earthquake early warning

The basic concept of earthquake early warning (EEW) is to warn of strong ground motion at a given point before its arrival. Unlike the traditional earthquake prediction, EEW does not concern earthquake occurrence but ground motion immediately after an earthquake occurs. Its lead time is usually short, less than a few tens of seconds. According to Allen et al. (2009), the first appearance of the idea of EEW can be found in the report by J. D. Cooper at *San Francisco Bulletin* on November 3, 1868. In this report, Cooper proposed the idea that just after the detection of an earthquake near its hypocenter, ringing a bell at a densely inhabited area by electronic current over a wire from the detected place. Cooper's idea is based on the difference in propagation velocity between seismic waves and electronic current. Because the propagation velocity of electronic current is much faster than that of seismic waves, it is expected to notify the occurrence of an earthquake immediately after its detection, if an appropriate detection analysis and communication systems exist. To implement the idea of EEW, not only waveform analysis techniques but also reliability of the equipment such as seismometers and communication line compose key issues. After about one hundred years later, thanks to the progress of both waveform analysis techniques and equipment, Cooper's idea has been realized.

Development of the EEW systems around the world

In Japan, studies on early detection and analysis of earthquakes have been conducted in the safe operation of railway transportation systems since 1950s. Nakamura & Saita (2007) summarizes the history of EEW for railway transportation systems. Nakamura's pioneering

work on the analysis of initial P-waves of earthquakes, including the estimates of both the backazimuth from their portion and the magnitude from their predominant period (Nakamura, 1988), had great influence on the following studies in real-time seismology such as Odaka et al. (2003), Wu & Kanamori (2005), Kanamori (2005) and so on.

After the implementation of the EEW system for the railway transportation system (e.g., Ashiya, 2004), the Japan Meteorological Agency (JMA) planned to issue the warning of ground motion for general public, and began to develop a new EEW system in cooperation with the Railway Technical Research Institute and the National Research Institute for Earth Science and Disaster Prevention (NIED; now the National Research Institute for Earth Science and Disaster Resilience). Utilizing some excellent works such as Odaka et al. (2003), Tsukada et al. (2004), Horiuchi et al. (2005) and other related works, JMA launched the operation of their EEW system for general public in October 2007 (Hoshiba et al., 2008; Doi, 2011).

Let us turn our attention to worldwide effort. In Mexico, the Michoacan earthquake (magnitude 8.1), which occurred on September 19, 1985, was the turning point for the EEW system (Espinosa-Aranda et al., 2009). Over ten thousand people were died in Mexico City, about 400 km away from the epicenter. After the event, the Centro de Instrumentación y Registro Sísmico (CIRES) began to develop the Sistema de Alerta Sísmica of Mexico City (SAS), an EEW system for Mexico City, in 1988, and its experimental operation started in 1991. Since 1993, SAS has issued warning information for general public in Mexico (Hoshiba, 2012). 14 years after the Michoacan earthquake, a magnitude 6.7 earthquake occurred near Oaxaca on June 15, 1999. This event has lead to the another EEW system development for Oaxaca, called Sistema de Alerta Sísmica de Oaxaca (SASO). SAS and SASO has been integrated into the Seismic Alert System of Mexico (SASMEX) by the CIRES. SASMEX has been working well for large earthquakes, including magnitude 8 class event occurred on September 8, 2017 (http://www.cires.org.mx/sasmex_historico_es.php).

Taiwan is located on the plate boundary and several destructive earthquakes have struck

there so that the EEW is expected to reduce the seismic hazard (Wu et al., 2013). According to Wu et al. (2013), initial experimental EEW system was constructed by the Central Weather Bureau (CWB) in 1993. The initial system was designed to issue ground motion warnings to the capital, Taipei, for the earthquakes occurred on central-eastern area (around Hualien) of Taiwan. Later, the CWB has begun to utilize the nationwide strong motion seismometer network for nationwide service of the EEW. At the same time, the CWB has continued to improve the EEW system, and currently the system called the Earthworm Based Earthquake Alarm Reporting (*e*BEAR) has been implemented for the EEW service (Chen et al., 2015). The CWB has begun to issue ground motion warning for general public at least since 2018 (e.g., <https://www.recordchina.co.jp/b567522-s0-c10-d0135.html>, last accessed August 2018), although no official announcements seem to be issued from CWB.

In Romania, the intermediate-depth earthquakes originating at the Vrancea region, about 140 km north of Bucharest are expected to be the major seismic hazard at the country (Böse et al., 2007). Ionescu et al. (2016) reports the status of the EEW service at Romania. According to Ionescu et al. (2016), the National Institute for Earth Physics (NIEP) has operated the Rapid Earthquake Early Warning System (REWS), which consists of two systems: one is based on the PProbabilistic and Evolutional early warning SysTem (PRESTo; Satriano et al., 2011) with 110 strong motion stations covering the whole country, and the other is developed by the NIEP specialized for earthquakes at the Vrancea area. The REWS has been in operational phase since 2013, and has been sending warning information for not only emergency authorities but also general public (Ionescu et al., 2017).

Basic methodology of current operational EEW systems

Current operational EEW algorithms can be classified into two types: (1) on-site and (2) network-based algorithms. The on-site algorithm utilizes seismic records from only one station, and usually predicts ground motions of S-waves from those of P-waves around a given

observation site (e.g., Kanamori, 2005; Böse et al., 2009; Parolai et al., 2017). One advantage of on-site algorithm is that the algorithm is self-completed to issue the alerts. In the area close to the hypocenter, on-site algorithm would be powerful because the algorithm does not have to wait the analysed results from several stations. The network-based algorithm utilizes seismic records from several stations. In general, the network-based algorithm is more reliable than the on-site algorithm and would issue warnings with long lead time when the hypocenter is not close to the target area. In this study, we will not discuss the on-site algorithm and will focus on EEW systems based on the network-based algorithm.

Basic methodologies of currently operating network-based EEW systems, such as JMA EEW system (Hoshiba & Ozaki, 2014), E-larmS/E-larmS2 (Allen, 2007; Kuyuk et al., 2014) and PRESTo (Satriano et al., 2011), are similar to each other. Each warning system consists of several techniques of waveform analysis:

1. Event detection as a result of appropriate grouping of triggered stations
2. Phase arrival picking (usually for P-waves) and hypocenter determination
3. Magnitude estimation from the maximum amplitude of each waveform with some empirical relationships
4. Ground motion prediction at a target site by utilizing a ground motion prediction equation (GMPE) with hypocentral parameters (location and magnitude)

The basic concept of each EEW system is in common; to determine hypocentral parameters then predicting ground motion from those parameters as quickly as possible. Because there are a strict limitation in calculation time, usually some selected stations of P-waves are used in these EEW systems.

1.2 The 2011 off the Pacific coast of Tohoku earthquake (Mw 9.0) and EEW

The Tohoku earthquake (e.g., Hirose et al., 2011) is one of the devastating earthquakes in and around Japan. According to Hoshiba et al. (2011), the JMA EEW system successfully issued a warning message before the arrival of strong ground motion for the Tohoku region (about 200 km from the epicenter), however, the warning was not issued for the Kanto region (about 200 to 350 km from the epicenter), where the maximum observed seismic intensity was 6 upper, the 2nd highest value of the JMA seismic intensity scale. One reason of missing alerts for the Kanto region was the underestimation of its earthquake magnitude at an early stage. The JMA EEW system estimated its magnitude in the JMA scale as 7.2 when the system issued the first warning (8.6 s after the first triggered time), and finally a magnitude of 8.1 at 105.0 s (Hoshiba & Ozaki, 2014). However, the seismic moment of this event was estimated to be in the order of 10^{22} Nm (corresponding moment magnitude is 9.0), and the duration of its source rupture time was up to 150 s (e.g., Yoshida et al., 2011). The JMA EEW system was forced to estimate its magnitude with a few P-wave portions while the rupture was still ongoing, resulted in the underestimation of its magnitude. In addition, point source assumption for GMPE was not appropriate for such a huge event. For example, Kurahashi & Irikura (2011) estimated five strong motion generation areas, and one of them was about 200 km far from the epicenter. In such a case, the assumption of point source was violated.

The seismicity around Japan after the Tohoku earthquake was more active than before (Hirose et al., 2011). Not only the aftershocks around the large focal area, the Tohoku earthquake induced many earthquakes over all the Japanese islands. Due to the large number of aftershocks and induced earthquakes, earthquakes occasionally occurred simultaneously, which confused the operation of the JMA EEW system. If several earthquakes occurred simultaneously, stations far from each other also triggered simultaneously and the system

might recognize these plural events as one event. Thus, the system estimates the hypocenter and magnitude improperly, resulting in an inappropriate ground motion prediction (Hoshiba et al., 2011; Hoshiba, 2014).

1.3 Improvement of the EEW methodologies after the Tohoku earthquake

EEW can be classified as a kind of ground motion predictions. In general, the observed ground motion is represented by the convolution of source, path, site amplification, and instrument response terms. For a precise ground motion prediction, careful treatment of the first three factors are required (e.g., chapter 26 of Hasegawa et al., 2015). This statement also holds for the EEW. In the EEW, observed waveforms can be used for prediction. In this way, using observed waveforms as an input is different from standard ground motion prediction problems in strong ground motion seismology. The conventional EEW methodologies (section 1.1) make use of this point to estimate source characteristics, such as its location and magnitude. As mentioned in the section 1.2, the Tohoku earthquake has revealed the two issues. Both issues are related to the estimation of source terms: the first is the underestimation of the magnitude and the second is the inappropriate treatment of the earthquakes occurred simultaneously. To tackle the first issue, utilizing the real-time GNSS data to estimate magnitude of a large earthquake have been proposed (e.g., Colombelli et al., 2013; Grapenthin et al., 2014). A novel grouping method based on bayesian inference (Tamaribuchi et al., 2014; Wu et al., 2015) can be the solution for the second issue.

Hoshiba (2013b) has devised a different approach for EEW from the above mentioned studies. The basic concept of Hoshiba (2013b) is the real-time monitoring of ground motions with a dense seismic network and ground motion prediction based on the physics of wave propagations. Hereafter we call the approach of Hoshiba (2013b) as the wavefield monitoring approach in this study. The wave equation (e.g., section 18.4 of Snieder, 2004) describes

that the time evolution of ground motions is determined by the spatial distribution of ground motions if there is no external excitation term. Hence, if the current distribution of ground motions is obtained, ground motions in the future can be predicted based on the wave equation. This approach of Hoshiha (2013b) also utilizes observed waveforms, but the usage is different from the approaches mentioned in section 1.1. Hoshiha & Aoki (2015) has implemented the concept of Hoshiha (2013b), by adopting a data assimilation technique to monitor ground motions in real time and the radiative transfer theory in its prediction stage. Despite the continuous efforts for the early estimation of source terms mentioned above, attempts to improve the prediction precision in the wavefield monitoring approach are rare.

1.4 Outline of this study

In this study, to improve the precision of ground motion prediction based on the wavefield monitoring approach, we focus on two of the three terms that affect the observed ground motions; site amplification and path terms. For the purpose of EEW, waveform analysis needs to be conducted in real time. We propose methodologies applicable to consider these two factors for the real-time prediction of ground motions.

In chapter 2, following the methodology of Hoshiha (2013a), we estimate site amplification factors of the Japanese dense strong ground motion networks called Kyoshin Network (K-NET) and Kiban Kyoshin Network (KiK-net) (Okada et al., 2004; Aoi et al., 2004), and approximate the frequency characteristics of site amplification in each station by time-domain filters. Then, we show how effective it is to introduce frequency dependency in site amplification through seismic intensity prediction experiments. Next, we focus on path terms, especially their attenuation effect. In chapter 3, we propose a new method to estimate intrinsic and scattering attenuations separately in three-dimensional (3D) space, and compare this new 3D attenuation model with the two-dimensional (2D) attenuation model estimated by a conventional method. Then, we examine the importance of heterogeneous attenuation structure in the real-time

ground motion prediction through the ground motion prediction simulation at chapter 4.
Finally, we summarize this study and future perspectives in chapter 5.

2 Frequency-dependent site amplification for real-time ground motion prediction

This section has been published as:

Ogiso, M., S. Aoki, and M. Hoshiya, Real-time seismic intensity prediction using frequency-dependent site amplification factors, *Earth Planets and Space*, **68**:83, 2016.

2.1 Introduction

It is often observed that the different maximum amplitudes of ground motion in time domain at two stations for a common earthquake even if their interstation distance is less than several hundred meters (e.g., Noguchi et al., 2005). This difference can be attributed to the velocity contrast in the subsurface structure (e.g., Bard & Bouchon, 1985), 2D and 3D basin shape (e.g., Sánchez-Sesma & Luzón, 1996), wave scattering and focusing effects due to the topography around stations (e.g., Bouchon, 1973), and so on, which is called site effects. The empirical site amplification factors derived from observed waveforms are widely used to model these complex effects in actual data. Because site amplification factors usually show frequency dependency (e.g., Satoh & Kawase, 2009), many studies have attempted to estimate amplification factors with frequency dependency (e.g., Phillips & Aki, 1986; Iwata & Irikura, 1986; Kawase & Matsuo, 2004; Ikeura & Kato, 2011; Takemoto et al., 2012; Nakano et al., 2015b). On the other hand, site amplification effects are corrected by simple scalar values in many ground motion prediction equations (GMPEs), where the prediction variable is the peak ground acceleration or peak ground velocity (e.g., Midorikawa, 2009). Although ground motion predictions with frequency-dependent site amplification have been conducted (e.g., Nozu et al., 2007; Hata et al., 2011), their use in any real-time (EEW) systems has yet to be investigated in detail.

Hoshiba (2013b) has recently proposed the scheme of real-time ground motion prediction where future ground motion is predicted directly from the observations in real time, avoiding difficulties in estimating hypocentral parameters. Correction of site amplification remains important in their approach, and Hoshiba (2013a) has proposed the use of digital causal filters that can be applied in a real-time setting for making the correction.

In this chapter, we explore the above new concept of Hoshiba (2013b) by predicting ground motion at a target station from the observed waveforms of its adjacent station. If their interstation distance is small compared with the epicentral distances, differences between ground motions of the two stations are mainly due to the difference in their site amplification factors. In such a case, we may be able to predict the ground motion at a target station from the observation at the adjacent station by applying their site amplification corrections. We conduct experiments on seismic intensity prediction at a target station from the waveforms of the adjacent station using two methods and compare their results. We use the seismic intensity scale defined by the JMA (JMA seismic intensity) as the index of ground motion. The JMA seismic intensity is defined as

$$I_{\text{JMA}} = \log_{10} A_c^2 + 0.94, \quad (2.1)$$

where A_c^2 is the sum of squares of band-pass filtered three-component accelerograms (Japan Meteorological Agency, 1996; Hoshiba & Ozaki, 2014).

2.2 Experimental methods of seismic intensity prediction

We have used two different methods to predict seismic intensity at a target station from observed waveforms of an adjacent station in a real-time manner. Our underlying assumption is that the differences between seismic intensities of the two stations arose mainly from differences in their site amplification factors. Our two prediction methods differ in their

correction for site amplification. The first method employs frequency-dependent correction. If we know the frequency-dependent site amplification factor at both stations, we can adjust the site amplification term of one station to that of the other station through deconvolution and convolution of the site amplification term. Correction of the site amplification term allows us to derive the predicted waveforms at the second station from those of the first. We can then calculate predicted seismic intensity from the predicted waveforms. Usually, deconvolution and convolution of the site amplification term are done in the frequency domain. However, Hoshiba (2013a) has proposed a real-time convolution and deconvolution technique employing digital causal filters in the time domain, which we have adopted for this study. The second method uses scalar correction. If many earthquakes have been recorded at both of two adjacent stations, we can calculate the average differences between the seismic intensity of the stations. To predict seismic intensity at one station, we simply add that average difference (a scalar value) to the seismic intensity at the other station.

Although the JMA seismic intensity originally requires a calculation in the frequency domain, we have adopted the procedure proposed by Kunugi et al. (2013) for a real-time calculation of the JMA seismic intensity in the time domain. This allowed both of our prediction methods to be applied in real time, as is required for EEW.

2.3 Estimation of frequency-dependent site amplification factors

For the frequency-dependent correction method described in section 2.2, we need relative site amplification between the station pairs. Let us consider the case when an earthquake is observed at two stations. If the epicentral distance is large compared with the interstation distance, we can neglect the small differences of radiation patterns at each station and attributed the differences in amplitude to the relative site amplification and differences in the path term. Ikeura & Kato (2011) adopted the above assumption to estimate site amplification factors. They calculated spectral ratios of direct S-waves for many station pairs, then estimated relative

site amplification factors with respect to a reference station by solving the relevant equation systems simultaneously. We took the similar approach in this study. Figure 2.1 shows the schematic illustration of the site amplification estimation.

The observed amplitude of a seismic waveform at the i -th station for the k -th earthquake in the frequency domain can be expressed as follows:

$$O_{ik}(f) = R_k(\theta_i, \phi_i) \cdot W_k(f) \cdot T_{ik}(f) \cdot G_i(f), \quad (2.2)$$

where $O_{ik}(f)$ is the observed amplitude, $R_k(\theta_i, \phi_i)$ the radiation pattern (θ_i and ϕ_i are incident angle and azimuth for station i , respectively), $W_k(f)$ the source term, $T_{ik}(f)$ the path term, and $G_i(f)$ the site amplification term at station i . Now we consider the case when two adjacent stations i and j observe the common earthquake k . If the hypocentral distances of each station are long enough compared with the inter-station distance, one can assume that the radiation pattern is almost the same, i.e. $\theta_i \approx \theta_j$ and $\phi_i \approx \phi_j$. In such a case, observed spectrum ratio between stations i and j can be expressed by the following path and site terms:

$$\frac{O_{ik}(f)}{O_{jk}(f)} = \frac{R_k(\theta_i, \phi_i) \cdot W_k(f) \cdot T_{ik}(f) \cdot G_i(f)}{R_k(\theta_j, \phi_j) \cdot W_k(f) \cdot T_{jk}(f) \cdot G_j(f)} \approx \frac{T_{ik}(f)}{T_{jk}(f)} \cdot \frac{G_i(f)}{G_j(f)}. \quad (2.3)$$

The ratio of path terms $T_{ik}(f)/T_{jk}(f)$ is expressed by

$$\frac{T_{ik}(f)}{T_{jk}(f)} = \frac{r_{jk}}{r_{ik}} \exp\left(-\frac{\pi f}{Q(f)\beta}(r_{ik} - r_{jk})\right), \quad (2.4)$$

where $Q(f)$ is the frequency-dependent intrinsic attenuation factor of the medium, β the seismic velocity of the medium, and r_{ik} , r_{jk} are the hypocentral distances from the k -th earthquake to each station. Substituting Eq. (2.4) into Eq. (2.3) and taking a common

logarithm, we can get the relationship between site amplification factors of two stations,

$$\log_{10} \frac{G_i(f)}{G_j(f)} = \log_{10} \left[\frac{O_{ik}(f)}{O_{jk}(f)} \frac{r_{ik}}{r_{jk}} \exp \left\{ \frac{\pi f}{Q(f) \beta} (r_{ik} - r_{jk}) \right\} \right]. \quad (2.5)$$

In this study, the seismic velocity of medium is supposed to be $\beta = 4.0$ km/s. To avoid the influence of the $Q(f)$ to the spectral ratios, Ikeura & Kato (2011) calculated a sophisticated weighted average of observed spectral ratios (right-hand side of Eq. 2.5) without any assumption of the $Q(f)$ value. In this study, however, we adopt the $Q(f)$ value based on Satoh et al. (1994):

$$Q(f) = \begin{cases} 110f^{0.69} & (f \geq 1.0 \text{ Hz}) \\ 110 & (f < 1.0 \text{ Hz}) \end{cases} \quad (2.6)$$

for the correction of the influences of the $Q(f)$. On the other hand, the relative site amplification factor for two station, $G_i(f)/G_j(f)$, can be rewritten by using the site amplification factors relative to a common reference station:

$$\frac{G_i(f)}{G_j(f)} = \frac{G_i(f)/G_{\text{ref}}(f)}{G_j(f)/G_{\text{ref}}(f)} = \frac{G_{i/\text{ref}}(f)}{G_{j/\text{ref}}(f)}, \quad (2.7)$$

where $G_{\text{ref}}(f)$ is the site amplification factor at the reference station and $G_{i/\text{ref}}(f)$, $G_{j/\text{ref}}(f)$ are the relative site amplification factors at stations i and j with respect to the reference station, respectively. Substituting Eq. (2.7) to Eq. (2.5), we get the linear relationship

$$\log_{10} G_{i/\text{ref}}(f) - \log_{10} G_{j/\text{ref}}(f) = \log_{10} \left[\frac{O_{ik}(f)}{O_{jk}(f)} \frac{r_{ik}}{r_{jk}} \exp \left\{ \frac{\pi f}{Q(f) \beta} (r_{ik} - r_{jk}) \right\} \right]. \quad (2.8)$$

The right-hand side of Eq. (2.8) is evaluated by using the observed waveforms, so that we can set up simultaneous equations with unknown parameters, $\log_{10} G_{i/\text{ref}}(f)$ and $\log_{10} G_{j/\text{ref}}(f)$, and solve these equations by a least-squares method.

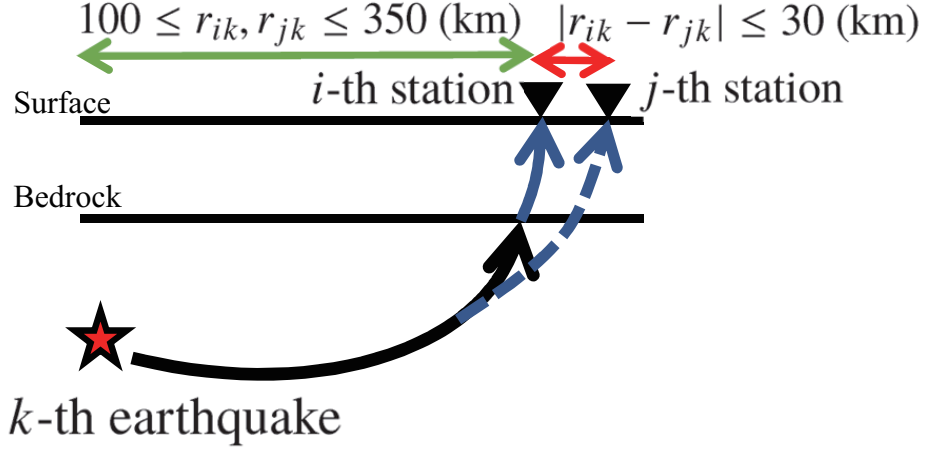


Figure 2.1 Schematic illustration of the relative site amplification estimation.

2.4 Estimation of digital filters

We have built digital filters that approximate site amplification characteristics ("site-correct filters" hereafter) based on Hoshiaba (2013a). Following Hoshiaba (2013a), site amplification characteristics are modeled by combinations of first- and second-order analog filters. Frequency response $F(s)$ of such a combination of analog filters is represented by

$$F(s) = G_0 \prod_{n=1}^{N_f} \left(\frac{\omega_{2n}}{\omega_{1n}} \right) \left(\frac{s + \omega_{1n}}{s + \omega_{2n}} \right) \cdot \prod_{m=1}^{M_f} \left(\frac{\omega_{2m}}{\omega_{1m}} \right)^2 \left(\frac{s^2 + 2h_{1m}\omega_{1m}s + \omega_{1m}^2}{s^2 + 2h_{2m}\omega_{2m}s + \omega_{2m}^2} \right), \quad (2.9)$$

where N_f and M_f are the numbers of first- and second-order filters, ω_{1n} , ω_{2n} , ω_{1m} , ω_{2m} are the angular frequencies and h_{1m} , h_{2m} are the damping factors that determine the frequency characteristics, G_0 is the gain factor, and $s = i(2\pi f)$ where i is the imaginary unit.

After modeling the site amplification characteristics by Eq. (2.9), digital filters are designed using bilinear transform and prewarping techniques. Digital filters which correspond to

Eq. (2.9) are expressed generally in the form:

$$F(z) = g_0 \prod_{n=1}^{N_f} \left(\frac{a_{0n} + a_{1n}z^{-1}}{1 + b_{1n}z^{-1}} \right) \prod_{m=1}^{M_f} \left(\frac{a_{0m} + a_{1m}z^{-1} + a_{2m}z^{-2}}{1 + b_{1m}z^{-1} + b_{2m}z^{-2}} \right), \quad (2.10)$$

where s and z satisfy the following equation:

$$s = \frac{1 - z^{-1}}{1 + z^{-1}}. \quad (2.11)$$

Coefficients of the digital filters a_{0n} , a_{1n} , b_{1n} , a_{0m} , a_{1m} , a_{2m} , b_{1m} , b_{2m} , and g_0 are calculated from the angular frequencies, damping factors, the gain factor in Eq. (2.9) and from the sampling interval of the time series, δT . Hoshiba (2013a) represents these coefficients of the digital filters explicitly using those of the analog filters in its appendix. The time-domain digital filters $F_{1n}(z)$ and $F_{2m}(z)$ ($n = 1, \dots, N_f$; $m = 1, \dots, M_f$) represents the convolution of site amplification at a given station. Inverse filters, which correspond to the deconvolution of site amplification, are derived from the reciprocal of Eq. (2.10). These site-correct filters satisfy causality and can be applicable in real time.

2.5 Data and procedures for the frequency-dependent site amplification

For the site amplification estimation, we used data from the strong-motion stations of the K-NET and KiK-net operated by the National Research Institute for Earth Science and Disaster Resilience (NIED, Okada et al., 2004; Aoi et al., 2004). KiK-net stations have two sensors, one at the surface and the other in a borehole, which are treated as different stations in this study. We collected waveforms from these two networks since May 1996 to May 2014, excluding March and April 2011 to avoid the high seismicity related to the Tohoku earthquake (Hirose et al., 2011). Figure 2.2 shows the epicentral distribution and magnitude-frequency relationship of the used earthquakes. Total number of earthquakes was 5574. While the

sampling frequency of a part of KiK-net records is 200 Hz, that for most other records is 100 Hz. We picked up station pairs that were less than 30 km apart, and calculated spectral ratios from the station records of earthquakes between 100 and 350 km from both stations, to satisfy the assumption of a shared source term (see Fig. 2.1). The records with a peak ground acceleration value over 100 gal (100 cm/s/s) were excluded to avoid influences of nonlinear behavior of medium to the observed waveforms. The time window for the direct S-wave record was 20.48 s, starting 2 s before the theoretical S-wave travel time as calculated from hypocentral information in the JMA unified earthquake catalogue and the JMA2001 travel time table (Ueno et al., 2002). After applying a 5% cosine taper to the resulting waveforms, we calculated Fourier spectra (with a 0.4 Hz Parzen window) for the time window from the three components of the accelerograms using fast Fourier transform. Note that spectra of the horizontal component are vector sums of the NS and EW components. We then calculated the path-corrected spectral ratios for each station pair (Eq. 2.8) with one reference station in each frequency. We selected a borehole sensor at IBRH19 (KiK-net Tsukuba, see Fig. 2.2) as the reference station, because the sensor on the borehole is located in the layer with the S-wave velocity of about 2.8 km/s. Site amplification at the reference station was set to be 1 for all frequency bands. We set up the simultaneous equation systems (Eq. 2.8) with the station pairs which shared at least one station with other pairs, then solved them by a least-squares method to derive relative site amplification factors relative to the reference station. We separately estimated the relative site amplification factors for vertical and horizontal components.

These site amplification factors were then modeled by using analog filters (Eq. 2.9). The number of filters, N_f and M_f , was restricted to values from 0 to 6, ruling out the case $N_f = M_f = 0$. We estimated the filter parameters, ω_{1n} , ω_{2n} , ω_{1m} , ω_{2m} , h_{1m} , h_{2m} , and G_0 by a least-squares method for each combination of N_f and M_f and selected the combination that had minimum residuals. Residuals were calculated for the frequency range from 0.05 to 20 Hz. Finally, the coefficients of the digital filters (Eq. 2.10) were calculated from the

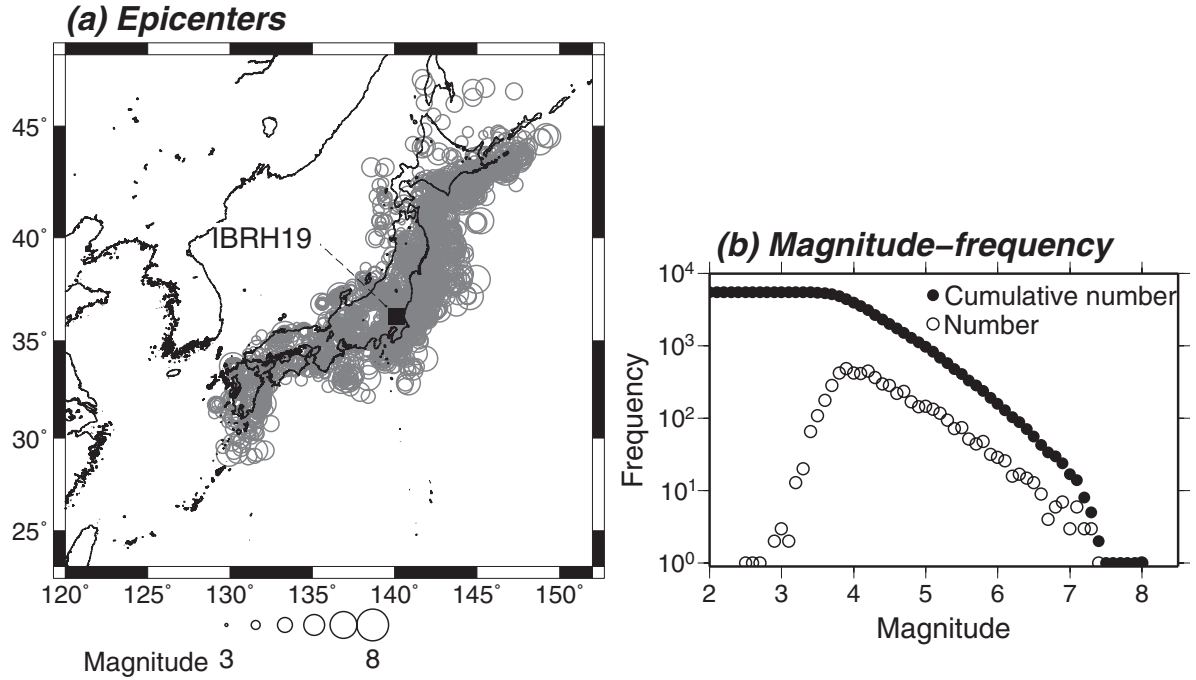


Figure 2.2 (a) Epicenter distribution and (b) Magnitude-frequency distribution of earthquakes used for site amplification estimation

parameters of the analog filters (Eq. 2.9).

Figure 2.3 shows the station distribution for which site amplification factors and corresponding site-correct filters successfully estimated and not estimated. Stations that we have estimated site amplification factors cover almost whole area of Japan, with the exception of northernmost Hokkaido, the northern part of Kyushu, and outlying islands due to the seismicity and long interstation distances.

2.6 Waveform prediction using site-correct filters

As an example of the predicted waveforms using site-correct filters, Fig. 2.4 shows the case for the large earthquake of 25 March 2007 (M6.9 in JMA scale), observed at stations FK1006 (K-NET Imajoh) and FK1H05 (KiK-net Tsuruga). The two stations are 21 km apart, and epicentral distance to FK1006 is about 166 km (Fig. 2.4a). The site amplification factors between these two stations, calculated from the site-correct filters and observed spectral

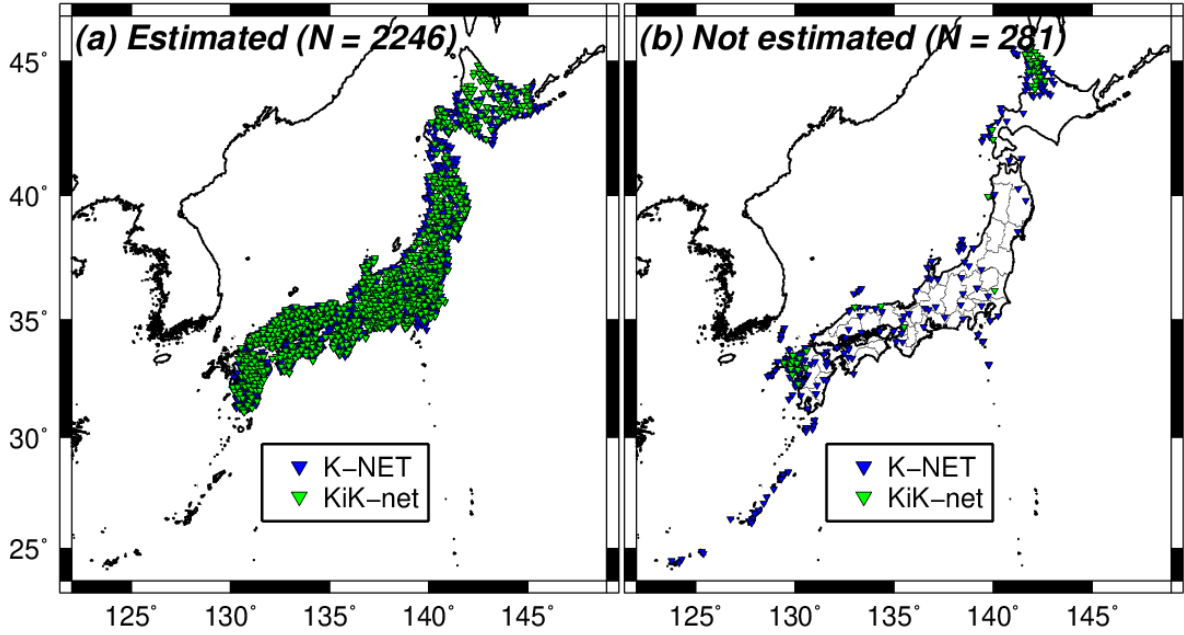


Figure 2.3 Distribution of stations for which site amplification factors and site-corrected filters have been (a) successfully estimated and (b) not estimated

ratios, are quite similar for both horizontal and vertical records (Fig. 2.4b). Note that the phase characteristics of the site-correct filters are determined by the condition of causality, so that phase characteristics between the two stations have not been included in this waveform calculation.

The observed waveforms of the 25 March 2007 earthquake at stations FKI006 and surface sensor of FKI05 are shown in Fig. 2.5a, b, respectively. Under the assumption that the source term is the same at the two stations, the waveforms reflect differences in the path and site terms. Figure 2.5c shows the waveforms at FKI006 after its site terms are replaced with those of surface sensor of FKI05, so that we can regard these waveforms as predicted waveforms of FKI05. The predicted waveforms are much more similar to the observed waveforms at FKI05, in both the time and frequency domains, than the observed waveforms at FKI006 even without correction for the differences in the path term.

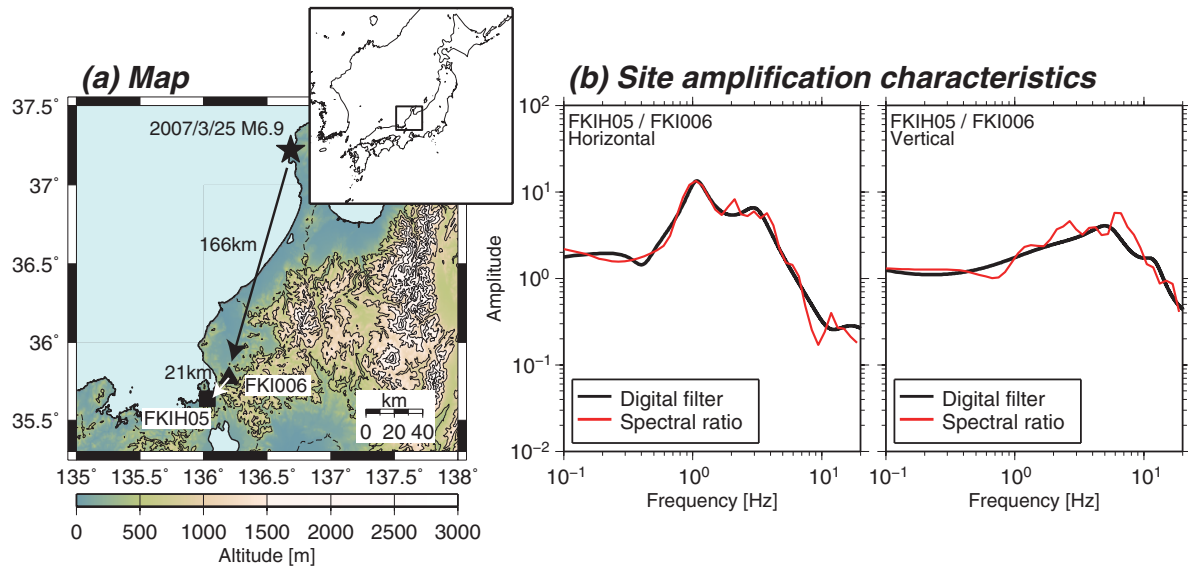


Figure 2.4 (a) Locations of the 25 March 2007 earthquake epicenter and two stations (FK1006: K-NET Imajoh and FK1H05: KiK-net Tsuruga). Contour interval of topography is 500 m. (b) Relative site amplification factors between FK1006 and surface sensor of FK1H05 by spectral ratio (red) and site-corrected filter methods (black).

