

INTEGRATED DETERMINISTIC AND STOCHASTIC FULL-WAVEFORM EARTH MODELS FOR BROADBAND WAVE PROPAGATION AND PREDICTION

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14. ABSTRACT This is a contract to develop integrated deterministic and stochastic models to better predict broadband waveforms. We took a comprehensive, three-pronged approach: (1) Computational Methods. We implemented a new and efficient algorithm for simulating wave propagation in anelastic media and developed a new passive-source reverse-time migration method. These new methods enable us to separate intrinsic attenuation from scattering attenuation and image scatterers and discontinuities. (2) Full-wave Tomography. We carried out a full-wave Pn tomography of the central and northern Tibetan plateau. To our knowledge, this is the first full-wave Pn tomography, which extends the traditional 2D Pn model to fully 3D. We calculated full-wave sensitivity kernels of Z/H Rayleigh wave amplitude ratios to better constrain the velocity and density structures of the crust and shallow mantle. We developed a joint gravity and full-wave seismic inversion and applied it to southern California. (3) Stochastic Models. We showed that small-scale heterogeneities have significant effects on traveltimes and amplitudes. We observed that the power spectral densities (PSDs) of the resolved spectra of western U.S., the local tomographic models and the stochastic model from modeling of scattered waves follow a linear trend consistent with the power-law distribution.					
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1. SUMMARY

The goals of this contract are to develop and apply full-waveform methods to improve velocity and attenuation models and to integrate deterministic (tomographic) earth models with stochastic models to better predict broadband seismic waveforms. We take a comprehensive, three-pronged approach to achieve these goals:

(1) Computational methods. We implemented a new finite-difference algorithm for efficient and accurate simulation of wave propagation in anelastic earth models (Wang et al., 2018a,b). This new code allows us to separate intrinsic attenuation from scattering attenuation and to carry out full-wave inversion for intrinsic attenuation (Wang et al., 2018a). In addition, we completed the development of a new 3D passive-source reverse time migration (3D PS-RTM) method (Li et al., 2018) for imaging seismic discontinuities and scatterers. As proven in exploration seismology, 3D RTM can improve resolution of velocity interfaces by an order of magnitude, from the size of the Fresnel zone to the scale of seismic wavelength. We applied our 3D PS-RTM method to image the mantle transition zone discontinuities beneath western U.S. The results show significantly sharper images of the mantle transition zone discontinuities compared to those by traditional receiver function methods and provide evidence for a mantle plume beneath the Yellowstone hotspot. Our 3D PS-RTM method can be used to more accurately map global and regional mantle transition zone discontinuity structures and scatterers, which are particularly important for simulating wave propagation in the epicentral distance range of the mantle-transition-zone triplicated waves (approximately 10-25 degrees).

(2) Full-waveform earth models. Our second effort involved three projects using full-wave tomography to obtain high-resolution velocity and attenuation models. In the first project, we carried out full-wave Pn tomography of the central and northern Tibetan plateau (Bao and Shen, 2017b). To our knowledge, this is the first full-wave Pn tomography, which extends traditional 2D Pn models to fully 3D. In the second project, we calculated the full-wave sensitivity kernels of the vertical-horizontal (Z/H) Rayleigh wave amplitude ratio (Bao and Shen, 2018) to better constrain the velocity and density structures of the crust and shallow mantle. Using the Z/H amplitude ratio kernels, we developed a joint gravity and full-wave seismic inversion method and applied it to southern California (Bao et al., 2018b; Guo et al., 2018). The sensitivities of seismic waves to density variations (Bao and Shen, 2018) provide a natural link between seismic and gravity data and the joint inversion significantly improves resolution of both density and velocity. The new model reveals low-density features in the upper and middle crust of southern California, which we interpret as granitic bodies. The well-constrained velocity and density structures can be converted into elastic moduli. The low density and low shear modulus granitic bodies appear to control the distribution of seismicity in southern California. In the third project, we carried out full-wave inversion of surface waves for velocity and intrinsic attenuation (Q) in northwestern U.S. (Wang et al., 2018a). Preliminary results show a relatively low Q (higher attenuation) beneath the Snake River hotspot track and parts of the Cascadia back-arc region.

(3) Integrated deterministic and stochastic models. We studied the distribution of velocity heterogeneities in southern California, where active seismicity, dense seismic stations and extensive tomographic studies yield some of the highest resolution regional velocity models anywhere in the world. We calculated the spectral distribution of velocity heterogeneities based on the existing models and determined the parameters of stochastic models using the relative

strength of scattered waves on the transverse component. We observed that the power spectral densities (PSDs) of the resolved spectra of the western U.S., the local tomographic models and the stochastic model from modeling of scattered waves follow a linear trend consistent with the power-law distribution (Bao et al., 2018a), providing a strong justification to extrapolate the power-law distribution of the resolved tomographic model to shorter length scales.

2. INTRODUCTION

Geological and geophysical observations suggest that heterogeneities in the Earth follow the ubiquitous power-law distribution (e.g., Wu et al., 1994; Sato and Fehler, 1998; Levander, 2003; Holliger and Goff, 2003). At large length scales (10s to 100s km and greater), the earth structure can be determined tomographically and thus deterministically, albeit with variable resolution and uncertainty depending on the data and inversion method. At short length scales (< 10s km), heterogeneities are beyond the resolution of regional seismic tomography. These small-scale heterogeneities and surface topography may have strong effects on scattering of short-period waves (>1 Hz for body waves and >0.1 Hz for surface waves). Scattering broadens wave pulses, distorts source radiation patterns, causes P-wave energy on the transverse component, and generates long-duration coda waves (Aki and Chouet, 1975; Sato, 1989; Kubanza et al., 2007; Takemura et al., 2009; Takemura and Furumura, 2013). Depending on wave frequency, propagation distance, and geology, scattering may contribute significantly to waveform and amplitude variations. Figure 1 shows some possible effects of small-scale heterogeneities on waves propagating from the North Korea nuclear test site to real and virtual seismic stations at regional distances. The presence of small-scale heterogeneities causes significant deviations in both the traveltimes and amplitudes of seismic waves (from those predicted by the deterministic model), highlighting the uncertainties in traveltime and amplitude predictions because of small-scale heterogeneities.

While small-scale heterogeneities are beyond the resolution of regional tomography, studies have shown that they can be represented with stochastic models and determined by comparing the observed and predicted wavefields. Because stochastic models can be described with a small number of variables (e.g., the correlation distance, root-mean-square value of variation and decay order in the von Karman distribution), determining regionalized and depth-dependent stochastic characteristics of Earth models is achievable, particularly with the ever-increasing power of high-performance computing (Korn, 1997; Takemura and Furumura, 2013; Furumura et al., 2014; Olsen et al., 2018).

Tomographic and stochastic models represent heterogeneities in the Earth at the long and short length scales, respectively, and are two complementary and essential components needed to fully characterize the Earth structure and predict broadband seismic wave propagation: Once the tomographic model for large-scale velocity and intrinsic attenuation structures and the stochastic model for small-scale heterogeneities are determined, they can be combined to better predict the effects of focusing/defocusing, scattering and intrinsic attenuation, as well as traveltime and amplitude uncertainties (Figure 1). Such integrated, physics-based models may provide more robust predictions of waveforms and amplitudes than traditional models for new source-receiver paths not present in the data used to construct the models (Maceira et al., 2017).

In the following, we document our comprehensive, three-pronged approach to develop, validate and apply full-waveform methods to improve velocity and attenuation models and to integrate deterministic earth models with stochastic models to better predict broadband seismic waveforms. Our three-pronged approach involves (1) development and application of new computational and imaging methods, (2) development and application of full-wave tomography for velocity, intrinsic attenuation and density, and (3) integration of deterministic and stochastic models.

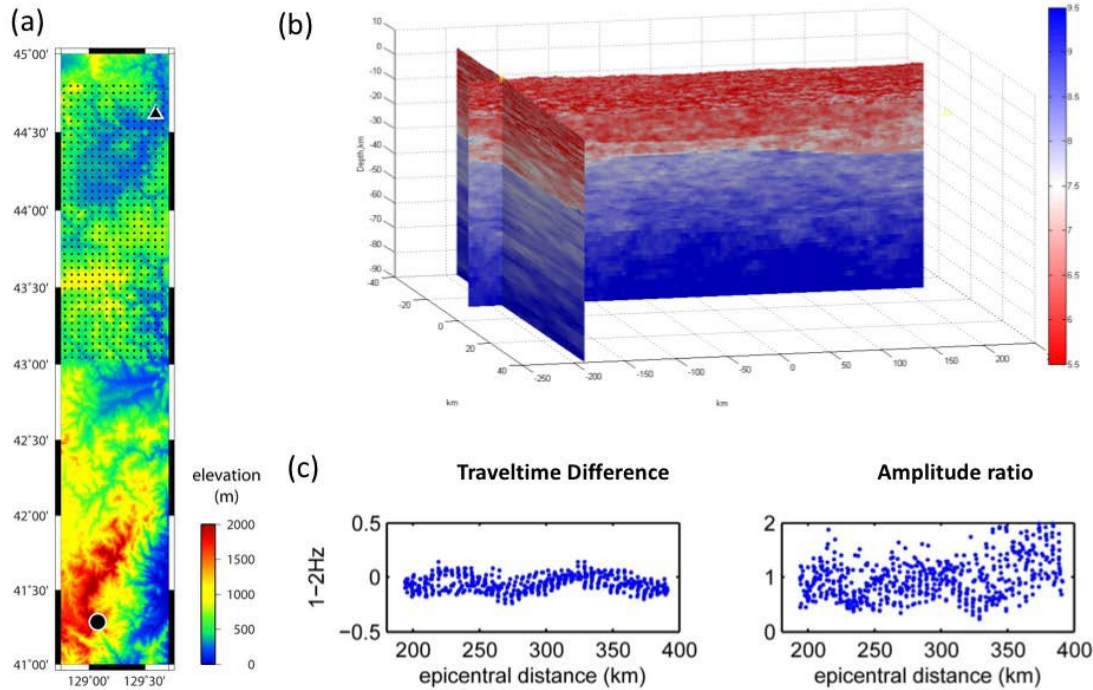


Figure 1. (a) Elevation map with the locations of the North Korea nuclear test site (large dot), station MDJ (triangle) and virtual receivers (small dots). (b) Small-scale velocity heterogeneity (stochastic model) superimposed on a large-scale 3D deterministic model (CRUST1.0). The test site is at the cross-section of the vertical profiles. The color bar shows the P-wave velocity. (c) Traveltime and amplitude variations at the virtual receivers caused by small-scale heterogeneity, highlighting the importance of the stochastic model.