2.7 Comparison of two intensity prediction methods

To compare the seismic intensity prediction results of the frequency-dependent and scalar correction methods, we selected station pairs that were separated by less than 10 km. We did not predict seismic intensity at sensors on the borehole of KiK-net stations in both methods. Although a practical application of this method is to predict seismic intensity at a target station using records from adjacent stations closer to the epicenter, i.e., front detection, we do not consider the status of front stations in this experiment. For the calculation of scalar correction value and the seismic intensity prediction experiments, we used almost the same data set as our site amplification estimation, except for the condition of maximum epicentral distance. An average difference between seismic intensity of two stations for scalar correction was calculated by at least seven earthquakes more than 100 km away from both stations. Comparison of two prediction methods was made with the earthquakes in the same

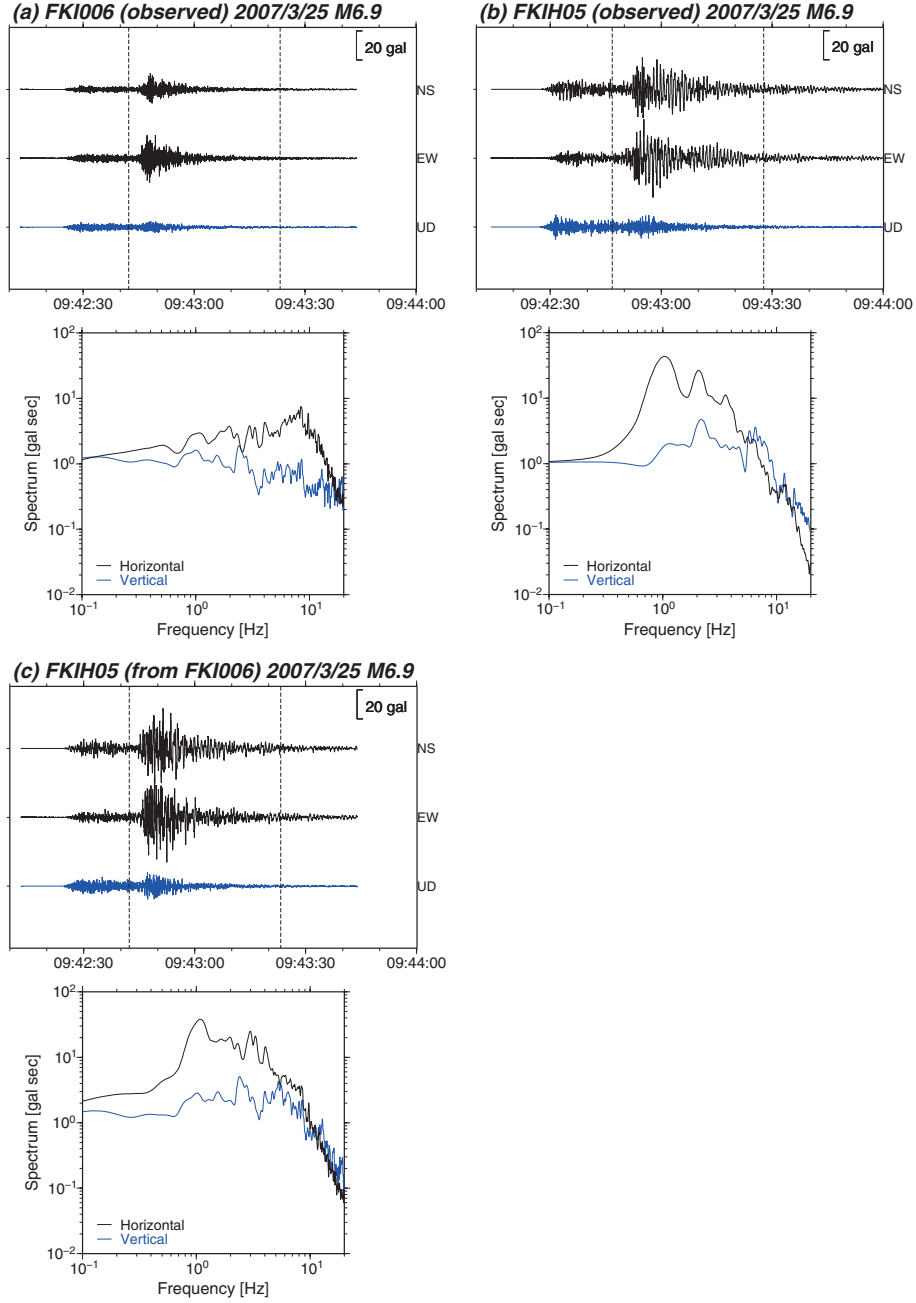


Figure 2.5 Waveforms and their spectra from (a) observations at station FK1006, (b) observation at the surface of station FK1H05, and (c) site-corrected waveforms at station FK1H05 based on observations at station FK1006. Dotted lines represent the 40.96 s time window used for spectrum calculation.

condition as the calculation of scalar correction values. Epicenters and magnitude-frequency distribution used for the seismic intensity prediction experiments are shown in Fig. 2.6. Number of earthquakes is 5972.

Figure 2.7 shows a map and relative site amplification factors for a surface station of TKCH01 (KiK-net Rikubetsu) and HKD088 (K-NET Rikubetsu), 5.3 km away. The site amplification characteristics calculated by spectral ratios show a peak (or trough) around 0.6 Hz in the horizontal component and around 1 Hz in the vertical component that has been replicated by the site-correct filters. (Fig. 2.7b).

Figure 2.8 shows the distribution of residuals (observed minus predicted value) and histograms of residuals of the seismic intensity predicted by both methods for each earthquake recorded by surface sensor of station TKCH01 and HKD088. The mean of residuals by scalar correction equals to zero, because we have used the same waveform data set in both calculating average seismic intensity differences and predicting seismic intensity with scalar correction. The histograms of residuals and root mean square (RMS) of residuals (right panels of Fig. 2.8) clearly show that the frequency-dependent correction prediction resulted in smaller residuals than the scalar correction prediction for this station pair. As an interesting feature, azimuthal dependency of prediction residuals has been found in both prediction methods. Here we only point out two possibilities for this feature, and we do not discuss the causes in detail: one is the azimuthal dependency of site amplification and the other is the differences in source characteristics. Azimuthal dependency of site amplification would affect both methods, while source characteristics would affect scalar correction method. Future work is required to investigate the azimuthal dependency of prediction residuals in both methods.

When we compare histograms of RMS of residuals for all 2417 station pairs separated by less than 10 km (Fig. 2.9), the mean RMS from the frequency-dependent correction method is about 27 % smaller than that from the scalar correction method. Figure 2.10, showing the RMS histograms in three dimensions, indicates that the frequency-dependent correction

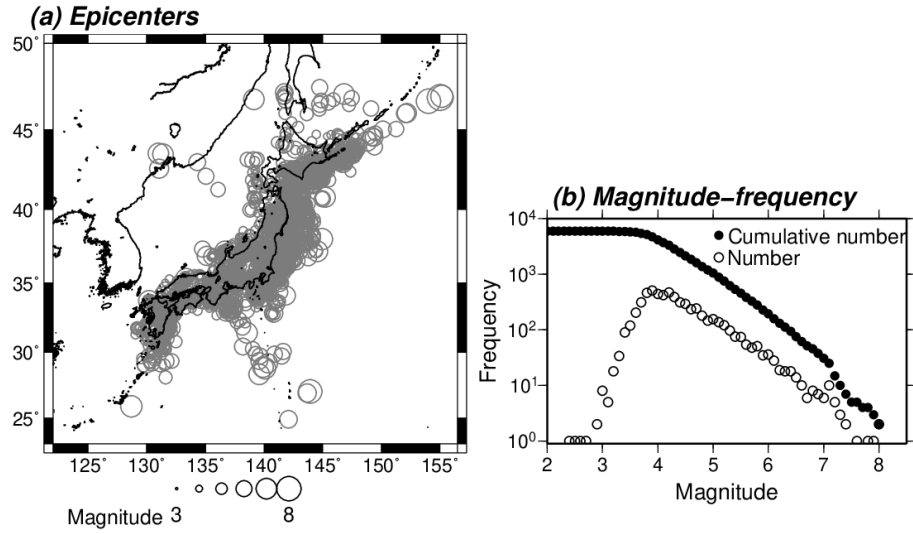


Figure 2.6 (a) Epicenters and (b) magnitude-frequency relation (open circle for number and black for cumulative number) of earthquakes used for seismic intensity prediction experiment.

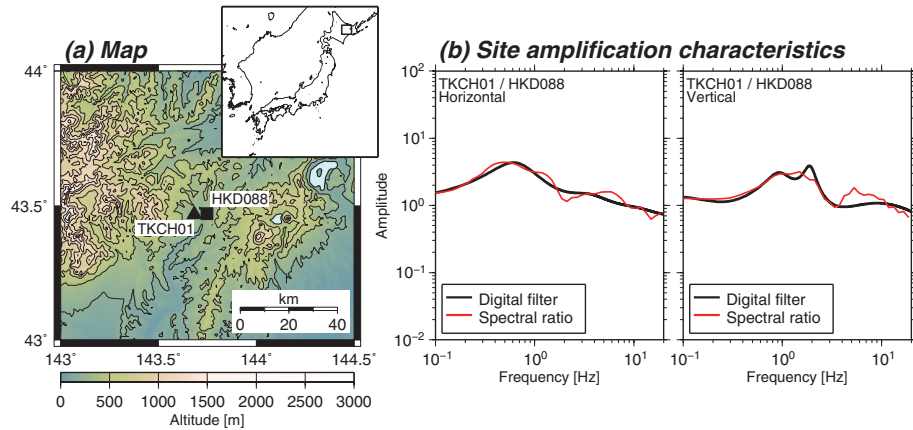
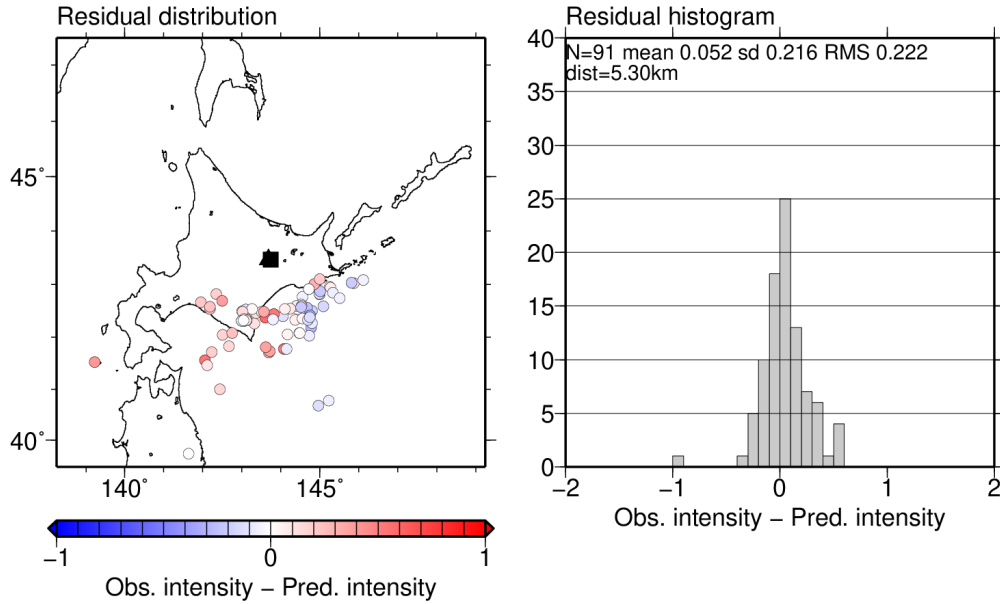


Figure 2.7 (a) Map of stations TKCH01 (KiK-net Rikubetsu) and HKD088 (K-NET Rikubetsu). Contour interval of topography is 200 m. (b) Relative site amplification factors between surface sensor of TKCH01 and HKD088 calculated by spectral ratio (red) and site-correct filters (black).

sometimes resulted in larger residuals than the scalar correction. Nevertheless, prediction using frequency-dependent correction has improved the RMS of prediction residuals in many cases.

(a) Frequency correction prediction from TKCH01[▲] to HKD088[■]



(b) Scalar correction prediction from TKCH01[▲] to HKD088[■]

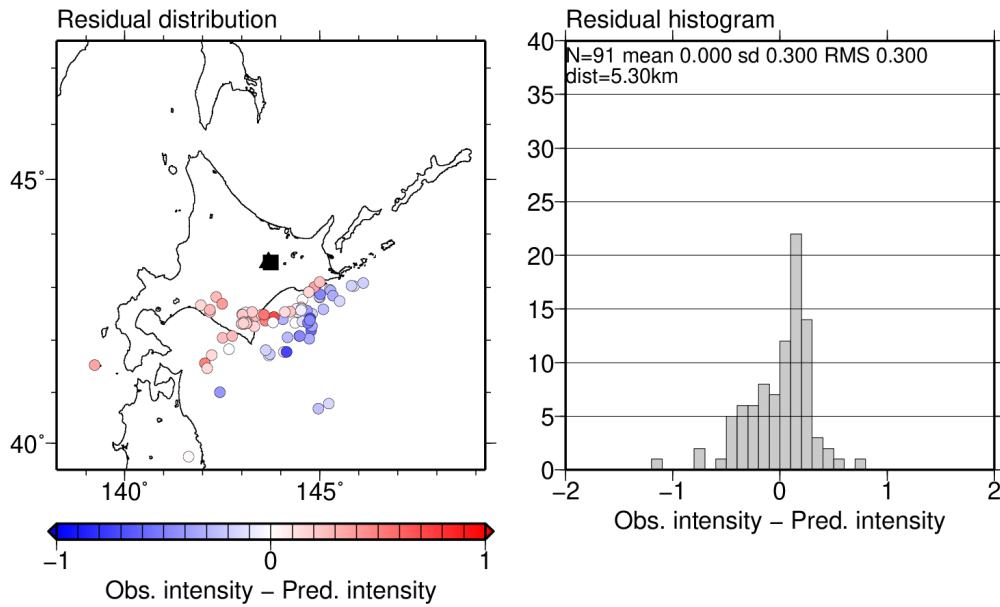


Figure 2.8 Predicted intensity residuals of earthquakes (left panel) and histograms of residuals (right panel) for station pair TKCH01 (surface) and HKD088 after (a) frequency-dependent correction and (b) scalar correction prediction.

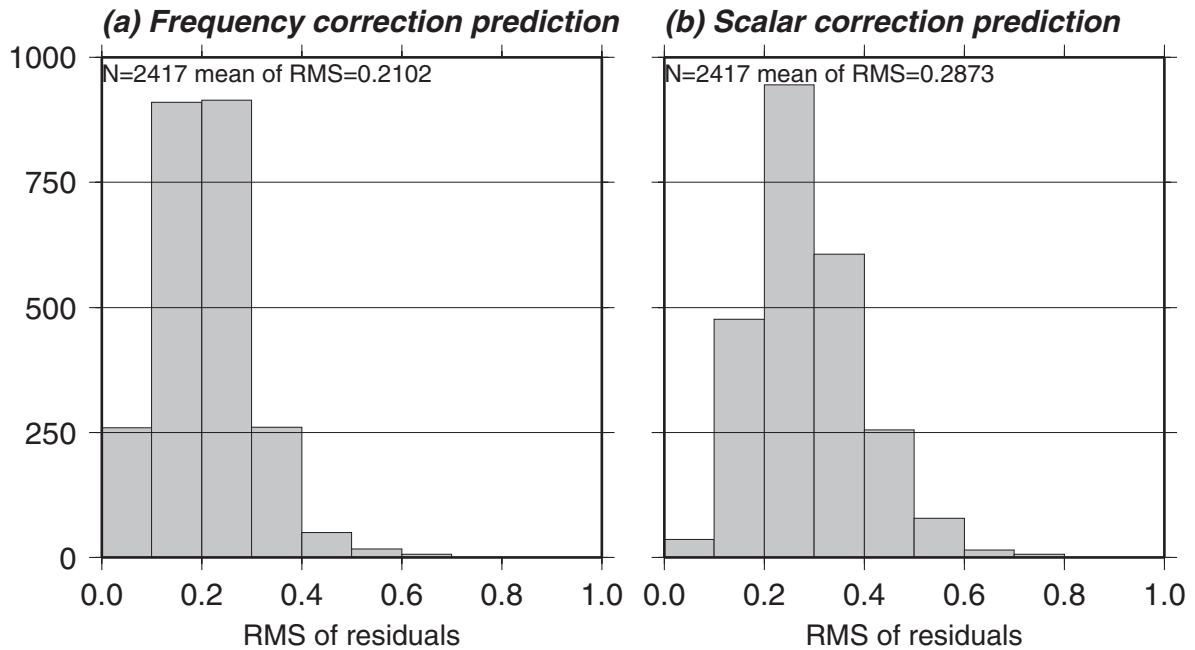


Figure 2.9 Histograms of RMS of residuals for all station pairs separated by less than 10 km after (a) frequency-dependent correction and (b) scalar correction prediction.

2.8 Discussion

We have shown that the frequency-dependent site amplification correction leads to the accurate seismic intensity prediction in our prediction experiment. In the viewpoint of long lead time for EEW, the interstation distance of 10 km seems to be too short, and we should use station pair with longer interstation distance. At interstation distances larger than the value of 10 km we have used, the source and path terms would surely have a greater influence on intensity predictions. We therefore evaluate the cases where interstation distances shorter than 20 km and shorter than 30 km. Histograms of the resulting RMS of residuals are shown in Fig. 2.11. The mean RMS of residuals of both prediction methods becomes large as the interstation distance increases, representing that the differences between seismic intensity of the two stations cannot be explained by the difference in site amplification term only, i.e., the effects of path and source term differences between two stations increase as the interstation distance

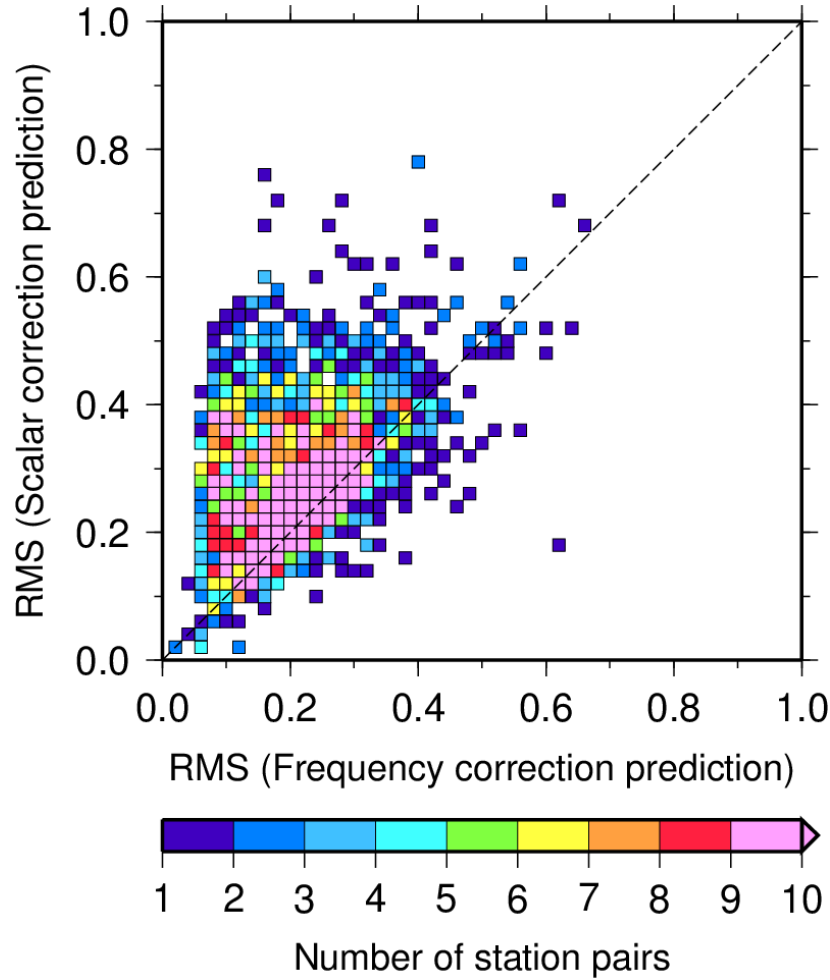


Figure 2.10 Three-dimensional histogram of RMS residuals for all station pairs by less than 10 km after frequency-dependent and scalar correction prediction. The number of station pairs was counted in bins with sides measuring 0.02.

increases. This fact suggests that the effects of path and source terms are important for practical use of our seismic intensity prediction concept with long lead time where we should use station pair with long interstation distances. The advantages of frequency-dependent correction become smaller as the interstation distance elongates, implying that the importance of frequency-dependent site amplification may be masked by the effects of path and source terms.

Hoshiba & Aoki (2015) have proposed the Numerical Shake Prediction scheme for EEW system. In their procedure, first current wavefield is estimated using data assimilation and

then ground motion is predicted by solving a wave equation in real time with the current wavefield as the initial condition. Same as our prediction concept, Hoshiba & Aoki (2015) aims to predict ground motion without any hypocentral parameters. Our prediction method assumes that the difference between seismic intensity of two stations is mainly caused by the difference in site amplification; however, the Numerical Shake Prediction scheme is based on the physics of wave propagation, so that effects of path term are included in the method. In addition, because the effect of seismic source should affect the amplitude distribution of current wavefield so that estimation of current wavefield incorporates the effect of source term. As a result, with the Numerical Shake Prediction scheme, the accuracy of seismic intensity prediction will be improved even if the interstation distance becomes longer or epicentral distance becomes shorter. Site-correct filters estimated in this study are indispensable in the scheme of the Numerical Shake Prediction, as shown by Hoshiba & Aoki (2015).

The improvement offered by frequency-dependent correction has not been found in all cases (see Fig. 2.10). An example is the case of the station pair YMTH06 (surface sensor of KiK-net Takahata, target station) and YMT015 (K-NET Yonezawa, observation station), in which the site-correct filters closely approximate the observed spectral ratios (Fig. 2.12). However, although the frequency-dependent correction has yielded a lower standard deviation of residuals than the scalar correction prediction, the mean of the residuals have deviated from zero, leading to a large RMS value (Fig. 2.13). One of the possible reasons of this result is an azimuthal dependency of site amplification. Earthquakes in the southeastern region from both stations (left panel of Fig. 2.13a) have large residuals, causing the large RMS value. Figure 2.14 shows the observed spectral ratios of each earthquake between YMT015 and YMTH06 drawn in different colors corresponding to the backazimuth. Site-correct filters (Fig. 2.12b) allow us to reproduce the general characteristics of spectral ratios well (Fig. 2.14); however, if we check the observed spectral ratios carefully, there may be a weak azimuthal dependency around 1 Hz in horizontal component in observed spectral ratios. We have not taken azimuthal

dependency of site amplification factor into consideration in designing site-correct filters in this study. If we design multiple site-correct filters regarding the azimuthal dependency of site amplification, we would be able to predict seismic intensity more precisely. Other reason is the influence of smoothing the site amplification. Although observed spectral ratios of each earthquake have many peaks and troughs, site-corrected filters do not represent such features (Fig. 2.14). The JMA seismic intensity is calculated from the peak value of the observed waveform envelope (Japan Meteorological Agency, 1996; Hoshihara & Ozaki, 2014). Site amplification correction with smoothed frequency characteristics (Figs. 2.12b and 2.14) might not reproduce the peak value of the envelope in the target station. More work in the future will be needed to reveal the relationship between the peak value of seismic waveforms and the frequency characteristics of both amplitudes and phases.

We have shown the importance of frequency-dependent site amplification term for ground motion (seismic intensity in this study) prediction. Because of dense seismic networks and highly active seismicity in and around Japan, we have been able to estimate site amplification factor with simple assumptions in source and path terms. In other area with relatively sparse seismic networks and low seismicity, it would be possible to estimate reliable site amplification factor if we adopt other noble method (e.g., coda normalization method: Phillips & Aki, 1986 and Takemoto et al., 2012 or generalized spectrum inversion: Kawase & Matsuo, 2004 and Nakano et al., 2015b) and careful treatment of source term.

2.9 Conclusion

We have conducted experiments predicting seismic intensity at a station based on the data from an adjacent station, comparing two different methods, frequency-dependent correction and scalar correction, to correct relative site amplification factors. To estimate frequency-dependent relative site amplification factors, we calculated spectral ratios of direct S-waves for many station pairs and solved simultaneous equation system to acquire site amplification

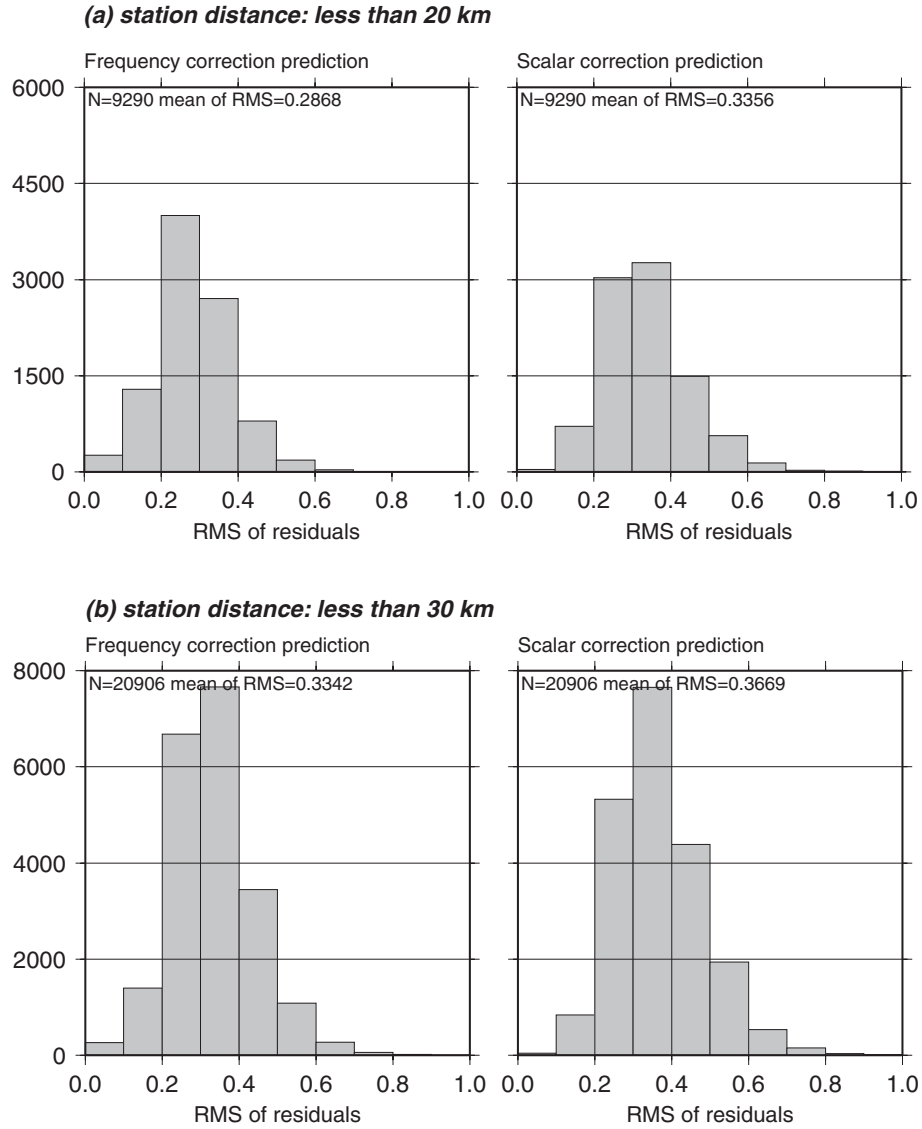


Figure 2.11 Histograms of RMS of residuals for all station pairs separated by (a) less than 20 km and (b) less than 30 km. Left side for frequency-dependent correction and right side for scalar correction prediction, respectively.

factors relative to a reference station. We then designed site-correct digital filters that approximate these site amplification characteristics of each station. Also, we calculated average differences between seismic intensity of two stations for scalar correction. In all but a few cases, seismic intensity prediction with frequency-dependent correction using site-correct filters resulted in smaller residuals than prediction with scalar correction, showing the impor-

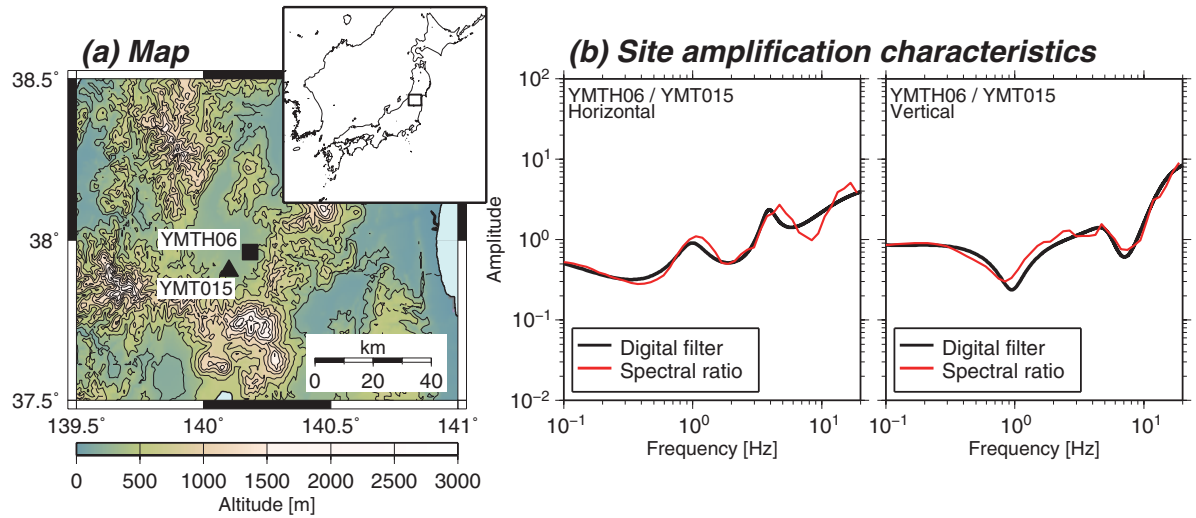
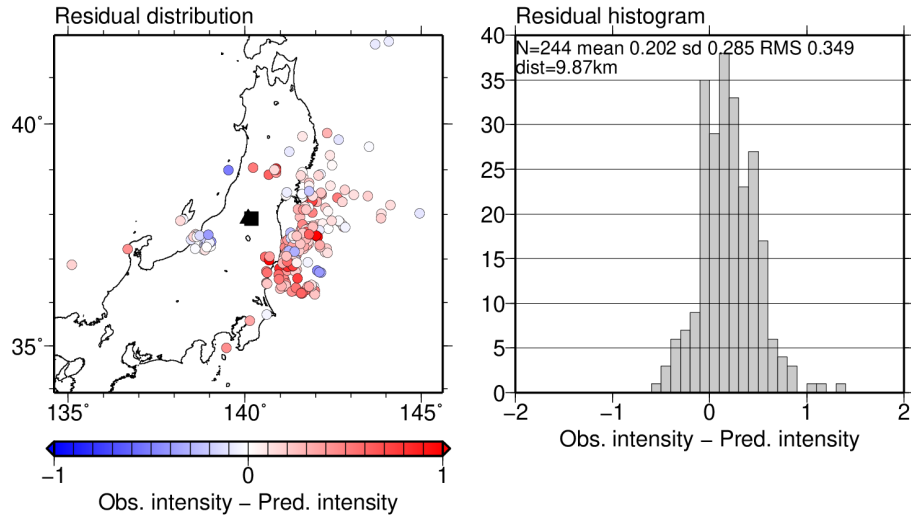


Figure 2.12 (a) Map of stations YMT015 (K-NET Yonezawa) and YMTH06 (KiK-net Takahata). Counter interval of topography is 200 m. (b) Relative site amplification factors between YMT015 and surface sensor of YMTH06 calculated by spectral ratio (red) and site-corrected filters (black).

tance of considering the frequency characteristics of site amplification in seismic intensity predictions. As interstation distance becomes longer, the differences between stations in the path and source terms increase until the site amplification correction offers only a marginal advantage for seismic intensity prediction.

(a) Frequency correction prediction from YMT015[▲] to YMTH06[■]



(b) Scalar correction prediction from YMT015[▲] to YMTH06[■]

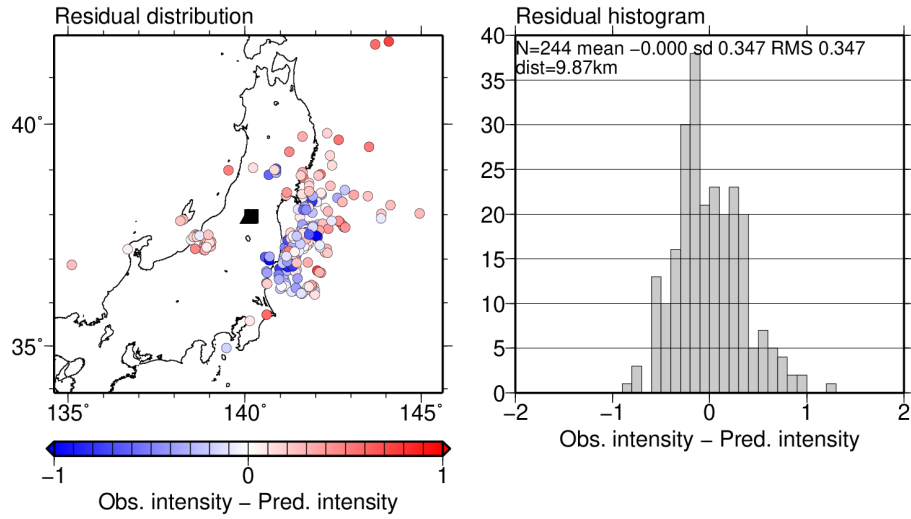


Figure 2.13 Predicted intensity residuals of earthquakes (left panel) and histograms of residuals (right panel) for station pair YMT015 and surface sensor of YMTH06 after (a) frequency correction and (b) scalar correction prediction.

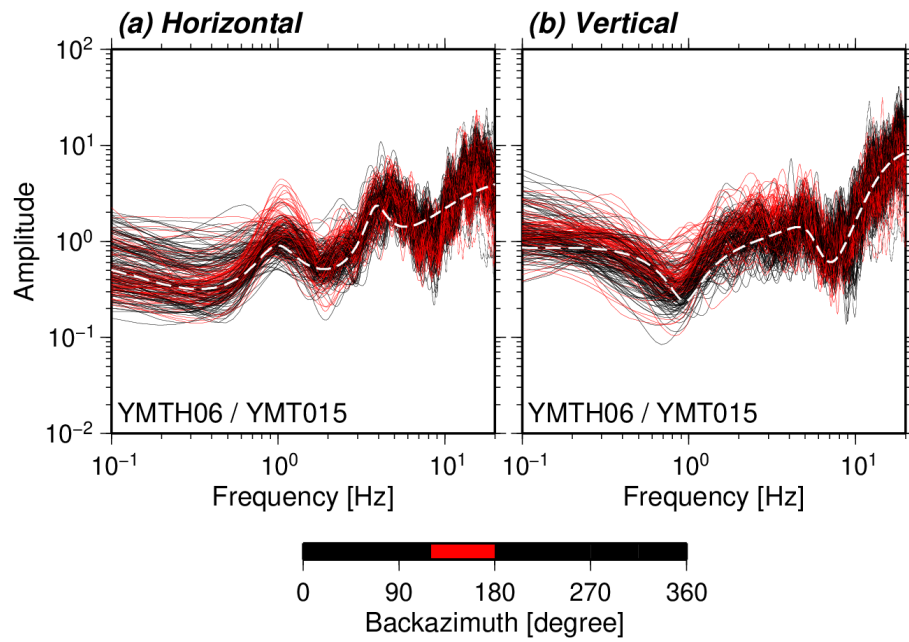


Figure 2.14 (a) Horizontal and (b) vertical component of observed spectral ratios of each earthquake between YMT015 and surface sensor of YMTH06. Earthquakes with the backazimuth angle of 120 to 180 degree are plotted in red, while those with the other angle are in black. Frequency characteristics of site-correct filters are plotted in white broken lines.

3 Estimation of heterogeneous attenuation structure

Section 3.2 has been published as a part of following paper:

Ogiso, M., M. Hoshihara, A. Shito, and S. Matsumoto, Numerical Shake Prediction for Earthquake Early Warning Incorporating Heterogeneous Attenuation structure: The Case of the 2016 Kumamoto Earthquake, *Bulletin of the Seismological Society of America*, **108** (6), 3457-3468, 2018, ©Seismological Society of America.

3.1 Introduction

Seismic attenuation is one of the basic seismic parameters of the medium in the Earth, as with seismic velocities of P- and S-waves. Attenuation parameters characterize the intrinsic properties of the medium. Recent dense seismic networks have enabled us to estimate detailed 3D attenuation structures in a variety of regions. Discussion about the relationship between attenuation structure and seismotectonics or volcanic activity is one of the important topics in seismology at present (e.g., Tomatsu et al., 2001; Nakajima et al., 2013; Komatsu & Oda, 2015; Kita et al., 2014). Detailed information of the 3D attenuation structure is also important for ground motion prediction. Nakamura & Uetake (2012) derived a frequency-dependent 3D attenuation structure of and around Japan, showing that the heterogeneous attenuation structure resulted in a better prediction of peak ground velocities in various frequency bands than a homogeneous structure.

The influence of the seismic scattering during seismic wave propagation depends on the wavenumber of seismic waves, characteristic scale length of the heterogeneity in the medium, and propagation distance (e.g., Fig. 13.11 of Aki & Richards, 1980). When the propagation distance is the order of 10^0 to 10^2 km, not only intrinsic attenuation but also seismic scattering plays an important role to alter the seismic amplitude at any stations in the high frequency range (roughly ≥ 1 Hz). Intrinsic attenuation, that is, energy absorption in the medium,

always reduces the amplitude of seismic waves while seismic scattering works on not only the decrease of amplitude but also the elongation of the duration of ground motion. Hence, the separate estimation of intrinsic absorption and seismic scattering should be important for precise ground motion prediction including both the maximum amplitude and the duration of ground shaking as well as understanding the structure of the Earth.

Recent developments of modeling seismogram envelopes, as summarized in Sato et al. (2012), have enabled us to estimate intrinsic and scattering attenuation parameters simultaneously from the observed seismogram envelopes. For example, Wu & Aki (1988) estimated intrinsic and scattering attenuations at the Hindu Kush region, based on the radiative transfer theory with the isotropic scattering model (Wu, 1985). They compared the travel distance dependency of seismic energy integrals between data and model predictions, using only one finite time window for each energy integration. Later, Fehler et al. (1992) proposed the multiple time window analysis (hereafter MLTWA) where they extended the approach of Wu & Aki (1988) to use three time windows in investigating the distance dependency of energy integrals. The numerical basis of Fehler et al. (1992) was provided by Hoshiba et al. (1991). Instead of using multiple stations in Fehler et al. (1992), Hoshiba (1993) used a single station to analyze distance dependency of energy integrals. Since Fehler et al. (1992) and Hoshiba (1993), the MLTWA has been widely used for a separate estimation of intrinsic and scattering attenuation parameters around the world, summarized in the Fig. 8.13 of Sato et al. (2012). Carcolé & Sato (2010) estimated intrinsic and scattering parameters for a given Hi-net stations (Okada et al., 2004) using the MLTWA, and then interpolated these parameters between stations to derive a detailed 2D distribution of intrinsic and scattering attenuation parameters in Japan. Eulenfeld & Wegler (2016) developed the *Qopen* package in which they estimated source amplitude, site amplification factors, intrinsic attenuation, and scattering coefficient simultaneously by the envelope fitting. Eulenfeld & Wegler (2017) applied *Qopen* to USArray network data to derive a 2D distribution of intrinsic attenuation and scattering coefficients in the United

States. Prudencio et al. (2013) and Del Pezzo et al. (2016) proposed a different approach with a space-weighting function to estimate a 2D distribution of attenuation parameters. The above previous studies attempted to estimate intrinsic and scattering attenuation parameters simultaneously in 2D space.

Nishigami (1991) formulated an inversion scheme of 3D distribution of relative scattering coefficients under the assumption of the single scattering. The scheme of Nishigami (1991) has been applied to 3D heterogeneities of scattering coefficients at San Andreas fault (Nishigami, 2000), Campi Flegrei caldera (Tramelli et al., 2006), and Mt. St. Helens (De Siena et al., 2014). Asano & Hasegawa (2004) improved the scheme of Nishigami (1991) to estimate 3D distribution of absolute scattering coefficients. Taira & Yomogida (2007) proposed a method to detect localized strong scattering regions in 3D space using seismic array observations, and Taira et al. (2007) applied the method to reveal the frequency-dependent 3D scattering structure around the Nagamachi-Rifu fault, near Sendai, Japan. These previous studies aimed at estimating 3D distribution of scattering coefficients were based on the single scattering model and intrinsic attenuation parameter was assumed.

Takahashi et al. (2009) estimated a 3D distribution of random velocity fluctuations in northeastern Japan from the analysis of peak delay times of seismogram envelopes. They used the Markov approximation (e.g., chapter 9 of Sato et al., 2012) to model seismogram envelopes, that is, a multiple scattering model was adopted. Takahashi (2012) estimated a 3D heterogeneous distribution of intrinsic attenuations with the correction of scattering attenuations. These pioneering studies of Takahashi and his colleagues focused on the seismogram envelopes of the direct S-wave and did not utilize scattered waves following the direct S-waves.