3. METHODS, ASSUMPTIONS, PROCEDURES, AND RESULTS

3.1 Computational Methods

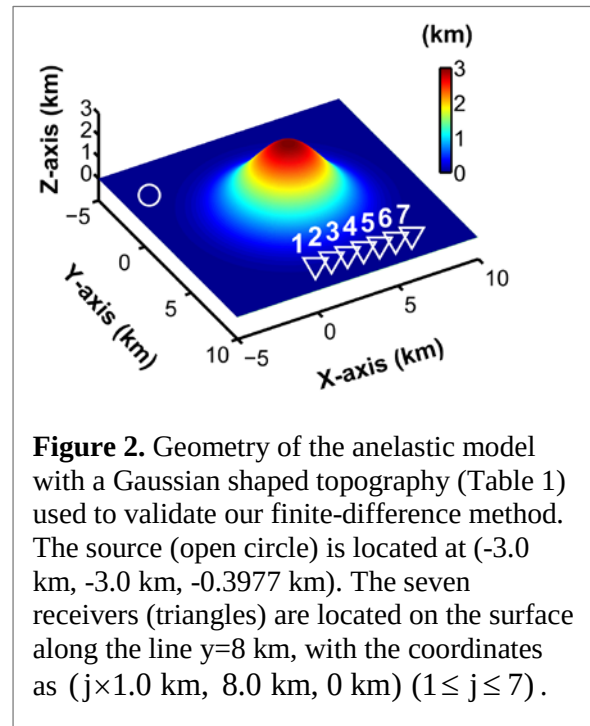
3.1.1 Finite-Difference Simulation of Wave Propagation in Anelastic Media

At the frequencies used in seismic monitoring, intrinsic attenuation of seismic waves is significant. To develop physics-based models (Maceira et al., 2017), it is essential to account for intrinsic attenuation and separate it from scattering caused by small-scale heterogeneities. Accurate and efficient forward simulation of seismic wave propagation in realistic 3D anelastic

Earth models is, however, a challenging task. In the presence of attenuation, wavefield simulation could be inaccurate or unstable over time if it is not well treated (Ammari et al., 2013; Blanc et al., 2016; Komatitsch et al., 2016).

To better simulate wave propagation in anelastic media, we implemented new numerical algorithms in the finite-difference curvilinear coordinate code for local/regional wave propagation in models with topography (Wang et al., 2018b) and the finite-difference spherical coordinate code for regional/global wave propagation (Wang et al., 2018a). To solve the anelastic wave equation, we choose the optimal strong stability preserving Runge-Kutta (SSPRK) method for temporal discretization, and apply the fourth order DRP/opt MacCormack scheme for spatial discretization. The combination of the SSPRK method and the DRP/opt MacCormack scheme provides a stable numerical algorithm and low numerical dispersion. We approximate the rheological behaviors of the Earth by using the generalized Maxwell body model and use an optimization procedure to calculate the anelastic coefficients determined by the $Q(\omega)$ -law. This optimization constrains positivity of the anelastic coefficients and ensures the decay of total energy with time, resulting in a stable viscoelastic system even in the presence of strong attenuation. We performed theoretical and numerical analyses of the SSPRK method, including the stability criteria and the numerical dispersion. We use the complex-frequency-shifted perfectly matched layer for the absorbing boundary conditions based on the auxiliary difference equation and employ the traction image method for the free-surface boundary condition on curvilinear grids representing the surface topography.

We performed several numerical experiments to demonstrate the accuracy of our anelastic modeling in the presence of surface topography (Wang et al., 2018b). Figure 2 shows the geometry of an anelastic model (Table 1) used to compare our finite-difference solution and the SPECFEM (Komatitsch, et al., 2005) solution. The synthetic waveforms from the two methods match very well (Figure 3). The time-frequency phase misfit (TFPM) and time-frequency envelope misfit (TFEM) (Kristekova et al., 2006, 2009) in the seismograms are within 1%. We note that because of an imperfection of the Perfectly Matched Layer (PML) absorbing boundary condition, SPECFEM requires an extended computational region to avoid artificial reflections from the model boundaries, resulting in a much higher computational cost (Wang et al., 2018b). The accuracy of the finite-difference solution verifies the efficacy of the SSPRK method and our implementation of the free surface boundary condition and the absorbing boundary conditions in the presence of surface topography.



These results were presented at the 2016 and 2018 Fall AGU Meetings (Wang et al., 2016, 2018a) and the 2017 Seismological Society of America (SSA) Meeting (Wang et al., 2017a) and published in Journal of Geophysical Research Solid Earth (Wang et al., 2018b).

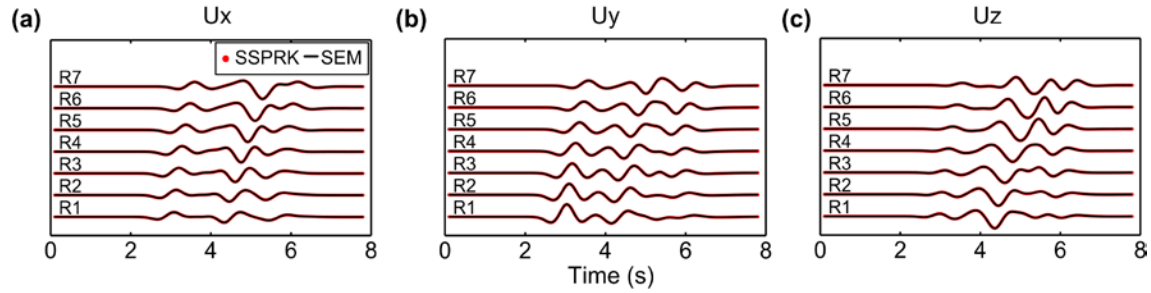


Figure 3. Comparison of the displacement seismograms in (a) the x direction, (b) the y direction, and (c) the z direction, at the seven receivers, between our finite-difference SSPRK method and the SEM solutions, for the Gaussian hill homogeneous model (Figure 2 and Table 1).

Table 1. Model and simulation parameters for the homogeneous model with a Gaussian-shaped hill surface.

ρ (kg m ⁻³)	V_P (km s ⁻¹)	V_S (km s ⁻¹)	Q_P	Q_S	f_r (Hz)	N	Δh (m)	Δt (s)
2100	6.0	3.5	160	80	1.0	4	50	0.001

3.1.2. Three-dimensional Passive-Source Reverse-Time Migration (RTM) Imaging

Seismic imaging of the earth structure falls into two main categories: **Tomographic inversion**, which depends on cumulative properties of seismic waves along the propagation path (e.g., traveltimes and amplitudes), and **seismic migration**, which relies on waves reflected/converted at seismic discontinuities and scatterers. Tomography is suited mostly for relatively large, volumetric structures and usually cannot resolve velocity interfaces and is insensitive to scattered waves. However, velocity interfaces, such as the Moho and mantle transition zone (MTZ) discontinuities, are key components of the earth model and have strong effects on regional seismic wave propagation. Developments of earth models in the seismic monitoring community have focused almost exclusively on tomographic models in the past two decades. We suggest that advanced, full-waveform-based seismic migration imaging of velocity discontinuities and scatterers should be an essential part of developing the physics-based earth models.

In earthquake seismology, the most commonly used method to study deep crust and mantle velocity discontinuities is the receiver function method (e.g., Langston, 1979). While the receiver function method has been widely and successfully used, it has limitations. For example,

because converted phases from crustal and mantle discontinuities are weak, receiver functions are usually stacked to enhance signal and suppress noise. The conventional common-conversion-point (CCP) stacking method assumes a flat or gently dipping interface and does not properly account for wave scattering by sharp corners of the interface and small-scale heterogeneities (Sheehan et al., 2000; Cheng et al., 2016).

In order to map complex velocity interfaces and possibly small-scale heterogeneities (strong scatterers), we have developed a 3D passive-source reverse time migration method (3D PS-RTM), taking advantage of increasingly dense seismic arrays and advances in high-performance computation (Li et al., 2018). We back-propagate the time-reversed, recorded wavefield into the velocity model numerically using a finite-difference method and separate the wavefields in the model into divergence and curl of the wavefields, which represent P- and S-wave modes respectively. An innovative data processing procedure and a new imaging condition are developed to use the separated P- and S-wavefields to map the velocity interfaces where P-to-S conversion occurs. Numerical experiments based on synthetic models show that the method can be used to greatly improve the resolution of the Moho, subducting slabs, MTZ and volumetric scatterers (Li et al., 2018).

The method can be extended to use many different combinations of forward-propagated (from the source) and/or backward-propagated (from the receiver) waves to image regional and global velocity discontinuities, including SS precursors at the mid-point between the source and receiver (e.g., Shearer, 1991) and P-to-S or S-to-P converted phases near the receiver. The main advantage of our new method is a potentially great improvement in resolution, particularly for the SS precursor type of waves, for which the Fresnel zone is large (Deuss, 2009), while the disadvantage is a relatively high computational cost.

For our first 3D PS-RTM application, we chose Yellowstone for several reasons: First, the Yellowstone hotspot is one of the most studied intraplate hotspots, yet there is on-going debate regarding whether the source of the Yellowstone hotspot comes from a deep mantle plume (e.g., Foulger et al., 2015; Leonard and Liu, 2016; Zhou et al., 2018; Zhou, 2018; Nelson and Grand, 2018); second, there have been previous studies of the MTZ in the region using traditional receiver function methods (e.g., Schmandt et al., 2012; Gao and Liu, 2014), so we can compare our RTM results with the previous results; finally, the USArray provides arguably the best coverage of any known hotspot for imaging the MTZ over a broad region.

The main MTZ interfaces, the 410- and 660-km discontinuities, are generally thought to correspond to the two phase transformations from olivine (α -phase) to spinel (β -phase) crystal structure and from ringwoodite (γ -phase) to perovskite plus magnesiowüstite, with a positive (pressure increases with temperature) and a negative (pressure decreases with temperature) Clapeyron slope, respectively (e.g., Bina and Helffrich, 1994), though the mineral phase assemblage and associated Clapeyron slope change at very high temperature (e.g., Hirose, 2002; Ishii et al., 2018). The opposite Clapeyron slopes of the two phase transformations predict that higher temperatures associated with a mantle thermal plume cause a deeper 410-km discontinuity and a shallower 660-km discontinuity within the plume conduit, and therefore a thinner MTZ thickness. Indeed, previous receiver functions studies reveal a thinner MTZ thickness beneath Iceland, which was suggested as evidence for a lower mantle origin of the Iceland hotspot (Shen et al., 1998). A thinner MTZ is also observed beneath Hawaii (Li et al., 2003; Shen et al., 2003;

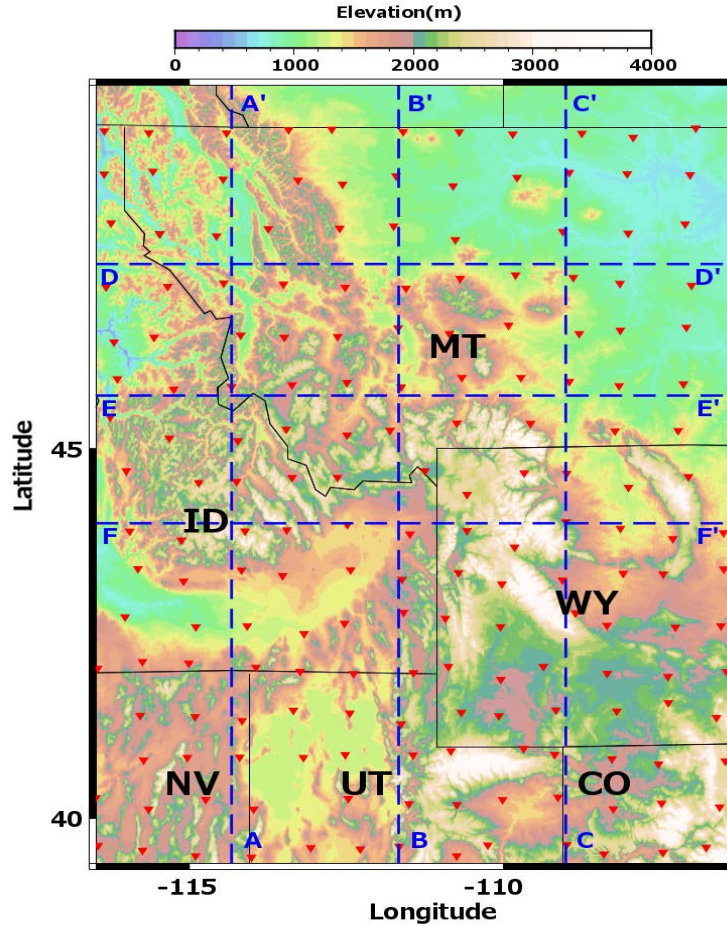


Figure 4. The topography map of the Yellowstone area. *The red triangles represent the USArray TA seismic stations. The dashed blue lines mark the locations of the vertical profiles in Figure 5.*

Agius et al., 2018), the Galapagos hotspot (Hooft et al., 2003), and some but not all other hotspots (e.g., Owens et al., 2000; Li et al., 2003; Tauzin et al., 2008; Vinnik et al., 2012; Reed et al., 2016; Wei and Chen, 2016).

Receiver functions have also been used to image the MTZ boundaries in the Yellowstone region with contradicting results. Using pre-USArray seismic data in the region, Fee and Dueker (2004) observed a thinner transition zone beneath Yellowstone, but since the discontinuity topographies are spatially uncorrelated, they concluded that their observation provided no evidence for a lower mantle plume currently beneath the hotspot. With data from the USArray plus long-term networks and other temporary arrays, Schmandt et al. (2012) showed a normal 410-km discontinuity and an anomalously shallow 660-km discontinuity that is uplifted by up to 18 km over an area of about 200 km wide. Hier-Majumder and Tauzin (2017) suggested a much broader (500-600 km in diameter) thinner transition zone beneath Yellowstone. Gao and Liu (2014), however, found no mantle transition zone anomaly beneath the Yellowstone hotspot. The latter three studies shared much of the same data in the Yellowstone region, so the different results are likely attributable to differences in data selection, processing, and the method of CCP stacking. The width of receiver function binning in the CCP method, for example, is a subjective choice that may affect the observation of the transition zone anomaly. Furthermore, the complex finite-

frequency wave phenomena such as multi-pathing and scattering, which may be significant in the Yellowstone region because of its highly heterogeneous three-dimensional velocity variations (e.g., Porritt et al., 2014; Schmandt and Lin, 2014), are not accounted for in those receiver function studies.

We used broadband earthquake signals recorded by the Transportable Array (TA) of the USArray and archived at the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (Li and Shen, in preparation). The stations used are within 38°N to 50°N, and -117°E to -105°E, a 12° by 12° area roughly centered on Yellowstone (Figure 4). We requested earthquake records between May 2nd, 2005 and Sep 5th, 2010 with moment magnitude (M_w) larger than 5.0 and epicentral distance between 28° to 98°, based on source information in the global CMT catalog (Ekström et al., 2012). We use the DNA13 velocity perturbation model (Porritt et al., 2014) added to the one-dimensional velocity model PREM (Dziewonski and Anderson, 1981) as our reference velocity model to back-propagate the signals. After initial data processing and quality control, we selected 192 earthquakes for back-propagation and RTM imaging (Li and Shen, in preparation).

Figure 5 shows 6 different vertical slices of the final RTM model along the latitudinal and longitudinal directions (AA' to FF' in Figure 4). The 410-km discontinuity can be clearly identified as a continuous feature on all the profiles, except south of 41.5°N on profile BB', where the image appears noisy. The 660-km discontinuity can also be identified and traced continuously along most of the profiles, though it appears as a slightly weaker feature compared to the 410-km discontinuity. In addition to the 410- and 660-km discontinuities, there is a feature at 500-520 km depth, which may correspond to the 520-km discontinuity (e.g., Shearer, 1996), and a feature at 200-260 km depth, which may be associated with the base of the asthenosphere or reverberation from a shallower mid-lithosphere discontinuity (e.g., Abt et al., 2010). Because of the ~70 km station spacing of the USArray, the crust and lithosphere structures are not illuminated continuously and by multiple stations. Thus we refrain from interpreting whether the feature at 200-260 km depth is a reverberation from a shallow depth or not. Herein we focus our discussion on the more prominent features in the mantle transition zone, the 410- and 660-km discontinuities.

A comparison of our RTM model with the result of plane-wave migration (Wang and Pavlis, 2016) indicates that our RTM method achieves a higher signal quality and more continuous velocity interfaces. For example, the 520-km discontinuity was not observed in Wang and Pavlis (2016). Our results thus cast doubt on Wang and Pavlis' conclusion about the roughness of the MTZ discontinuities.

The MTZ discontinuity topography map (Figure 6) shows an apparent depression in the 410-km discontinuity and an upward deflection of the 660-km discontinuity near Yellowstone. The thinner-than-normal transition zone near the Yellowstone hotspot supports the notion of a mantle thermal plume tilted to the northwest from the surface to the MTZ depth. The uncertainty of the discontinuity depth measurement is relatively high south of 42°N and along the eastern edge of the model. Both the 410- and 660-km discontinuities appear shallower along the northern edge of the model near the Canadian border. However, the transition zone thickness there is not abnormal, indicating that the apparent shallow discontinuities there are most likely due to velocity variations above the 410-km discontinuity. It is possible that the apparent shallower

discontinuities near the northern edge of the RTM model is caused by an underestimation of high velocities in the north due to a lack of station coverage in the tomographic inversion (Porritt et al., 2014).

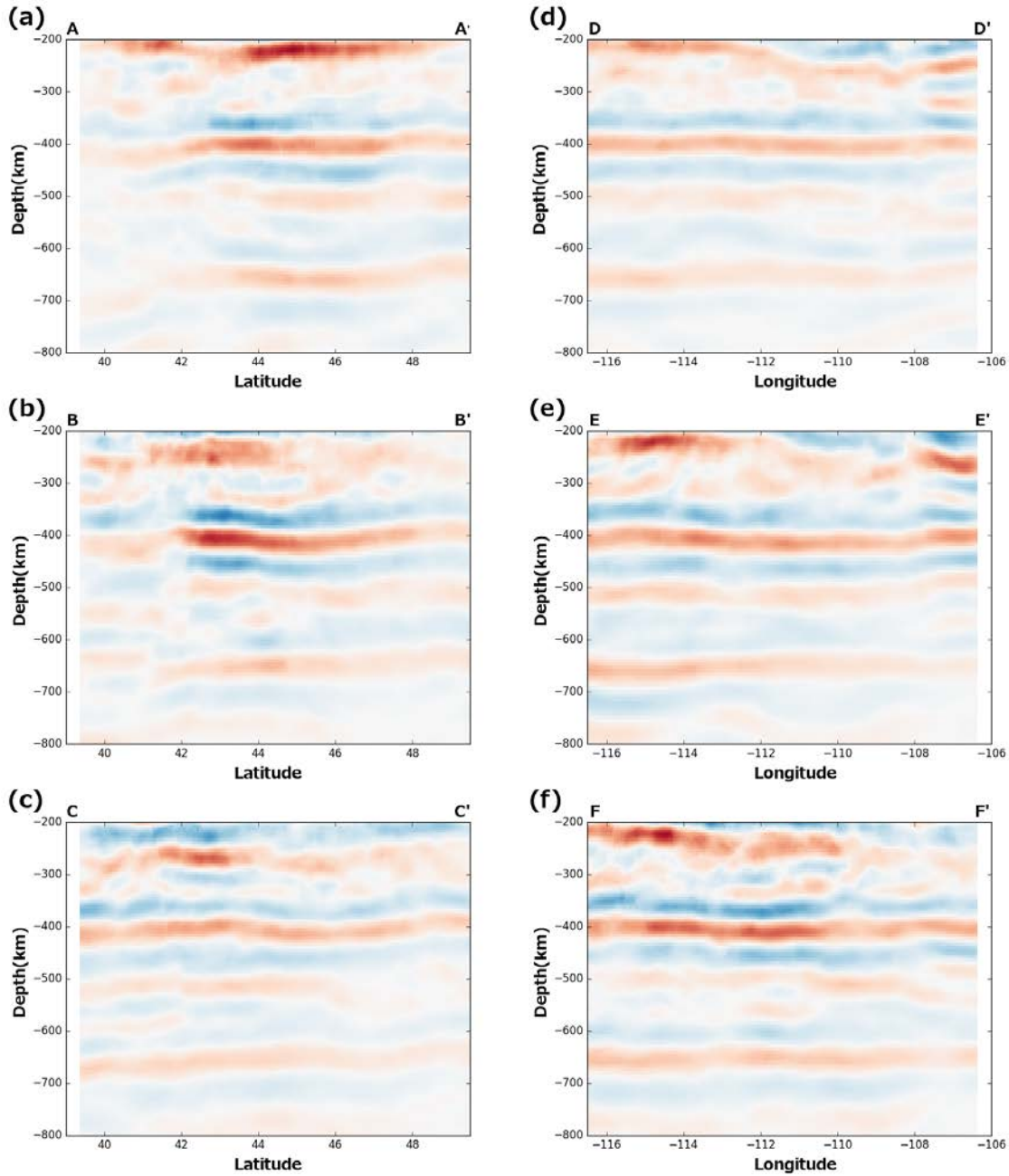


Figure 5. Stacked RTM model obtained with 10-s period waves, the DNA13 velocity model and PCA (Principal Component Analysis) components that contain 95% energy. (a), (b) and (c) represent three vertical slices of the RTM model along the latitudinal direction at longitude of -114.333° , -111.667° and -109° , respectively. (d), (e) and (f) represent three vertical slices of the RTM model along the longitudinal direction at latitude of 47.333° , 45.676° and 44.018° , respectively. Warm colors represent velocity interfaces with a positive (increasing with depth) velocity contrast.