In this chapter, we estimate heterogeneous distributions of intrinsic and scattering attenuation parameters in southwestern Japan by two methods. First, we estimate 2D distribution of intrinsic attenuation and scattering coefficients by the MLTWA. Second, we propose the new

method of estimating 3D distribution of the intrinsic attenuation and the strength of random velocity fluctuation simultaneously from the whole of seismogram envelopes, then apply the proposed method to southwestern Japan. Finally, we discuss the features of the estimated heterogeneous attenuation structures.

3.2 2D heterogeneous attenuation structure of southwestern Japan estimated by the Multiple Lapse Time Window Analysis

Seismic scattering distorts the waveform shape of direct waves and excites wave trains following the direct waves (e.g., Aki, 1969). The wave trains excited by the seismic scattering have been widely called seismic coda. The MLTWA utilizes the difference between the contribution to the temporal decay character of seismic energy of intrinsic and scattering attenuations (Hoshihara et al., 1991). The coda energy immediately after the direct waves (in most cases, direct S-waves) depends on the total attenuation defined by the sum of intrinsic and scattering attenuations. The contribution of seismic scattering to the coda energy gradually increases as time elapses. By investigating the travel distance dependency of the coda energy in multiple time windows, the separation of intrinsic and scattering attenuation parameters turns to be possible (Hoshihara et al., 1991).

In this section, we estimate a 2D distribution of intrinsic attenuation and scattering coefficient of southwestern Japan by the MLTWA under the assumption of the isotropic scattering model, following the procedure of Carcolé & Sato (2010). We used 293 stations and 3117 earthquakes in this analysis. Figure 3.1 shows the distribution of stations and epicenters used in the MLTWA. The hypocentral depths of earthquakes are less than 30 km. First, we collected seismic waveforms of the events at the stations shown in Fig. 3.1 where their hypocentral distances are less than 50 km for the Kyushu and 100 km for the other regions. Because we will later use the estimated attenuation structure in this section for the JMA

seismic intensity prediction in chapter 4, we focus on the heterogeneous attenuation structure in a frequency range of 1-2 Hz. After applying a 1-2 Hz band-pass filter to each waveform, we calculated the squared sum of the three-component velocity seismograms to derive a time series of its energy density in the time window of 45 s starting 1 s before the direct S-wave arrival. Hence, the estimated intrinsic attenuation and scattering coefficient were for S waves. We divided this time window into three sub-windows of 15 s (Fig. 3.2a), then integrated the energy density in each sub-window. Although we adopted the time window of 15 s for the energy integration, a shorter time window would be acceptable in the MLTWA (e.g., Hoshiba, 1993; Yoshimoto & Okada, 2009). Geometrical spreading effect can be compensated by multiplying these integrated energy densities by its hypocentral distance. To eliminate the effects of the source and site amplification terms, the coda normalization technique was applied to each time series of energy density. We adopted a window of 15 s starting 57.5 s after the origin time as the coda normalization window. This time window for the coda normalization is suitable for observations with hypocentral distance less than 100 km (Hoshiba et al., 1991). We calculated the integral of energy density in the coda normalization window, and divided the energy densities of the above three 15 s sub-windows by those of the coda normalization window. Then, we made a scatter diagram of the coda-normalized energy densities of the three sub-windows versus hypocentral distances for each station (Fig. 3.2c). The distance dependency of the coda-normalized energy densities in these three sub-windows can be modeled using an analytical equation derived by Paasschens (1997). The time series of energy density $E(r, t)$ is represented by:

$$E(r, t) = W_s \exp\left(-\frac{2\pi ft}{Q_i}\right) \times \left[\frac{\exp(-g_0 V_0 t)}{4\pi V_0 r^2} \delta\left(t - \frac{r}{V_0}\right) + \frac{\{1 - r^2/(V_0 t)^2\}^{1/8}}{\{4\pi V_0 t/(3g_0)\}^{3/2}} \exp(-g_0 V_0 t) G\left(g_0 V_0 t \left(1 - \frac{r^2}{V_0^2 t^2}\right)^{3/4}\right) \Theta\left(t - \frac{r}{V_0}\right) \right], \quad (3.1)$$

where $E(r, t)$ is the time trace of energy density at the hypocentral distance r on the time t measured from its earthquake origin time, f the frequency, W_s the magnitude of source energy, V_0 the medium velocity, Θ the step function, Q_i the intrinsic attenuation factor, g_0 the scattering coefficient which relates to the probability of scattering. The function G is defined as

$$G(x) = 8 \cdot (3x)^{-3/2} \sum_{k=1}^{\infty} \left\{ \frac{\Gamma(3/4k + 3/2)}{\Gamma(3/4k)} \cdot \frac{x^k}{k!} \right\}, \quad (3.2)$$

in which Γ denotes the gamma function. V_0 in Eq. (3.1) can be calculated from the hypocentral distance and the arrival times of direct S-waves. With given values of Q_i^{-1} and g_0 , we calculated a time series of energy density of a given source-station pair using Eqs. (3.1) and (3.2) and energy density distributions in three time windows for comparison with the observed data (Fig. 3.2b, c). We conducted a grid search ranging from 0.001 to 0.1 with an increment of 0.0001 for both Q_i^{-1} and g_0 to find their optimal values at each station, then derived a 2D distribution by interpolation. The grid interval of this interpolation was 0.02° in both longitude and latitude.

Figure 3.3 shows the interpolated distribution of Q_i^{-1} and g_0 , together with the estimated values at each station. Both intrinsic and scattering attenuations are strongly heterogeneous in the central and southern part of Kyushu, yet weakly heterogeneous in Chugoku and Shikoku.

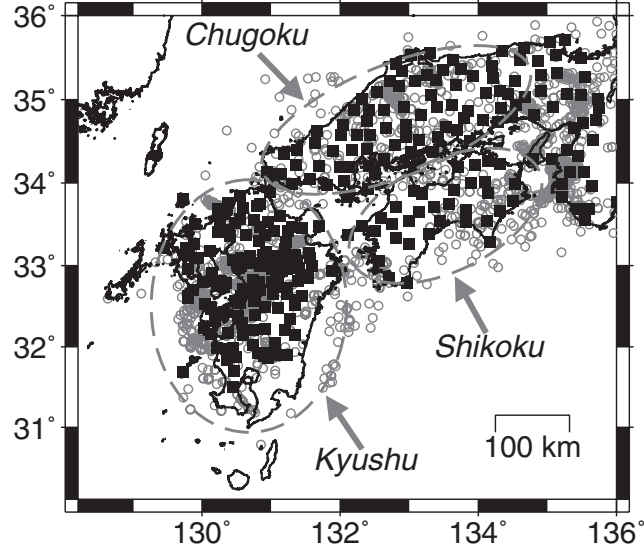


Figure 3.1 Distribution of stations (black squares) and earthquakes (gray circles) for the MLTWA. The hypocentral depths are less than 30 km. The Chugoku, Shikoku, and Kyushu regions are outlined by gray dashed ovals.

3.3 Estimating 3D distribution of intrinsic and scattering attenuations in southwestern Japan

3.3.1 Scattering regime

Here we briefly review the scattering regime for our 3D attenuation structure estimation, following Sato et al. (2012). If a medium has random velocity fluctuations, the velocity at a certain location $\mathbf{x}$, $V(\mathbf{x})$, may be represented by

$$V(\mathbf{x}) = V_0\{1 + \xi(\mathbf{x})\}, \quad (3.3)$$

where V_0 is the mean velocity of the medium and $\xi(\mathbf{x})$ represents the random fluctuations in the velocity at the location $\mathbf{x}$. We assume that $\langle V(\mathbf{x}) \rangle = V_0$ and $\langle \xi(\mathbf{x}) \rangle = 0$ where $\langle \cdots \rangle$ denotes the ensemble average. Autocorrelation of velocity fluctuations at spatial lag $\mathbf{y}$, $R(\mathbf{y}, \mathbf{x})$, is represented by

$$R(\mathbf{y}, \mathbf{x}) = \langle \xi(\mathbf{x}) \xi(\mathbf{x} + \mathbf{y}) \rangle. \quad (3.4)$$

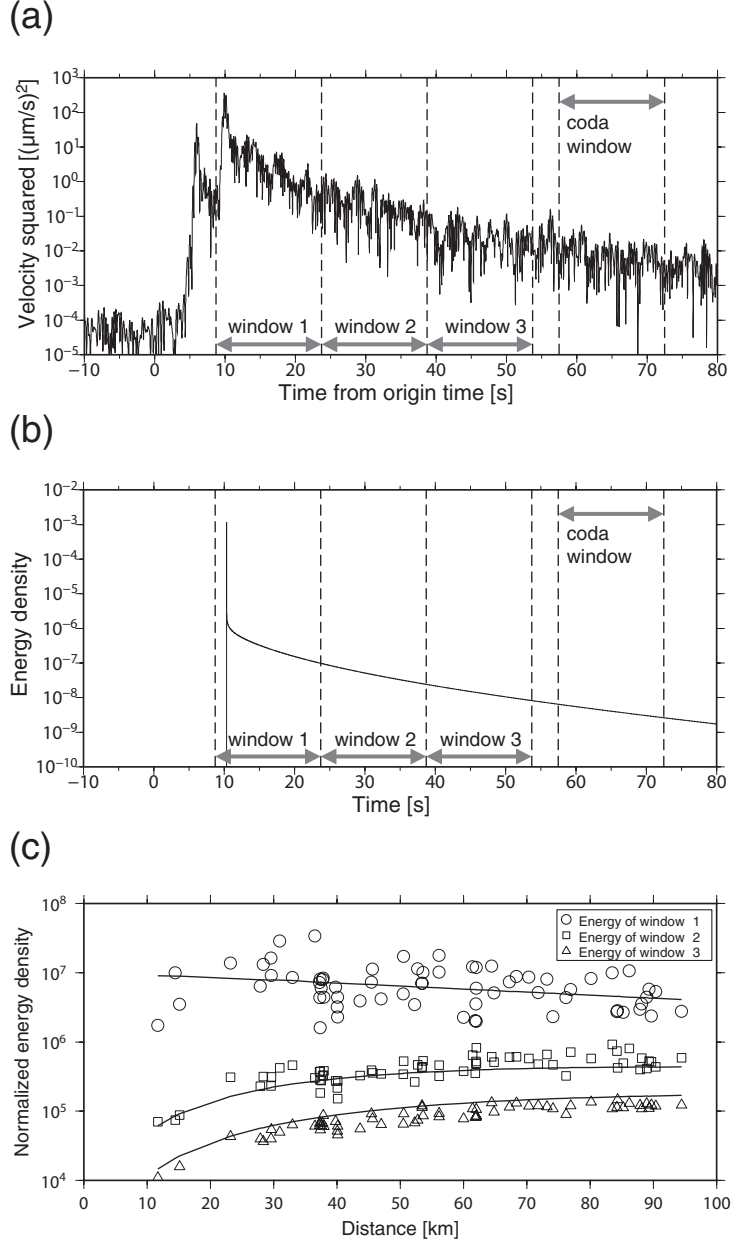


Figure 3.2 Example of MLTWA at station N.DWAH (location indicated in Fig. 3.3). (a) Observed envelope for an earthquake on 31 July, 2011, calculated after band-pass filtering. (b) Calculated energy density envelope using equations of Paasschens (1997). (c) Comparison of coda-normalized energy from calculations (curves) and observations (symbols) for time windows 1 (circles), 2 (squares), and 3 (triangles) plotted against hypocentral distance. Curves are the calculated energy distribution in each window using best-fit values of Q_i and g_0 .

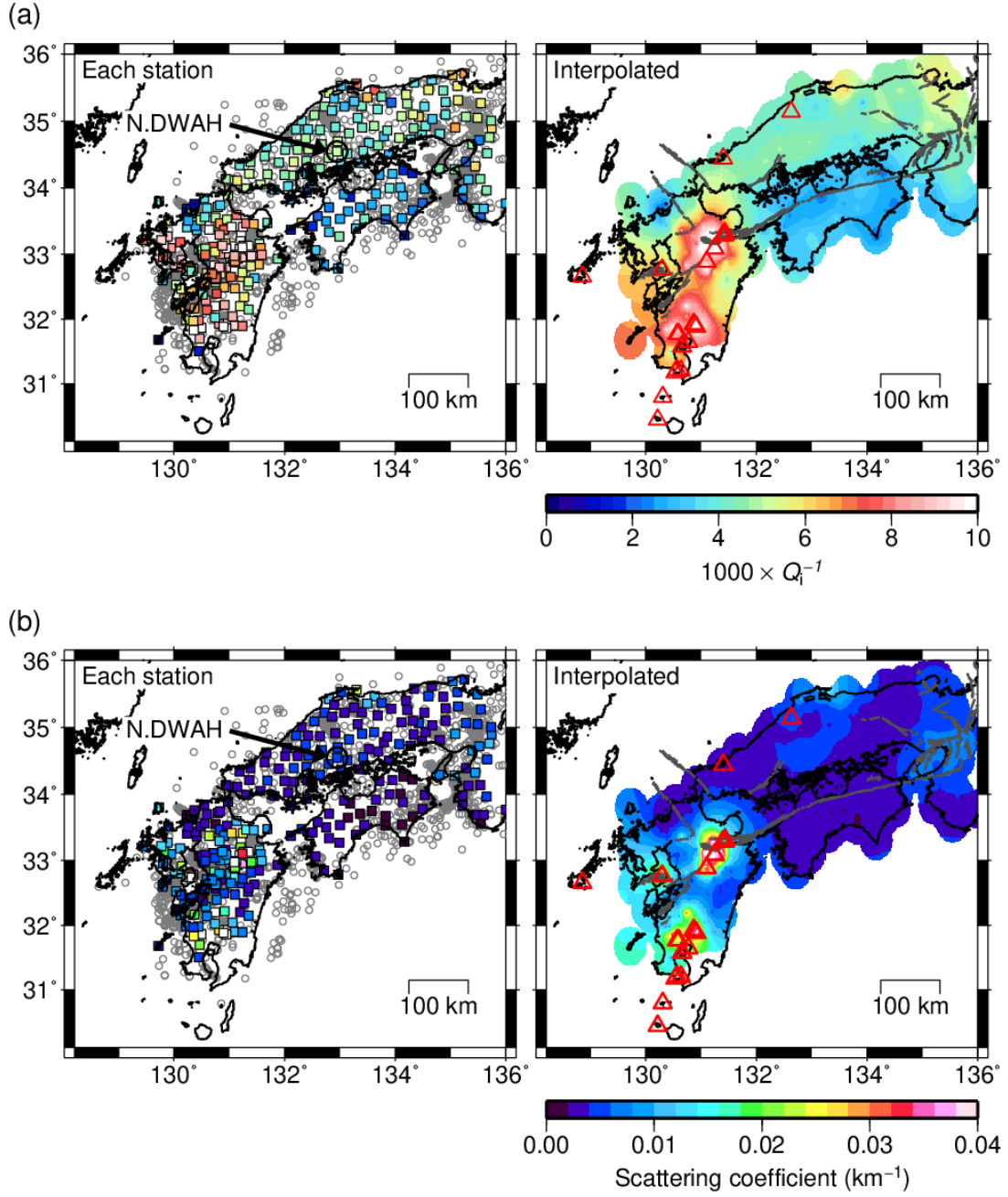


Figure 3.3 Distributions of (a) inverse of intrinsic absorption Q_i^{-1} and (b) scattering coefficient g_0 derived by MLTWA. Left panels show values at each station and right panels show interpolated values. Gray circles in the left panels are epicenters of events used for MLTWA. Gray lines in the right panels represent active faults, and red triangles represent active volcanoes.

Assuming the velocity fluctuation is isotropic and stationary, $R(\mathbf{y}, \mathbf{x})$ does not depend the location $\mathbf{x}$, therefore it becomes a function only of lag distance $r \equiv |\mathbf{y}|$. The strength of fractional fluctuations of the velocity is given by mean square of fractional fluctuations ε^2 , which is defined as

$$\varepsilon^2 \equiv R(0). \quad (3.5)$$

In an exponential-type random medium, $R(r)$ is represented by

$$R(r) = \varepsilon^2 \exp\left(-\frac{r}{a_r}\right), \quad (3.6)$$

and its power spectrum density $P(m)$ is represented by

$$P(m) = \frac{8\pi\varepsilon^2 a_r^3}{(1 + a_r^2 m^2)^2}, \quad (3.7)$$

where a_r is the representative correlation distance and m is the wavenumber.

If the condition $\varepsilon^2 a_r^2 k_0^2 \ll 1$ is satisfied for an incident wave with wavenumber k_0 , the Born approximation is applicable. In this case, the non-isotropic (Born) scattering coefficient $g(\psi; k_0)$ for the scalar waves can be calculated as

$$g(\psi; k_0) = \frac{k_0^4}{\pi} P\left(2k_0 \sin \frac{\psi}{2}\right), \quad (3.8)$$

where ψ is the direction angle of the scattered wave measured from the direction of incident wave (called the scattering angle, see Fig. 4.1 of Sato et al., 2012). If g does not depend on the scattering angle ψ , the scattering is called isotropic. Otherwise, g varies according to the scattering angle ψ , which is called non-isotropic scattering.

Energy propagation of scalar waves in a 3D medium, including scattered waves, can be described by the radiative transfer theory (Przybilla et al., 2006; Wegler et al., 2006; Sato

et al., 2012). The radiative transfer equation in this case becomes:

$$\begin{aligned} \frac{\partial}{\partial t} f(\mathbf{x}, t; \mathbf{q}) + V_0 \mathbf{q} \frac{\partial}{\partial \mathbf{x}} f(\mathbf{x}, t; \mathbf{q}) = \\ - \left\{ g_0(k_0) + k_0 Q_i^{-1} \right\} V_0 f(\mathbf{x}, t; \mathbf{q}) + \frac{V_0}{4\pi} \oint d\Omega_{q'} g(k_0 \mathbf{q} - k_0 \mathbf{q}') f(\mathbf{x}, t; \mathbf{q}'), \quad (3.9) \end{aligned}$$

where $f(\mathbf{x}, t; \mathbf{q})$ is the energy at time t on location $\mathbf{x}$ with propagation direction $\mathbf{q}$, $g_0(k_0)$ is the total scattering coefficient, Q_i^{-1} is the inverse of intrinsic attenuation factor, $g(k_0 \mathbf{q} - k_0 \mathbf{q}')$ means the scattering coefficient for changing propagation direction from $\mathbf{q}'$ to $\mathbf{q}$, and $\oint d\Omega_{q'}$ is the integral over solid angles. The total scattering coefficient, $g_0(k_0)$, is calculated by taking the average of g over solid angles, i.e., $g_0(k_0) = (1/4\pi) \oint d\Omega_{q'} g(k_0 \mathbf{q} - k_0 \mathbf{q}')$. Note that Eq. (3.1) is an approximate solution of Eq. (3.9) for the isotropic scattering case.

3.3.2 Monte Carlo simulation of the radiative transfer theory

The solution of Eq. (3.9) is given by Zeng (1991) for the isotropic scattering case, and by Sato (1995) for the non-isotropic scattering case. However, it is difficult to treat complex heterogeneities of the Earth, e.g., depth dependency of velocity, scattering coefficient or intrinsic attenuation, by any analytical solutions. Monte Carlo simulation is another approach to derive the solution of Eq. (3.9) (e.g., Gusev & Abubakirov, 1987; Abubakirov & Gusev, 1990; Hoshiba, 1991; Yoshimoto, 2000; Shearer & Earle, 2004; Sanborn et al., 2017). Because Monte Carlo simulation can take such a complex heterogeneities of the medium into consideration, we have adopted Direct Simulation Monte Carlo (DSMC) of Yoshimoto (2000) to solve Eq. (3.9).

In the DSMC, the energy distribution of wavefield is discretized as many particles, and the propagation of energy is represented by the movement of the particles. Each particle is emitted from the source and its movement is independent from each other, or with no interactions. The heterogeneity of medium velocity, scattering coefficient and intrinsic attenuation (V_0 , g , and Q^{-1}) in the Eq. (3.9) are treated as the function of location $\mathbf{x}$ of the each particle. Here

we assume the source radiation to be isotropic. Initial azimuthal angle φ and takeoff angle ϑ of each particle are therefore assigned by

$$\begin{aligned}\varphi &= 2\pi U_1 \\ \vartheta &= \cos^{-1}(1 - 2U_2),\end{aligned}\tag{3.10}$$

where U_1 and U_2 are uniform random numbers between 0 and 1. Each particle is located at $\mathbf{x} = (r_c, \theta, \phi)$ where r_c is the distance from the center of the Earth, θ is the azimuth, ϕ is the colatitude. It moves in each time step, according to the following relationships (Koketsu, 1991):

$$\left\{ \begin{aligned} \frac{dr_c}{d\tau} &= -V(\mathbf{x}) \cos \vartheta \\ \frac{d\theta}{d\tau} &= -\frac{V(\mathbf{x})}{r_c} \sin \vartheta \cos \varphi \\ \frac{d\phi}{d\tau} &= \frac{V(\mathbf{x})}{r_c \sin \theta} \sin \vartheta \cos \varphi \\ \frac{d\vartheta}{d\tau} &= \left(-\frac{\partial V(\mathbf{x})}{\partial r_c} + \frac{V(\mathbf{x})}{r_c} \right) \sin \vartheta - \cos \vartheta \left(-\frac{\cos \varphi}{r_c} \frac{\partial V(\mathbf{x})}{\partial \theta} + \frac{\sin \varphi}{r_c \sin \theta} \frac{\partial V(\mathbf{x})}{\partial \phi} \right) \\ \frac{d\varphi}{d\tau} &= \frac{1}{\sin \vartheta} \left(-\frac{\sin \varphi}{r_c} \frac{\partial V(\mathbf{x})}{\partial r_c} - \frac{\cos \varphi}{r_c \sin \theta} \frac{\partial V(\mathbf{x})}{\partial \phi} \right) + \frac{V(\mathbf{x})}{r_c} \sin \vartheta \sin \varphi \cot \theta \end{aligned} \right.\tag{3.11}$$

where τ is the travel time. In a small time step Δt , each particle moves by the distance of $V(\mathbf{x})\Delta t$. Scattering probability along the path of each particle is represented by $1 - \exp(-g_0(\mathbf{x})V(\mathbf{x})\Delta t)$ (section 3.1 of Sato et al., 2012). Hence, wave scattering occurs when the following condition is satisfied:

$$1 - \exp(-g_0(\mathbf{x})V(\mathbf{x})\Delta t) > U_3,\tag{3.12}$$

where U_3 is the uniform random number between 0 and 1. In the case of isotropic scattering, new azimuthal and takeoff angles are determined by the other two uniform random numbers and Eq. (3.10) upon the wave because the scattering angle does not depend on the propagation

direction of the incident waves. In the case of the non-isotropic scattering adopted later in this chapter, a new scattering angle ψ measured from the propagation direction of the incident waves is determined so as to obey the probability density function of Eq. (3.8), then new azimuthal and takeoff angles are calculated from those of incident waves and ψ .

3.3.3 Step 1: estimating the attenuation parameters of each envelope by the envelope fitting method

To estimate attenuation structure, we utilize amplitude information from the observed waveform. In general, observed seismic amplitude is expressed by the convolution of source, path, and site amplification terms. Attenuation structure relates to the path term. Other two terms, source and site amplification terms, also affect the amplitude of observed waveforms. In this section, we first separate attenuation parameters from other terms in each observed seismogram envelope by the envelope fitting method.

The observed narrow band-pass-filtered seismogram envelope of the i -th earthquake at station j located at $\mathbf{x}_j$ is represented by

$$W_{ij}^{\text{obs}}(t) = W_{ij}^s E_i(\varepsilon_{ij}; \mathbf{x}_j, t) \exp(-b_{ij}t), \quad (3.13)$$

where $W_{ij}^{\text{obs}}(t)$ is the observed envelope at time t , W_{ij}^s is the constant that includes source amplitude, site amplification and the effect of radiation pattern, $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$ is the Green's function of the envelope with the strength of a fractional velocity fluctuation ε_{ij} , and b_{ij} is equivalent to $2\pi f Q_{ij}^{-1}$ where Q_{ij}^{-1} is the inverse of intrinsic attenuation factor and f is the frequency. The Green's function of the envelope $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$ is the solution of the radiative transfer equation (Eq. 3.9) with a delta function as an initial condition. In the calculation of $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$, exponential-type random media described in section 3.3.1 is assumed and the correlation distance, a_r , in Eq. (3.7) is set to be constant. Although Sawazaki et al. (2011) have shown that the non-isotropic radiation pattern affects the shape of seismogram envelopes,

the complex effect of seismic wave scattering makes it difficult to model such an influence of radiation pattern deterministically in each observed envelopes (e.g., Takemura et al., 2017). Instead, we assume an isotropic source radiation and introduce the path-dependent constant term W_{ij}^s which depends on both earthquake and station locations for simplified consideration of non-isotropic radiation pattern.

To estimate optimal values of W_{ij}^s , ε_{ij} and b_{ij} , we calculate Green's function $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$ with an assumed value of ε_{ij} . Taking a natural logarithm of both sides of Eq. (3.13), we derive the following linear relation:

$$\log W_{ij}^{\text{obs}}(t) - \log E_i(\varepsilon_{ij}; \mathbf{x}_j, t) = \log W_{ij}^s - b_{ij}t. \quad (3.14)$$

A linear equation (3.14) can be solved in a least-squares sense for each earthquake–station pair if there are enough number of observed data W_{ij}^{obs} . The optimal ε_{ij} for each earthquake–station pair is estimated by the grid search of ε_{ij} .

3.3.4 Step 2: mapping attenuation parameters into 3D space

Waveform inversion is one of the powerful tools to investigate the internal structure of the Earth. Generally speaking, an initial structure model is assumed in waveform inversion. Difference between observed and calculated waveforms with the initial structure is used to update the initial (or assumed) structure. Sensitivity kernels connect the differences in waveforms with the modification amount of structural parameters (e.g., Yoshizawa, 2005). In this section, we estimate 3D distributions of two attenuation parameters, the inverse of intrinsic attenuation Q_i^{-1} and the strength of fractional velocity fluctuation ε , using sensitivity kernels.

Takeuchi (2016) has proposed a calculation procedure of 3D sensitivity kernels between differential envelope and attenuation parameters (Q_i^{-1} and ε) using a Monte Carlo simulation.

According to Eq. (11) of Takeuchi (2016), a differential envelope $\Delta W(t)$ is expressed as

$$\Delta W(t) = \left\{ - \int_l \frac{2\delta\varepsilon}{\varepsilon} g_0 dl + \sum_k \frac{2\delta\varepsilon(\mathbf{x}_k)}{\varepsilon(\mathbf{x}_k)} - \int_l k_0 \delta Q^{-1} dl \right\} \exp \left(-k_0 \int_l Q_i^{-1} dl \right) W_0, \quad (3.15)$$

where δQ^{-1} is the perturbation of the inverse of intrinsic attenuation factor, $\delta\varepsilon$ is that of the strength of fractional velocity fluctuation, g_0 the total scattering coefficient, k_0 the wavenumber and W_0 the energy emitted from the source. $\int_l dl$ denotes the integral along a propagation path of each particle in the Monte Carlo simulation. Subscription k denotes the k -th scattering event of each particle at location $\mathbf{x}_k$, and $\sum_k$ denotes the summation over all scattering events of each particle. Equation (3.15) represents the relationship between the differential envelope in the left-hand side and the perturbation of attenuation parameters, δQ^{-1} and $\delta\varepsilon$, in the right. Because Takeuchi (2016) has relied on the Monte Carlo simulation, sensitivity kernels derived from Eq. (3.15) are basically numerical and approximated solutions. However, if we use many particles in the Monte Carlo simulation, the difference between numerical solution and analytical one (e.g., Mayor et al., 2014; Margerin et al., 2016 for 2D space) can be negligible. One advantage of the approach of Takeuchi (2016) is that we can derive sensitivity kernels in fully 3D space, including realistic velocity structures such as depth dependency in velocity. Note that Eq. (3.15) does not contain the perturbation of correlation distance a_r . As shown in Fig. 6 of Gaebler et al. (2015), a_r and ε strongly correlate each other so that the separate estimation of a_r and ε from observational data is generally difficult. Same as the assumption of Takeuchi (2016) and section 3.3.3 in this study, we fix the value of a_r in calculating sensitivity kernels using Eq. (3.15).

Although observed seismogram envelopes are usually used in calculating Eq. (3.15), they contain several features difficult to model precisely. For example, a site amplification factor may have azimuthal dependency (e.g., Nakano et al., 2015a). Such an issue makes it difficult to separate the magnitude of source energy and site amplification from W_{ij}^s of Eq. (3.14), then affects W_0 in Eq. (3.15). Alternatively, we utilize the envelopes synthesized in section 3.3.3 as

our observed data. Because the synthesized envelopes should represent the estimated intrinsic attenuation factor and the strength of random velocity fluctuation, separation of source energy and site amplification in W_{ij}^s is not necessary for mapping the estimated attenuation parameters into 3D space. In this way, Eq. (3.15) becomes

$$\frac{W_{ij}^{\text{cal}}(t)}{W_{ij}^s} - E_i(Q_{\text{ini}}^{-1}, \varepsilon_{\text{ini}}; \mathbf{x}_j, t) = \left\{ - \int_l \frac{2\delta\varepsilon}{\varepsilon_{\text{ini}}} g_0 dl + \sum_k \frac{2\delta\varepsilon(\mathbf{x}_k)}{\varepsilon_{\text{ini}}(\mathbf{x}_k)} - \int_l k_0 \delta Q^{-1} dl \right\} \exp\left(-k_0 \int_l Q_{\text{ini}}^{-1} dl\right), \quad (3.16)$$

in which $W_{ij}^{\text{cal}}(t)$ is the synthesized envelope in the first step and $E_i(Q_{\text{ini}}^{-1}, \varepsilon_{\text{ini}}; \mathbf{x}_j, t)$ is the Green's function of envelope between the i -th earthquake and the j -th station with structural parameters Q_{ini}^{-1} and ε_{ini} . In Eq. (3.16), we explicitly indicate the initial assumed value of Q_i^{-1} and ε by the subscript "ini".

The integral along a raypath $\int_l dl$ in Eq. (3.16) is discretized by the relationship $dl \approx V(\mathbf{x})\Delta t$ where Δt is a small time step. Since Eq. (3.16) becomes a linear equation with respect to perturbations δQ^{-1} and $\delta\varepsilon$, it can be rewritten in a well-known matrix form with many observed data:

$$\mathbf{d} = \mathbf{G}\mathbf{m}, \quad (3.17)$$

where $\mathbf{d}$ is the data vector, $\mathbf{m}$ the model parameter vector, and $\mathbf{G}$ the matrix composed by sensitivity kernels for both δQ^{-1} and $\delta\varepsilon$.

The matrix $\mathbf{G}$ is large and sparse in an application of the waveform inversion, resulted in a difficulty of solving Eq. (3.17) in a straightforward way in most cases. Bayesian-type algebraic reconstruction technique (ARTB) of Herman et al. (1979) is one of the effective techniques to solve large and sparse matrix inversion problems. In seismology, Hirahara (1988) first utilized ARTB in seismic tomography followed by Kamiya & Kobayashi (2000) and Nakamura et al. (2006). In this study, we also adopted ARTB to solve Eq. (3.17).

Let d_i is the i -th component of vector $\mathbf{d}$ and $\mathbf{g}_i$ the i -th row vector of $n \times m$ matrix $\mathbf{G}$ of Eq. (3.17) where the numbers of components of data vector $\mathbf{d}$ and model vector $\mathbf{m}$ are n and m , respectively. The algorithm of ARTB can be summarized as follows (Hirahara, 1988):

Algorithm 1 Algorithm ARTB

Input: $\mathbf{d}$: data vector, $\mathbf{G}$: matrix which consists of sensitivity kernels

Output: $\mathbf{m}$: model vector

Initialization: $\mathbf{u}^{(1)} = 0$ ($\mathbf{u}$ is the n -th dimensional vector), $\mathbf{m}^{(1)} = \mathbf{m}_0$

for $p = 1$ **to** P (p -th global iteration step) **do**

for $i = 1$ **to** n (i -th data iteration step) **do**

$k = n(p - 1) + i$

$c^{(k)} = \lambda^{(k)} \frac{1/\sigma_{di} (d_i - \mathbf{g}_i \cdot \mathbf{m}^{(k)}) - u_i^{(k)}}{1 + |\overline{\mathbf{g}}_i|^2}$

$\mathbf{u}^{(k+1)} = \mathbf{u}^{(k)} + c^{(k)} \mathbf{e}_i$

$\mathbf{m}^{(k+1)} = \mathbf{m}^{(k+1)} + c^{(k)} \overline{\overline{\mathbf{g}}_i}$

end for

end for

where $\overline{\mathbf{g}}_i$ and $\overline{\overline{\mathbf{g}}_i}$ are defined as

$$\overline{\mathbf{g}}_i = \frac{1}{\sigma_{di}} (\sigma_{m1} G_{i1}, \sigma_{m2} G_{i2}, \dots, \sigma_{mm} G_{im})^t, \quad (3.18)$$

and

$$\overline{\overline{\mathbf{g}}_i} = \frac{1}{\sigma_{di}} (\sigma_{m1}^2 G_{i1}, \sigma_{m2}^2 G_{i2}, \dots, \sigma_{mm}^2 G_{im})^t. \quad (3.19)$$

$(\sigma_{m1}, \sigma_{m2}, \dots, \sigma_{mm})$ are the m -th column components of the model covariance matrix $\sigma_{\mathbf{m}}$, σ_{di} the standard deviation for the i -th data, and $\lambda^{(k)}$ the relaxation parameter for the k -th step. The ratio of $\sigma_{\mathbf{m}}$ and σ_{di} works to stabilize the inversion results. If σ_{di} is large, i.e., the quality of the data is low, or the diagonal components of matrix $\sigma_{\mathbf{m}}$ are small, i.e., the initial model is expected to be close to the true model, the solution $\hat{\mathbf{m}}$ is strongly constrained to the initial model value $\mathbf{m}_0$. The square of off-diagonal components of $\sigma_{\mathbf{m}}$ represents the covariance between each model parameters. In this chapter, we set the off-diagonal components of $\sigma_{\mathbf{m}}$ to be zero, and σ_{di} is set to be constant regardless of i . After sufficient global iteration steps,

$\mathbf{m}^{(k)}$ converge to the final solution in the least-squares sense denoted by $\hat{\mathbf{m}}$. As shown above, ARTB does not require the whole components of matrix $\mathbf{G}$ so that it can be applied in a large equation system such as an envelope inversion in this study.

3.3.5 Data and target region

To apply the method proposed in sections 3.3.3 and 3.3.4 to southwestern Japan, we collected three-component velocity seismograms of 328 earthquakes recorded at 236 stations of Hi-net (Okada et al., 2004). Station and hypocenter distributions are shown in Fig. 3.4 (a) and (b), respectively. Magnitude range of earthquakes covered from 3.0 to 4.5 in the JMA magnitude scale, which was sufficiently small to neglect the influence of source time functions. Epicentral distance of seismograms was restricted to be less than 100 km, and the depth range of earthquakes was down to 90 km. Because the DSMC can treat depth-dependent velocity structure, we adopted the 1D velocity structure shown in Fig. 3.4 (c). Our velocity structure was similar to the S-wave velocity structure of the JMA2001 model (Ueno et al., 2002). After applying a 1-2 Hz or 2-4 Hz band-pass filter for each seismic record, we calculated its mean-square (MS) envelope by taking a vector sum of each three-component seismograms. Then, each MS envelope was resampled from 100 Hz to 1 Hz after taking a moving average with a time window of 1.5 s. Irrelevant envelopes such as contaminated by other small earthquakes or any malfunctions of seismometers were excluded by visual inspection.

We assumed the exponential-type random medium with $a_r = 1$ (km) in calculating $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$ with Eq. (3.9). The above value of the correlation distance a_r was selected following the crustal structure model proposed by Kobayashi et al. (2015).

3.3.6 Result (1): Attenuation parameters for each earthquake-station pair

In application of the method proposed in section 3.3.3 to the actual data described in section 3.3.5, we set the range of our grid search for ε_{ij} from 0.016 to 0.136. The Green's function of seismogram envelope $E_i(\varepsilon_{ij}; \mathbf{x}_j, t)$ was calculated using the DSMC with the 1D velocity

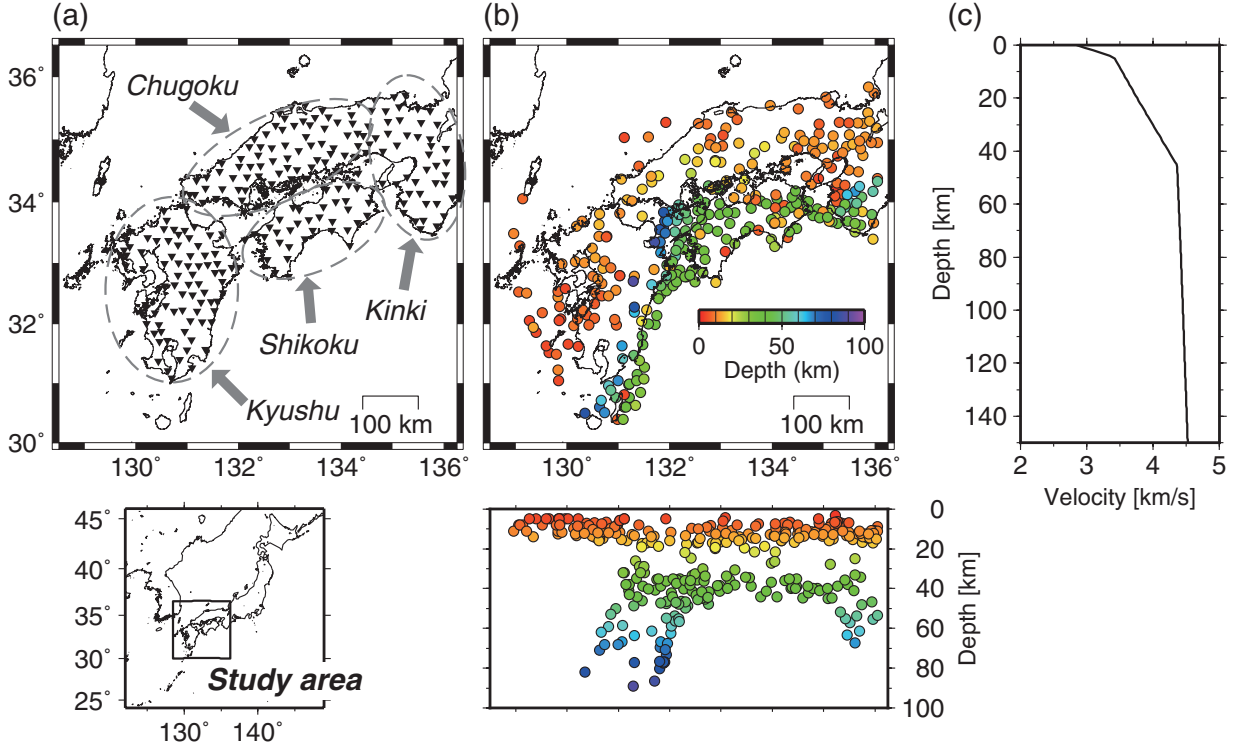


Figure 3.4 (a) Stations, (b) hypocenters and (c) 1D velocity structure used for the analysis of section 3.3. Gray ovals in (a) represent approximate regions called Kinki, Chugoku, Shikoku, and Kyushu, respectively.

structure of Fig. 3.4 (c). The time window for envelope fitting (Eq. 3.14) was set to be 60 s measured from the arrival time of the corresponding direct S-wave. After searching the global minimum of b_{ij} and ε_{ij} for each observed envelope, we calculated a correlation coefficient between observed and synthesized envelopes, and excluded them if the correlation coefficient was less than 0.8. Figure 3.5 shows the scatter diagrams of intrinsic attenuation (Q_{ij}^{-1}) and velocity fluctuation (ε_{ij}) together with their histograms in two different frequency ranges. Both Q_{ij}^{-1} and ε_{ij} at 2-4 Hz are clearly smaller than those at 1-2 Hz. Positive correlation between Q_{ij}^{-1} and ε_{ij} is obvious in both frequency range in Fig. 3.5. Modal values of Q_{ij}^{-1} and ε_{ij} are about 0.005 and 0.05 at 1-2 Hz while 0.00275 and 0.036 at 2-4 Hz, respectively. The estimation error of each Q_{ij}^{-1} at each frequency range was calculated by the Jackknife method (e.g., chapter 3 of Wolter, 1985).