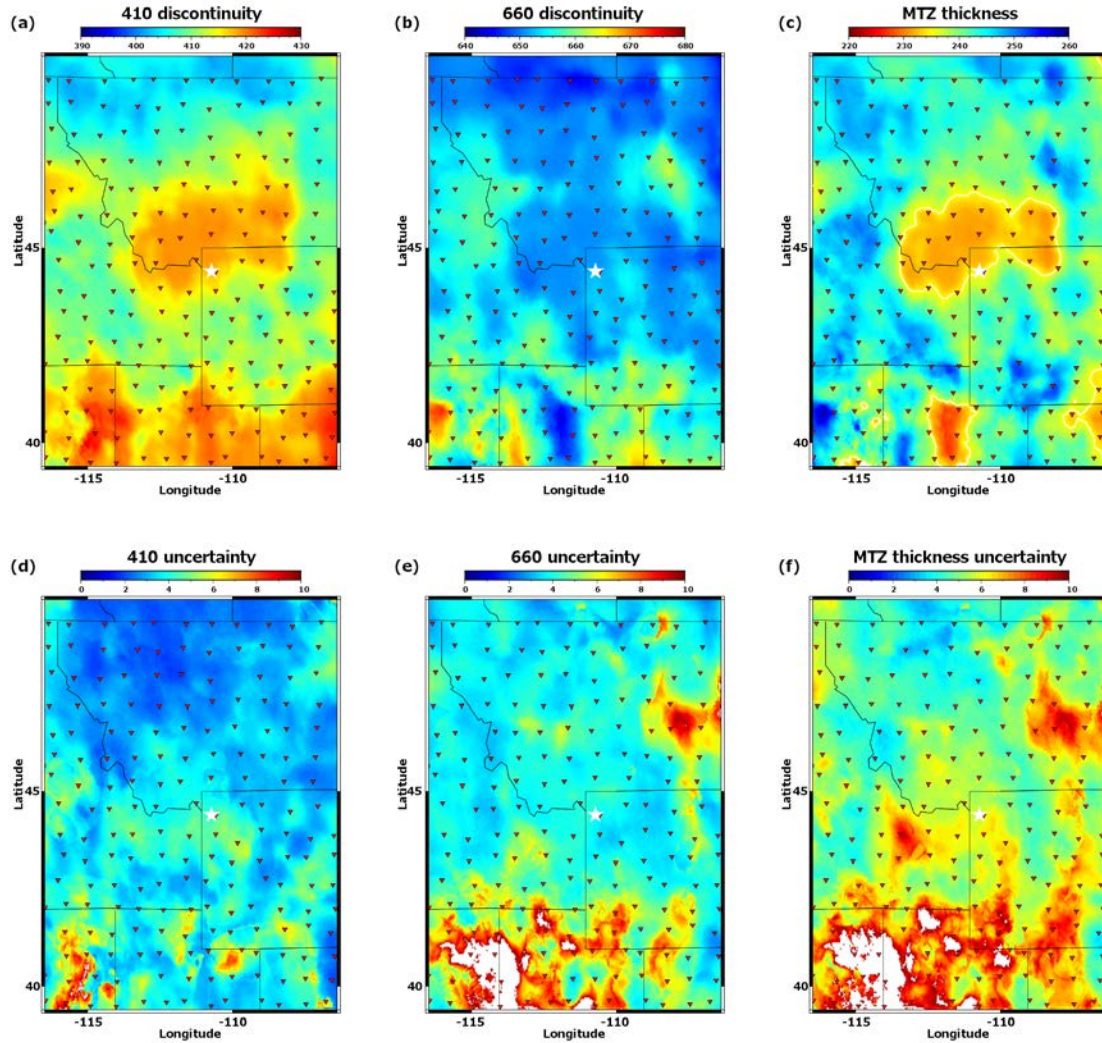


Figure 6. (a) and (b) are the topography map of the 410- and 660-km discontinuities respectively; (c) is the MTZ thickness, the white line mark the 235 km MTZ thickness contour; (d) and (e) are the uncertainty (one standard deviation) of the 410- and 660-km discontinuities, respectively; (f) is the uncertainty (one standard deviation) of the MTZ thickness.

The 3D PS-RTM method was presented at the 2015, 2016 and 2017 SSA meetings (Li et al., 2015, 2016, 2017) and published in JGR-Solid Earth (Li et al., 2018). A manuscript on the application of the method to the Yellowstone hotspot is in preparation (Li and Shen, in preparation). We note that this development took several years and was partially supported by previous AFRL contracts. It would not have been possible without the continuous support of AFRL.

3.2 Full-Wave Tomography

In order to use the power-law distribution of velocity heterogeneities discussed in section 3.3, we need to develop and validate high resolution velocity and attenuation models. To this end, we carried out three full-wave tomographic projects. Among them, the full-wave P_n tomography is a continuation of an effort of an earlier AFRL contract.

3.2.1 Full-wave P_n Tomography

As the first and typically prominent arrival at regional epicentral distances ($\sim 2^\circ$ - 15°), the P_n phase has been widely used in earthquake studies and seismic monitoring. Since the upper mantle V_p structure is one of the important seismic constraints for the physical and compositional properties of the mantle, P_n travel-time tomography has been an active topic of study for Tibet and the surrounding regions in the past two decades (e.g., Hearn et al., 2004; Liang et al., 2004). These previous studies shared a common assumption that P_n waves mainly travel right beneath the Moho like head waves. Consequently P_n tomography was traditionally carried out as a two-dimensional (2D) inverse problem involving lateral velocity variation only. To approximately account for P_n waves that sample the shallow and deep parts of the mantle lithosphere, some studies carried out separate inversions for P_n arrivals in short- and long-epicentral distance groups (Hearn et al., 2004) or solved for mantle velocity gradients (Myers et al., 2010). Nevertheless, the ray assumption does not accurately account for finite-frequency P_n wave propagation in 3D Earth structures (Zhang et al., 2007), and therefore the accuracy and resolution of traditional ray-based P_n tomography are hindered by its theoretical limitation.

We selected $M_w = 4\sim 6$ crustal earthquakes that occurred between 2007 and 2010 in Tibet and the surrounding regions and collected waveform data at permanent and temporary broadband

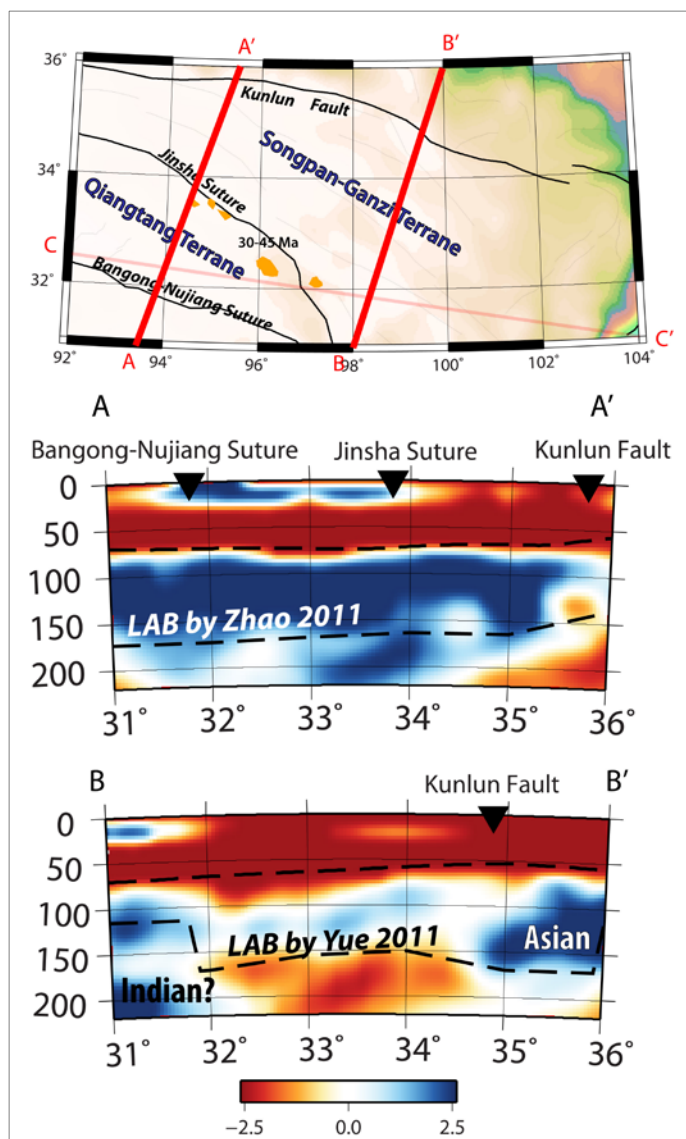


Figure 7. (upper panel) Topography of the Tibetan plateau and main tectonic sutures. Lines AA' and BB' mark the locations of the profiles. (middle panel) Full-wave P_n model along profile AA'. The shallow dashed line represent the Moho and the deep one the lithosphere-asthenosphere boundary (LAB) determined by Zhao et al. (2011). (lower panel) Same as the middle panel, except along profile BB' and the LAB is from Yue et al. (2012).

seismic stations in the central and northeastern Tibetan Plateau from the IRIS DMC. After initial data quality inspection, 59 earthquakes recorded at 101 seismic stations in the study area were selected and used in inversion. We use P_n waveforms in the period range of 5-20 s. At the long-period end of our data and depending on the epicentral distance, the first arrival may overlap in time with the crustal P arrivals and Moho reflection and is sometimes referred to as P_{nl} . For brevity, we use P_n to represent both P_n and P_{nl} .

The resulting three-dimensional (3D) model exhibits similarities to and notable differences from the previous models based on ray theory. Figure 7 shows two vertical profiles of the full-wave P_n model. The base of the high-velocity mantle lid matches the lithosphere-asthenosphere boundary (LAB) determined from S-to-P receiver functions (Zhao et al., 2011; Yue et al., 2012). Lateral variations in the mantle velocity show evidence for possible on-going and previous mantle delamination. To our knowledge, this is the first full-wave P_n tomography, which extends the traditional 2D P_n models to a fully 3D model. The model is validated against ground-truth seismic events. The results show that the new model significantly improves predictions of P_n waves.

The results were presented at the 2016 GIG workshop (Bao and Shen, 2016a), the 2016 Fall AGU meeting (Bao and Shen, 2016b), and the 2017 Fall AGU meeting (Bao and Shen, 2017b). A manuscript documenting the results is in preparation (Bao and Shen, in preparation).

3.2.2 Joint Gravity and Full-wave Seismic Inversion

Previous studies of Rayleigh wave vertical-to-horizontal amplitude ratios (Z/H ratios) have demonstrated that the Z/H ratios provide powerful constraints on the crust and mantle structure (e.g., Lin et al., 2012; Yuan et al., 2016). However, these studies assumed one-dimensional velocity and density structures beneath the receiver, neglecting the effects of three-dimensional (3D) heterogeneities on wave amplitudes. The azimuthal averaging of the Z/H ratios common in those studies means reduced sensitivities to the 3D structure and thus the Z/H ratios' potential resolving power. To fully utilize the unique constraints of Z/H ratios on 3D structures near the receiver, we derived equations for the 3D full-wave sensitivity kernels of the Z/H ratio and differential traveltime of the vertical and radial components of Rayleigh and P waves, to perturbations in shear and compressional velocities and density (Bao and Shen, 2018).

We calculate the sensitivity kernels based on finite-difference modeling of wave propagation in 3D structures and the scattering-integral method. The 3D full-wave sensitivity kernels are verified by comparing the predictions from the kernels with the measurements from numerical simulations of wave propagation in models with various small-scale perturbations. In contrast to the 1D depth sensitivity kernels, our 3D full-wave kernels exhibit patterns that vary with back-azimuths and distances to the receiver, indicating the resolving power not only vertically, but also laterally. In places with dense seismic stations, these kernels provide a powerful tool to obtain accurate and high-resolution velocity and density models.

Figure 8 shows an example of the full-wave sensitivity kernels of Rayleigh waves to the vertical-radial (Z/H) component-differential traveltimes (phase delays) and amplitudes at a single seismic station (Bao and Shen, 2018). Density clearly has significant effects on both traveltimes and amplitudes of Rayleigh waves. The relatively strong sensitivity of Z/H ratios and differential-

traveltimes with respect to ρ indicates that the resulting V_s model from those observables can be significantly biased if the density structure is not accounted for in inversion. This leads to the realization of the necessity of a joint gravity and full-wave seismic inversion.

There are two possible approaches in joint seismic and gravity inversion. A common practice is to employ the *a priori* empirical relationship of P- and/or S-wave velocity-density (e.g., Maceira and Ammon, 2009; Moulik & Ekström, 2016; Syracuse et al., 2017). In those studies, density is not solved for explicitly as a separate variable and gravity observations are linked to velocity via velocity-density scaling. A caveat of this practice is that the velocity-density relation varies substantially with rock types (Gardner et al., 1974), so the empirical relationships are likely not universally applicable. Alternatively, one can use the full-wave seismic sensitivity kernels to density variations (Bao and Shen, 2018) to link gravity and seismic observations and solve for both velocity and density as separate variables in the inversion.

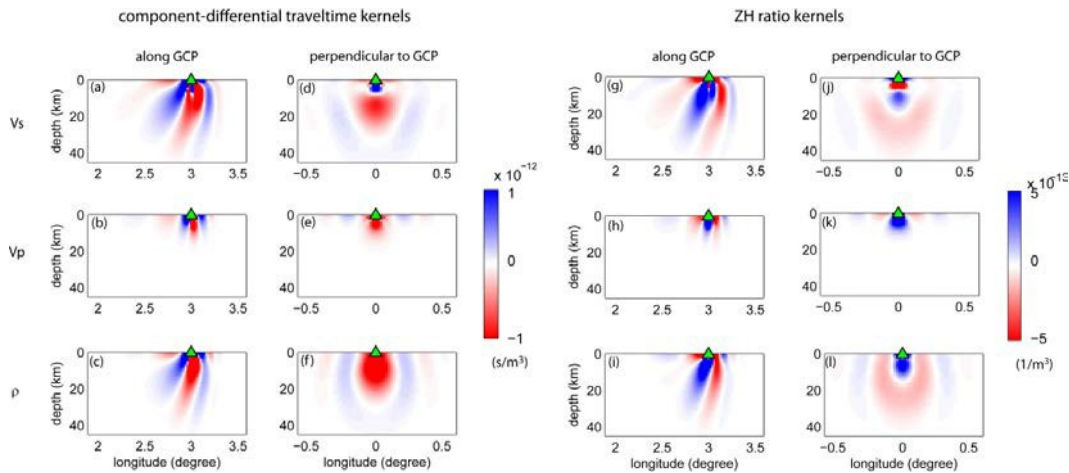


Figure 8. Component-differential traveltime (a-f, the two left columns) and ZH amplitude ratio (g-l, the two right columns) kernels for 10-20 s period Rayleigh waves on the vertical profile along the great circle path (a-c, and g-i) and the vertical profile perpendicular to the great circle path at the station (d-f, and j-l). The green triangle marks the receiver is on the surface. *The top subpanels (a, d, g, j) are the sensitivity to V_s , the middle subpanels (b, e, h, k) are sensitivities to V_p , and the bottom subpanels (c, f, i, l) are the sensitivities to density, respectively.*

We have developed a joint gravity and full-wave seismic inversion method and applied it to Southern California (Bao et al., 2018b; Guo et al., 2018). We collected waveform data from 206 local earthquakes between 2000 and 2018, and 55 permanent broadband seismic stations located in Southern California (Figure 9). We acquired the complete Bouguer gravity anomaly observations from the USGS. The 3D Earth model CVMH15.1 (Shaw et al. 2015) is used as the starting model. Numerical simulations of wave propagation are carried out with a 3D collocated-grid finite-difference method on curvilinear grids that follow surface topography. The 3D seismic kernels are calculated using the scattering-integral method. The gravity kernels are calculated based on the curvilinear grids used in waveform simulations.

Figure 10 shows the resolution of an input density anomaly at two different depths in the center of the model, demonstrating the enhanced resolving power of the joint inversion. With only the gravity constraint, the inverted density anomaly is smeared vertically (Figure 10b). The joint gravity and full-wave seismic inversion significantly reduces vertical smearing (Figure 10c). Gravity constraints, on the other hand, help reduce horizontal smearing of velocity anomalies (not shown).

We have carried out the first three iterations of the joint inversion of gravity anomaly and P- and Rayleigh-wave traveltimes. A comparison of the gravity anomaly predicted by the starting model and the inverted model is shown in Figure 11. The starting model (Shaw et al., 2015) fails to predict the observed Bouguer gravity anomalies (Figure 11b), because density in the starting model is based on the empirical relationship between velocity and density (Brocher, 2005). In contrast, the model from joint inversion significantly improves the prediction of Bouguer gravity (Figure 11c).

The joint inversion reveals low density bodies in the upper and mid crust (Figure 12). Based on their relations with surface geology and velocity anomalies (not shown), we interpret the low-density bodies in the mid crust as granitic roots. Most earthquakes in southern California (e.g., Lin et al., 2007) are located within low-density zones or in the transition zones between high-and-low density, suggesting that the quartz-rich granitic low-density (and low-shear-strength) bodies play an important role in crustal deformation.

The full-wave sensitivity kernels for Rayleigh wave Z/H ratio were presented at the 2017 AGU meeting (Shen and Bao, 2017) and published in JGR Solid Earth (Bao and Shen, 2018). The results of a joint gravity and full-wave seismic inversion were presented at the 2018 Fall AGU meeting (Bao et al., 2018b; Guo et al. 2018). A manuscript is in preparation (Bao, Guo and Shen, in preparation).

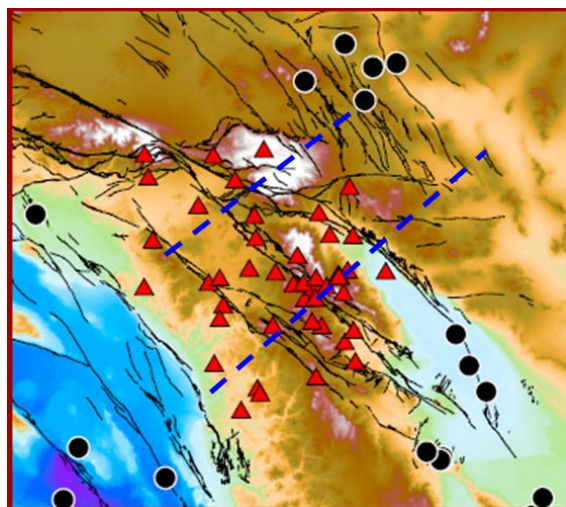


Figure 9. Map showing the topography of southern California, the seismic stations (red triangles) and earthquakes (black dots) used in the joint gravity and full-wave seismic inversion. The blue dashed lines mark the locations of the vertical profiles shown in Figure 12.

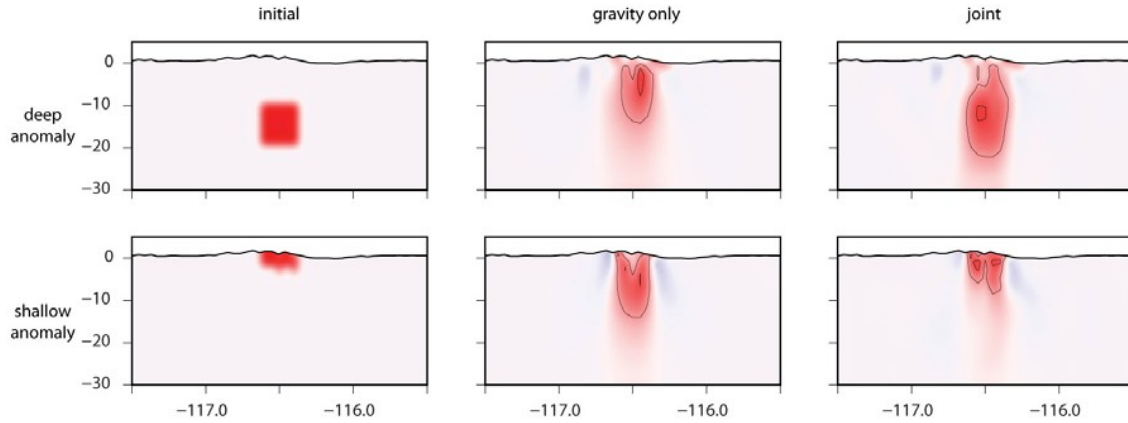


Figure 10. Density resolution showing (a) the input density anomaly, (b) the recovered density using only gravity constraint, and (c) the recovered density with both gravity and seismic constraints.