Figures 3.6-3.9 show spatial distributions of Q_{ij}^{-1} and ε_{ij} , scatter diagrams of Q_{ij}^{-1} and ε_{ij} , and comparisons of observed and synthesized envelopes for several stations of a common earthquake in the frequency range of 1-2 Hz. Figure 3.6 is for the earthquake that has occurred in the crust of the Chugoku region while Fig. 3.7 for the one in the Kyushu region. In both cases, large Q_{ij}^{-1} and ε_{ij} values are found in earthquake-station pairs with short hypocentral distances. These results imply that both Q_{ij}^{-1} and ε_{ij} in the shallow part of the crust would be large. In addition, both Q_{ij}^{-1} and ε_{ij} values tend to be larger in the Kyushu region than in the Chugoku region. The same relationship can be found in the results of other crustal earthquakes. Hence, both intrinsic attenuation and velocity fluctuation were expected to be larger in the Kyushu region than the other regions.

Figures 3.8 and 3.9 show the result for the intraplate earthquake that occurred in the Kinki and Kyushu regions, respectively. In Fig. 3.8, we can see weak hypocentral distance dependency in both Q_{ij}^{-1} and ε_{ij} , although the absolute values of Q_{ij}^{-1} and ε_{ij} are smaller, that is, weak intrinsic attenuation and scattering, than the case of crustal earthquakes (Figs. 3.6 and 3.7). In summary, Figs. 3.6-3.9 suggest the existence of complex 3D heterogeneity in both Q^{-1} and ε in the target region shown in Fig. 3.4(a).

3.3.7 Result (2): 3D heterogeneous structure of intrinsic attenuation and velocity fluctuation

To apply the mapping method described in section 3.3.4 to the result shown in section 3.3.6, we discretized the subsurface structure of the target region by small blocks. Size of each block was 0.4° in the latitudinal and longitudinal directions and 20 km in the depth direction, except for the shallowest layer whose thickness was set to be 10 km. ARTB was applied to the envelopes synthesized in section 3.3.6, except the envelopes whose ε_{ij} lay on the edge values of the parameter space: 0.016 or 0.136.

ARTB was applied sequentially to each earthquake. First, we applied ARTB with the initial model of attenuation and the first earthquake to calculate δQ^{-1} and $\delta \varepsilon$. The calculated δQ^{-1}

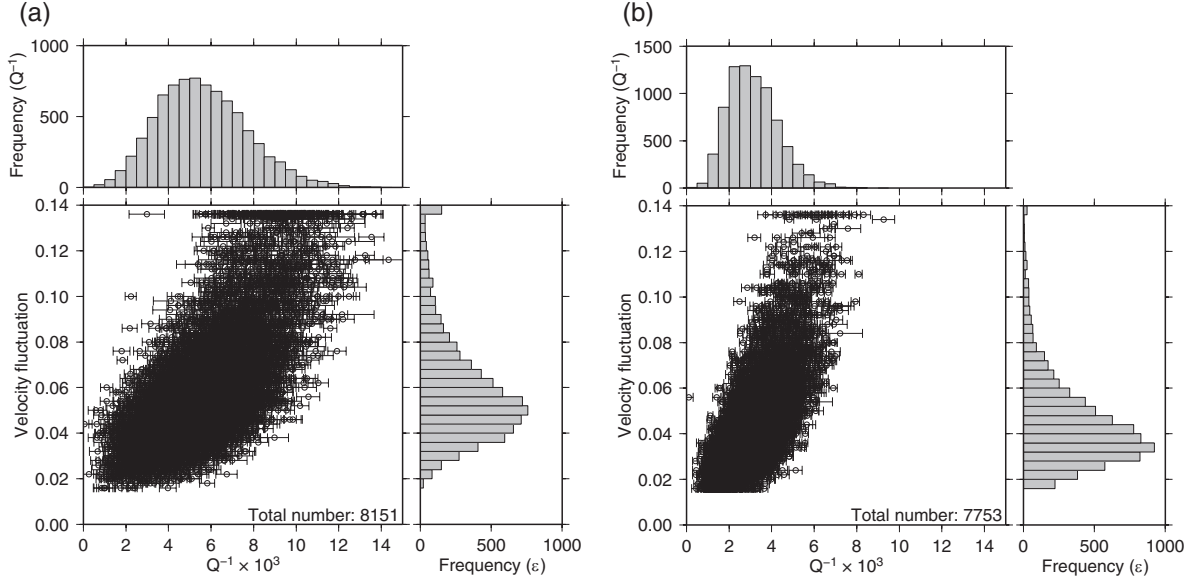


Figure 3.5 Scatter diagram of Q_{ij}^{-1} and ε_{ij} for all the earthquake-station pairs using frequency range of (a) 1-2 Hz and (b) 2-4 Hz. Histograms of Q_{ij}^{-1} and ε_{ij} are upper and right panel of scatter diagram, respectively. The bins of Q_{ij}^{-1} and ε_{ij} are 0.0005 and 0.004, respectively. Each error bar denotes estimation error of Q_{ij}^{-1} inferred from the Jackknife method.

and $\delta\varepsilon$ were then used to modify the initial model. The modified model was used in the sensitivity kernel calculation for the second earthquake, and the first modified model was modified again by the values of δQ^{-1} and $\delta\varepsilon$ calculated for the second earthquake. Repeating the above procedure for all the earthquakes shown in Fig. 3.4(b), we obtained our final 3D heterogeneous distribution of Q^{-1} and ε of the target area. The initial attenuation parameters were taken from the modal values shown in Fig. 3.5: $(Q_{ini}^{-1}, \varepsilon_{ini})$ was $(5.0 \times 10^{-3}, 0.05)$ at the frequency range of 1-2 Hz and $(2.75 \times 10^{-3}, 0.036)$ at 2-4 Hz, respectively. To avoid the attenuation parameters to be inappropriate, we set the lower limits for Q^{-1} and ε ; Q^{-1} should be larger than 0.6667×10^{-3} or $Q \leq 1500$ in the frequency range of 1-2 Hz and 0.25×10^{-3} or $Q \leq 4000$ in 2-4 Hz, respectively, and ε was assumed to be larger than 0.01 in both frequency ranges. We set the standard deviation for model parameters (i.e., diagonal elements of σ_m) to be 1.0 for δQ^{-1} , and 0.1 for $\delta\varepsilon$ in each block and adjusted standard deviation for the data to stabilize the inversion described in section 3.3.4. Because seismic wave scattering is a

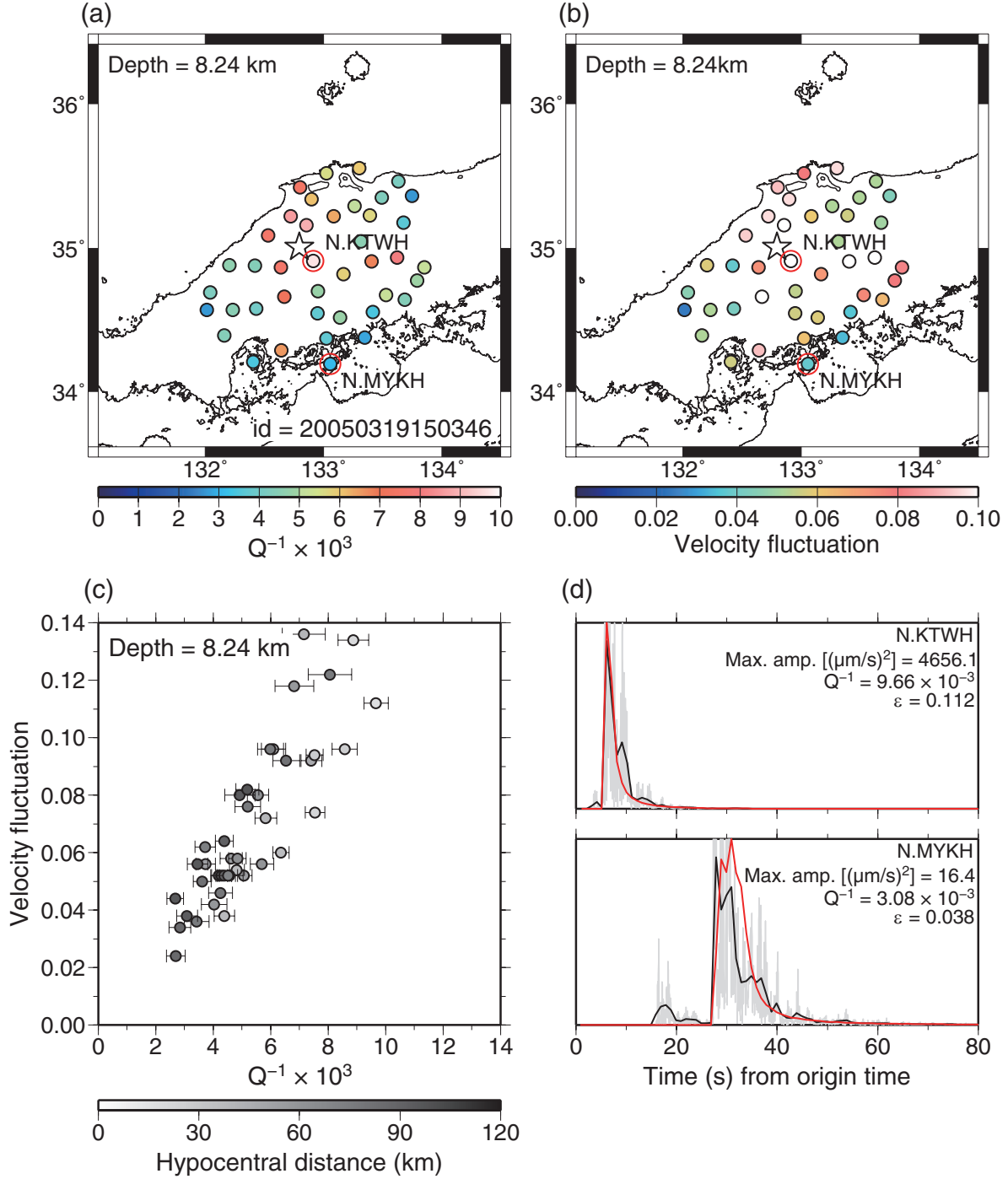


Figure 3.6 Spatial distribution of (a) Q_{ij}^{-1} and (b) ε_{ij} , (c) scatter diagram of Q_{ij}^{-1} and ε_{ij} , and (d) comparison of observed (gray line for non-resampled and black for resampled) and synthesized (red) envelopes at stations N.KTWH and N.MYKH for the earthquakes occurred at 19:15:03 on March 19, 2005. Stars in (a) and (b) represent epicenter and hypocentral depth is plotted at upper left of each panel. Shades of symbols in (c) denote hypocentral distance. Red circles in (a) and (b) show locations of stations shown in (d). Vertical axis of (d) is in linear scale with which maximum value is shown in top right.

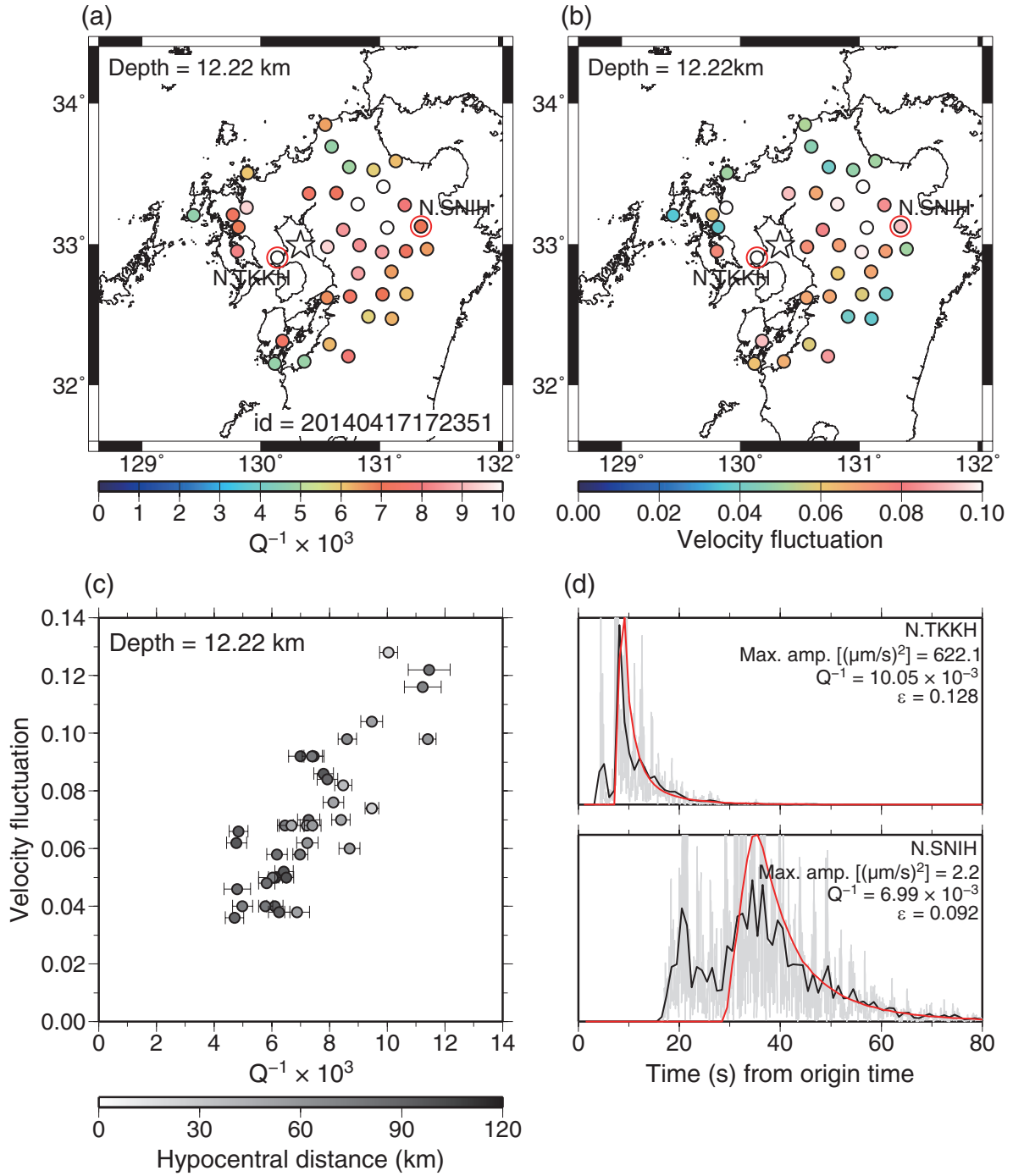


Figure 3.7 Same as Fig. 3.6 but for earthquake occurred at 17:23:51 on April 17, 2014 and stations N.TKKH and N.SNIH.

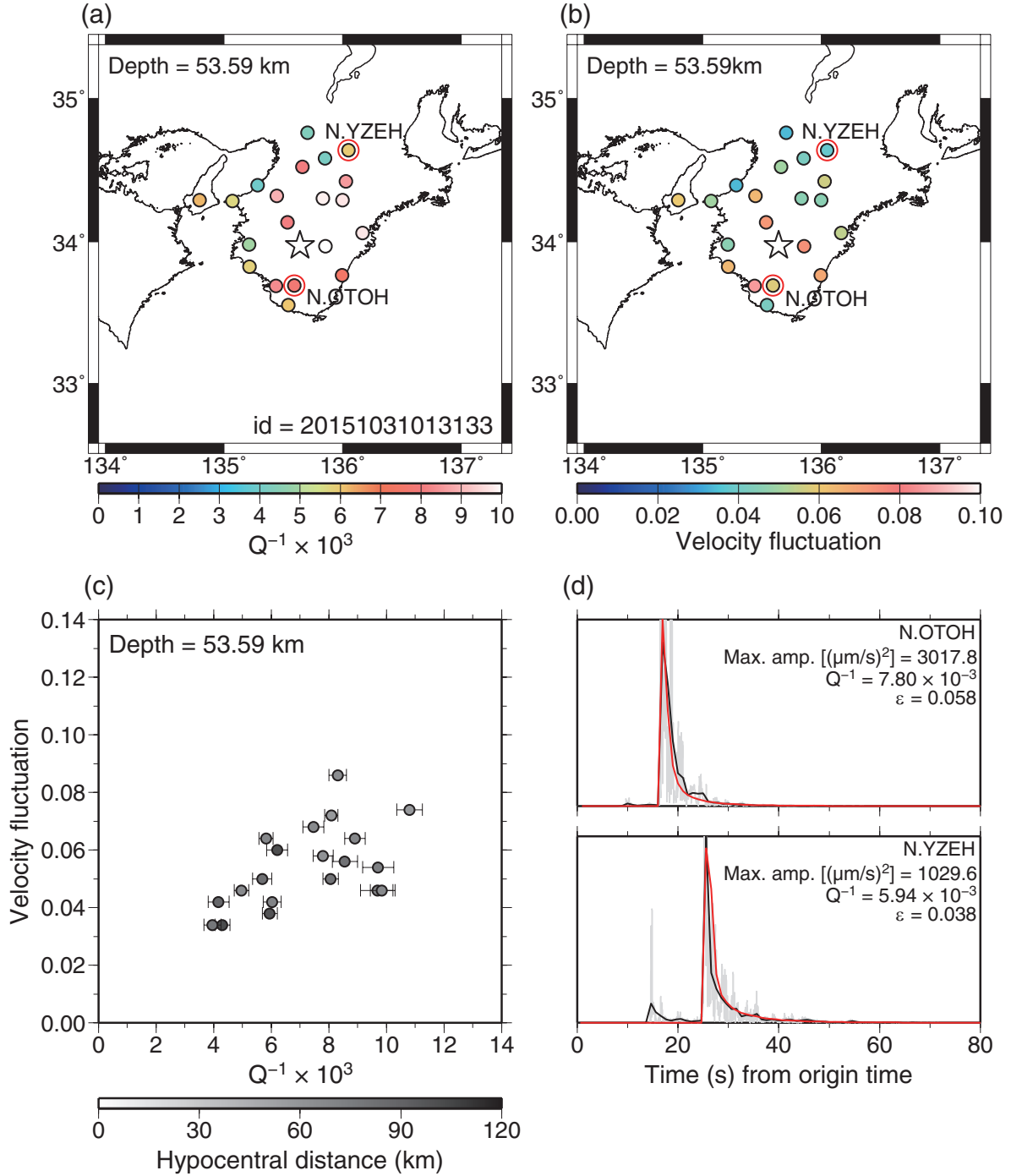


Figure 3.8 Same as Fig. 3.6 but for earthquake occurred at 01:31:33 on October 31, 2015 and stations N.OTOH and N.YZEH.

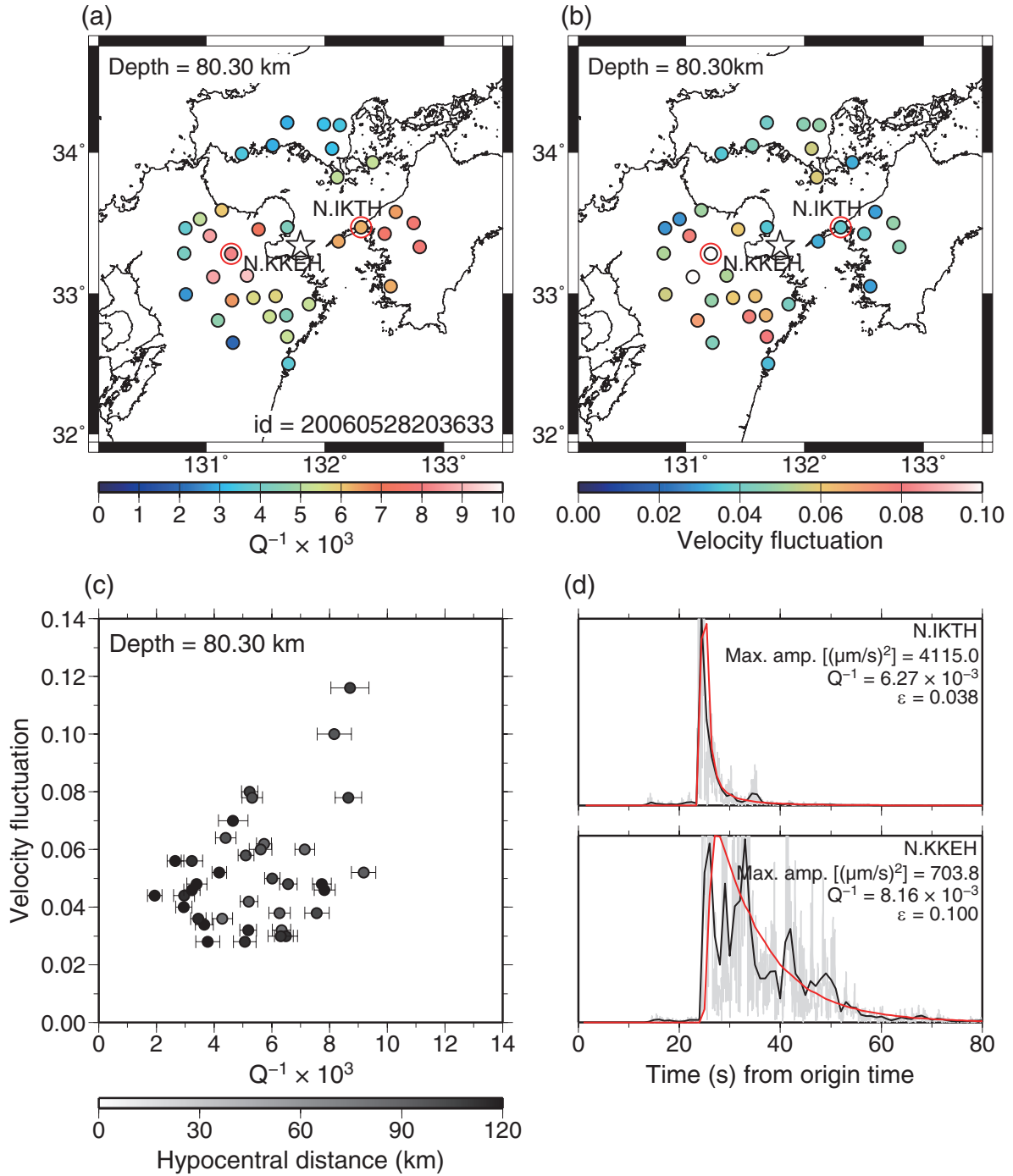


Figure 3.9 Same as Fig. 3.6 but for earthquake occurred at 20:36:33 on May 28, 2006 and stations N.IKTH and N.KKEH.

stochastic process, each observed seismogram envelope is a realization which obeys a certain probability distribution function. Hence, estimated structural parameters in the step 1 (section 3.3.6) basically had a scatter. In other words, structural parameters might be different from each other in the two seismogram envelopes even if each raypath is similar to each other. Thus, we assumed various values in the standard deviation for the data, to find acceptable inversion result. Appropriate value of standard deviation for the data was determined by the trade-off curve between data residual and variance of model perturbation. We adopted 8000 for standard deviation for the data (σ_{di}) for all the data at 1-2 Hz and 11000 at 2-4 Hz, respectively.

To investigate the resolution of the inverted attenuation distribution model, we conducted a checkerboard resolution test, assuming the checkerboard pattern with (Q^{-1}, ε) to be (0.01, 0.073) and (0.003, 0.031). The velocity structure for the checkerboard resolution test was the same as that of our inversions (Fig. 3.4c). Seismogram envelopes in the assumed checkerboard pattern were synthesized using the DSMC. The data set of envelopes was the same as that in the 1-2 Hz frequency range. Figure 3.10 shows the result of the checkerboard resolution test. Because we needed a large value for the standard deviation of the data in the inversion process to obtain the stable results, the amplitudes of the inverted anomalies of both intrinsic attenuation and velocity fluctuation in the resolution test were smaller than those of the assumed ones. Nevertheless, assumed checkerboard pattern has been reconstructed well for the first and second layers (layers 1 and 2). Because of the distribution of earthquakes used in this study (Fig. 3.4b), assumed pattern has been recovered only in the limited area of the Kinki, Shikoku, and Kyushu region of the third layer (layer 3). Although we should pay attention to the amplitude of anomalies in the inverted attenuation structure, we conclude that the regional and depth dependencies of both attenuation parameters can be discussed.

Figures 3.11 and 3.12 show the estimated 3D distributions of the intrinsic attenuation and the strength of velocity fluctuation in the frequency ranges of 1-2 Hz and 2-4 Hz, respectively.

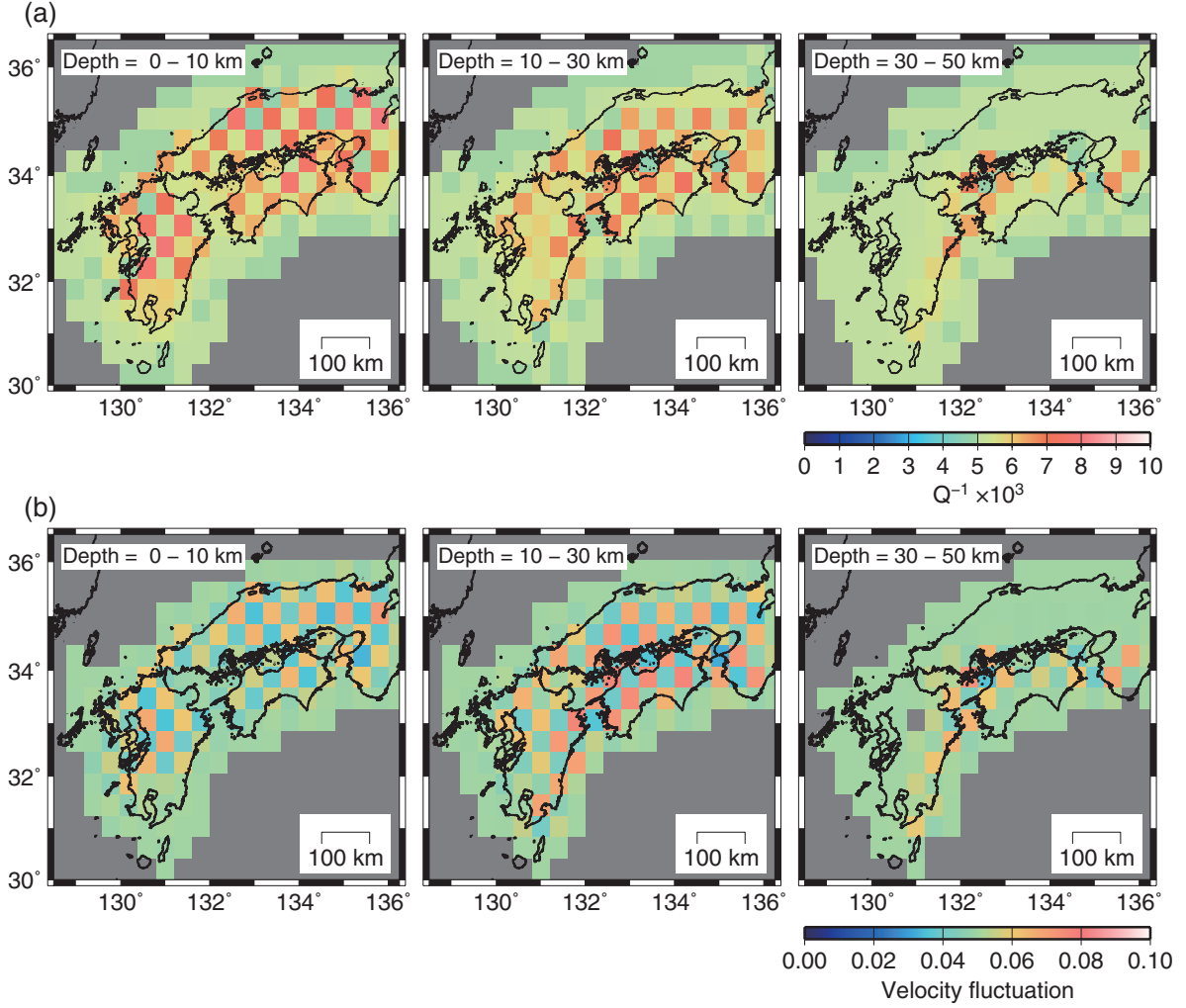


Figure 3.10 Inverted anomalies of (a) intrinsic attenuation and (b) strength of fractional velocity fluctuation of the checkerboard resolution test. From left to right, the 1st layer (0 to 10 km depth), 2nd layer (10 to 30 km depth), and 3rd layer (30 to 50 km depth), respectively.

Interesting features of our estimated attenuation parameters are as follows: (1) larger values of Q_i^{-1} and ε in the first layer of the Kyushu region than the other regions, (2) their smaller values in the layer 2 than the layer 1, which is obvious in the Chugoku region, and (3) an area of small Q_i^{-1} in the central Shikoku between the two areas of large Q_i^{-1} in the layer 3 of eastern and western Shikoku regions. These features can be seen in the frequency ranges of both 1-2 and 2-4 Hz, although their absolute values of attenuation parameters at 2-4 Hz are smaller than those at 1-2 Hz. To estimate the error of the inverted attenuation parameters (Figs. 3.11

and 3.12), we have conducted a Jackknife test for them. In the Jackknife test, we first sorted the used earthquakes into 10 groups. Each group contained 90% of the total earthquakes, that is, the removed earthquakes at each group were about 10% of the total earthquakes. The removed earthquakes were selected using the uniform random variables without duplication. Then, we conducted the same inversion process described in section 3.3.4 for each group, and combined the 10 inverted results to calculate the estimation error of Q_i^{-1} and ε . Figs. 3.13 and 3.14 show the spatial distributions of estimated errors of Q_i^{-1} and ε in the frequency ranges of 1-2 Hz and 2-4 Hz, respectively. The maximum values of the estimation error for Q_i^{-1} were 4.46×10^{-3} at 1-2 Hz and 2.31×10^{-3} at 2-4 Hz while 0.021 and 0.019 for ε . The errors in our inversions do not affect the discussions in the following section.

3.4 Discussion

As a previous result, the 2D intrinsic and scattering attenuation structure of southwestern Japan for the frequency range of 1-2 Hz estimated by Carcolé & Sato (2010) have the following features. First, their intrinsic attenuation in the Kyushu region is stronger than the other regions. Second, a region of large scattering is found around the volcanic front in the Kyushu region, and weak in the other regions. Takahashi et al. (2013) have estimated the strong seismic scattering at a shallow depth beneath active volcanoes in the Kyushu region from the analysis of peak delay times of the direct S-waves. Both 2D (Fig. 3.3) and 3D (Figs. 3.11 and 3.12) attenuation models in this study have the same features as these previous studies. The 2D structure estimated by the MLTWA in this study (Fig. 3.3) indicates the two regions of strong intrinsic and scattering attenuation regions: central and southern Kyushu. The strong attenuation area in southern Kyushu is not clear in Carcolé & Sato (2010), probably due to the difference between used stations of theirs and ours. We have utilized more stations in the southern part of Kyushu than Carcolé & Sato (2010) so that its strong attenuation around the volcanoes has been clearly estimated. Although the horizontal resolution of our

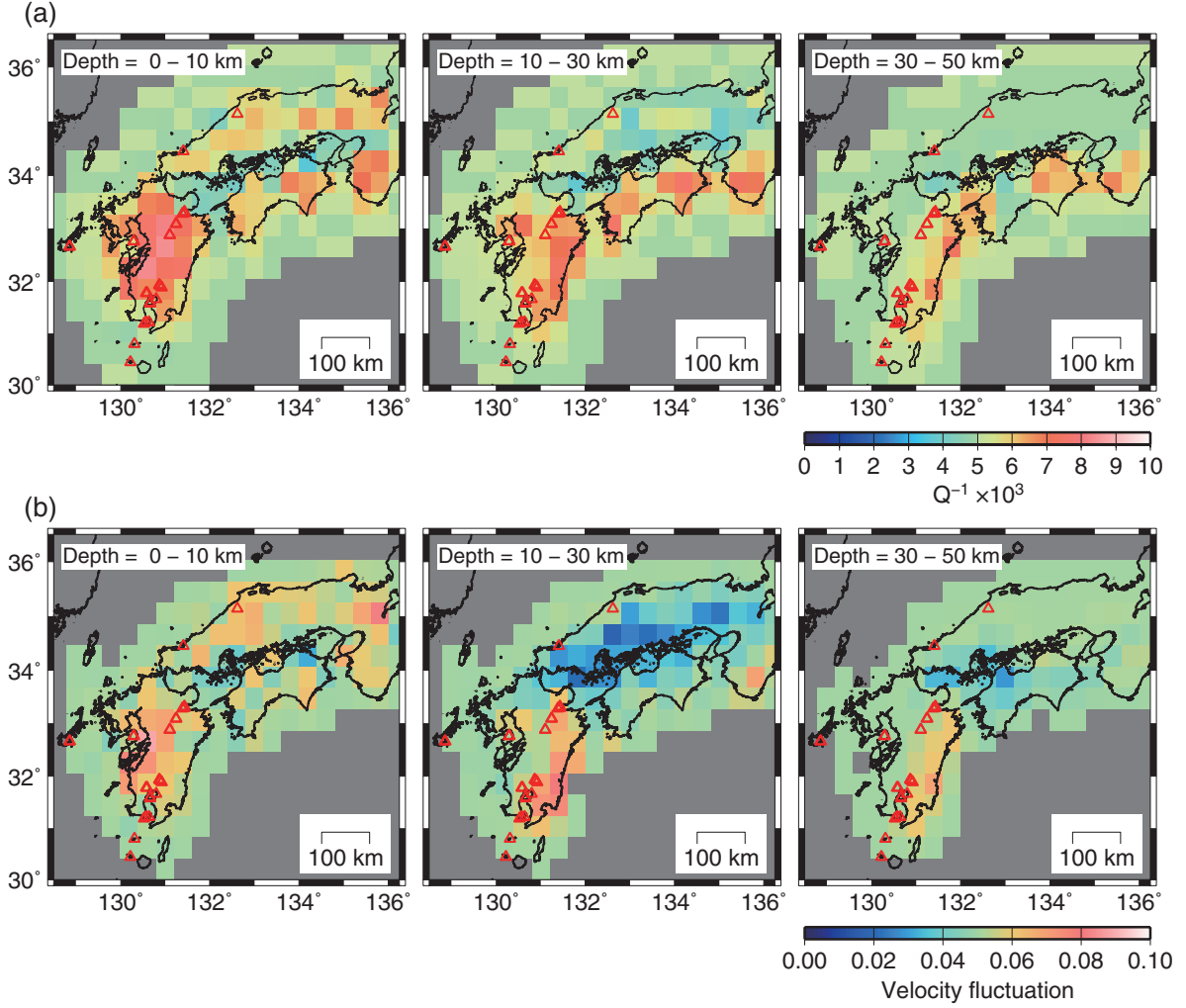


Figure 3.11 Estimated 3D distribution of (a) intrinsic attenuation and (b) strength of fractional velocity fluctuation using 1-2 Hz frequency range. From left to right, the 1st layer (0 to 10 km depth), 2nd layer (10 to 30 km depth), and 3rd layer (30 to 50 km depth), respectively. Red triangles denote active volcanoes listed in Japan Meteorological Agency (2013). Area with no perturbed attenuation parameters is shaded by gray.

3D attenuation structure (Figs. 3.11 and 3.12) is still coarse in comparison of the attenuation structure with volcanoes in detail, the strong intrinsic and scattering attenuation in layer 1 of the Kyushu region found in our 3D model would relate to the active volcanoes in the region.