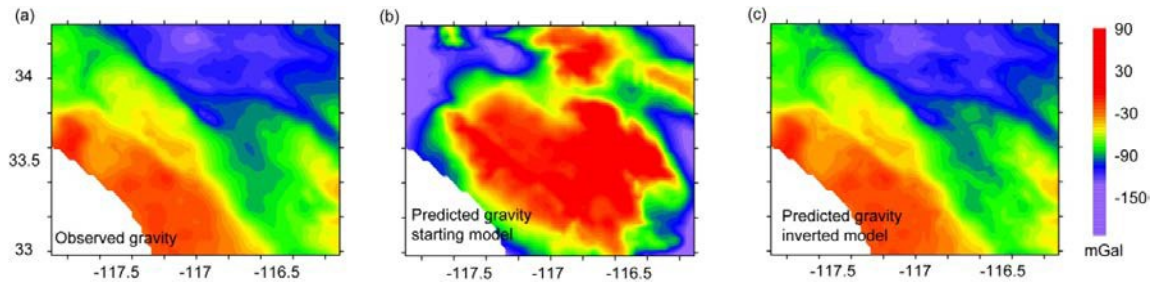


Figure 11. (a) Observed complete Bouguer gravity anomalies of southern California. (b) The gravity anomaly predicted based on the starting model CVMH15.1 (Shaw et al., 2015). (c) The gravity anomaly predicted based on the model from joint inversion after the first iteration.

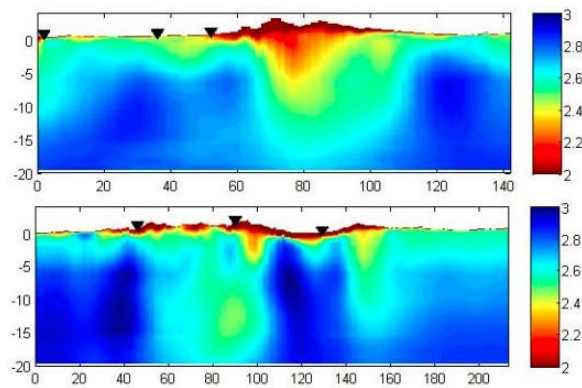


Figure 12. Density structures along the northern (upper panel) and southern (lower panel) profiles shown in Figure 9. *The inverted triangles mark the intersections of the three major fault zones in the study area (from left to right, the Elsinore, San Jacinto, and San Andreas faults). The density unit is in g/cm^3 .*

3.2.3 Full-Waveform Attenuation Tomography

To date, development of seismic attenuation models of the Earth has lagged behind those of elastic velocity models. This is, in part, due to the more complex factors that affect wave amplitude, the primary observable in attenuation studies. In addition to intrinsic attenuation, amplitudes are affected by wave focusing/defocusing, uncertainties in earthquake source excitation, site amplification, and errors in instrument calibration at the receiver (e.g., Romanowicz and Mitchell, 2007; Bao et al., 2016a,b).

Different theories have been used to invert for the attenuation structure, including the ray theory (e.g., Romanowicz, 1995; Dalton et al., 2008), the first-order perturbation theory (or Rayleigh's principle) (Smith and Dahlen, 1973), and the finite frequency theory (e.g., Tromp et al., 2005; Zhao et al., 2005; Komatitsch et al., 2016).

Building on the work on forward modeling for anelastic media described in section 3.1.1, we have developed an anelastic full waveform tomography method, in which we apply the optimal fourth-order Strong Stability Preserving Runge-Kutta method to solve the anelastic wave equation and the scattering-integral method to calculate the velocity and attenuation sensitivity kernels (Wang et al., 2017a,b; Wang et al., 2018a). We use the generalized Maxwell body model to approximate the rheological behaviors of the Earth (Moczo and Kristek, 2005) and perform forward and backward simulations in the anelastic model, taking into account the effects of both physical dispersion and dissipation. We have shown the accuracy and validity of this method through several synthetic tests. Moreover, we validate the sensitivity kernels by comparing the measurement and the prediction in homogeneous anelastic models with velocity/attenuation heterogeneity perturbations.

Figure 13 shows an example of Rayleigh wave amplitude sensitivity kernels to shear-wave attenuation, Q_β , perturbations for a station-event pair in the western U.S. A model combining the 3D velocity structure of Chai et al. (2015) and the 1D attenuation structure of PREM (Dziewonski and Anderson, 1981) is used in the wave simulation.

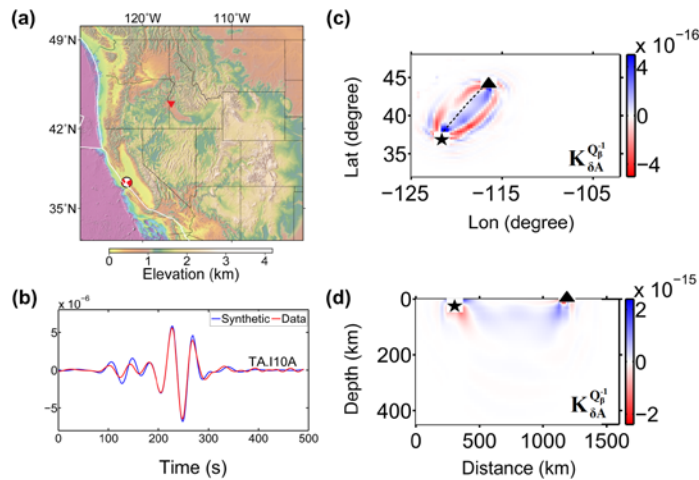


Figure 13. Map of the western United States and the locations of an earthquake and a station TA.I10A (a), comparison of the observed and synthetic seismograms at the station (b), and horizontal (c) and vertical (d) views of the Rayleigh wave sensitivity kernels in units of $1/m^3$ to attenuation, Q_β , perturbations at [0.015 0.03] Hz. The star represents the moment source and the triangle denotes the station.

We collected seismic waveforms from a total of 725 stations in northwestern U.S. in 2005-2017, including the USArray TA stations, regional seismic networks, and other temporary stations. We inspected 170 events with magnitude larger than 5.0 ($M_w > 5.0$) (from GCMT website) along the Juan de Fuca ridge and related transform faults and in northern California, for which 119 events are used for inversion and 51 events (~30% of 170 events) are selected as test events for verification of the tomographic model. Preliminary results show relatively low Q_β (high attenuation) beneath the Snake River hotspot track and parts of the Cascadia back-arc region (Wang et al., 2018a).

The development of the sensitivity kernels for attenuation and preliminary attenuation tomography of the northwestern U.S. were presented at the 2017 SSA meeting (Wang et al., 2017a) and the 2016, 2017, 2018 AGU meetings (Wang et al., 2016, 2017b, 2018a).

3.3 Integrated Deterministic and Stochastic Model

3.3.1 Power-law Distribution of Earth's Heterogeneity

One pervasive organizing principle that has emerged from geological and geophysical observations of crustal heterogeneities in the past several decades is that heterogeneities exhibit spatially self-affine, or power-law scaling, often described by the van Karman distribution (Levander, 2003). For example, a power-law distribution was found for the KTB deep wells in Germany (Wu et al., 1994) and the YT2 well in Japan (Sato and Fehler, 1998). Holliger and Goff (2003) suggested that velocity perturbation in the upper crust follows a ubiquitous $1/f$ power-law distribution, where f is the wavenumber (the inverse of wavelength). Holliger and Levander (1994) suggested that these scaling relations are powerful constraints that allow us to interpolate stochastically between observations made at different scales and to extrapolate observations made at a certain scale to other scales.

In section 3.2, we discussed full-wave tomography that yields intermediate- and large-scale regional models. As heterogeneities get smaller, they move beyond the resolution of conventional tomography at some point. However, studies have shown that small-scale heterogeneities can be represented with stochastic models and determined by comparing the observed and predicted wavefields (e.g., Wu et al., 1994). The transition from the deterministic to stochastic scales depends on resolution and thus varies from place to place and from model to model. Nevertheless we can utilize tomographic models and the scaling relations as an *a priori* constraint to determine the stochastic model for small-scale heterogeneities: The developments of full-wave tomographic models improve the model resolution and accuracy and provide the necessary constraints to interpolate stochastically to small-scale heterogeneities. This is one of the innovative aspects that differentiate our research from previous studies. Previous tomographic models did not include stochastic small-scale heterogeneities and previous stochastic models were derived without the *a priori* constraints of the deterministic models.

Figure 14 shows the power spectral density (PSD) of the upper crust in three different tectonic areas of the velocity model RSTT, a $1^\circ \times 1^\circ$ grid 3-D model for the crust and uppermost mantle generated by using regional P and S travel time data (e.g., Myers et al., 2010). At long wavelengths ($> \sim 300$ km), the structure can be considered deterministic, albeit with variable

resolutions and uncertainties for the different tectonic regions. The Western U.S. and Iran have relatively high PSDs, reflecting large velocity variations that are consistent with active tectonics in both regions. In contrast, the North American craton has the lowest PSD, indicating a more uniform upper crust that is consistent with the tectonically stable core of the North American continent. At wavelengths shorter than 200 km, the velocity structure is unresolved in RSTT.

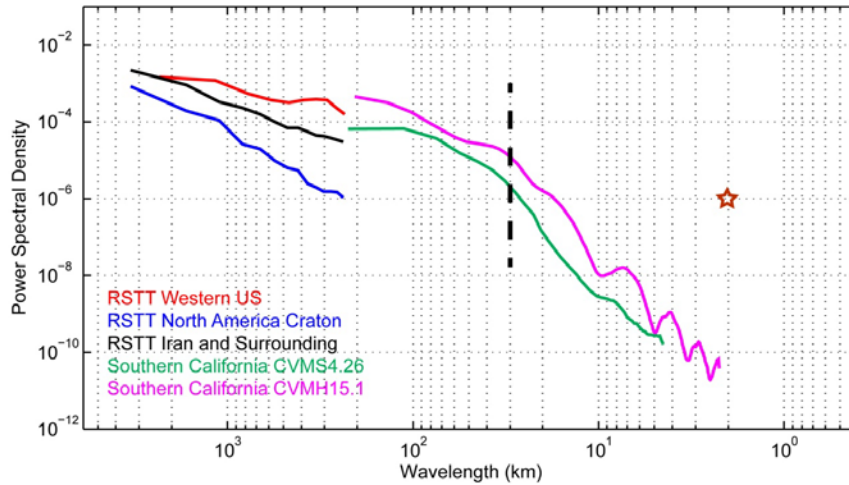


Figure 14. PSDs of the velocity structure in the upper crust. *The western U.S., Iran and surrounding areas, and North American Craton are from the RSTT velocity model. CVMS4.26 and CVMH15.1 are high-resolution velocity models of southern California. Vertical black dashed line marks the wavelength where the resolution of the local models starts to diminish towards shorter wavelengths. Brown star shows the PSD at wavelength of 2 km of the stochastic model that best predicts the observed PEP_T (see Figure 15).*

Also presented in Figure 14 are two high-resolution full-wave velocity models for southern California (CVMS4.26 and CVMH15.1; Lee et al., 2014; Plesch et al., 2009; Shaw et al., 2015), a part of the Western U.S. In these local models, the velocity structures are considered well resolved at wavelengths greater than ~ 30 km. The obvious deepening of the PSDs around 30 km wavelength in both models reflects a diminished resolution at smaller scales. The difference in the levels of the PSDs of these two models is likely due to the choices of regularization in tomography, thus reflecting the level of uncertainties of the models, and consequently the uncertainty in the extrapolation of the power-law distribution to shorter wavelengths (< 30 km). Our joint gravity and full-wave seismic inversion (Bao et al., 2018b; Guo et al., 2018) yields significant changes in density (section 3.2.2), but the relatively small perturbations to the initial velocity model (CVMH15.1, Shaw et al., 2015) do not change the velocity spectra significantly.

It is worth emphasizing several important observations in Figure 14: (1) Different tectonic regions have different levels of heterogeneities as shown by the PSDs. This defines the likely range of parameters for the stochastic model at shorter wavelengths for a given tectonic region. (2) Within the resolved spectra of the models, the PSDs follow a $(1/f)$ power-law distribution, consistent with previous observations of other geological and geophysical features (Holliger and Goff, 2003). This provides a strong justification to extrapolate the power-law distribution of the resolved model to shorter wavelengths. (3) The higher the resolution and the more accurate the

velocity model, the lower the uncertainty of the extrapolation of the power-law distribution to small scales for the stochastic model. (4) The PSDs of the resolved spectra of the western U.S., the local tomographic model (CVMH15.1) and the stochastic model follow a linear trend in Figure 14, which is consistent with the power-law distribution.

By comparing observed waveforms and full-wave synthetics of the deterministic models, we can further assess the uncertainties in the parameters of the scaling distribution (i.e., a model that predicts well-matched synthetics to the observations has low uncertainties in the scaling distribution). Thus, as we invert for the stochastic model by comparing observed and simulated scattered waves, the scaling relation from the deterministic portion can be used as our *a priori* constraint. In addition, by including the deterministic model in wave simulations as in Furumura et al. (2014), one can account for wave focusing/defocusing due to the deterministic structure and thus better isolate the effects of scattering due to small-scale heterogeneities. The resulting integrated model will have a continuous PSD from the deterministic part to the stochastic part. Such an integrated model can be used to predict travel times and long-period waveforms deterministically and high-frequency scattered wave coda statistically.

3.3.2 Stochastic Models

Previous studies have shown that small-scale heterogeneities can be described with stochastic models based on statistical properties of the medium instead of the exact location of each scatterer (e.g., Frankel and Clayton, 1986) and the observed and predicted wave fields can be statistically compared (e.g., Takemura and Furumura, 2013; Olsen et al., 2018). Because stochastic models generally involve a few variables, determination of regionalized stochastic Earth models is achievable (e.g., Kubanza et al., 2007).

Superimposed on a deterministic velocity background, a stochastic random velocity fluctuation can be incorporated into a 3-D volume. This can be described as

$$V(x, y, z) = V_0(x, y, z)[1 + \xi(x, y, z)]. \quad (1)$$

where $V_0(x, y, z)$ is deterministic and obtained from seismic tomography or other methods and $\xi(x, y, z)$ is the stochastic part. The property of velocity fluctuation is commonly characterized by three types of correlation functions: the Gaussian, the exponential, and the von Karman. Previous studies demonstrated that the von Karman-type PSD can describe the fluctuation of velocity structures in the crust and upper mantle well (e.g., Frankel and Clayton, 1986). The PSD of the von Karman correlation function is

$$P(k) = \frac{8\pi^{3/2}\varepsilon^2 a^2 \Gamma(\kappa+3/2)}{\Gamma(\kappa+3/2)(1+k^2 a^2)^{\kappa+3/2}} \quad (2)$$

in 3D, where ε is the strength of fluctuation, a the correlation distance, k the wavenumber, and κ the decay rate or the Hurst number.

Determining Stochastic Models Using Individual Three-component Stations

Scattering observed at individual stations experiences the cumulative path effects from the source to the receiver, and thus reflects regional small-scale heterogeneities. By examining the P-wave energy on the transverse component of short-period seismograms in Japan, described as

$$PEP_T = \frac{E_T}{E_Z + E_R + E_T}, \quad (3)$$

Takemura and Furumura (2013), for example, found that scattering in the crust and upper mantle in Japan can be characterized with a von Karman PSD having $a=3-10$ km, $\varepsilon=0.02-0.09$, and $\kappa=0.3-0.8$. By using PEP_T , the effect of intrinsic attenuation along propagation is canceled.

It has been suggested that the stochastic distribution of heterogeneities changes with depth (e.g., Kennett, 1987; Olsen et al., 2018). Thus, we divide the lithosphere into three layers (the upper crust, the lower crust and the mantle) and carry out the inversion for the stochastic model of each layer in three corresponding steps:

- (1) Determine the stochastic model parameters of the upper crust by using the PEP_T measurements from the P-waves propagating within only the upper crust. We first select the earthquake sources located in the upper crust. Then we use ray tracing to determine the P-wave turning point for each source-station ray path and select the source-station pair with the turning point in the upper crust. Using the PEP_T measurements from the selected source-station pairs, we follow a Monte Carlo inversion (Savran and Olsen, 2016, and see below for more detail) to search for the optimal stochastic parameters of the upper crust, including the horizontal and vertical correlation lengths, the velocity fluctuation, and the decay rate (Hurst number). The tomographic model of the region is used to constrain the range of the stochastic model parameter search and reduce computational costs.
- (2) Select the source-station pairs with P-wave turning points in the lower crust and use the new PEP_T measurements to invert for the stochastic parameters of the lower crust while keeping the stochastic model of the upper crust from step 1 unchanged.
- (3) Similar to (2) to invert for the stochastic parameters of the mantle while keeping the crustal stochastic models from the steps 1 and 2 unchanged.

We have tested this method in Southern California. We performed a series of waveform simulations for 10 shallow earthquakes recorded by seismic stations in the region. Figure 15(a) and 15(b) show the V_p in the reference tomographic model CVMH15.1 and an integrated model of CVMH15.1 and stochastic heterogeneities. A Monte Carlo inversion (Savran and Olsen, 2016) is used to search for the stochastic model that maximizes the variance reduction of PEP_T :

$$VR(\%) = 100 - 100 \times \frac{\sum (PEP_T^{obs} - PEP_T^{syn})^2}{\sum (PEP_T^{obs})^2}, \quad (4)$$

where the superscripts *obs* and *syn* denote the PEP_T measured from recorded and synthetic seismograms, respectively. We search the upper crust stochastic model with the Hurst number κ within 0.1-0.5, the horizontal correlation length a_x within 2.5-20 km, and velocity fluctuation ε

within 0.02-0.3. We don't search for the vertical correlation length a_z , since the vertical grid spacing in our model is already larger than 54 m estimated from sonic logs in the Los Angeles Basin by Savran and Olsen (2016). Tests show that increasing a_z typically reduces variance reduction of PEP_T . We observed that PEP_T variance reduction is strongly influenced by the value of velocity fluctuation. We find that a velocity fluctuation of 0.12 provides a maximum PEP_T variance reduction. A grid search of different horizontal correlation length a_x and the Hurst number κ is performed to determine the optimal stochastic model. A stochastic model with the von Karman distribution of $\kappa=0.3$, $a=15$ km, and $\varepsilon=0.12$ yields a maximum PEP_T variance reduction of 69%. This solution is close to the estimation from sonic log analysis and V30 measurements in which $a=5$ -10 km and $\kappa=0.3$ (Savran and Olsen, 2016).

Figure 15(d) shows the power spectra of the upper crust in the deterministic model CVMH15.1, the power spectra of the stochastic models generated from the Monte Carlo inversion, and the power spectra of the stochastic model that provides the maximum PEP_T variance reduction (the optimal model). We find that the power spectra of the optimal model follow the same trend of the deterministic model at long wavelengths (i.e., larger than ~ 100 km, and Figure 14). The power spectra of the models yielding PEP_T variance reductions above 60% are located around that of the optimal model, suggesting the robustness of the optimal stochastic model. This validates our strategy of using long wavelength spectra to predict short wavelength spectra.

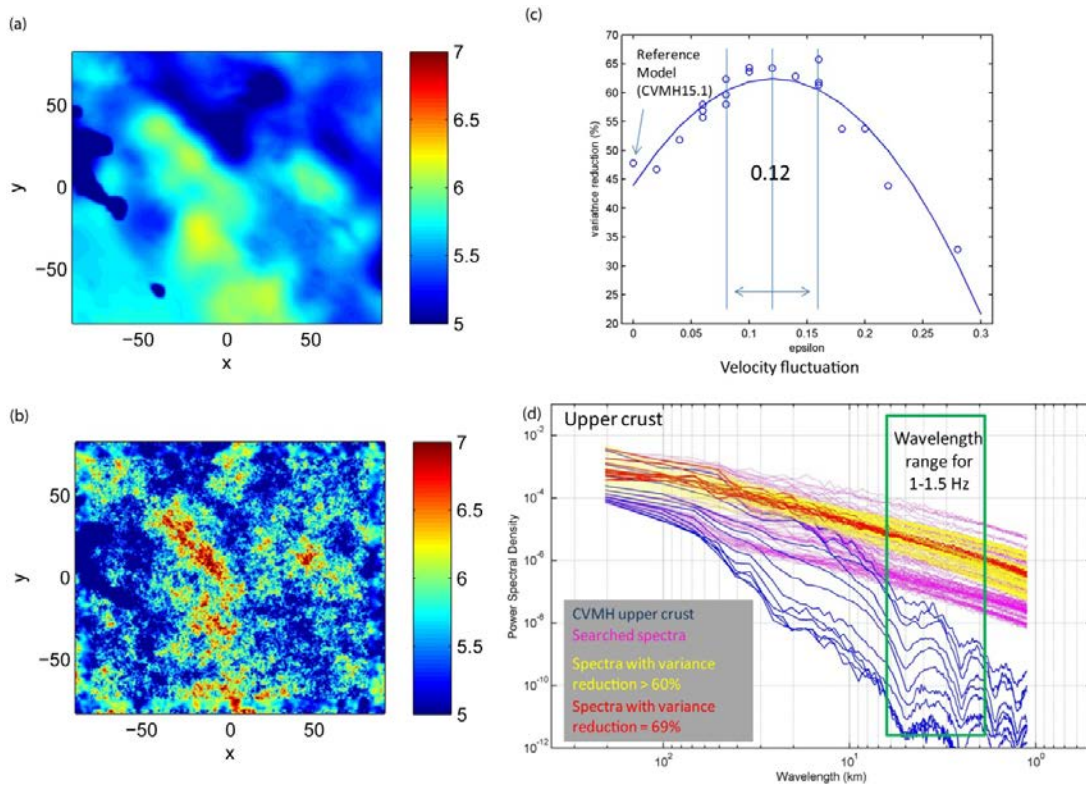


Figure 15. (a) Lateral variation of V_p at 3 km depth in the tomography model CVMH15.1 of southern California. (b) Same as (a) but for an integrated model of CVMH15.1 plus a stochastic model with small-scale heterogeneities. (c) PEP_T variance reduction versus velocity fluctuation ε added into CVMH15.1.