Next, we discuss the depth dependency of attenuation parameters found by our 3D analysis. Comparing the attenuation parameters between layers 1 and 2 (Figs. 3.11 and 3.12), we can see that both intrinsic and scattering attenuations in layer 2 are weaker than those

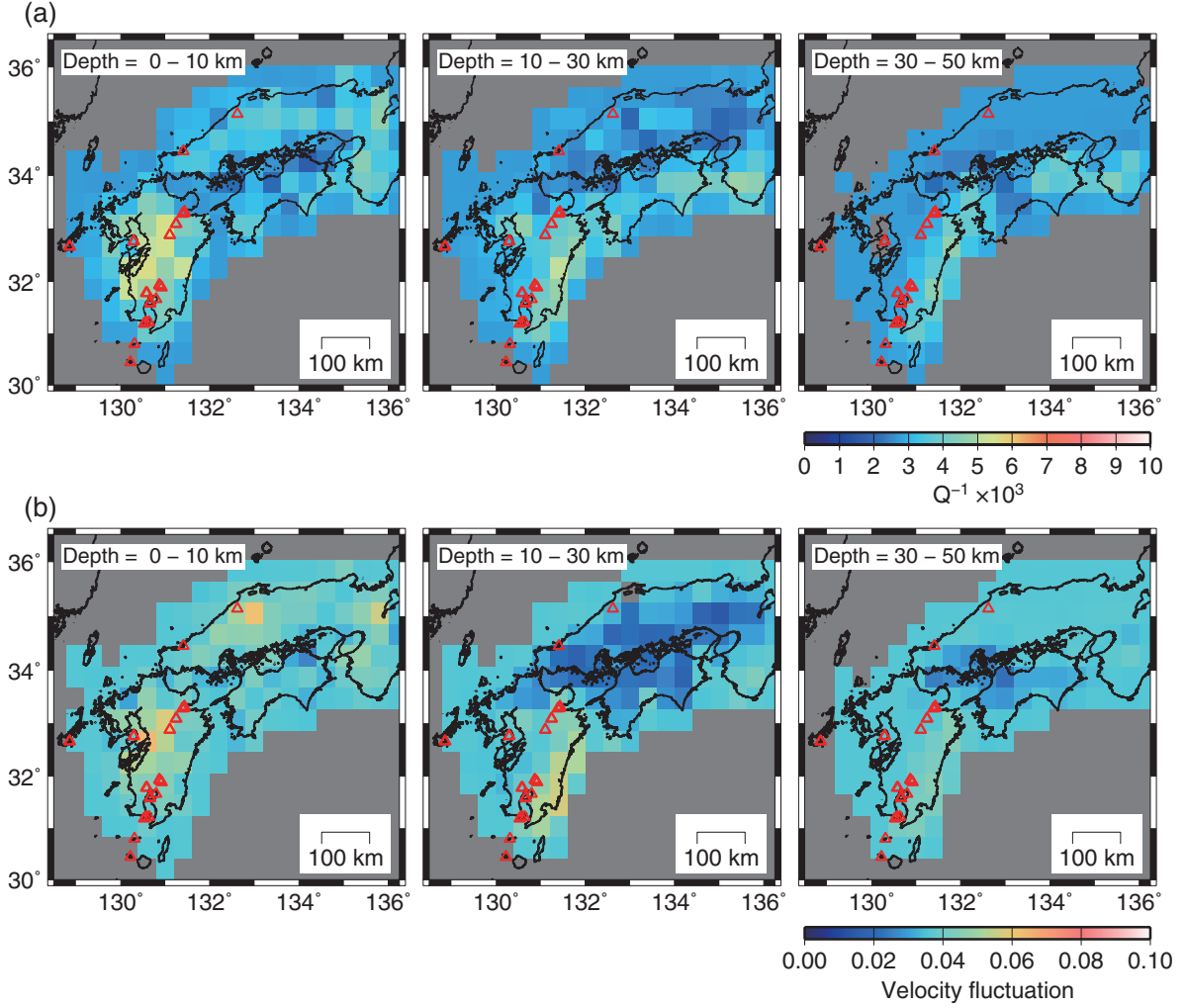


Figure 3.12 Same as Fig. 3.11 but using the frequency range of 2-4 Hz.

in layer 1, particularly in the Chugoku region. By considering the checkerboard resolution test (Fig. 3.10), this feature can be evaluated to be reliable. Similar depth dependency of attenuation parameters has been derived in Southern California (Wang & Shearer, 2017). Wang & Shearer (2017) has estimated the depth-dependent 1D intrinsic and scattering attenuation structures by the forward modeling of the stacked seismogram envelopes in the frequency range of 2-4 Hz. Takemura & Furumura (2013) analyzed the energy partition of direct P-waves of crustal earthquakes in Japan and found that the energy partition in the transverse component was increased as the propagation distance increased in high frequency range

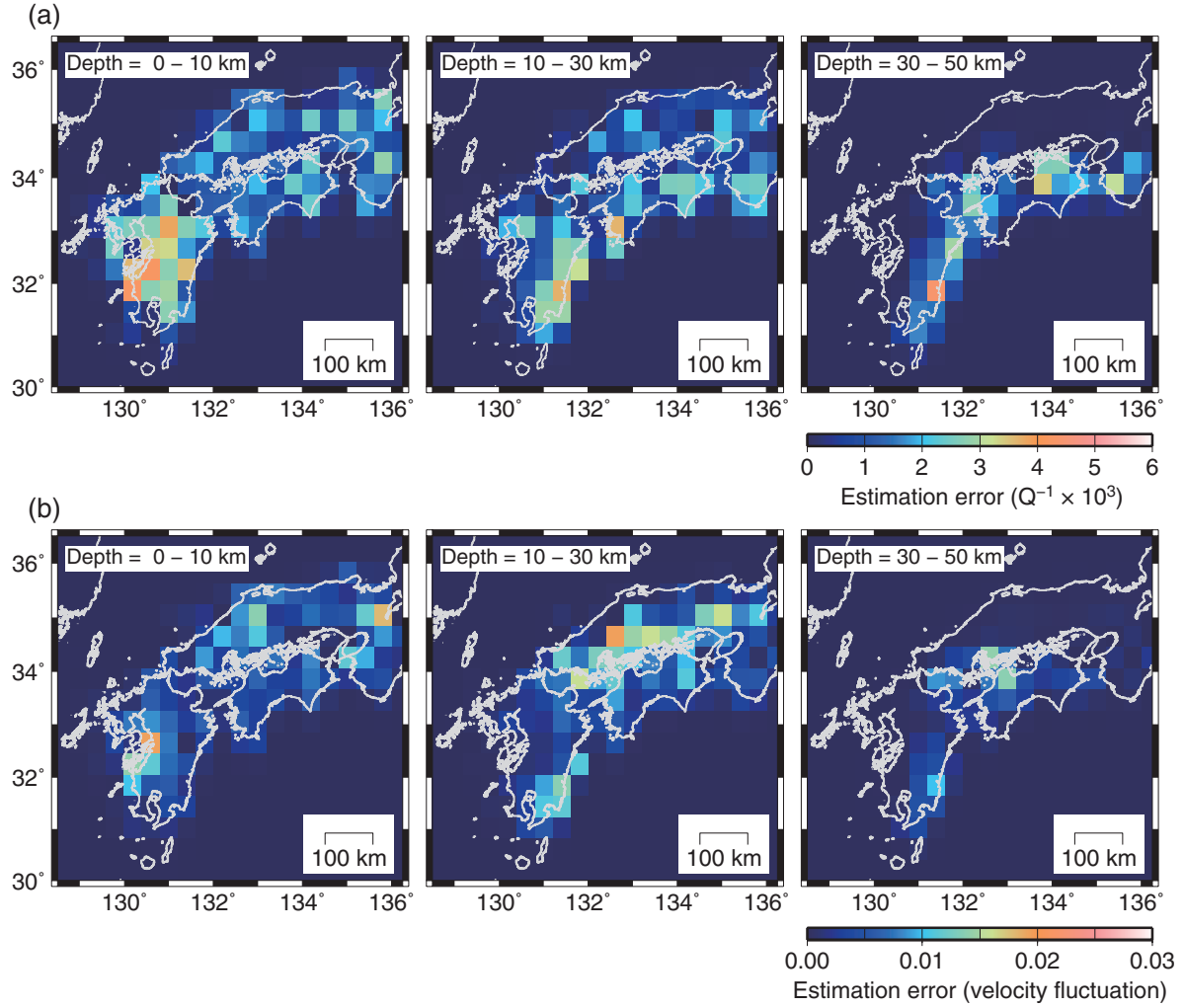


Figure 3.13 Estimated 3D distribution of errors for the (a) inverse of intrinsic attenuation factor and (b) strength of fractional velocity fluctuation inferred from the Jackknife test using the frequency range of 1-2 Hz.

(roughly $\geq 8\text{Hz}$). Takemura & Furumura (2013) has proposed the crustal structure model whose strength of random velocity fluctuations increases as the depth increases through the forward modeling of the waveform of initial P-waves by the finite difference calculation. Meanwhile, they also pointed out that their proposed model could not explain the frequency dependency in the energy partition of the initial P-waves. Their observed energy partition in the transverse component in both frequency range of 1-2 and 2-4 Hz seemed independent of the propagation distance (Figs. 6 and 7 of Takemura & Furumura, 2013), implying that

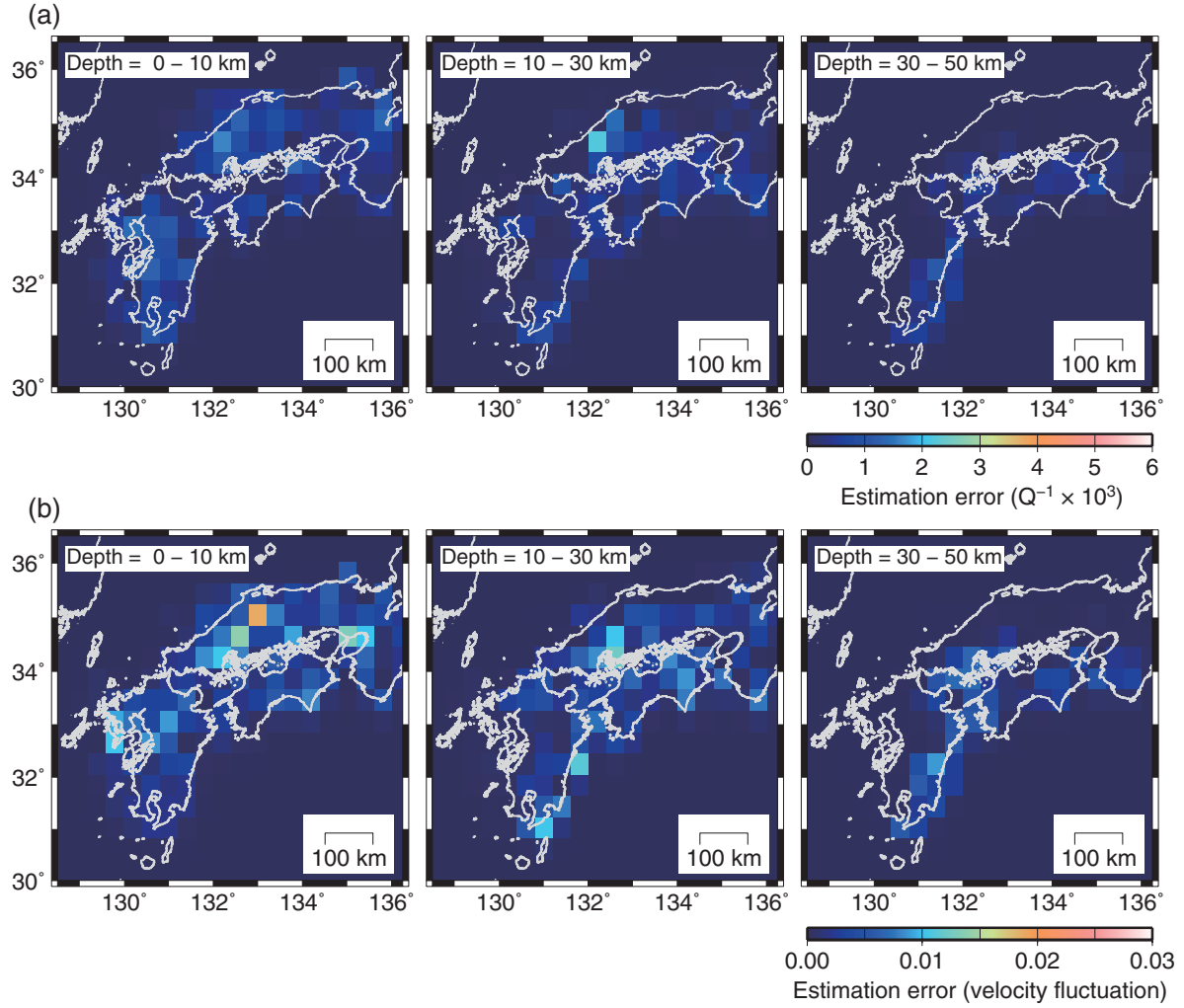


Figure 3.14 Same as Fig. 3.13 but using the frequency range of 2-4 Hz.

their crustal structure model may be inconsistent with the observations in the low frequency range. Since the same two frequency ranges (1-2 and 2-4 Hz) have been used in this section, the derived scattering attenuation parameters in this section may be consistent with the above discussion of Takemura & Furumura (2013). Note that Takahashi et al. (2013) have proposed the structure model where the power spectrum density of the random media decreased at depth, seems to be inconsistent with the above mentioned model of Takemura & Furumura (2013). To reveal the attenuation characteristics of the medium in a wide frequency range, it is important to compare the estimated structure models by the various analysis techniques in

a wide frequency range in the future. Our proposed structure estimation method relies on the Born approximation so that it is difficult to apply it directly to the seismogram envelopes at a higher frequency range than this study (i.e., ≥ 4 Hz), and further theoretical developments are required for modeling high-frequency seismogram envelopes.

Now, let us discuss the intrinsic attenuation properties in the Shikoku region. The area of low intrinsic attenuation can be found at the depth of 10-50 km in the central Shikoku region, surrounded by high intrinsic attenuation areas in west and east, as shown in Figs. 3.11 and 3.12. One interesting feature is that the strength of random velocity fluctuation in the Shikoku region is weak and almost homogeneous at the above depth while its intrinsic attenuation parameter shows obvious regional dependency. Kita & Matsubara (2016) have estimated a 3D heterogeneous total attenuation structure, i.e., sum of intrinsic and scattering attenuations, in and around the Shikoku region and found an area of low attenuation at the depth of 10-20 km at its certain region. Considering the separate estimation by our 3D analysis, the low attenuation area of Kita & Matsubara (2016) can be attributed to not scattering but intrinsic attenuation. The central and western Shikoku regions are known as a focal area of long-term slow slip events, and the activity of the long-term slow slip events shows spatiotemporally heterogeneous (Kobayashi, 2014). Because the spatial resolution of our 3D structure is lower than that of Kita & Matsubara (2016), our discussions may not go into detailed relationship between the attenuation structure and the focal areas of long-term slow slip events, as Kita & Matsubara (2016) have done. Here, we only point out the possibility that the spatiotemporal heterogeneities of the slow slip events (Kobayashi, 2014) may have some interactions with the heterogeneous intrinsic attenuation in the Shikoku region.

Finally, we compare the ranges of intrinsic and scattering attenuation parameters in this study with those in several previous studies that have discussed frequency-dependent attenuation parameters. Sawazaki et al. (2011) estimated intrinsic attenuation and strength of random velocity fluctuation in the Chugoku and Shikoku regions by modeling the envelope broaden-

ing effect of direct S-waves in several frequency ranges of 2-16 Hz. Takahashi et al. (2013) estimated a 3D distribution of power spectral densities of random media from the analysis of peak delay times of the direct S-waves at 4-32 Hz. In addition, Takahashi et al. (2014) estimated 3D distribution of intrinsic attenuation from the maximum amplitude of direct S-waves. On the other hand, Takemura et al. (2017) focused on the distortion of the radiation pattern in direct P- and S-waves at 0.5-4 Hz and estimated intrinsic attenuation and strength of random velocity fluctuation in the Chugoku region by the forward modeling of waveforms with the finite difference method. Each of these previous studies have proposed a model to characterize frequency dependency of intrinsic attenuation or power spectrum density of the random media. Figure 3.15 compares the intrinsic attenuations and power spectrum densities estimated in this study with the models proposed by the previous studies mentioned above. There are two models from Takahashi et al. (2013) and Takahashi et al. (2014) because they have modeled each parameter for two regions (weak and strong) separately. Although analysis methods and frequency ranges are different, the ranges of our estimated attenuation parameters are consistent with the previous studies, implying that the attenuation characteristics of southwestern Japan can be modeled by these previous and our studies.

3.5 Conclusion

In this chapter, we have estimated the 2D and 3D intrinsic and scattering attenuation structures of the southwestern Japan. The 2D attenuation structure has been estimated using the MLTWA under the assumption of isotropic scattering. For estimating the 3D heterogeneous attenuation structure, we have proposed a new method consisting of two steps. The first step estimates intrinsic attenuation and strength of random velocity fluctuation for each earthquake–station pair by the envelope fitting method, and the second maps the above attenuation parameters to 3D space using sensitivity kernels. A depth-dependent velocity structure has been considered by using the Monte Carlo simulation in modeling the seismogram envelopes at each step. We

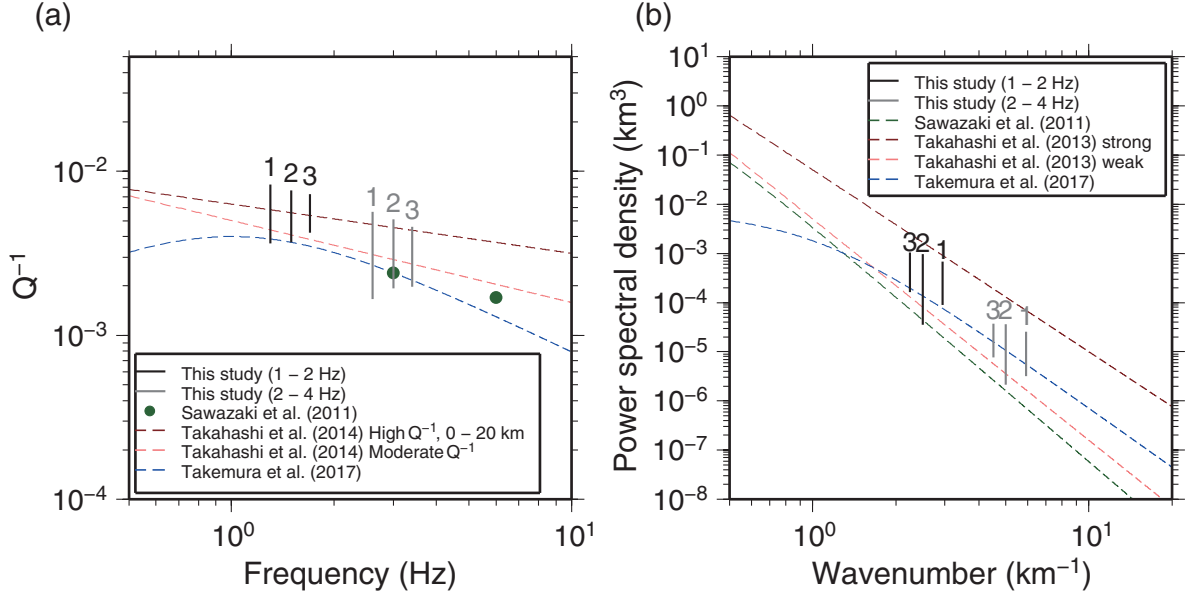


Figure 3.15 Comparison of (a) intrinsic attenuation and (b) power spectral density function of random media between this study and previous studies. Black solid lines denote the maximum and minimum values for each layer using the frequency range of 1-2 Hz while gray solid lines for 2-4 Hz. Numbers on top of the solid lines represent the depth of the layer: 1, 2, and 3 for 0-10, 10-30, and 30-50 km depth, respectively. Circles and broken lines show estimated and modeled intrinsic attenuation and power spectral density function in the previous studies. Note that the intrinsic attenuation of the layer 1 and 3 are plotted with some shifts from the actual frequency (1.5 or 3.0 Hz) to avoid overlapping with those of the layer 2.

have applied the proposed method to estimate the attenuation structure in southwestern Japan. The estimated 3D attenuation structure has been compared with the 2D attenuation structure derived by the MLTWA. The resulted overall feature is in common, that is, strong intrinsic and scattering attenuation in the Kyushu region where tectonic activities are higher than the other regions. In addition, we have found non-negligible depth dependency of intrinsic attenuation and random velocity fluctuation in our 3D structure. Such depth dependency of attenuation parameters is particularly noticeable in the Chugoku region. Furthermore, horizontal variation of intrinsic attenuation has been found at the depth of 10-50 km beneath the Shikoku region where random velocity fluctuation is relatively weak and uniform. These features of horizontal and depth dependency of attenuation parameters cannot be retrieved by any conventional analyses in 2D space. The ranges of values of intrinsic attenuation and

power spectrum density are consistent with previous studies, indicating that the structural parameters of the southwestern Japan can be reliably modeled by the previous studies as well as this study.

4 Numerical shake prediction with heterogeneous attenuation structure

Sections 4.2, 4.3 and 4.4.1 has been published as a part of following paper:

Ogiso, M., M. Hoshiba, A. Shito, and S. Matsumoto, Numerical Shake Prediction for Earthquake Early Warning Incorporating Heterogeneous Attenuation structure: The Case of the 2016 Kumamoto Earthquake, *Bulletin of the Seismological Society of America*, **108** (6), 3457-3468, 2018, ©Seismological Society of America.

4.1 Introduction

Ground motion prediction in most of currently operational EEW systems (e.g., E-larmS: Allen, 2007, PRESTo: Satriano et al., 2011, and the JMA EEW system: Hoshiba & Ozaki, 2014) relies on a ground motion prediction equation (GMPE). In constructing GMPE, a kind of regression analyses is usually conducted. The influence of a complex attenuation structure in the observed ground motion is suppressed through such regression analyses, making it difficult to consider the effect of the heterogeneous attenuation structure in any operational EEW systems.

Hoshiba & Aoki (2015) have proposed the Numerical Shake Prediction scheme, where a ground motion is predicted by the wave propagation simulation instead of GMPE. Because the scheme of Hoshiba & Aoki (2015) adopts the wave propagation simulation for prediction, it is possible to consider heterogeneous attenuation structure in its prediction stage. In this chapter, we extend the prediction scheme of Hoshiba & Aoki (2015) to consider the heterogeneous attenuation structure in the wave simulation propagation, and investigate how effective it is on the ground motion prediction.

4.2 Ground motion prediction scheme: Numerical Shake Prediction

The Numerical Shake Prediction scheme (Hoshiaba & Aoki, 2015) predicts a ground motion with wave propagation simulation from a current distribution of ground motions. Because calculation time is at a premium in EEW, Hoshiaba & Aoki (2015) have adopted the radiative transfer theory in 2D space (e.g., Sato et al., 2012) for their wave propagation simulation, rather than computationally expensive approaches such as finite-difference and boundary integral methods. In the radiative transfer theory, wave propagation is approximated by energy propagation without phase terms. In a 2D medium with a constant velocity V_0 , the space and time evolution of the energy distribution $f(\mathbf{x}, t; \mathbf{q})$ is represented by the following equation (Sato et al., 2012; Hoshiaba & Aoki, 2015):

$$\frac{\partial}{\partial t} + V_0 \mathbf{q} \frac{\partial}{\partial \mathbf{x}} f(\mathbf{x}, t; \mathbf{q}) = - \left\{ g_0(k_0) + k_0 Q_i^{-1} \right\} V_0 f(\mathbf{x}, t; \mathbf{q}) + \frac{V_0}{2\pi} \int_{-\pi}^{\pi} d\theta' g(k_0 \mathbf{q} - k_0 \mathbf{q}') f(\mathbf{x}, t; \mathbf{q}') \quad (4.1)$$

where $\mathbf{x}$ is the location vector, $\mathbf{q}$ and $\mathbf{q}'$ are the direction vectors of energy propagation (propagation directions are θ and θ'), and k_0 is the wavenumber. $g(k_0 \mathbf{q} - k_0 \mathbf{q}')$ is the scattering coefficient for changes in propagation direction from $\mathbf{q}'$ to $\mathbf{q}$, and $g_0(k_0)$ is the total scattering coefficient, that is, $g_0(k_0) = (1/2\pi) \int_{-\pi}^{\pi} g(k_0 \mathbf{q} - k_0 \mathbf{q}') d\theta$, where $\int_{-\pi}^{\pi} d\theta$ is the integral over all the scattering directions. Q_i^{-1} is the inverse of intrinsic attenuation factor. This equation holds for both P- and S-waves; V_0 is either V_P or V_S , P- and S-wave velocities, respectively. Similarly, Q_i^{-1} is either Q_P^{-1} or Q_S^{-1} , the inverse of intrinsic attenuation factors for P- and S-waves. Given the initial distribution of seismic energy $f_{\text{ini}}(\mathbf{x}, t; \mathbf{q})$, we can calculate future energy distribution with Eq. (4.1). We assume no phase conversions between P and S waves during each scattering process.

Same as chapter 3, we adopted the Monte Carlo simulation based on the radiative transfer

theory (e.g., Yoshimoto, 2000) to calculate energy propagation with Eq. (4.1). In the Monte Carlo simulation, energy distribution is discretized by many particles having a certain energy. Energy propagation process is approximated by the movement of each particles. For the prediction variable, we adopt the JMA seismic intensity I_{JMA} (Japan Meteorological Agency, 1996; Hoshiba & Ozaki, 2014) as an index of the strength of ground motion. As shown in Eq. (2.1), the JMA seismic intensity is related to the logarithm of seismic energy. The radiative transfer theory describes the energy transport of the wavefield so that our results can easily be translated into I_{JMA} .

The wave propagation starts from the current energy distribution, or initial condition, $f_{\text{ini}}(\mathbf{x}, t; \mathbf{q})$, which can be estimated from the observations by data assimilation technique. Because the energy propagation direction θ is difficult to estimate in practice without special equipment such as seismic arrays, we instead estimate the energy distribution without the propagation direction, $F_{\text{ini}}(\mathbf{x}, t)$, as calculated from $f_{\text{ini}}(\mathbf{x}, t; \mathbf{q})$ by:

$$F_{\text{ini}}(\mathbf{x}, t) = \int_{-\pi}^{\pi} d\theta f_{\text{ini}}(\mathbf{x}, t; \mathbf{q}). \quad (4.2)$$

Following the technical terms of the numerical weather prediction, $f_{\text{ini}}(\mathbf{x}, t; \mathbf{q})$ and $F_{\text{ini}}(\mathbf{x}, t)$ are also called as an analysis (e.g., Kalnay, 2003).

To reduce our computational cost, we adopt a simple data assimilation technique called the optimal interpolation (e.g., Kalnay, 2003), to estimate $F_{\text{ini}}(\mathbf{x}, t)$ in the analysis step. The analysis can be represented by the combination of the energy distribution predicted at time t from one time step earlier, $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$, and the observed energy distribution at time t , $F_{\text{obs}}(\mathbf{x}_{\text{obs}}, t)$ as follows:

$$F_{\text{ini}}(\mathbf{x}, t) = F_{\text{pred}}(t - \Delta t; \mathbf{x}, t) + \mathbf{B}\mathbf{H}^T \left(\mathbf{R} + \mathbf{H}\mathbf{B}\mathbf{H}^T \right)^{-1} \left\{ F_{\text{obs}}(\mathbf{x}_{\text{obs}}, t) - \mathbf{H}F_{\text{pred}}(t - \Delta t; \mathbf{x}, t) \right\}, \quad (4.3)$$

where $\mathbf{B}$ is the background error covariance matrix, $\mathbf{R}$ the observational error covariance

matrix, and $\mathbf{H}$ the observation matrix that connects the model space $\mathbf{x}$ and observation space $\mathbf{x}_{\text{obs}}$ (e.g., Kalnay, 2003). Here we assume n observation stations, and the model space $\mathbf{x}$ is discretized into m grids. In this study, we define each component of matrix $\mathbf{H}$, h_{ig} ($i = 1, 2, \dots, n; g = 1, 2, \dots, m$) as

$$h_{ig} = \frac{1}{d_i} \exp\left(-\frac{r_{ig}^2}{a^2}\right). \quad (4.4)$$

in which r_{ig} is the distance between the i -th station and the g -th grid, a is the correlation distance, and d_i is the normalization coefficient defined as,

$$d_i = \sum_{g=1}^m \exp\left(-\frac{r_{ig}^2}{a^2}\right). \quad (4.5)$$

The size of $\mathbf{H}$ becomes $n \times m$. The observation matrix $\mathbf{H}$ works as the projection operator of the model space $\mathbf{x}$ into observational space $\mathbf{x}_{\text{obs}}$. The $m \times n$ matrix $\mathbf{BH}^T$ denotes the background error covariance matrix between the g -th grid in the model space $\mathbf{x}$ and the i -th station in the observational space $\mathbf{x}_{\text{obs}}$. The (gi) -th element of $\mathbf{BH}^T$, b_{gi} , is represented by:

$$b_{gi} = \mu_{gi}^{\text{bkg}} \sigma_g^{\text{bkg}} \sigma_i^{\text{bkg}}, \quad (4.6)$$

where μ_{gi}^{bkg} is the correlation of background error between the g -th grid and the i -th station, σ_g^{bkg} and σ_i^{bkg} are the standard deviation of background error at the g -th grid and the i -th station, respectively. The $n \times n$ matrix $\mathbf{HBH}^T$ denotes the projection of the model space onto the observational space. The (ij) -th ($i, j = 1, 2, \dots, n$) element of $\mathbf{HBH}^T$, b_{ij}^{H} , is represented by:

$$b_{ij}^{\text{H}} = \mu_{ij}^{\text{bkg}} \sigma_i^{\text{bkg}} \sigma_j^{\text{bkg}}. \quad (4.7)$$

Similarly, the observational error covariance matrix $\mathbf{R}$ has $n \times n$ elements with its (ij) -th

element R_{ij}^{obs} is represented by:

$$R_{ij}^{\text{obs}} = \mu_{ij}^{\text{obs}} \sigma_i^{\text{obs}} \sigma_j^{\text{obs}}, \quad (4.8)$$

where μ_{ij}^{obs} is the correlation of the observational error between the i -th and j -th stations, σ_i^{obs} and σ_j^{obs} are the standard deviation of observational error at stations i and j , respectively. By introducing the $m \times n$ matrix $\mathbf{W}$ defined by $\mathbf{W} = \mathbf{B}\mathbf{H}^T(\mathbf{R} + \mathbf{H}\mathbf{B}\mathbf{H}^T)^{-1}$, its (gj) -th element w_{gj} satisfies:

$$\sum_{j=1}^N w_{gj} (R_{ij}^{\text{obs}} + b_{ij}^{\text{H}}) = b_{gi}. \quad (4.9)$$

Substituting Eqs. (4.6-4.8) to (4.9) and assuming that the standard deviations of both background and observational error do not depend on the locations, i.e., $\sigma_g^{\text{bkg}} = \sigma_i^{\text{bkg}} = \sigma_j^{\text{bkg}} = \sigma^{\text{bkg}}$ and $\sigma_i^{\text{obs}} = \sigma_j^{\text{obs}} = \sigma^{\text{obs}}$, the following system of equations for w_{gj} are obtained:

$$\sum_{j=1}^n w_{gj} \left\{ \mu_{ij}^{\text{obs}} \left(\frac{\sigma^{\text{obs}}}{\sigma^{\text{bkg}}} \right)^2 + \mu_{ij}^{\text{bkg}} \right\} = \mu_{gi}^{\text{bkg}}. \quad (4.10)$$

μ_{ij}^{obs} , μ_{ij}^{bkg} , and $\sigma^{\text{obs}}/\sigma^{\text{bkg}}$ determine the behavior of the optimal interpolation.

In every time step, we calculate the predicted energy distribution at time t , $f_{\text{pred}}(t - \Delta t; \mathbf{x}, t; \mathbf{q})$ and its integration over all the propagation direction $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$, with the Monte Carlo simulation following Eq. (4.1). Then, we calculate $F_{\text{ini}}(\mathbf{x}, t)$ from $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$ and $F_{\text{obs}}(\mathbf{x}_{\text{obs}}, t)$ using Eq. (4.3). If $F_{\text{ini}}(\mathbf{x}, t)$ is larger than $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$ at certain grid cell $\mathbf{x}$, we add new energy there. To do so, we assign more particle to that particular cell and adjust the energy of each particle such that the sum of the energy in the cell equals $F_{\text{ini}}(\mathbf{x}, t)$. The propagation direction of each new particle is set to follow the distribution of propagation directions in the predicted energy distribution $f_{\text{pred}}(t - \Delta t; \mathbf{x}, t; \mathbf{q})$. If $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$ equals zero, that is, there is no observed seismic waveform at time $t - \Delta t$ on grid cell $\mathbf{x}$, the propagation direction of the new particles are assigned based on a uniform random number

between $-\pi$ and π . The wave type (P or S) of each new particle is also set to follow the energy ratio of P- and S-waves in $f_{\text{pred}}(t - \Delta t; \mathbf{x}, t; \mathbf{q})$. If $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$ equals zero, the wave type of the particle is assigned such that S-wave energies are $3/2(V_p/V_s)^5$ times larger than those of P-waves (Eq. 6.10 of Sato et al., 2012) in the corresponding cell. If $F_{\text{ini}}(\mathbf{x}, t)$ is smaller than $F_{\text{pred}}(t - \Delta t; \mathbf{x}, t)$, the energy at cell $\mathbf{x}$ is adjusted to $F_{\text{ini}}(\mathbf{x}, t)$ by randomly removing particles from the cell. By combining $F_{\text{ini}}(\mathbf{x}, t)$ and energy propagation direction of $f_{\text{pred}}(t - \Delta t; \mathbf{x}, t; \mathbf{q})$, we can estimate the initial condition at time t , or $f_{\text{ini}}(\mathbf{x}, t; \mathbf{q})$, to calculate the predicted energy distribution based on Eq. (4.1). Because all the seismic stations are at the surface (or at very shallow depths compared to the hypocentral depth), the analysis represents the 2D distribution of ground motions, and we conduct the wave propagation simulation in 2D space. The difference between geometrical spreading of 2D and 3D space may result in overprediction in our 2D computation. To account for this effect, energy decay for each particle is set to be proportional to $r^{-1.5}$, where r is the travel distance of each particle from its origin in the wave propagation simulation. The Numerical Shake Prediction scheme is schematically shown in Fig. 4.1. Table 4.1 lists the parameters used for the simulation of ground motion predictions, including those to construct matrices $\mathbf{W}$ and $\mathbf{H}$.

The observed waveform retains the complex effects of source and path terms, therefore the estimated initial condition (Eq. 4.3) also contain their effects: however, adopting a homogeneous attenuation structure in the Numerical Shake Prediction scheme precludes a full consideration of the effect of heterogeneous attenuation structures. To consider the heterogeneous attenuation structure, the inverse of the intrinsic attenuation Q_i^{-1} and scattering coefficient $g(k_0\mathbf{q} - k_0\mathbf{q}')$ in Eq. (4.1) should depend on the location vector $\mathbf{x}$:

$$\begin{aligned} Q_i^{-1} &\longrightarrow Q_i^{-1}(\mathbf{x}), \\ g(k_0\mathbf{q} - k_0\mathbf{q}') &\longrightarrow g(k_0\mathbf{q} - k_0\mathbf{q}'; \mathbf{x}). \end{aligned} \tag{4.11}$$

Note that the medium velocity V_0 also depends on the location, however, we assumed the

V_0 to be constant in this study. Not only intrinsic and scattering attenuation but also the heterogeneous velocity structure affects the observed amplitude due to the focusing and defocusing effect. However, we assumed that the velocity was horizontally homogeneous in estimating the attenuation structure in chapter 3. Hence, estimated heterogeneous attenuation structure in chapter 3 has already contained the influence of amplitude fluctuation due to the variation of velocity structure.

4.3 Target earthquake, data and preprocessing for ground motion prediction experiments

In this study, we employed the mainshock (M_w 7.0) of the 2016 Kumamoto earthquake which began at 01:25:05 on 16 April 2016 (in Japan Standard Time: UTC + 9 hours). This event triggered a M 5.7 event at 32 s after the origin time (e.g., Miyazawa, 2016; Nakamura & Aoi, 2017; Suzuki et al., 2017). The Numerical Shake Prediction scheme (Hoshiba & Aoki, 2015; section 4.2) is effective for earthquakes that occurred simultaneously or very close in time, and the 2016 Kumamoto earthquake sequence is a good example to demonstrate that capability. Figure 4.2 shows the epicenter and focal mechanism of the target event as well as the K-NET and KiK-net station distributions (Okada et al., 2004; Aoi et al., 2004) used for the present Numerical Shake Prediction exercise. The target area of the prediction in this study includes almost whole southwestern Japan. We applied the site-correct filters (Ogiso et al., 2016; chapter 2) to the observed accelerograms in a real-time manner to produce waveforms in which the site amplification term at each station referenced that of an arbitrary reference station; we selected station IBR011 (K-NET Tsukuba), about 50 km northeast of Tokyo as the reference station. Then, we calculated time series of the JMA seismic intensity (Japan Meteorological Agency, 1996; Hoshiba & Ozaki, 2014) in time domain at each station using Kunugi et al. (2013)’s technique. We adopted these site-corrected time series of seismic intensity as the observations in this study.

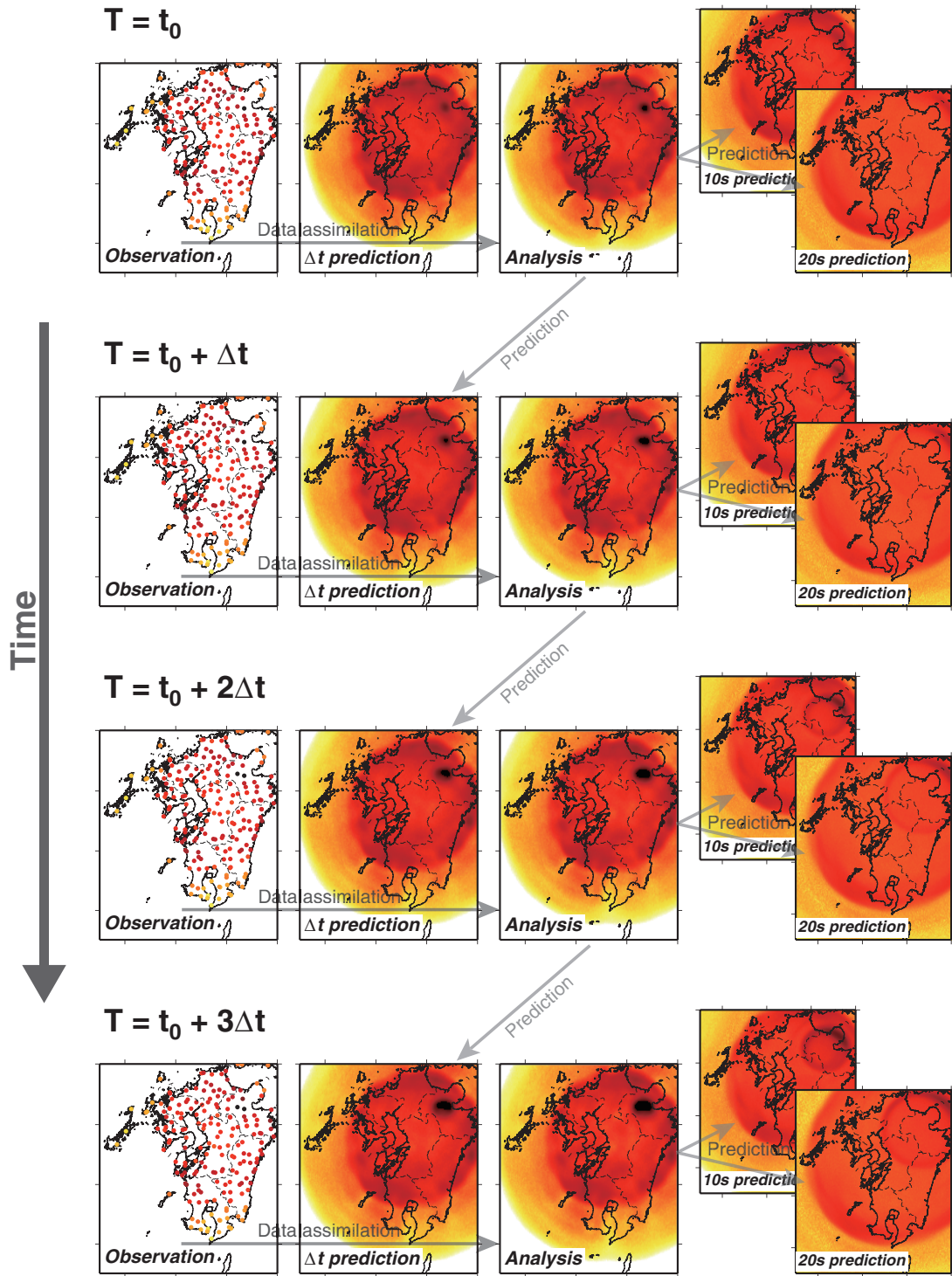


Figure 4.1 Schematic illustration of the Numerical Shake Prediction scheme. In each time step, observation and prediction of corresponding time is combined by the data assimilation to make analysis. The analysis is then used as the initial condition for the wave propagation simulation to predict future ground motion.

Table 4.1 Parameters for the Numerical Shake Prediction.

Parameter description	Value
Grid interval (longitudinal and latitudinal directions)	0.02°
Time interval for data assimilation	1 s
S-wave velocity (V_S)	3.5 km/s
V_P/V_S ratio	1.73
Correlation distance of the background error (observation matrix and optimal interpolation): a	10 km
Correlation type of the background error (optimal interpolation): μ_{ij}^{bkg}	$\exp(-r_{ij}^2/a^2)$ r_{ij} is the distance between the i -th grid (or station) and the j -th station
Correlation of the observational error between i -th and j -th stations (optimal interpolation): μ_{ij}^{obs}	0 if $i \neq j$, otherwise 1
Ratio of observational and background error (optimal interpolation): $\sigma^{\text{obs}}/\sigma^{\text{bkg}}$	1.0
Number of the particles used in the simulation	8,000,000

4.4 Results of the Numerical Shake Prediction with a heterogeneous attenuation structure

4.4.1 Isotropic scattering case: the attenuation structure derived from the MLTWA

We first show results using the heterogeneous structure estimated in section 3.2 under the assumption of isotropic scattering. The adopted heterogeneous structure is shown in the right panels of Fig. 3.3. We compare results with homogeneous structure of $Q_1^{-1} = 0.0047$ and $g_0 = 0.0067 \text{ km}^{-1}$ for S-waves. These values are averaged values over all the stations used in the MLTWA, and are also used for areas where no stations exist around a particular grid cell in the case of heterogeneous structure. Q_1^{-1} for P-waves is set to be a half of that for S-waves, and g_0 for P-waves the same as that for S-waves.

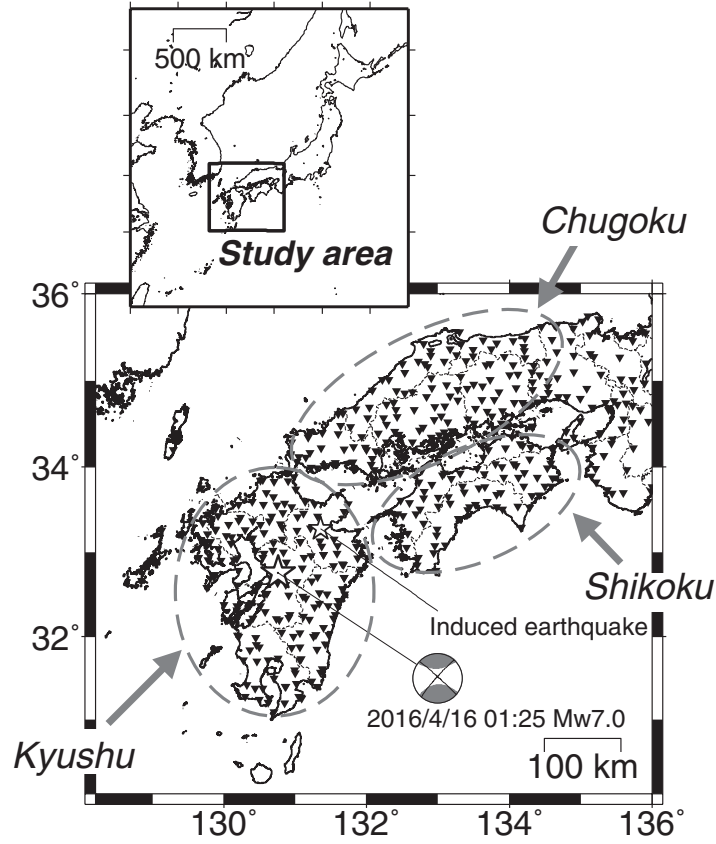


Figure 4.2 Map of the ground motion prediction experiments. The largest star is the epicenter of the largest earthquake (M_w 7.0) of the 2016 Kumamoto earthquake sequence (plotted with the JMA centroid moment tensor solution); the small star is the epicenter of the M 5.7 induced earthquake. Inverted triangles mark strong-motion stations (K-NET and KiK-net) used for the ground motion prediction simulation. The Chugoku, Shikoku, and Kyushu regions are outlined by gray dashed ovals.

Figure 4.3 shows snapshots of the Numerical Shake Prediction simulation for the homogeneous and heterogeneous structures. Because the observational data are common in both structure cases, the analyses are similar to each other in Fig. 4.3. At a glance, there is no obvious difference in the distributions of the 10 s prediction and residual in Fig. 4.3. One reason for this is that the focal area of the target event was in central Kyushu, an area of strong intrinsic and scattering attenuation. We expect that differences between the two structures in the predicted wavefield and residual distribution would be obvious in this strongly heterogeneous area. However, because central Kyushu is also the focal region of the Kumamoto

earthquake sequence, differences between the homogeneous and heterogeneous structures are masked by underprediction due to the small hypocentral distance.

Although the distributions of the residuals are similar for both structure cases, the values of the residuals (shown by the depth of color in Fig. 4.3), appear to be slightly smaller in the heterogeneous structure case. Figure 4.4 compares the observed and predicted time series of seismic intensities using both structure cases. The heterogeneous structure provides a smaller prediction residuals than the homogeneous structure at stations OITH07 and MYZH12 (Fig. 4.4b, c). At station YMGH10 (Fig. 4.4d), discrepancies in the attenuation structure around the station are small, resulting in almost the same prediction for both structure cases. Both cases underpredict the maximum intensity at station OITH11 (Fig. 4.4e), since the induced earthquake (small star of Fig. 4.4a) produces the maximum intensity there.

Because an important aspect of EEW is the precision of the predicted maximum ground motion, we also compare the maximum values of the predicted and observed seismic intensities. Figure 4.5 shows the residuals at each station for predictions 10 s and 20 s ahead, as well as the RMS values of prediction residuals from a selected set of stations: Figure 4.6 shows the histograms of the residuals. These residuals are plotted versus hypocentral distance for the 10 s and 20 s predictions in Figs. 4.7 and 4.8, respectively. When calculating RMS of residuals, we excluded stations for which intensities were unavoidably underpredicted (i.e., at short and long hypocentral distances): these stations are shown as small circles of Fig. 4.5 and are shaded in gray in Figs. 4.7 and 4.8. Stations close to the epicenter of the target earthquake (within 50 km for the 10 s and 70 km for the 20 s prediction) were excluded because the wavefront of the predicted direct S-wave had already passed by the prediction time. The predicted seismic intensity at these stations was determined by the coda excited by the predicted direct S-wave. Because coda wave energy is spatially homogeneous (e.g., Sato et al., 2012), the predicted intensities at these stations are almost identical, and smaller than the observed intensities. This also explains the underpredicted maximum intensity at station

OITH11 (Fig. 4.4e). OITH11 is about 13 km west of the epicenter of the induced earthquake, too short an epicentral distance for a 10 s ahead prediction. Stations at hypocentral distances greater than about 300 km were prone to underpredict because of the effect of *Lg* waves. We assumed that the wave packets propagated as body waves, and the estimated attenuation structure (Fig. 3.3) was also based on body wave propagation. As shown in Furumura & Kennett (2001), *Lg* waves are dominated at hypocentral distance greater than about 300 km in this region. The geometrical spreading factors are different between body and *Lg* waves, and both the intrinsic and scattering attenuation structures estimated in this study (Fig. 3.3) may not be appropriate for *Lg* waves.

In the 10 s prediction for a homogeneous structure (left panel, Fig. 4.5a), overpredicted stations (shades of red) are obvious in most of Kyushu. In addition, the residual versus distance plot in Fig. 4.7b clearly shows that more stations have negative residuals than positive residuals at distances less than about 150 km. The same is true in the heterogeneous structure case, but the residuals are close to zero for most stations (right panel, Fig. 4.5b). This improvement in predicted intensity is especially evident at stations in the area of strong intrinsic and scattering attenuation south of the epicenter (Fig. 3.3: shown in cyan in Fig. 4.7c). Residuals of the two stations (one was OITH11) located to the northeast of the target earthquake become large when using the heterogeneous structure (Fig. 4.5a). As shown in Fig. 4.4 (e), the observed maximum intensity at those stations is caused by the induced earthquake whereas the 10 s or 20 s prediction was performed solely for the target earthquake. Comparing the observed intensity from the target earthquake and our predictions, the predictions in both structure cases are acceptable. By inference from Fig. 3 of Suzuki et al. (2017), the influence of the induced earthquake on the observed maximum intensity has been expected at only two or three stations. Hence, we do not consider the influence of the induced earthquake in this RMS calculation.

The RMS of the 10 s prediction residuals is 0.34 for the homogeneous structure and 0.30

for the heterogeneous structure, a relative improvement of about 12% although 0.34 is a small RMS residual with little room for improvement. The standard deviation of prediction residuals become small when we adopt the heterogeneous structure (Fig. 4.6a). In the 20 s prediction with the homogeneous structure, stations about 100 km from the hypocenter are obviously overpredicted (Figs. 4.5b and 4.7b). Residuals at these stations are clearly improved under the heterogeneous structure (Figs. 4.5b and 4.7c); the RMS residuals are decreased from 0.45 for the homogeneous structure to 0.38 for the heterogeneous structure. Although the RMS of residuals is larger for the 20 s prediction than the 10 s prediction, the relative improvement was about 15%, greater than that for the 10 s prediction case. For 5 s predictions (figures not presented), the relative improvement is about 7%.

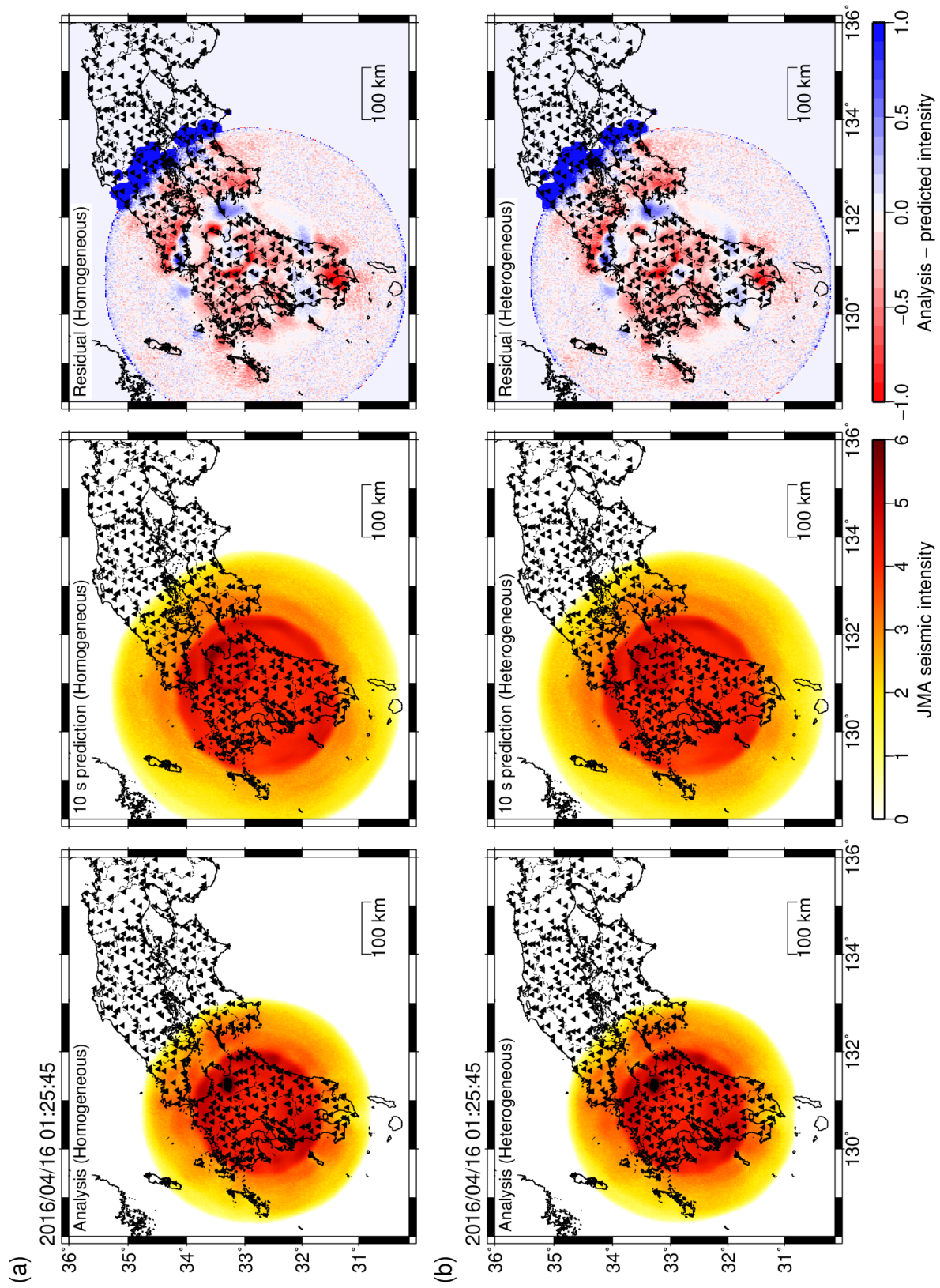


Figure 4.3 Snapshots of the Numerical Shake Prediction simulation for (a) homogeneous and (b) heterogeneous attenuation structures. From left to right are shown the analysis (estimated current wavefield) at 30 s after the origin, the prediction for 10 s later (40 s) and the residual (the analysis at 40 s minus the prediction for 30 s).

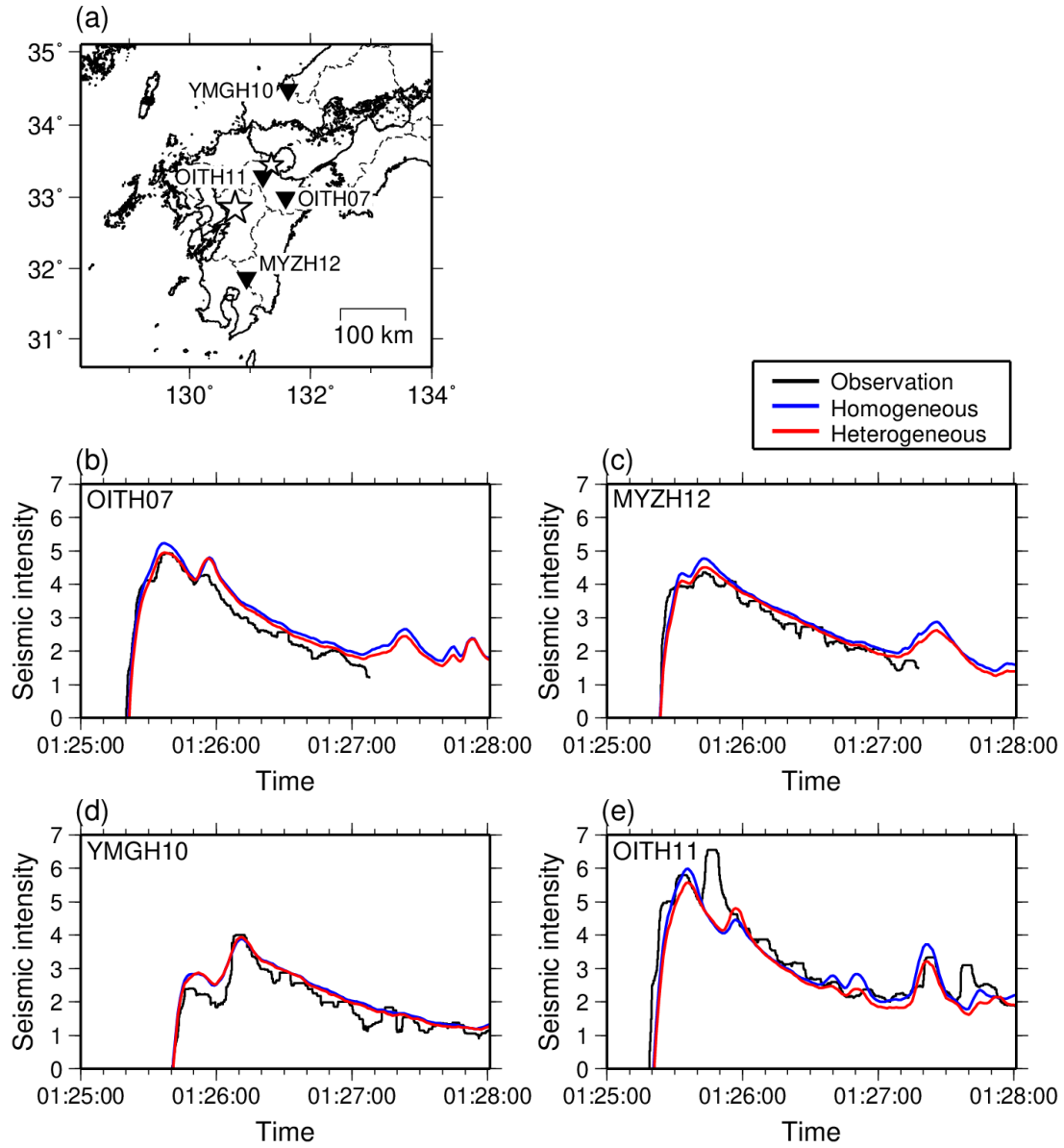


Figure 4.4 (a) Map of four stations (inverted triangles) and the epicenters of the target earthquake (large star) and induced earthquake (small star). Plots of the observed time series of seismic intensity (black lines), and 10 s prediction using homogeneous and heterogeneous structures (blue and red lines, respectively) for stations (b) OITH07, (c) MYZH12, (d) YMGH10, and (e) OITH11. Note that predicted envelopes are shifted by 10 s to overlap the observed time series.

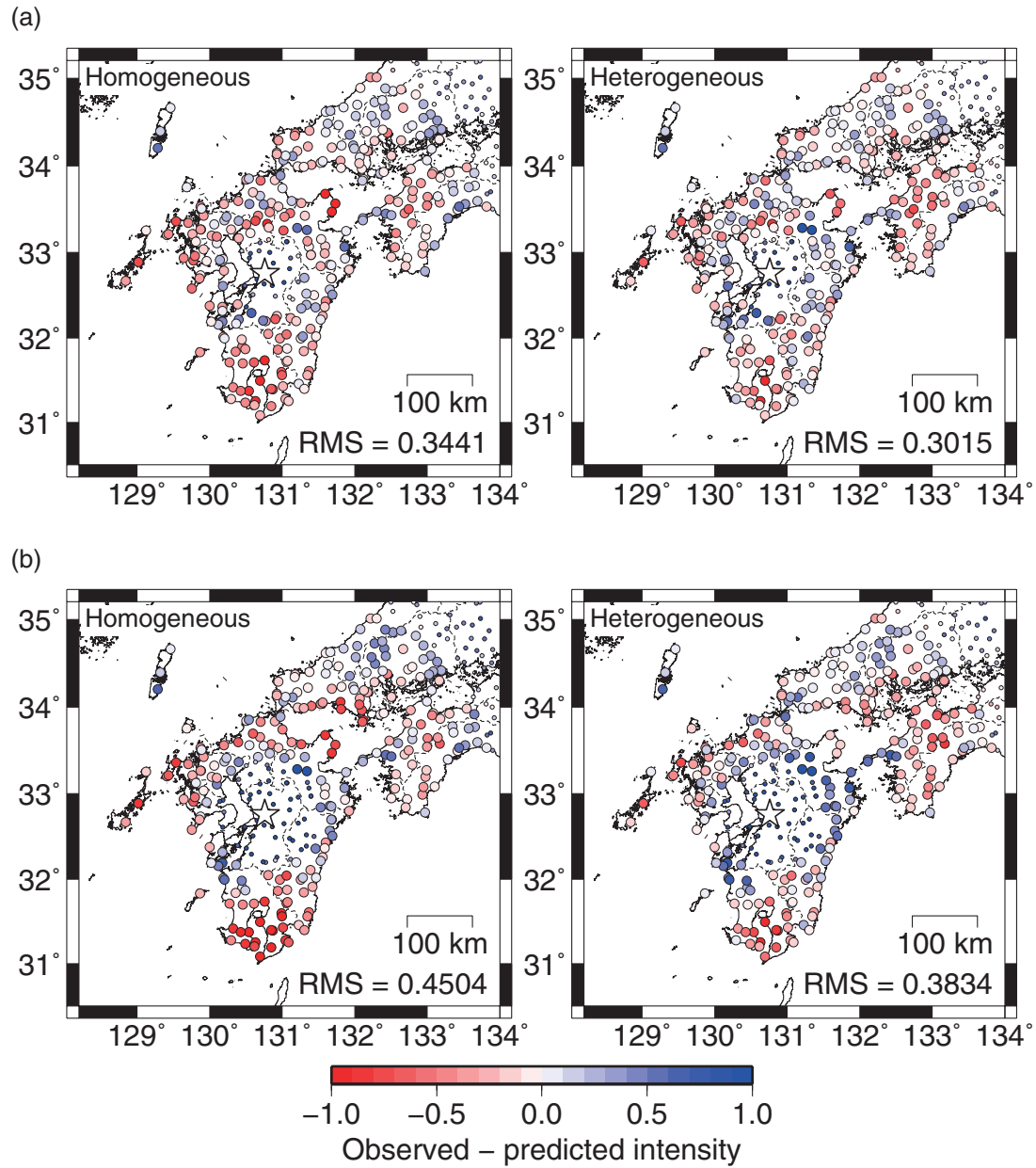


Figure 4.5 Distributions of intensity prediction residuals for the (a) 10 s and (b) 20 s prediction using homogeneous (left) and heterogeneous (right) attenuation structures. Shades of red (negative residuals) represent overpredictions and shades of blue (positive residuals) underpredictions. The RMS of residuals is shown in the lower right corner of each panel. The star is the epicenter of the target earthquake. Stations shown as small circles were excluded from RMS calculations of prediction residuals.

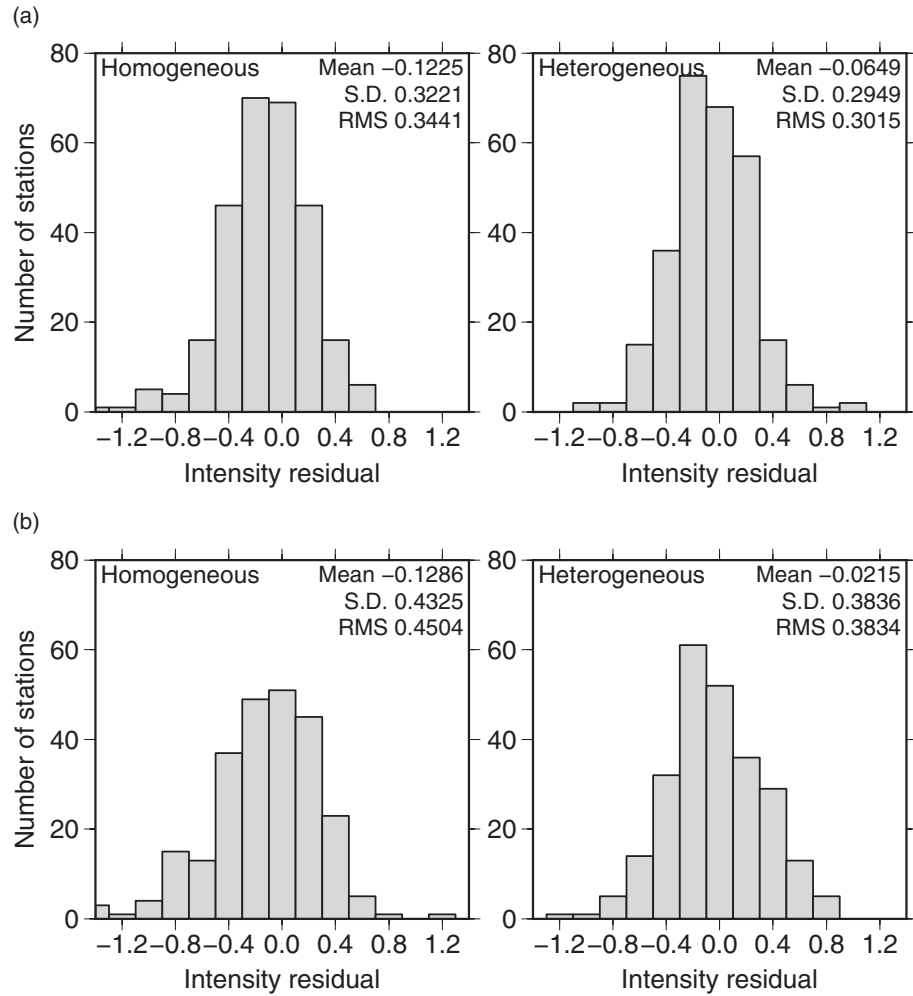


Figure 4.6 Histograms of intensity residuals for the (a) 10 s and (b) 20 s predictions using homogeneous (left) and heterogeneous (right) attenuation structures. Negative and positive residuals represent over- and underpredictions, respectively.

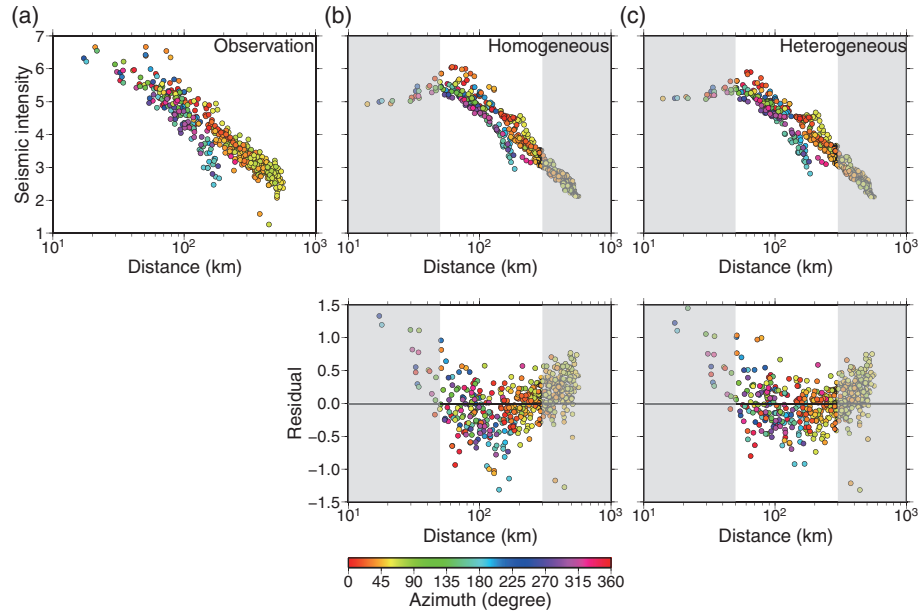


Figure 4.7 Plots of observed seismic intensities, seismic intensities predicted 10 s ahead, and their residuals versus hypocentral distance. Symbol colors represent the station azimuth from the epicenter. (a) Observed seismic intensities, (b) predicted seismic intensities (upper panel) and residuals (lower panel) from a simulation using a homogeneous attenuation structure, and (c) predicted seismic intensities and residuals from a simulation using a heterogeneous attenuation structure. Shaded areas represent hypocentral distances excluded from the calculation of RMS residuals.

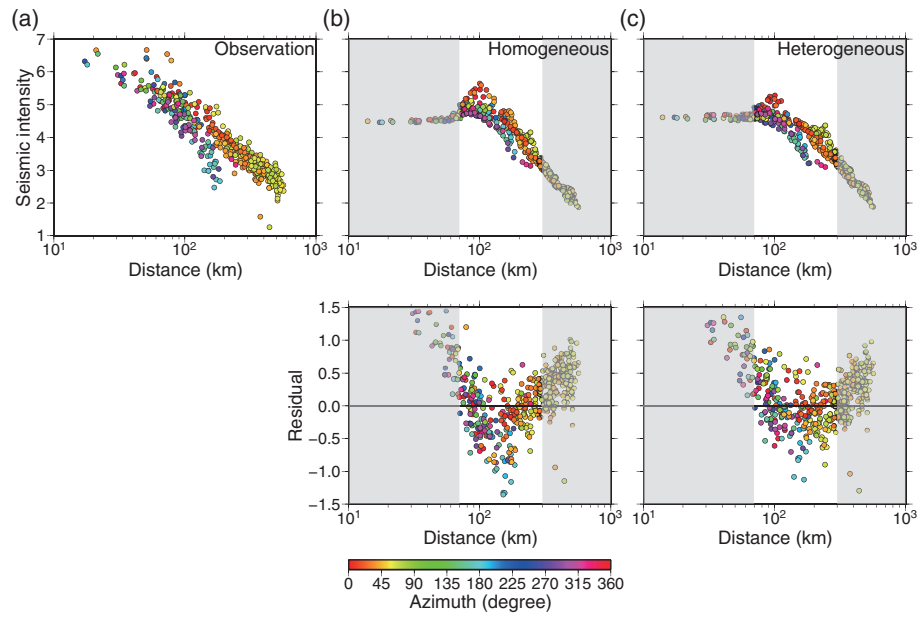


Figure 4.8 Plots of observed seismic intensities, seismic intensities predicted 20 s ahead, and their residuals versus hypocentral distance. See Fig. 4.7 for details.

4.4.2 Non-isotropic scattering case: the attenuation structure estimated in this study

Introducing non-isotropic scattering coefficient in Eq. (4.1) enables us to treat non-isotropic scattering effect in the Numerical Shake Prediction scheme. A non-isotropic scattering coefficient can be calculated based on the Born approximation. According to Sato et al. (2012), the power spectral density of the 2D exponential-type isotropically random media is given by

$$P_{2D}(m) = \frac{2\pi\varepsilon^3 a_r^2}{(1 + a_r^2 m^2)^{3/2}} \quad (4.12)$$

where m is the wavenumber, ε is the RMS of random fluctuation of medium velocity and a_r is the correlation distance of randomness. The wavenumber of incident wave is denoted by k_0 , then, similar to the 3D case described in section 3.3.1, the non-isotropic scattering coefficient $g(\theta; k_0)$ for the scalar wave in 2D media is given by

$$g(\theta; k_0) = k_0^3 P_{2D} \left(2k_0 \sin \frac{\theta}{2} \right), \quad (4.13)$$

where θ is the scattering angle measured from the direction of incident wave (Wegler et al., 2006).

We first assumed a homogeneous structure as $Q_i^{-1} = 0.005$, $a_r = 1$ (km), and $\varepsilon = 0.05$, in the frequency range of 1-2 Hz (Fig. 3.5a), and calculated the non-isotropic scattering coefficient (Eq. 4.13) to introduce the non-isotropic scattering effect to our Numerical Shake Prediction scheme. Other parameters for the scheme were same as listed in Table 4.1. Figure 4.9 shows the distribution of maximum intensity prediction residuals for 10 s and 20 s predictions using isotropic and non-isotropic scattering models. The attenuation structure is homogeneous for each scattering model: hence, the isotropic scattering model in Fig. 4.9 is the same as the homogeneous structure case in Fig. 4.5. RMS of maximum intensity

prediction residuals is almost same or slightly large if the non-isotropic scattering model is adopted. This result indicates that there would be more appropriate attenuation parameters for the non-isotropic scattering model. Attenuation parameters of $(Q_i^{-1}, \varepsilon) = (0.005, 0.05)$ used for the homogeneous structure include the effect of 3D wave propagation, that is, a depth-dependent velocity structure, so that these values would not be appropriate in the 2D wave propagation model.

Non-isotropic scattering model based on the scattering coefficient derived by the Born approximation can model full seismogram envelopes (Przybilla et al., 2006; Wegler et al., 2006). To confirm such a feature, we compare the observed time series of real time seismic intensities (i.e., observed seismogram envelopes) with the predicted ones. We calculate residuals of the predicted envelopes at the time when both observed and predicted seismic intensities are larger than 0.0, then RMS of them at each station. Hereafter we call this RMS as the RMS residual of whole envelope. Figure 4.10 shows the distribution of RMS residuals of whole envelope for 10 s and 20 s predictions. So far as this case, isotropic and non-isotropic scattering models yield nearly the same predicted envelopes. The correlation distance a used in the optimal interpolation (Table 4.1) determines how much the optimal interpolation broadens the wavefront of the direct waves in space. The longer the correlation distance becomes, the wider the area observation at each station affects around the station. Hence, the wavefront in our analysis would be strongly broadened in space if a large value of correlation distances were adopted for the optimal interpolation. This process works as the equivalent to the envelope broadening in time domain caused by strong forward scattering.

Next, to investigate the heterogeneous attenuation structure in the Numerical Shake Prediction scheme with non-isotropic scattering model, we adopted attenuation parameters of the layer 2 using the frequency range of 1-2 Hz estimated in section 3.3 (Fig. 3.11). As mentioned before, the estimated heterogeneous attenuation structure in section 3.3 was in a 3D domain. To utilize this 3D attenuation structure for present 2D wave propagation calculation, the atten-

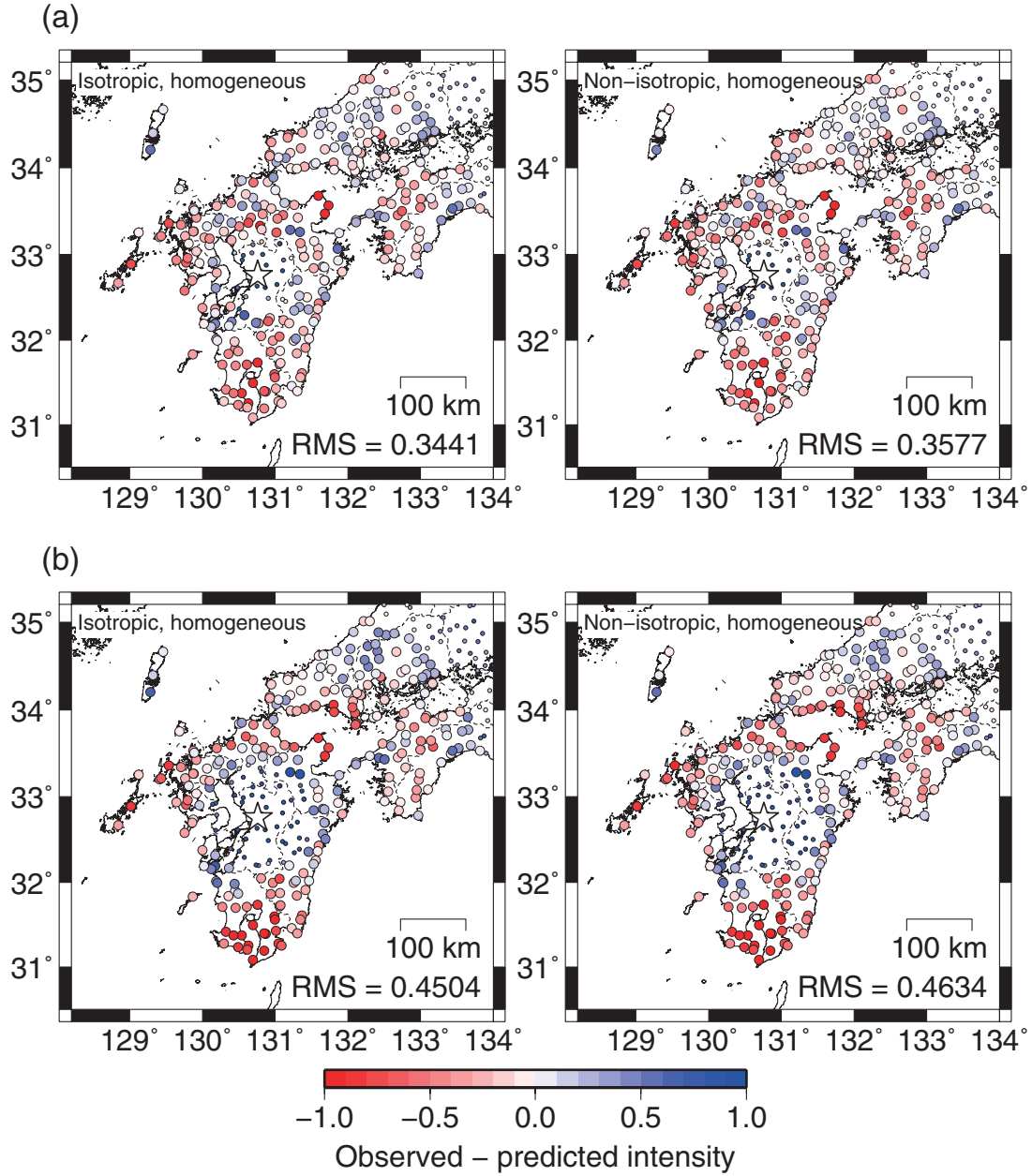


Figure 4.9 Same as Fig. 4.5 but for using isotropic (left) and non-isotropic (right) scattering model.

uation structure should be transformed from 3D to 2D. Making an appropriate transformation for all the earthquakes would be difficult because the influence of each structural layer to the observed amplitude of seismogram depends on the wave propagation path, i.e., hypocenter and station locations. The purpose of this section is to investigate the effect of heterogeneous attenuation structure in the Numerical Shake Prediction scheme. On the premise that there

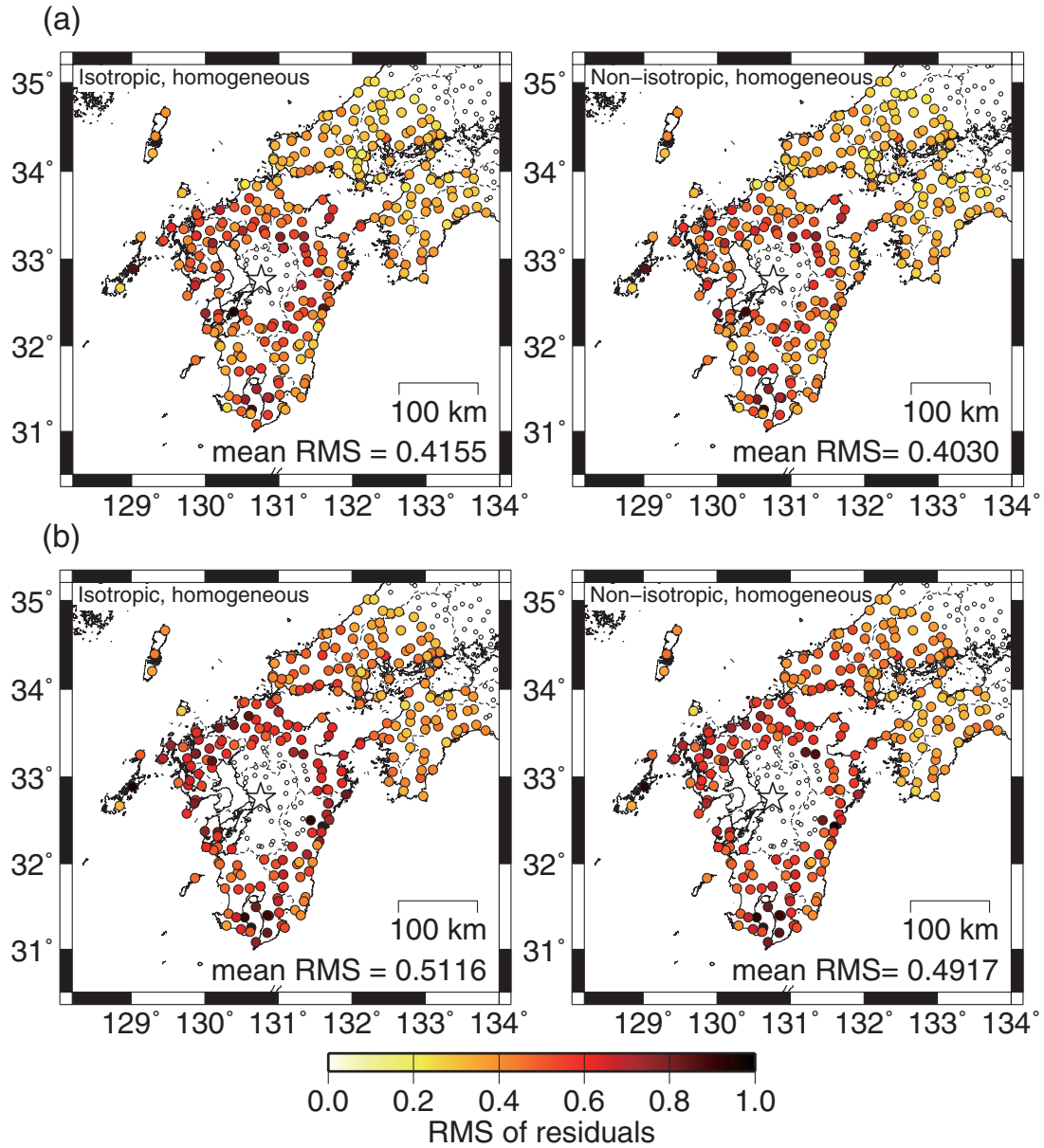


Figure 4.10 Distribution of RMS residuals of whole envelope for (a) 10 s and (b) 20 s ahead prediction using homogeneous attenuation structure. In (a) and (b), prediction based on isotropic scattering model was on the left panel while based on non-isotropic scattering model was on the right panel. The star is the epicenter of the target earthquake. Stations plotted in small circles were excluded in calculating a mean value of RMS which was plotted in right lower of the each panel.

would be more appropriate 2D attenuation structure for non-isotropic scattering model, we adopted the attenuation structure of layer 2 of Fig. 3.11 as the heterogeneous structure in this experiment, and compared the predictions of the heterogeneous structure with those of homogeneous one.

Figures 4.11 and 4.12 show the distributions of prediction residuals for maximum seismic intensity and RMS residuals of whole envelope, respectively. In the 10 s prediction case, relative improvement is about 10% for both RMS residuals of maximum intensity prediction and mean RMS residuals of whole envelope when we use the heterogeneous attenuation structure (Figs. 4.11a and 4.12a). In the 20 s prediction case, the relative improvement of RMS is about 9% for maximum intensity prediction and 11% for mean RMS residuals of whole envelope, respectively (Figs. 4.11b and 4.12b). These results are slightly different from those based on isotropic scattering model (section 4.4.1), that is, almost the same relative improvement for both 10 s and 20 s predictions. Nevertheless, prediction residuals of maximum intensity has been improved when we adopt the heterogeneous attenuation structure in non-isotropic scattering model. Figures 4.14 and 4.15 show the histograms of RMS residuals of whole envelope for 10 s and 20 s predictions based on four prediction models. Comparing the homogeneous and heterogeneous structure cases (a and b of Figs. 4.14 and 4.15 for the isotropic scattering case, or c and d for for non-isotropic scattering case), the frequency peak of each histogram is slightly shifted to left with the heterogeneous structure. Focusing on the differences between the isotropic and non-isotropic scattering models with heterogeneous attenuation structure (b and d of Figs. 4.14 and 4.15), we can see that the number of stations with RMS values larger than 0.5 has been reduced in case of the non-isotropic scattering model than the isotropic scattering. Although the heterogeneous attenuation structure with the isotropic scattering model yields the smallest RMS of residuals for maximum intensity predictions (Figs. 4.5 and 4.11), the heterogeneous attenuation structure with non-isotropic scattering model results in the smallest RMS residuals of whole envelope (Figs. 4.10, 4.12

and 4.13) in both 10 s and 20 s predictions among four prediction models. Figures 4.16 and 4.17 show the plots of the observed versus four predicted envelopes (with the homogeneous or heterogeneous attenuation structure and the isotropic or non-isotropic scattering model) of 10 s and 20 s predictions at four stations, respectively. In both 10 s and 20 s predictions, maximum intensities are almost the same in these four prediction models. Focused on the coda part, we can see that the predictions based on the non-isotropic scattering model with heterogeneous attenuation structure yields the closest prediction to the observation among the four models. We conclude that the non-isotropic scattering model is appropriate in wave propagation simulations, and that the non-isotropic scattering model has potential to reproduce the whole seismogram envelope with any appropriate attenuation structure, as already shown by Przybilla et al. (2006) and Wegler et al. (2006).

4.5 Discussion and conclusion

In this chapter, we first extended the Numerical Shake Prediction scheme to incorporate the heterogeneous attenuation structures estimated in chapter 3. Next, we conducted ground motion prediction simulations, using the extended Numerical Shake Prediction scheme for the largest event of the 2016 Kumamoto earthquake sequence to investigate the effect of a heterogeneous attenuation structure to the prediction accuracy of seismic intensity. The simulations were conducted for four prediction models: homogeneous or heterogeneous attenuation structure based on either isotropic or non-isotropic scattering model. In the isotropic scattering model, we adopted the 2D heterogeneous structure retrieved by the MLTWA (section 3.2) while the layer 2 of the 3D attenuation structure developed in section 3.3 for the non-isotropic scattering model. With the isotropic scattering model, the heterogeneous attenuation structure predicted residuals of maximum intensity 12% smaller for the 10 s and 15% smaller for the 20 s prediction than the homogeneous structure. Relative improvements of residuals in maximum intensity with heterogeneous attenuation structure with the non-isotropic scattering

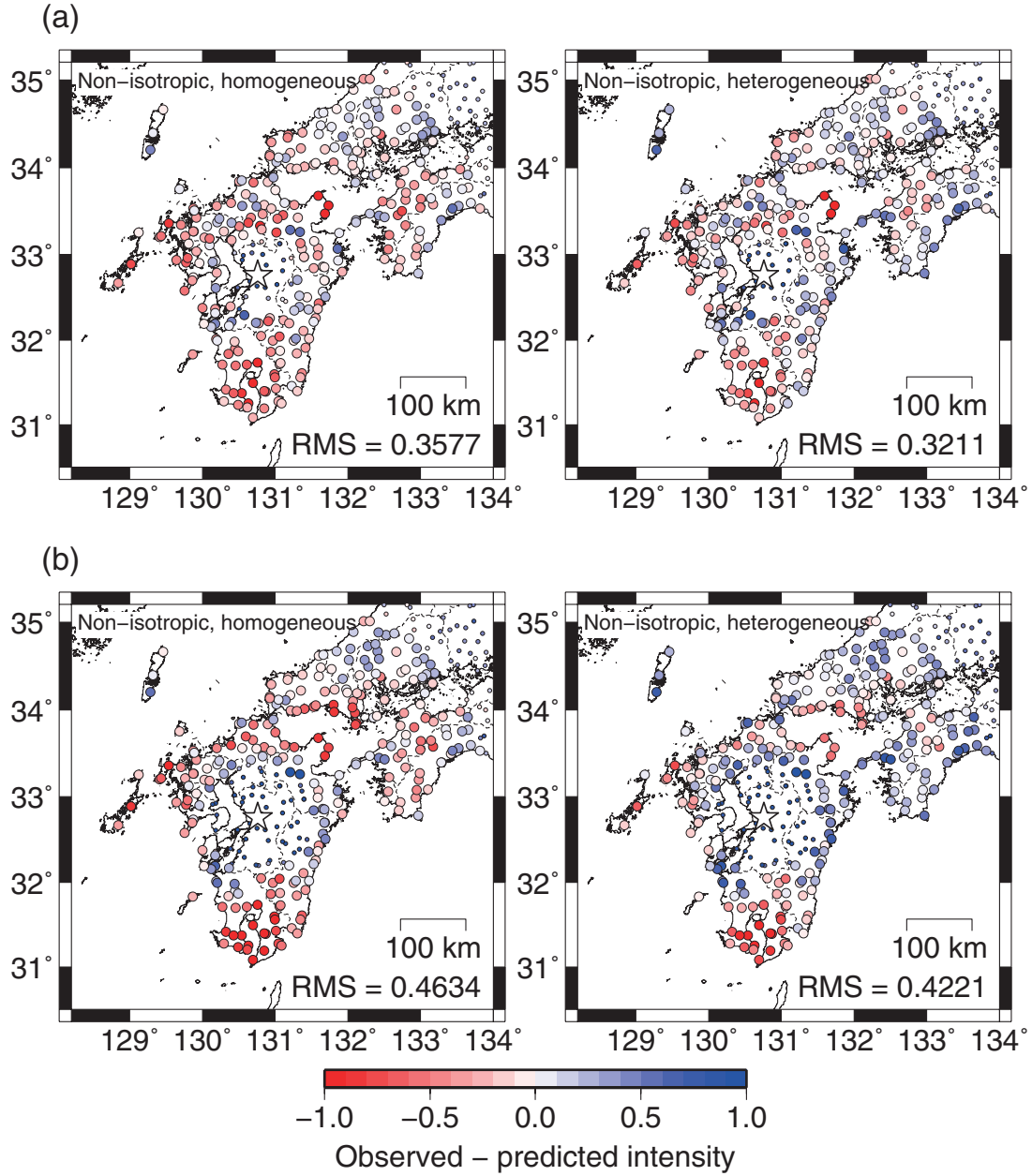


Figure 4.11 Same as Fig. 4.5 but with non-isotropic scattering model.

model was about 10% smaller for both 10 s and 20 s predictions, slightly small compared with the isotropic scattering model. The non-isotropic scattering model with the heterogeneous attenuation structure predicted whole seismogram envelopes with the smallest residuals among four prediction models. In summary, heterogeneous attenuation structures have improved the prediction residuals of the Numerical Shake Prediction scheme with both isotropic and

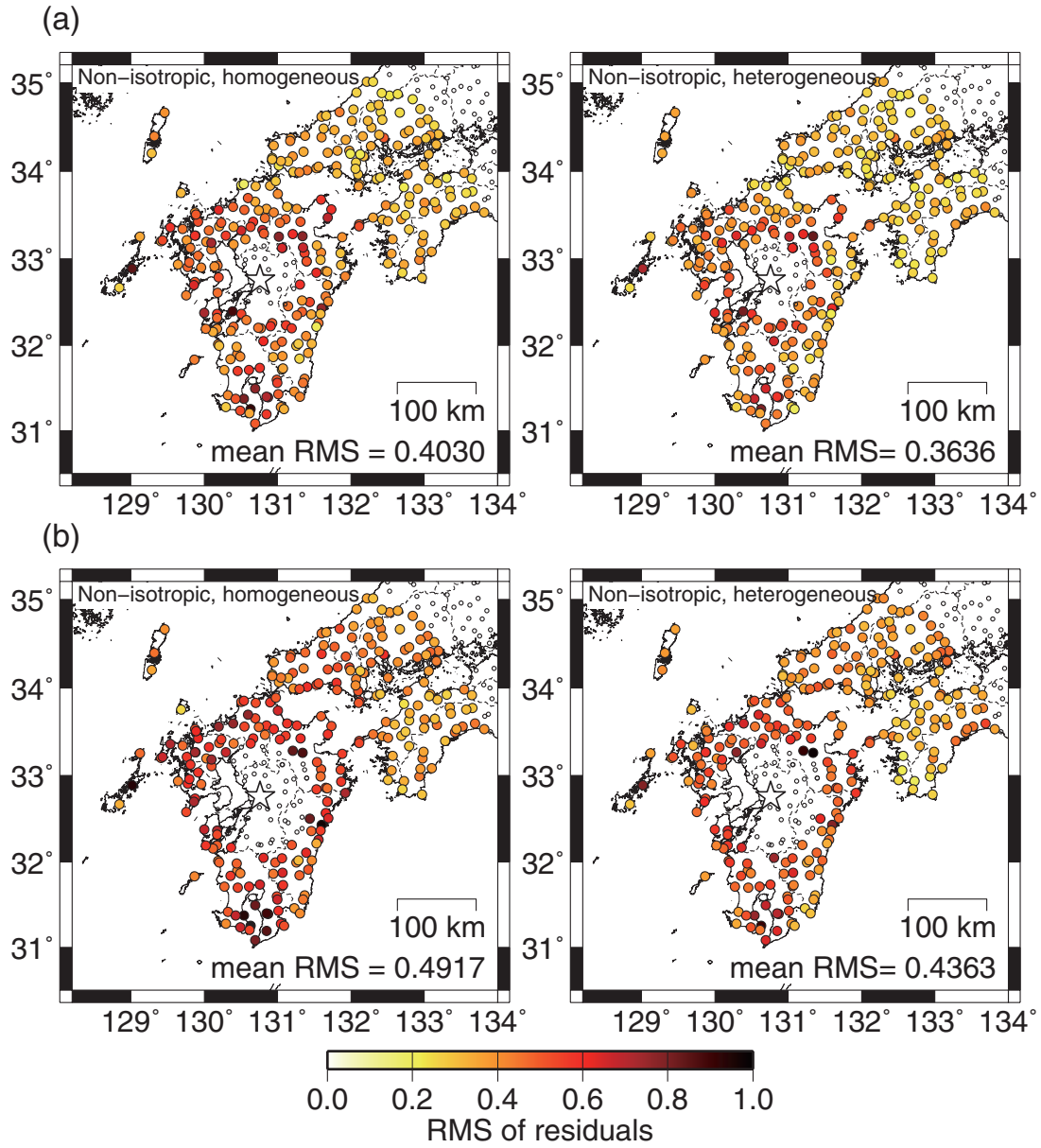


Figure 4.12 Distribution of RMS residuals of whole envelope for (a) 10 s and (b) 20 s ahead prediction using homogeneous (left) and heterogeneous (right) attenuation structures with non-isotropic scattering model. Mean value of RMSs is shown in the lower right corner of each panel. The star is the epicenter of the target earthquake. Stations shown as small circles were excluded from mean value calculation of RMS.

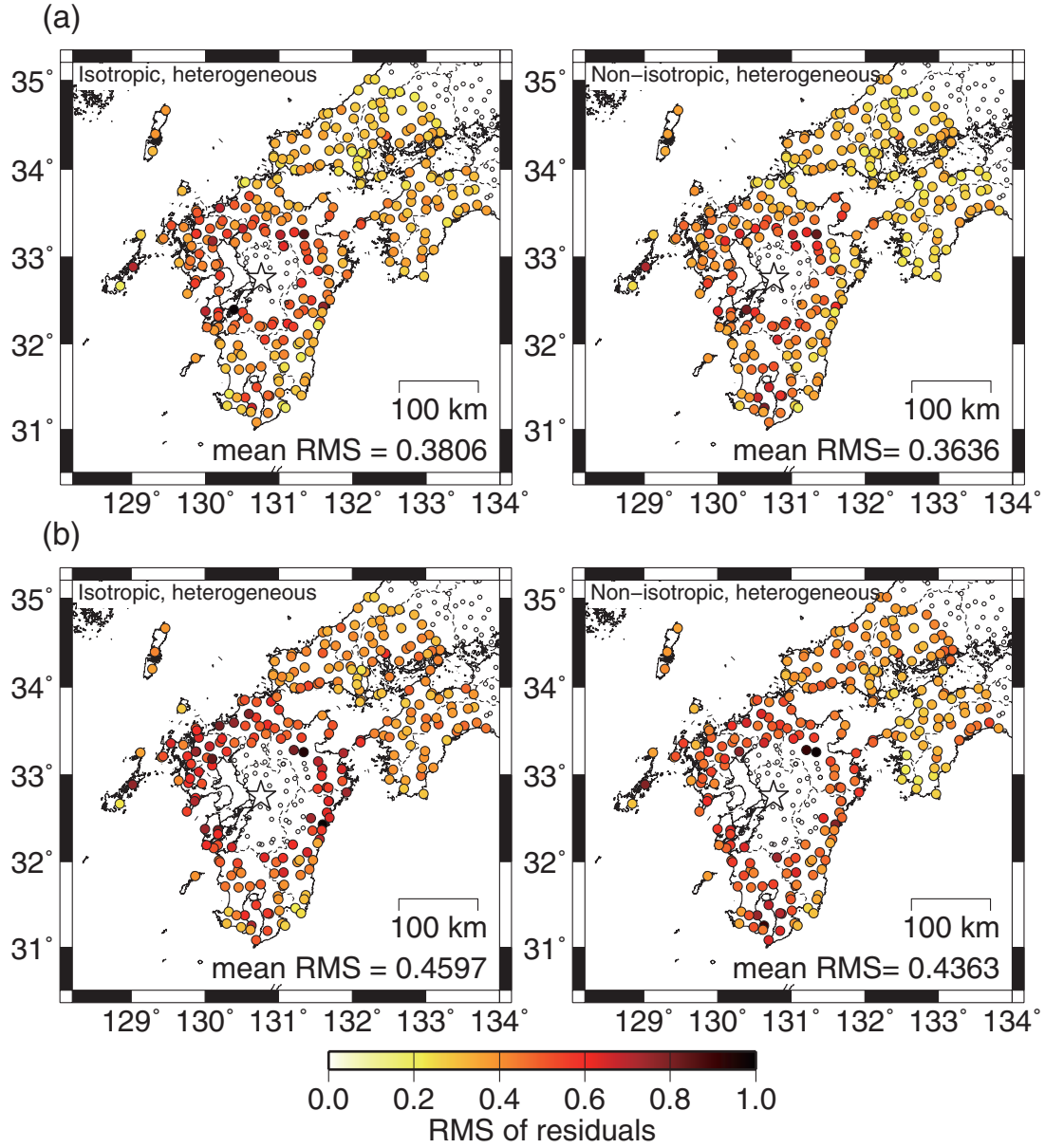


Figure 4.13 Same as Fig. 4.12 but comparison of isotropic (left) and non-isotropic (left) scattering model using heterogeneous attenuation structure.

non-isotropic scattering models.

Let us now discuss some aspects in the present experiments. First, detailed tendency of prediction improvement is different between isotropic and non-isotropic scattering models. The effect of heterogeneous attenuation is expected to be large for a long lead time, because the total attenuation (intrinsic plus scattering attenuation) affects seismic amplitude as

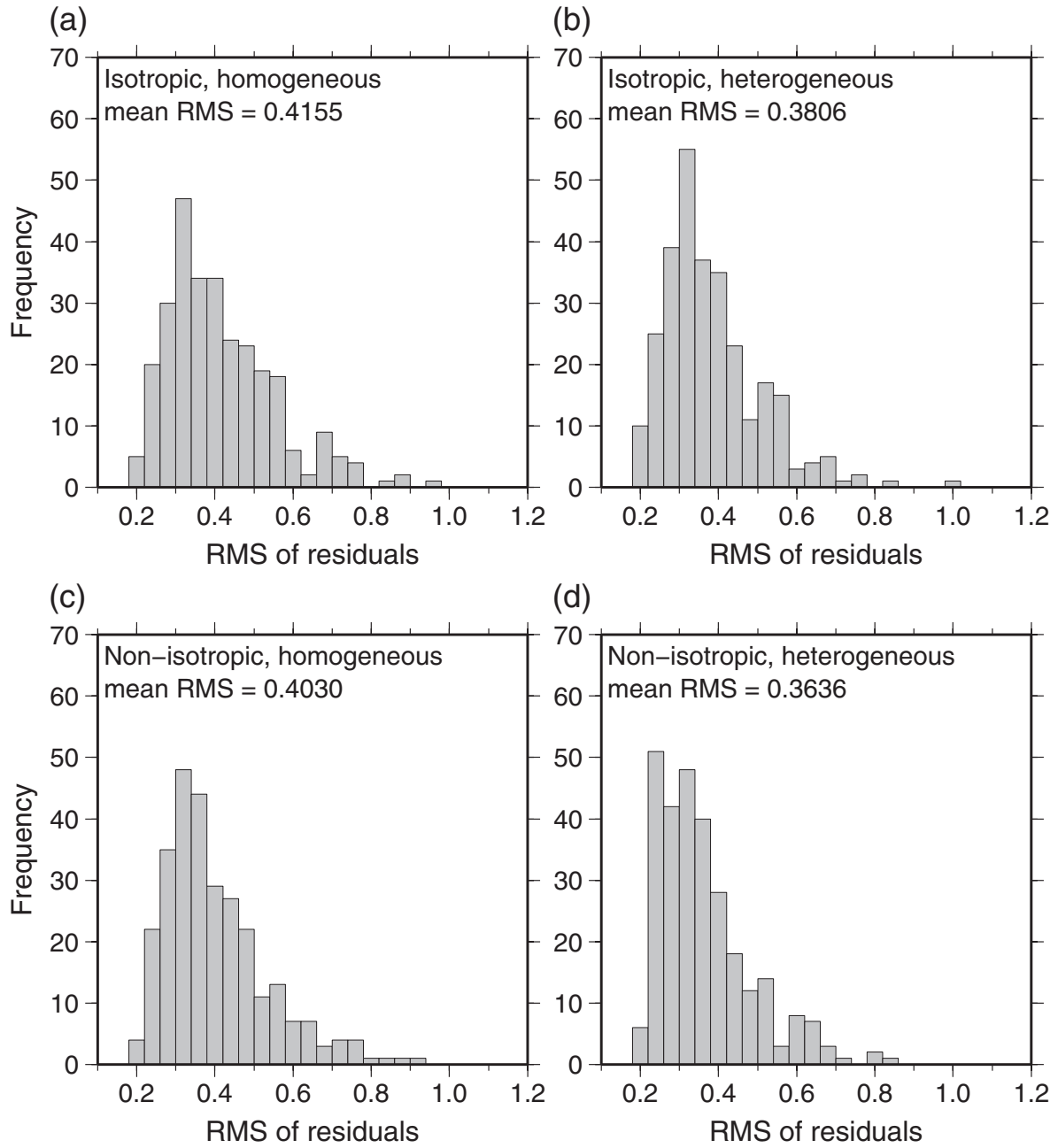


Figure 4.14 Histogram of RMS residuals of whole envelope for 10 s prediction with (a) homogeneous structure with isotropic scattering model, (b) heterogeneous structure with isotropic scattering model, (c) homogeneous structure with non-isotropic scattering model, and (d) heterogeneous structure with non-isotropic scattering model.

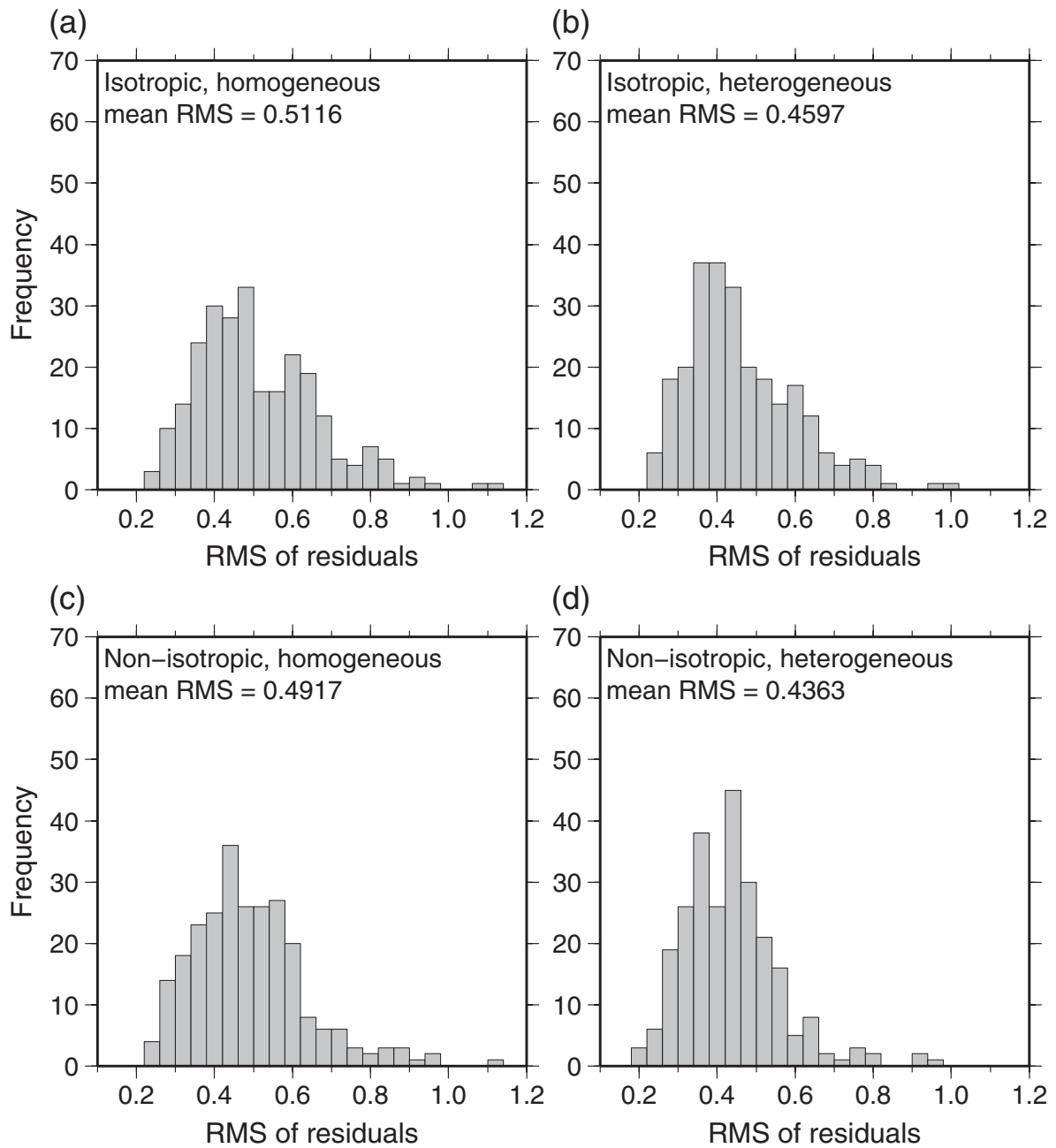


Figure 4.15 Same as Fig. 4.14 but for 20 s prediction.

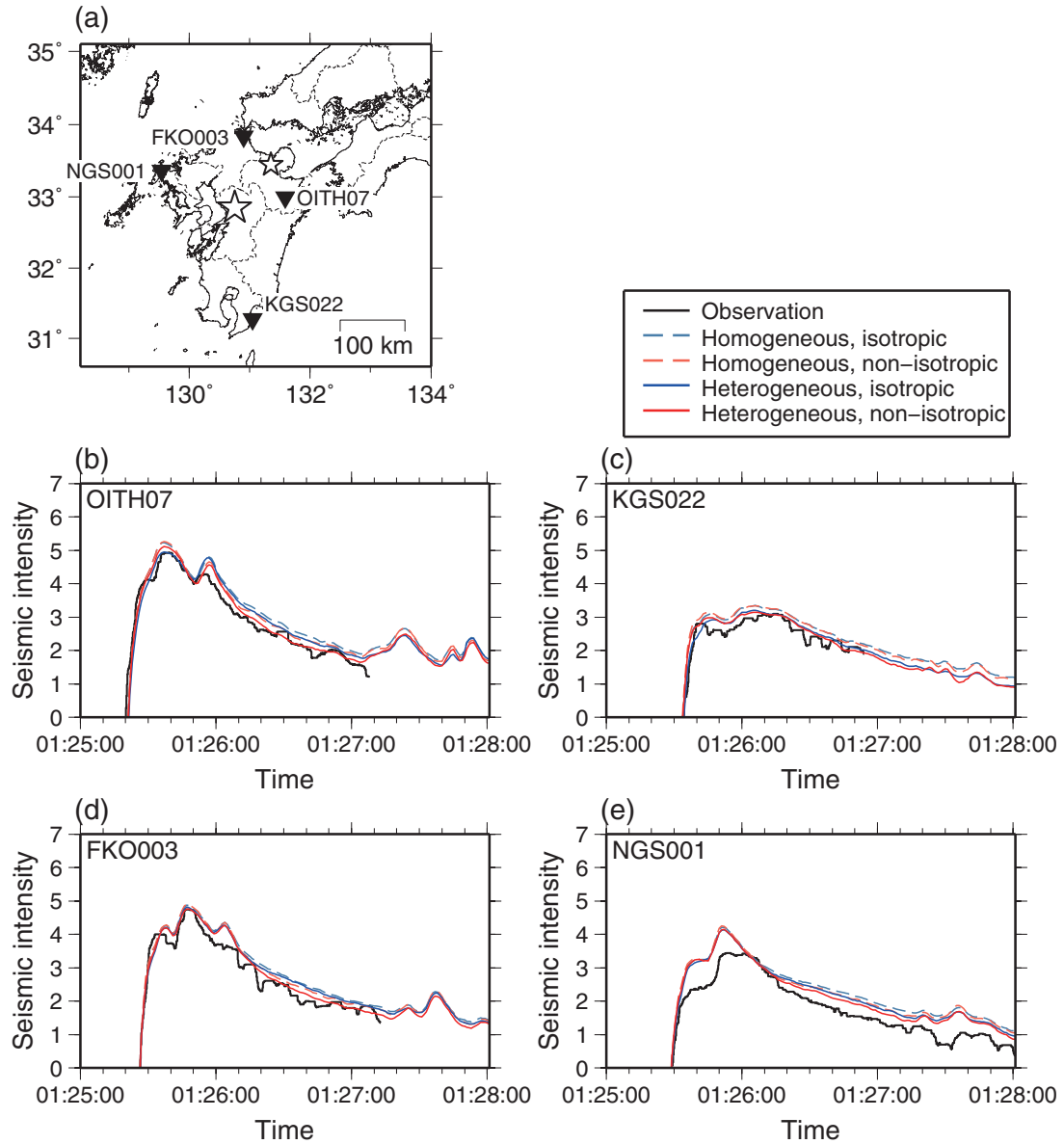


Figure 4.16 (a) Map of four stations (inverted triangles) and the epicenters of the target earthquake (large star) and induced earthquake (small star). Plots of the observed time series of seismic intensity (black lines), 10 s prediction using homogeneous and heterogeneous structure with isotropic scattering model (broken light blue and broken light red lines respectively), and 10 s prediction using homogeneous and heterogeneous structure with non-isotropic scattering model (blue and red lines, respectively) for stations (b) OITH07, (c) KGS022, (d) FKO003, (e) NGS001. Note that predicted envelopes are shifted by 10 s to overlap the observed time series.

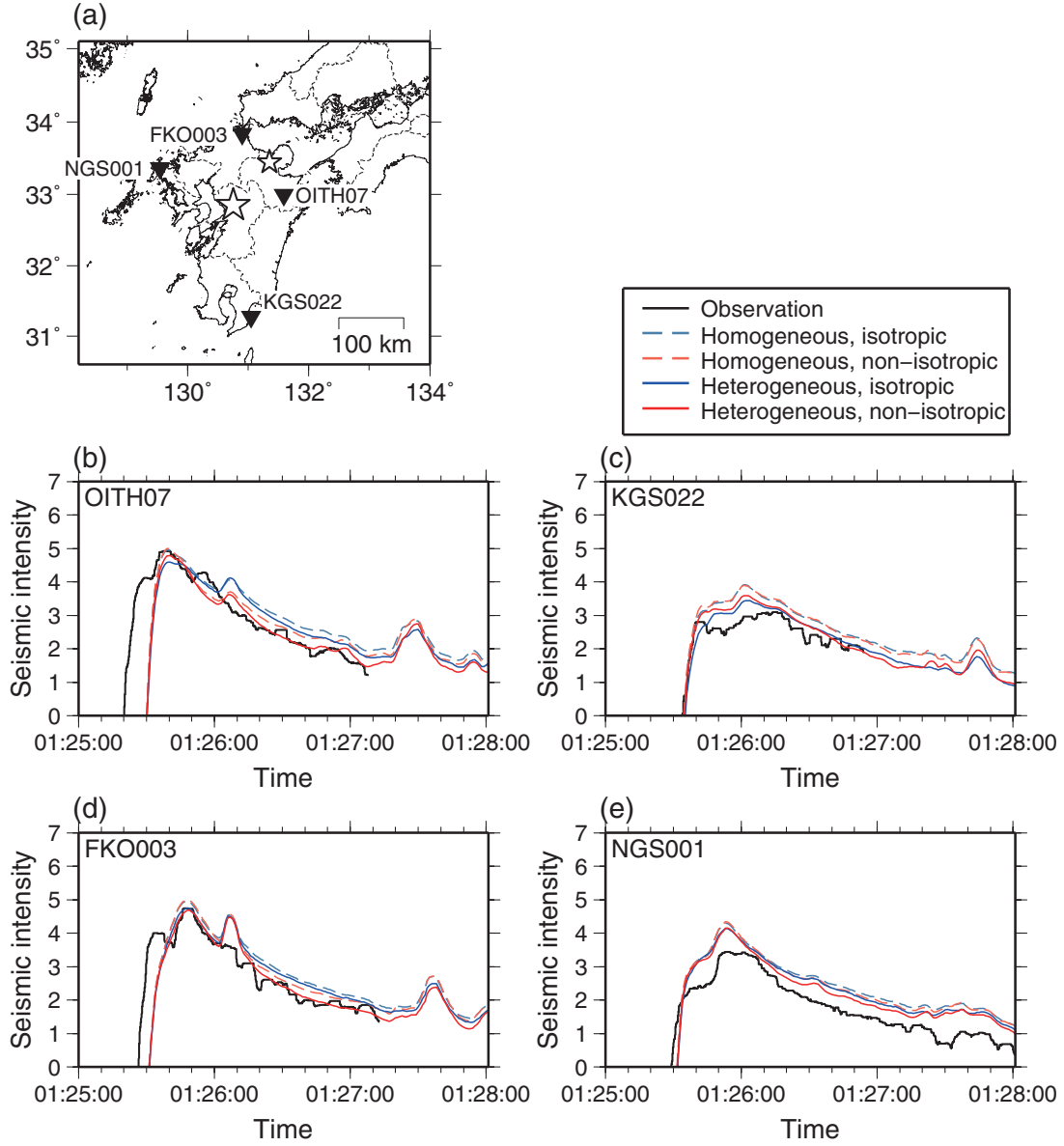


Figure 4.17 Same as Fig. 4.16 but for 20 s prediction.

$\exp(-\pi f t / Q)$ where t is the time, f is the frequency, and Q is the total attenuation. In the isotropic scattering model, this hypothesis was confirmed, however, not in the non-isotropic scattering model. This fact would be attributed to how the waves are affected by the heterogeneous attenuation structure. In section 3.2, we assumed homogeneous velocity structure so that the effect of ray bending caused by velocity heterogeneity was not included in the estimated structure (Fig. 3.3). On the other hand, the attenuation structure shown in Fig. 3.11

was fully in three dimensional. As mentioned in section 4.4.2, the dimension of the structure should be reduced from 3D to 2D. The influence of attenuation in each layer to the observed amplitude in the 3D space varies depending on hypocenter and station locations, meaning a simple dimension reduction used in this study would not always be appropriate for all the hypocenter–station pairs.

Second, we point out the importance of frequency dependency of an attenuation structure for more precise prediction. The frequency characteristics of the band-pass filter used for the JMA seismic intensity has a peak frequency around 1 Hz (Japan Meteorological Agency, 1996; Hoshiba & Ozaki, 2014), however, the peak frequency of the observed seismogram at a station is not always around 1 Hz. Hence, the observed seismic intensity at arbitrary stations is not caused by the seismic wave energy around 1 Hz in every case. Hence, attenuation structure of the frequency range of 1-2 Hz is not always appropriate for all the earthquakes. For more appropriate predictions using the Numerical Shake Prediction scheme, frequency dependency of an attenuation structure should be considered. One solution to realize such an idea is to run the schemes parallel with various frequency ranges, and a prediction is conducted by combining the results of all the frequency ranges.

Finally, we mention the computational cost to adopt non-isotropic scattering process in the Numerical Shake Prediction scheme. In a Monte Carlo simulation, the occurrence of scattering is judged by a random number. If a scattering occur, a new propagation direction of a given particle is determined by other random numbers. In the isotropic scattering model, the new propagation direction can be determined directly from the random numbers (e.g., Eq. 3.10 for 3D case). In the non-isotropic scattering process, one additional procedure is required to determine the new propagation direction, because the new propagation direction should obey the probability distribution function calculated from the non-isotropic scattering coefficient (Eq. 4.13). Pre-computed tables for scattering angles and binary search techniques (e.g., section 3.4 of Press et al., 1992) would greatly reduce the computational cost (Shearer

& Earle, 2004). An additional calculation load for the non-isotropic scattering model can be negligible using such a technique with modern computers. For more sophisticated real time ground motion predictions, that is, not only the maximum amplitude but also the duration of ground motions, non-isotropic scattering model should be adopted in the Numerical Shake Prediction scheme.

5 Conclusion and future perspectives

5.1 Summary of this study

The Tohoku earthquake and its following aftershock activity have revealed the limitation of operational EEW systems that are based on early estimation of source parameters. New EEW (or real-time ground motion prediction) methodologies are required for a robust and precise ground motion prediction in real time even in such a great earthquake sequence. Wavefield monitoring approach, where ground motion is predicted directly from observations, has been proposed to tackle the issues arisen from the Tohoku earthquake. The effort to improve the prediction precision in the wavefield monitoring approach has been demanded. In this study, we have focused on the effect of site amplification term and path term, especially heterogeneous attenuation structure, to investigate how these terms improve the real-time ground motion prediction based on the wavefield monitoring approach.

In chapter 2, we have estimated frequency-dependent site amplification terms and corresponding minimum-phase time-domain filters which approximate frequency characteristics of the site amplification at each K-NET and KiK-net strong motion station. The stations whose site amplification and corresponding time-domain filters estimated successfully in this scheme distributes almost whole of Japan, except the northern part of Hokkaido, the northwestern part of Kyushu, and outlying islands. The estimated time-domain filters enabled site amplification correction in a real-time manner. We investigated the effects of frequency-dependent site amplification correction by the seismic intensity prediction experiments where seismic intensities at a target station was predicted by the real-time site amplification correction of those at an adjacent station. These experiments were conducted for the station pairs with three interstation distances, less than 10, 20, and 30 km. In all the cases, the frequency-dependent site amplification correction by the estimated time-domain filters showed smaller prediction residuals than the scalar correction using averaged seismic intensity differences. Prediction

residuals became smaller as an interstation distances decreased. These results imply that the influence other than a site amplification term becomes large with its interstation distance, hence the path term should be carefully considered for a prediction with a long interstation distance, that is, a prediction with a long lead time.

The Numerical Shake Prediction scheme is one of the real-time ground motion prediction scheme based on a wavefield monitoring approach. This scheme requires the site amplification correction for estimating the current wavefield from observed ground motions. The time-domain filters estimated in chapter 2 can be used for this request. Since the ground motion prediction in the scheme is conducted by the wave propagation simulation, the effect of path terms is naturally taken into account in the scheme. Hence, the scheme is expected to solve the issue in the prediction with long lead time shown in chapter 2. Careful treatment of the detailed heterogeneous attenuation structure in the scheme is expected to improve the prediction precision. In chapter 3, we have estimated the heterogeneous attenuation structure of southwestern Japan by two methods. We first have estimated 2D heterogeneous distribution of intrinsic attenuation and scattering coefficients in southwestern Japan by the conventional multiple lapse time window analysis based on the assumption of isotropic scattering. Next, we have proposed the method to estimate 3D distribution of intrinsic attenuation factors and the strength of random velocity fluctuations from the seismogram envelopes and then applied it in southwestern Japan. The method consists of the following two steps. In step 1, we estimate the above two parameters for each earthquake-station pair by the envelope fitting method. In step 2, we map all the estimated attenuation parameters in step 1 into a 3D space using 3D sensitivity kernels of these parameters. The attenuation structures estimated by both the MLTWA and the new method showed strong intrinsic and scattering attenuation in the Kyushu region than the other regions such as Kinki, Chugoku and Shikoku, which is consistent with previous studies. We also found that the intrinsic attenuation factors and the strength of random velocity fluctuation are smaller in the lower crust than those in the upper crust, and

strong horizontal variations of intrinsic attenuation factors in the focal area of episodic slow slips beneath the Shikoku region.

In chapter 4, we have extended the Numerical Shake Prediction scheme to deal with a heterogeneous attenuation structure in its ground motion prediction stage, and investigated its effect for prediction precision thorough ground motion prediction simulations for the largest event of the 2016 Kumamoto earthquake sequence. First, the 2D heterogeneous attenuation structure derived by the MLTWA was included in the scheme. The relative improvement of seismic intensity prediction residuals, calculated from the mean RMS of maximum intensity prediction residuals, became 10% small for the 10 s prediction and 15% small for the 20 s one in the heterogeneous structure. This result indicates that the heterogeneous structure should be considered for predictions, and its effect becomes large as the lead time increases. Next, the 2nd layer of the estimated 3D heterogeneous attenuation structure in chapter 3 was utilized in the scheme with a non-isotropic scattering model. The relative improvements of maximum intensity prediction was about 10% for both 10 s and 20 s predictions. Although the heterogeneous attenuation structure with the isotropic scattering model showed the smallest residuals in the maximum intensity prediction, the heterogeneous attenuation structure with the non-isotropic scattering model resulted in the smallest RMS residuals of whole envelope. A non-isotropic scattering model should be adopted for real-time predictions of not only maximum amplitude but also whole of the ground motion, and the heterogeneous attenuation structure helps us to improve the prediction precision in the Numerical Shake Prediction scheme.

5.2 Perspectives for future studies

Here we state some issues for future studies of real-time ground motion prediction schemes based on a wavefield monitoring approach for nationwide service.

A dense station network with real time data transmission will be demanded. In Japan,

several dense strong motion station networks have already existed, however, most of the data acquisition of K-NET stations are not in real time. Dense seismic stations are indispensable for wavefield monitoring. To conduct ground motion prediction for a certain area from observations at adjacent stations with frequency-dependent site amplification corrections, reliable time-domain site-correct filters are required for all the target areas, even if there are no stations. The estimation procedure of site amplification adopted in chapter 2 depends on observed waveforms. The site amplification derived from the observed waveform is expected to be more reliable than that from the other information, e.g., estimated by the subsurface velocity structure. However, low seismicity will be problematic when one try to use observed waveforms. Another method to estimate reliable site amplification without observed waveforms, such as the one based on local subsurface velocity structure, is required especially for regions with low seismicity as well as with no stations.

Another consideration of the site amplification term is also required in the Numerical Shake Prediction scheme. In chapter 4, we regarded the time series of seismic intensity with the site amplification unified to one reference station as the observation. Hence, the site amplification of predicted ground motion was also unified to the reference station. Convolution of site amplification is necessary for proper ground motion prediction of an arbitrary area. Time-domain filters like those estimated in chapter 2 can be applied for seismic waveforms, but not for seismogram envelopes such as the output of the Numerical Shake Prediction. Establishment of techniques to convolve the site amplification with seismogram envelopes is required.

Treating the effect of 3D structure is necessary for full consideration of path terms in the Numerical Shake Prediction scheme. At first, 3D attenuation structure which covers all the target areas for prediction should be estimated. In this study, we proposed a method to separate estimation of intrinsic and scattering attenuation structure in 3D space in chapter 3. Seismicity is an important factor for estimating attenuation structures, since weak and localized seismicity reduces the reliability in the estimation of attenuation structure. We

utilized seismogram envelopes to estimate intrinsic and scattering attenuations. Not only intrinsic and scattering attenuation factors but also velocity structure should affect the shape of seismogram envelopes. The velocity structure adopted in this study may not be suitable for envelope calculations. Not a traveltime-based but a waveform-based heterogeneous velocity structure is required for more detailed estimation of intrinsic and scattering attenuations. Next, to utilize the full information of a 3D heterogeneous attenuation structure in the Numerical Shake Prediction scheme, wavefield calculations in fully 3D space is necessary. To do so, an initial ground motion distribution in 3D space will be needed for 3D wavefield calculations. However, because almost all seismic stations are located at the surface or very shallow depth compared with usual hypocentral depths, it is essentially difficult to estimate the 3D distribution of ground motions by simple techniques such as optimal interpolation. Any advanced data assimilation technique such as Kalman filter or adjoint method in 4D (3D space domain plus time domain), may resolve the difficulty in estimating the 3D distribution of ground motions. In addition, we should incorporate the heterogeneous velocity structure in the scheme.

Finally, we point out the importance of developing effective operational systems. In general earthquake information systems, observed waveforms are usually transmitted to the central servers at first, then analyses are conducted to make information, and finally the public will receive its information via broadcast, social networking service, etc. For an EEW purpose, these procedures should be executed as fast as possible, because strong ground motions may hit less than a few seconds after its origin time in near-source areas. Time delay of data acquisition and information transfer for the general public should be minimized, and calculation time of an analysis system should also be small. For minimizing the calculation time of the Numerical Shake Prediction scheme, parallel computing may be one of the solutions. Parallel computing is also important for nationwide service of the scheme in Japan. The overall system of EEW service should run automatically, without any assistance from operators. Hence, fail safe

design and enhancement of the availability are the other important aspects required in the EEW systems.

References

- Abubakirov, I., & Gusev, A. (1990). Estimation of scattering properties of lithosphere of Kamchatka based on Monte-Carlo simulation of record envelope of a near earthquake. *Physics of the Earth and Planetary Interiors*, 64, 52–67. doi:10.1016/0031-9201(90)90005-I.
- Aki, K. (1969). Analysis of the seismic coda of local earthquakes as scattered waves. *Journal of Geophysical Research*, 74, 615–631. doi:10.1029/JB074i002p00615.
- Aki, K., & Richards, P. G. (1980). *Quantitative Seismology Volume 1 and 2*. (1st ed.). San Francisco: W. H. Freeman and Company.
- Allen, R. M. (2007). The ElarmS Earthquake Early Warning Methodology and Application across California. In P. Gasparini, G. Manfredi, & J. Zschau (Eds.), *Earthquake Early Warning Systems 2003* (pp. 21–43). Berlin, Heidelberg: Springer Berlin Heidelberg. doi:10.1007/978-3-540-72241-0_3.
- Allen, R. M., Gasparini, P., Kamigaichi, O., & Böse, M. (2009). The Status of Earthquake Early Warning around the World: An Introductory Overview. *Seismological Research Letters*, 80, 682–693. doi:10.1785/gssrl.80.5.682.
- Aoi, S., Kunugi, T., & Fujiwara, H. (2004). STRONG-MOTION SEISMOGRAPH NETWORK OPERATED BY NIED: K-NET AND KiK-net. *Journal of Japan Association for Earthquake Engineering*, 4, 65–74. doi:10.5610/jaee.4.3_65.
- Asano, Y., & Hasegawa, A. (2004). Imaging the fault zones of the 2000 western Tottori earthquake by a new inversion method to estimate three-dimensional distribution of the scattering coefficient. *Journal of Geophysical Research: Solid Earth*, 109, 1–13. doi:10.1029/2003JB002761.
- Ashiya, K. (2004). EARTHQUAKE ALARM SYSTEMS IN JAPAN RAILWAYS. *Journal of Japan Association for Earthquake Engineering*, 4, 112–117. doi:10.5610/jaee.4.3_112.

- Bard, P. Y., & Bouchon, M. (1985). The two-dimensional resonance of sediment-filled valleys. *Bulletin of the Seismological Society of America*, 75, 519–541.
- Böse, M., Hauksson, E., Solanki, K., Kanamori, H., Wu, Y.-M., & Heaton, T. H. (2009). A New Trigger Criterion for Improved Real-Time Performance of Onsite Earthquake Early Warning in Southern California. *Bulletin of the Seismological Society of America*, 99, 897–905. doi:10.1785/0120080034.
- Böse, M., Ionescu, C., & Wenzel, F. (2007). Earthquake early warning for Bucharest, Romania: Novel and revised scaling relations. *Geophysical Research Letters*, 34, L07302. doi:10.1029/2007GL029396.
- Bouchon, M. (1973). Effect of topography on surface motion. *Bulletin of the Seismological Society of America*, 63, 615–632.
- Carcolé, E., & Sato, H. (2010). Spatial distribution of scattering loss and intrinsic absorption of short-period S waves in the lithosphere of Japan on the basis of the Multiple Lapse Time Window Analysis of Hi-net data. *Geophysical Journal International*, 180, 268–290. doi:10.1111/j.1365-246X.2009.04394.x.
- Chen, D.-Y., Hsiao, N.-C., & Wu, Y.-M. (2015). The Earthworm Based Earthquake Alarm Reporting System in Taiwan. *Bulletin of the Seismological Society of America*, 105, 568–579. doi:10.1785/0120140147.
- Colombelli, S., Allen, R. M., & Zollo, A. (2013). Application of real-time GPS to earthquake early warning in subduction and strike-slip environments. *Journal of Geophysical Research: Solid Earth*, 118, 3448–3461. doi:10.1002/jgrb.50242.
- De Siena, L., Thomas, C., Waite, G. P., Moran, S. C., & Klemme, S. (2014). Attenuation and scattering tomography of the deep plumbing system of Mount St. Helens. *Journal of Geophysical Research: Solid Earth*, 119, 8223–8238. doi:10.1002/2014JB011372.
- Del Pezzo, E., Ibañez, J., Prudencio, J., Bianco, F., & De Siena, L. (2016). Absorption and scattering 2-D volcano images from numerically calculated space-weighting functions.

- Geophysical Journal International*, 206, 742–756. doi:10.1093/gji/ggw171.
- Doi, K. (2011). The operation and performance of Earthquake Early Warnings by the Japan Meteorological Agency. *Soil Dynamics and Earthquake Engineering*, 31, 119–126. doi:10.1016/j.soildyn.2010.06.009.
- Espinosa-Aranda, J. M., Cuellar, A., Garcia, A., Ibarrola, G., Islas, R., Maldonado, S., & Rodriguez, F. H. (2009). Evolution of the Mexican Seismic Alert System (SASMEX). *Seismological Research Letters*, 80, 694–706. doi:10.1785/gssrl.80.5.694.
- Eulenfeld, T., & Wegler, U. (2016). Measurement of intrinsic and scattering attenuation of shear waves in two sedimentary basins and comparison to crystalline sites in Germany. *Geophysical Journal International*, 205, 744–757. doi:10.1093/gji/ggw035.
- Eulenfeld, T., & Wegler, U. (2017). Crustal intrinsic and scattering attenuation of high-frequency shear waves in the contiguous United States. *Journal of Geophysical Research: Solid Earth*, 122, 4676–4690. doi:10.1002/2017JB014038.
- Fehler, M., Hoshiba, M., Sato, H., & Obara, K. (1992). Separation of scattering and intrinsic attenuation for the Kanto-Tokai region, Japan, using measurements of S-wave energy versus hypocentral distance. *Geophysical Journal International*, 108, 787–800. doi:10.1111/j.1365-246X.1992.tb03470.x.
- Furumura, T., & Kennett, B. L. N. (2001). Variations in Regional Phase Propagation in the Area around Japan. *Bulletin of the Seismological Society of America*, 91, 667–682. doi:10.1785/0120000270.
- Gaebler, P. J., Eulenfeld, T., & Wegler, U. (2015). Seismic scattering and absorption parameters in the W-Bohemia/Vogtland region from elastic and acoustic radiative transfer theory. *Geophysical Journal International*, 203, 1471–1481. doi:10.1093/gji/ggv393.
- Grapenthin, R., Johanson, I. A., & Allen, R. M. (2014). Operational real-time GPS-enhanced earthquake early warning. *Journal of Geophysical Research: Solid Earth*, 119, 7944–7965. doi:10.1002/2014JB011400.

- Gusev, A., & Abubakirov, I. (1987). Monte-Carlo simulation of record envelope of a near earthquake. *Physics of the Earth and Planetary Interiors*, 49, 30–36. doi:10.1016/0031-9201(87)90130-0.
- Hasegawa, A., Sato, H., & Nishimura, T. (2015). *Seismology*. Tokyo: Kyoritsu Shuppan.
- Hata, Y., Nozu, A., & Ichii, K. (2011). A practical method to estimate strong ground motions after an earthquake, based on site amplification and phase characteristics. *Bulletin of the Seismological Society of America*, 101, 688–700. doi:10.1785/0120100142.
- Herman, G. T., Hurwitz, H., Lent, A., & Lung, H.-P. (1979). On the Bayesian approach to image reconstruction. *Information and Control*, 42, 60–71. doi:10.1016/S0019-9958(79)90160-8.
- Hirahara, K. (1988). Detection of three-dimensional velocity anisotropy. *Physics of the Earth and Planetary Interiors*, 51, 71–85. doi:10.1016/0031-9201(88)90025-8.
- Hirose, F., Miyaoka, K., Hayashimoto, N., Yamazaki, T., & Nakamura, M. (2011). Outline of the 2011 off the Pacific coast of Tohoku Earthquake (Mw 9.0) —Seismicity: foreshocks, mainshock, aftershocks, and induced activity—. *Earth, Planets and Space*, 63, 513–518. doi:10.5047/eps.2011.05.019.
- Horiuchi, S., Negishi, H., Abe, K., Kamimura, A., & Fujinawa, Y. (2005). An Automatic Processing System for Broadcasting Earthquake Alarms. *Bulletin of the Seismological Society of America*, 95, 708–718. doi:10.1785/0120030133.
- Hoshiba, M. (1991). Simulation of multiple-scattered coda wave excitation based on the energy conservation law. *Physics of the Earth and Planetary Interiors*, 67, 123–136. doi:10.1016/0031-9201(91)90066-Q.
- Hoshiba, M. (1993). Separation of scattering attenuation and intrinsic absorption in Japan using the multiple lapse time window analysis of full seismogram envelope. *Journal of Geophysical Research*, 98, 15809. doi:10.1029/93JB00347.
- Hoshiba, M. (2012). Earthquake Early Warning System in Mexico: Current Status and

- Prospects -Invitation of Juan Manuel Espinosa Aranda, General Director of CIRES-. *Quarterly Journal of Seismology*, 75, 107–111.
- Hoshiba, M. (2013a). Real-time correction of frequency-dependent site amplification factors for application to earthquake early warning. *Bulletin of the Seismological Society of America*, 103, 3179–3188. doi:10.1785/0120130060.
- Hoshiba, M. (2013b). Real-time prediction of ground motion by Kirchhoff-Fresnel boundary integral equation method: Extended front detection method for Earthquake Early Warning. *Journal of Geophysical Research: Solid Earth*, 118, 1038–1050. doi:10.1002/jgrb.50119.
- Hoshiba, M. (2014). Review of the Nationwide Earthquake Early Warning in Japan during Its First Five Years. In M. Wyss (Ed.), *Earthquake Hazard, Risk and Disasters* (pp. 505–529). Elsevier. doi:10.1016/B978-0-12-394848-9.00019-5.
- Hoshiba, M., & Aoki, S. (2015). Numerical shake prediction for earthquake early warning: Data assimilation, real-time shake mapping, and simulation of wave propagation. *Bulletin of the Seismological Society of America*, 105, 1324–1338. doi:10.1785/0120140280.
- Hoshiba, M., Iwakiri, K., Hayashimoto, N., Shimoyama, T., Hirano, K., Yamada, Y., Ishigaki, Y., & Kikuta, H. (2011). Outline of the 2011 off the Pacific coast of Tohoku Earthquake (Mw 9.0) —Earthquake Early Warning and observed seismic intensity—. *Earth, Planets and Space*, 63, 547–551. doi:10.5047/eps.2011.05.031.
- Hoshiba, M., Kamigaichi, O., Saito, M., Tsukada, S., & Hamada, N. (2008). Earthquake Early Warning Starts Nationwide in Japan. *Eos, Transactions American Geophysical Union*, 89, 73. doi:10.1029/2008EO080001.
- Hoshiba, M., & Ozaki, T. (2014). Earthquake Early Warning and Tsunami Warning of the Japan Meteorological Agency, and Their Performance in the 2011 off the Pacific Coast of Tohoku Earthquake (Mw 9.0). In F. Wenzel, & J. Zschau (Eds.), *Early Warning for Geological Disasters Advanced Technologies in Earth Sciences* (pp. 1–28). Berlin, Heidelberg: Springer Berlin Heidelberg. doi:10.1007/978-3-642-12233-0_1.

- Hoshiba, M., Sato, H., & Fehler, M. (1991). Numerical basis of the separation of scattering and intrinsic absorption from full seismogram envelope. A monte-carlo simulation of multiple isotropic scattering. *Papers in Meteorology and Geophysics*, 42, 65–91. doi:10.2467/mripapers.42.65.
- Ikeura, T., & Kato, K. (2011). Evaluation of relative site amplification factors by combining average spectral ratios of strong ground motions simultaneously observed at adjacent two sites—application to K-net and KiK-net sites in the pacific coast side of the tohoku district. *Journal of Japan Association for Earthquake Engineering*, 11, 48–67.
- Ionescu, C., Marmureanu, A., & Marmureanu, G. (2016). Rapid Earthquake Early Warning (REWS) in Romania: Application in Real Time for Governmental Authority and Critical Infrastructures. In R. Vacareanu, & C. Ionescu (Eds.), *The 1940 Vrancea Earthquake. Issues, Insights and Lessons Learnt* (pp. 441–449). Springer, Cham. doi:10.1007/978-3-319-29844-3_31.
- Ionescu, C., Marmureanu, A., Marmureanu, G., & Cioflan, C. O. (2017). Assessment of early warning system performance and improvements since it is in operational phase in Romania title. In *Geophysical Research Abstracts* (pp. EGU2017–16408). European Geophysical Union.
- Iwata, T., & Irikura, K. (1986). Separation of Source, Propagation and Site Effects from Observed S-Waves. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 39, 579–593. doi:10.4294/zisin1948.39.4_579.
- Japan Meteorological Agency (1996). *Seismic Intensity*. Tokyo: Gyosei.
- Japan Meteorological Agency (2013). National catalogue of the active volcanoes in Japan (the fourth edition, English version). URL: http://www.data.jma.go.jp/svd/vois/data/tokyo/STOCK/souran_eng/menu.htm.
- Kalnay, E. (2003). *Atmospheric modeling, data assimilation and predictability*. Cambridge: Cambridge University Press. doi:10.1017/CBO9780511802270.

- Kamiya, S., & Kobayashi, Y. (2000). Seismological evidence for the existence of serpentinized wedge mantle. *Geophysical Research Letters*, 27, 819–822. doi:10.1029/1999GL011080.
- Kanamori, H. (2005). REAL-TIME SEISMOLOGY AND EARTHQUAKE DAMAGE MITIGATION. *Annual Review of Earth and Planetary Sciences*, 33, 195–214. doi:10.1146/annurev.earth.33.092203.122626.
- Kawase, H., & Matsuo, H. (2004). Separation of Source, Path, and Site Effects based on the Observed Data by K-NET, KiK-net, and JMA Strong Motion Network. *Journal of Japan Association for Earthquake Engineering*, 4, 33–52. doi:10.5610/jaee.4.33.
- Kita, S., & Matsubara, M. (2016). Seismic attenuation structure associated with episodic tremor and slip zone beneath Shikoku and the Kii peninsula, southwestern Japan, in the Nankai subduction zone. *Journal of Geophysical Research: Solid Earth*, 121, 1962–1982. doi:10.1002/2015JB012493.
- Kita, S., Nakajima, J., Hasegawa, A., Okada, T., Katsumata, K., Asano, Y., & Kimura, T. (2014). Detailed seismic attenuation structure beneath Hokkaido, northeastern Japan: Arc-arc collision process, arc magmatism, and seismotectonics. *Journal of Geophysical Research: Solid Earth*, 119, 6486–6511. doi:10.1002/2014JB011099.
- Kobayashi, A. (2014). A long-term slow slip event from 1996 to 1997 in the Kii Channel, Japan. *Earth, Planets and Space*, 66, 9. doi:10.1186/1880-5981-66-9.
- Kobayashi, M., Takemura, S., & Yoshimoto, K. (2015). Frequency and distance changes in the apparent P-wave radiation pattern: Effects of seismic wave scattering in the crust inferred from dense seismic observations and numerical simulations. *Geophysical Journal International*, 202, 1895–1907. doi:10.1093/gji/ggv263.
- Koketsu, K. (1991). Three-Dimensional Ray Tracing in a Subduction Zone. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 44, 165–176. doi:10.4294/zisin1948.44.3_165.
- Komatsu, M., & Oda, H. (2015). Three-dimensional P-wave Attenuation Structure beneath

- Southwest Japan. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 67, 105–124. doi:10.4294/zisin.67.105.
- Kunugi, T., Aoi, S., Nakamura, H., Suzuki, W., Morikawa, N., & Fujiwara, H. (2013). An Improved Approximating Filter for Real-Time Calculation of Seismic Intensity. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 65, 223–230. doi:10.4294/zisin.65.223.
- Kurahashi, S., & Irikura, K. (2011). Source model for generating strong ground motions during the 2011 off the Pacific coast of Tohoku Earthquake. *Earth, Planets and Space*, 63, 571–576. doi:10.5047/eps.2011.06.044.
- Kuyuk, H. S., Allen, R. M., Brown, H., Hellweg, M., Henson, I., & Neuhauser, D. (2014). Designing a network-based earthquake early warning algorithm for California: ElarmS-2. *Bulletin of the Seismological Society of America*, 104, 162–173. doi:10.1785/0120130146.
- Margerin, L., Planès, T., Mayor, J., & Calvet, M. (2016). Sensitivity kernels for coda-wave interferometry and scattering tomography: Theory and numerical evaluation in two-dimensional anisotropically scattering media. *Geophysical Journal International*, 204, 650–666. doi:10.1093/gji/ggv470.
- Mayor, J., Margerin, L., & Calvet, M. (2014). Sensitivity of coda waves to spatial variations of absorption and scattering: Radiative transfer theory and 2-D examples. *Geophysical Journal International*, 197, 1117–1137. doi:10.1093/gji/ggu046.
- Midorikawa, S. (2009). Ground Motion Attenuation Relations. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 61, 471–477. doi:10.4294/zisin.61.471.
- Miyazawa, M. (2016). An investigation into the remote triggering of the Oita earthquake by the 2016 Mw 7.0 Kumamoto earthquake using full wavefield simulation. *Earth, Planets and Space*, 68, 205. doi:10.1186/s40623-016-0585-z.
- Nakajima, J., Hada, S., Hayami, E., Uchida, N., Hasegawa, A., Yoshioka, S., Matsuzawa, T., & Umino, N. (2013). Seismic attenuation beneath northeastern Japan: Constraints on mantle dynamics and arc magmatism. *Journal of Geophysical Research: Solid Earth*, 118,

- 5838–5855. doi:10.1002/2013JB010388.
- Nakamura, R., Satake, K., Toda, S., Uetake, T., & Kamiya, S. (2006). Three-dimensional attenuation (Qs) structure beneath the Kanto district, Japan, as inferred from strong motion records. *Geophysical Research Letters*, 33, 3–7. doi:10.1029/2006GL027352.
- Nakamura, R., & Uetake, T. (2012). Attenuation structure at crust and upper mantle —Estimation of Qs-value by tomographic analysis and its applications—. *BUTSURI-TANSA(Geophysical Exploration)*, 65, 67–77. doi:10.3124/segj.65.67.
- Nakamura, T., & Aoi, S. (2017). Source location and mechanism analysis of an earthquake triggered by the 2016 Kumamoto, southwestern Japan, earthquake. *Earth, Planets and Space*, 69, 6. doi:10.1186/s40623-016-0588-9.
- Nakamura, Y. (1988). On the urgent earthquake detection and alarm system (UrEDAS). In *9th World Conference on Earthquake Engineering* (pp. 673–678).
- Nakamura, Y., & Saita, J. (2007). UrEDAS, the Earthquake Warning System: Today and Tomorrow. In P. Gasparini, G. Manfredi, & J. Zschau (Eds.), *Earthquake Early Warning Systems* (pp. 249–281). Berlin, Heidelberg: Springer Berlin Heidelberg. doi:10.1007/978-3-540-72241-0_13.
- Nakano, K., Kawase, H., & Matsushima, S. (2015a). Statistical properties of strong ground motions based on the spectral inversion method, part II: Focus on the characteristics of strong ground motions estimated by Fourier spectra. *Journal of Japan Association for Earthquake Engineering*, 15, 1_38–1_59. doi:10.5610/jaee.15.1_38.
- Nakano, K., Matsushima, S., & Kawase, H. (2015b). Statistical properties of strong ground motions from the generalized spectral inversion of data observed by K-NET, KiK-net, and the JMA Shindokei network in Japan. *Bulletin of the Seismological Society of America*, 105, 2662–2680. doi:10.1785/0120140349.
- Nishigami, K. (1991). A new inversion method of Coda waveforms to determine spatial distribution of Coda scatterers in the crust and uppermost mantle. *Geophysical Research*

- Letters*, 18, 2225–2228. doi:10.1029/91GL02823.
- Nishigami, K. (2000). Deep crustal heterogeneity along and around the San Andreas fault system in central California and its relation to the segmentation. *Journal of Geophysical Research: Solid Earth*, 105, 7983–7998. doi:10.1029/1999JB900381.
- Noguchi, T., Nishida, R., & Hayashi, K. (2005). Determination of 2D subsurface structure of Sakaiminato city in Tottori prefecture, Japan. *Proceedings of the JSCE Earthquake Engineering Symposium*, 28, 154. doi:10.11532/proee2005a.28.154.
- Nozu, A., Nagao, T., & Yamada, M. (2007). Site Amplification Factors for Strong-Motion Sites in Japan Based on Spectral Inversion Technique and Their Use for Strong-Motion Evaluation. *Journal of Japan Association for Earthquake Engineering*, 7, 215–234. doi:10.5610/jaee.7.2_215.
- Odaka, T., Ashiya, K., Tsukada, S., Sato, S., Ohtake, K., & Nozaka, D. (2003). A New Method of Quickly Estimating Epicentral Distance and Magnitude from a Single Seismic Record. *Bulletin of the Seismological Society of America*, 93, 526–532. doi:10.1785/0120020008.
- Ogiso, M., Aoki, S., & Hoshiba, M. (2016). Real-time seismic intensity prediction using frequency-dependent site amplification factors. *Earth, Planets and Space*, 68. doi:10.1186/s40623-016-0467-4.
- Okada, Y., Kasahara, K., Hori, S., Obara, K., Sekiguchi, S., Fujiwara, H., & Yamamoto, A. (2004). Recent progress of seismic observation networks in Japan —Hi-net, F-net, K-NET and KiK-net—. *Earth, Planets and Space*, 56, xv–xxviii. doi:10.1186/BF03353076.
- Paasschens, J. (1997). Solution of the time-dependent Boltzmann equation. *Physical Review E*, 56, 1135–1141. doi:10.1103/PhysRevE.56.1135.
- Parolai, S., Oth, A., & Boxberger, T. (2017). Performance of the GFZ Decentralized On-Site Earthquake Early Warning Software (GFZ-Sentry): Application to K-NET and KiK-Net Recordings, Japan. *Seismological Research Letters*, 88, 1480–1490. doi:10.1785/0220170048.

- Phillips, W. S., & Aki, K. (1986). Site Amplification of Coda Waves From Local Earthquakes From Local Earthquakes in Central California. *Bulletin of the Seismological Society of America*, 76, 627–648.
- Press, W. H., Flannery, B. P., Teukolsky, S. A., & Vetterling, W. T. (1992). *Numerical Recipes in Fortran 77 The Art of Scientific Computing*. (2nd ed.). New York, NY: Cambridge University Press.
- Prudencio, J., Del Pezzo, E., García-Yeguas, A., & Ibáñez, J. M. (2013). Spatial distribution of intrinsic and scattering seismic attenuation in active volcanic islands-I: Model and the case of tenerife Island. *Geophysical Journal International*, 195, 1942–1956. doi:10.1093/gji/ggt361.
- Przybilla, J., Korn, M., & Wegler, U. (2006). Radiative transfer of elastic waves versus finite difference simulations in two-dimensional random media. *Journal of Geophysical Research: Solid Earth*, 111, 1–13. doi:10.1029/2005JB003952.
- Sanborn, C. J., Cormier, V. F., & Fitzpatrick, M. (2017). Combined effects of deterministic and statistical structure on high-frequency regional seismograms. *Geophysical Journal International*, 210, 1143–1159. doi:10.1093/gji/ggx219.
- Sánchez-Sesma, F. J., & Luzón, F. (1996). Can horizontal P waves be trapped and resonate in a shallow sedimentary basin? *Geophysical Journal International*, 124, 209–214. doi:10.1111/j.1365-246X.1996.tb06364.x.
- Sato, H. (1995). Formulation of the multiple non-isotropic scattering process in 3-D space on the basis of energy transport theory. *Geophysical Journal International*, 121, 523–531. doi:10.1111/j.1365-246X.1995.tb05730.x.
- Sato, H., Fehler, M. C., & Maeda, T. (2012). *Seismic Wave Propagation and Scattering in the Heterogeneous Earth : Second Edition*. Berlin, Heidelberg: Springer Berlin Heidelberg. doi:10.1007/978-3-642-23029-5.
- Satoh, T., & Kawase, H. (2009). Site Responses of Sediments. *Zisin (Journal of the Seismo-*

- logical Society of Japan. 2nd ser.*), 61, 455–470. doi:10.4294/zisin.61.455.
- Satoh, T., Kawase, H., & Sato, T. (1994). STATISTICAL SPECTRAL CHARACTERISTICS FOR ENGINEERING BEDROCK WAVES IN WHICH LOCAL SITE EFFECTS OF SURFACE GEOLOGY ARE REMOVED : Based on the ground motion records of small and medium earthquakes observed in the boreholes in Sendai. *Journal of Structural and Construction Engineering (Transactions of AIJ)*, 59, 79–89. doi:10.3130/aijs.59.79_2.
- Satriano, C., Elia, L., Martino, C., Lancieri, M., Zollo, A., & Iannaccone, G. (2011). PRESTo, the earthquake early warning system for Southern Italy: Concepts, capabilities and future perspectives. *Soil Dynamics and Earthquake Engineering*, 31, 137–153. doi:10.1016/j.soildyn.2010.06.008.
- Sawazaki, K., Sato, H., & Nishimura, T. (2011). Envelope synthesis of short-period seismograms in 3-D random media for a point shear dislocation source based on the forward scattering approximation: Application to small strike-slip earthquakes in southwestern Japan. *Journal of Geophysical Research: Solid Earth*, 116, 1–18. doi:10.1029/2010JB008182.
- Shearer, P. M., & Earle, P. S. (2004). The global short-period wavefield modelled with a Monte Carlo seismic phonon method. *Geophysical Journal International*, 158, 1103–1117. doi:10.1111/j.1365-246X.2004.02378.x.
- Snieder, R. (2004). *A Guided Tour of Mathematical Methods*. Cambridge: Cambridge University Press. doi:10.1017/CBO9780511983962.
- Suzuki, W., Aoi, S., Kunugi, T., Kubo, H., Morikawa, N., Nakamura, H., Kimura, T., & Fujiwara, H. (2017). Strong motions observed by K-NET and KiK-net during the 2016 Kumamoto earthquake sequence. *Earth, Planets and Space*, 69, 19. doi:10.1186/s40623-017-0604-8.
- Taira, T., & Yomogida, K. (2007). Imaging of crustal heterogeneous structures using a slowness-weighted back-projection with effects of scattering modes: 1. Theory. *Journal of Geophysical Research*, 112, B06311. doi:10.1029/2006JB004381.

- Taira, T., Yomogida, K., Kuwahara, Y., Imanishi, K., & Ito, H. (2007). Imaging of crustal heterogeneous structures using a slowness-weighted back-projection with effects of scattering modes: 2. Application to the Nagamachi-Rifu fault, Japan, area. *Journal of Geophysical Research*, *112*, B06312. doi:10.1029/2006JB004382.
- Takahashi, T. (2012). Three-dimensional attenuation structure of intrinsic absorption and wide-angle scattering of S waves in northeastern Japan. *Geophysical Journal International*, *189*, 1667–1680. doi:10.1111/j.1365-246X.2012.05438.x.
- Takahashi, T., Obana, K., Yamamoto, Y., Nakanishi, A., Kodaira, S., & Kaneda, Y. (2013). The 3-D distribution of random velocity inhomogeneities in southwestern Japan and the western part of the Nankai subduction zone. *Journal of Geophysical Research: Solid Earth*, *118*, 2246–2257. doi:10.1002/jgrb.50200.
- Takahashi, T., Obana, K., Yamamoto, Y., Nakanishi, A., Kodaira, S., & Kaneda, Y. (2014). S wave attenuation structure on the western side of the Nankai subduction zone: Implications for fluid distribution and dynamics. *Journal of Geophysical Research: Solid Earth*, *119*, 7805–7822. doi:10.1002/2014JB011103.
- Takahashi, T., Sato, H., Nishimura, T., & Obara, K. (2009). Tomographic inversion of the peak delay times to reveal random velocity fluctuations in the lithosphere: Method and application to northeastern Japan. *Geophysical Journal International*, *178*, 1437–1455. doi:10.1111/j.1365-246X.2009.04227.x.
- Takemoto, T., Furumura, T., Saito, T., Maeda, T., & Noguchi, S. (2012). Spatial- and frequency-dependent properties of site amplification factors in Japan derived by the coda normalization method. *Bulletin of the Seismological Society of America*, *102*, 1462–1476. doi:10.1785/0120110188.
- Takemura, S., & Furumura, T. (2013). Scattering of high-frequency P wavefield derived by dense Hi-net array observations in Japan and computer simulations of seismic wave propagations. *Geophysical Journal International*, *193*, 421–436. doi:10.1093/gji/ggs127.

- Takemura, S., Kobayashi, M., & Yoshimoto, K. (2017). High-frequency seismic wave propagation within the heterogeneous crust: Effects of seismic scattering and intrinsic attenuation on ground motion modelling. *Geophysical Journal International*, 210, 1806–1822. doi:10.1093/gji/ggx269.
- Takeuchi, N. (2016). Differential Monte Carlo method for computing seismogram envelopes and their partial derivatives. *Journal of Geophysical Research: Solid Earth*, 121, 3428–3444. doi:10.1002/2015JB012661.
- Tamaribuchi, K., Yamada, M., & Wu, S. (2014). A New Approach to Identify Multiple Concurrent Events for Improvement of Earthquake Early Warning. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 67, 41–55. doi:10.4294/zisin.67.41.
- Tomatsu, T., Kumagai, H., & Dawson, P. B. (2001). Tomographic inversion of P-wave velocity and Q structures beneath the Kirishima volcanic complex, Southern Japan, based on finite difference calculations of complex traveltimes. *Geophysical Journal International*, 146, 781–794. doi:10.1046/j.1365-246X.2001.00491.x.
- Tramelli, A., Del Pezzo, E., Bianco, F., & Boschi, E. (2006). 3D scattering image of the Campi Flegrei caldera (Southern Italy). *Physics of the Earth and Planetary Interiors*, 155, 269–280. doi:10.1016/j.pepi.2005.12.009.
- Tsukada, S., Odaka, T., Ashiya, K., Ohtake, K., & Nozaka, D. (2004). Analysis of the Envelope Waveform of the Initial Part of P-waves and its Application to Quickly Estimating the Epicentral Distance and Magnitude. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 56, 351–361. doi:10.4294/zisin1948.56.4_351.
- Ueno, H., Hatakeyama, S., Aketagawa, T., Funasaki, J., & Hamada, N. (2002). Improvement of hypocenter determination procedures in the Japan Meteorological Agency. *Quarterly Journal of Seismology*, 65, 123–134.
- Wang, W., & Shearer, P. M. (2017). Using direct and coda wave envelopes to resolve the scattering and intrinsic attenuation structure of Southern California. *Journal of Geophysical*

- Research: Solid Earth*, 122, 7236–7251. doi:10.1002/2016JB013810.
- Wegler, U., Korn, M., & Przybilla, J. (2006). Modeling full seismogram envelopes using radiative transfer theory with born scattering coefficients. *Pure and Applied Geophysics*, 163, 503–531. doi:10.1007/s00024-005-0027-5.
- Wolter, K. (1985). *Introduction to Variance Estimation*. Statistics for Social and Behavioral Sciences. New York, NY: Springer. doi:10.1007/978-0-387-35099-8.
- Wu, R.-S. (1985). Multiple scattering and energy transfer of seismic waves – separation of scattering effect from intrinsic attenuation – I. Theoretical modelling. *Geophysical Journal International*, 82, 57–80. doi:10.1111/j.1365-246X.1985.tb05128.x.
- Wu, R.-S., & Aki, K. (1988). Multiple scattering and energy transfer of seismic waves— Separation of scattering effect from intrinsic attenuation II. Application of the theory to Hindu Kush region. *Pure and Applied Geophysics*, 128, 49–80. doi:10.1007/BF01772590.
- Wu, S., Yamada, M., Tamaribuchi, K., & Beck, J. L. (2015). Multi-events earthquake early warning algorithm using a Bayesian approach. *Geophysical Journal International*, 200, 791–808. doi:10.1093/gji/ggu437.
- Wu, Y.-M., Hsiao, N.-C., Chin, T.-L., Chen, D.-Y., Chan, Y.-T., & Wang, K.-S. (2013). Earthquake Early Warning System in Taiwan. In *Encyclopedia of Earthquake Engineering* (pp. 1–12). Berlin, Heidelberg: Springer Berlin Heidelberg. doi:10.1007/978-3-642-36197-5_99-1.
- Wu, Y.-M., & Kanamori, H. (2005). Experiment on an Onsite Early Warning Method for the Taiwan Early Warning System. *Bulletin of the Seismological Society of America*, 95, 347–353. doi:10.1785/0120040097.
- Yoshida, Y., Ueno, H., Muto, D., & Aoki, S. (2011). Source process of the 2011 off the Pacific coast of Tohoku Earthquake with the combination of teleseismic and strong motion data. *Earth, Planets and Space*, 63, 565–569. doi:10.5047/eps.2011.05.011.
- Yoshimoto, K. (2000). Monte Carlo simulation of seismogram envelopes in scat-

- tering media. *Journal of Geophysical Research: Solid Earth*, 105, 6153–6161. doi:10.1029/1999JB900437.
- Yoshimoto, K., & Okada, M. (2009). Frequency-dependent attenuation of S-waves in the Kanto region, Japan. *Earth, Planets and Space*, 61, 1067–1075. doi:10.1186/BF03352957.
- Yoshizawa, K. (2005). Probing into the Upper Mantle Using Surface Waves. *Zisin (Journal of the Seismological Society of Japan. 2nd ser.)*, 57, 393–408. doi:10.4294/zisin1948.57.3_393.
- Zeng, Y. (1991). Compact Solutions for Multiple Scattered Wave Energy in Time Domain. *Bulletin of The Seismological Society of America*, 81, 1022–1029.