Each blue circle denotes the PEP_T variance reduction from one stochastic distribution. Blue curve is the polynomial-fitting curve, which has a peak at the velocity fluctuation of 0.12. The two light blue vertical lines outline the range of velocity fluctuation yielding PEP_T variance reductions greater than 60%. (d) Power spectral density distribution of V_p in the upper crust of CVMH15.1 (blue curves, above 16 km depth) and those of the integrated model. Magenta lines represent the spectra of the searched models; Yellow lines denote models with the PEP_T variance reduction greater than 60%. The best-fitting models (red lines) have a PEP_T variance reduction of ~69%. The green rectangle outlines the wavelength range corresponding to the frequency band in which PEP_T is measured.

Determining Stochastic Models Using Data Recorded by Seismic Arrays

Previous studies of stochastic models mostly relied on local or regional earthquakes (e.g., Takemura and Furumura, 2013). This limits their applicability to regions of active seismicity. Kubanza et al. (2007) analyzed the amplitude of P waves on the transverse component assuming a vertically incident plane wave. They concluded that variations in the small-scale heterogeneities correlate with tectonic settings. Korn (1997) studied the effects of small-scale heterogeneities on teleseismic P coda envelopes with an energy-flux model. With full waveform simulation, Korn (1997) concluded that the energy-flux model yields reliable results. Both Korn (1997) and Kubanza et al. (2007) did not consider the effects of surface topography.

In order to broaden the investigation to both seismic and aseismic regions and to understand the effects of surface topography, we investigated whether energy partition of teleseismic P-waves on the transverse component (PEP_T) along a linear array provides us information about the stochastic nature of the media beneath the stations (Wei et al., 2018). Figure 16 shows a linear seismic array across the eastern margin of the Tibetan Plateau, where the highly deformed mountain belts are juxtaposed with the stable Yangtze Craton to the east. Across the boundary between the craton and the plateau, there is an abrupt change in the PEP_T values: the values are low within the craton (except for one station) and high on the plateau. Such an abrupt transition in waveform characteristics cannot be attributed to the source-side scattering. Simulations of teleseismic wave propagation across the array for a model with high resolution topography show that scattering by topography can account for most of the PEP_T variations in the craton, but is insufficient to explain the level of scattering on the plateau, suggesting that small-scale heterogeneities are needed to account for the higher level of scattering on the plateau. The results match with our expectations for the stable craton and the highly deformed mountain ranges. Similar to the study in southern California (Figure 15), stochastic models for the plateau and the craton can be determined by fitting observed and synthetic PEP_T .

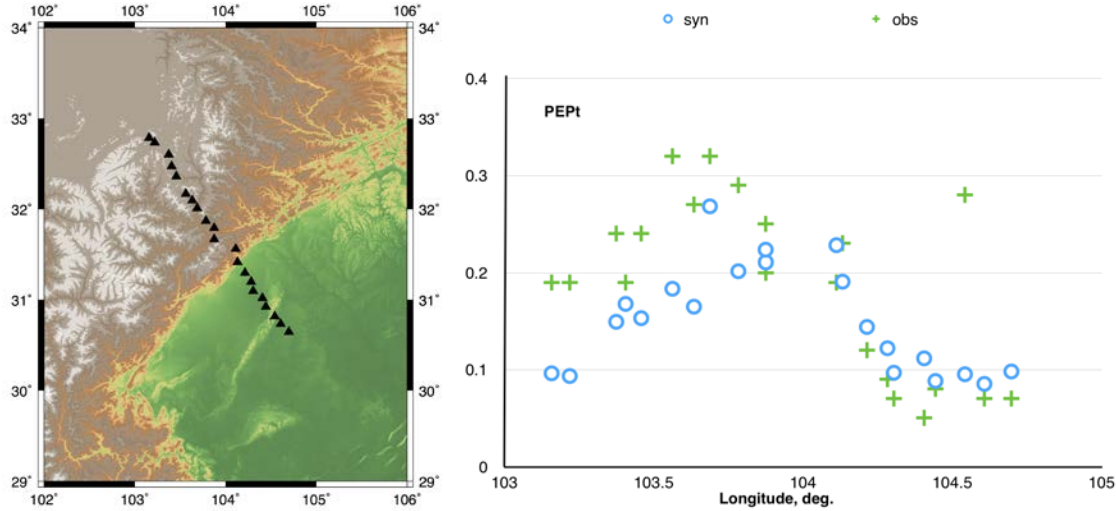


Figure 16. (left panel) Location of the linear seismic array across the eastern margin of the Tibetan plateau. (right panel) Observed (crosses) and simulated (circles) PEP_T values.

In places with 2D seismic arrays, another way to determine the level of small-scale heterogeneities is analysis of waveform differences among nearby stations in the array, which reflect local small-scale heterogeneities beneath the array. A seismic survey in 2014 near Sweetwater, TX (Barklage et al., 2014) included 25 broadband sensors and ~ 2600 short-period sensors. The array recorded multiple earthquakes in Oklahoma. Visual inspection of the waveforms found that the Pn arrivals are nearly identical, though scattering yields observable differences in Pn coda waves. To characterize the differences, we apply an Empirical Orthogonal Function analysis (EOF), which highlights the principle elements in data sets. We calculate the first and second eigenvalues of a sample matrix, which includes all the waveforms of all stations near a center station. The vector corresponding to the first singular value represents the average waveform, while that of the second singular value represents the first-order dissimilarities of the waveforms, which is considered as a measure of scattering strength. Figure 17 shows the lateral variation of the second singular value for the real data and the synthetics from LITHO1.0 (Pasyanos et al., 2014) and LITHO1.0 plus random heterogeneities, respectively. Our goal is not to match the observed and predicted singular values point-by-point, but to have a predicted variation that matches the observed statistically. We note that the average 2nd singular value increases with the level of random heterogeneities and that the strength of scattering in synthetics may approach that in observations when the magnitude of heterogeneities is increased.

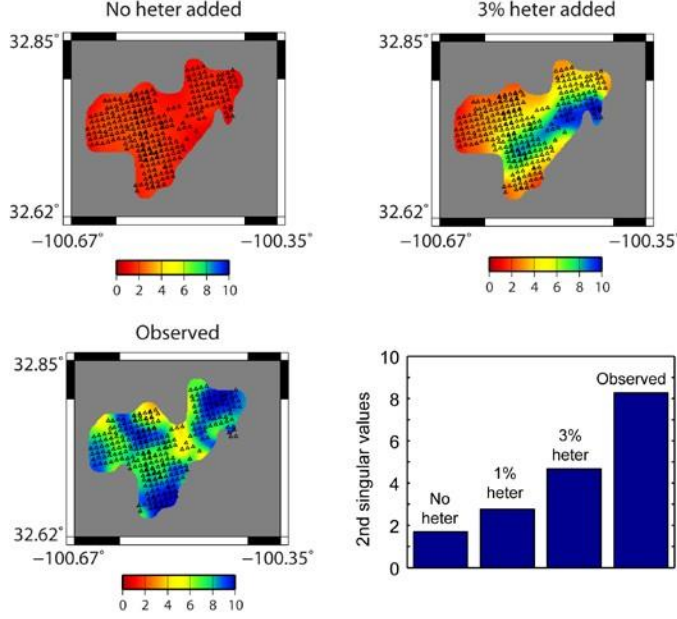


Figure 17. Distribution of the 2nd singular value for the synthetics of the model LITHO 1.0 (a), LITHO 1.0 plus 3% small-scale heterogeneities (b), the observed (c), and the average values (d). Comparing to the strong lateral variation measured for real data, LITHO1.0 yields very weak lateral variations. With the added random heterogeneities, the model produces a stronger lateral variation in the second singular value, but the value is still weaker than that of real data.

When the singular values from the EOF analysis for multiple earthquakes are compared, an optimal estimation for the stochastic parameters will be determined by minimizing the error

$$E(\kappa, a_m, \varepsilon) = \frac{\sum_{j=1}^M \sum_{i=1}^N [S_1^{obs}(i,j) - S_1^{syn}(i,j)]^2}{\sum_{j=1}^M \sum_{i=1}^N [S_1^{obs}(i,j)]^2} + W \frac{\sum_{j=1}^M \sum_{i=1}^N [S_2^{obs}(i,j) - S_2^{syn}(i,j)]^2}{\sum_{j=1}^M \sum_{i=1}^N [S_2^{obs}(i,j)]^2}, \quad (5)$$

where S_1 and S_2 are the first and second singular values, respectively. M and N are the total number of sources and stations. W is a weighting factor. The subscript m denotes the three components of correlation length, where we assume $a_x = a_y$ but both are larger than a_z . The EOF method provides a robust way to measure the magnitude of scattering for the whole network.

The results were presented at the 2017 and 2018 SSA meetings (Bao and Shen, 2017a; Bao et al., 2018a) and the 2018 AGU meeting (Wei et al., 2018).

4. CONCLUSIONS

We took a comprehensive, three-pronged approach to develop, validate and apply full-waveform methods to improve velocity and attenuation models and to integrate deterministic earth models with stochastic models to better predict broadband seismic waveforms. We developed a new computationally efficient and accurate finite-difference method (Wang et al., 2018b) to calculate wave propagation in anelastic media to better account for intrinsic and scattering attenuation. Preliminary full-wave attenuation tomography of the western U.S. reveals relatively high attenuation beneath the Snake River hotspot track and part of the Cascadia back-arc (Wang et al., 2018a). We also developed a passive-source reverse time migration method (Li et al., 2018) to image velocity discontinuities and possibly strong scatterers in order to better characterize the regional and global mantle transition zones and small-scale heterogeneities. A first application of this full-wave migration method results in greatly improved images of the mantle transition zone in the Yellowstone region (Li and Shen, in preparation). We carried out a full-wave Pn

tomography of the central and northern Tibetan plateau (Bao and Shen, 2016a,b, 2017b), extending Pn tomography from the traditional 2D approach to 3D. Full-wave 3D sensitivity kernels for Rayleigh ellipticity are developed to better image shallow crust and mantle structures (Bao and Shen, 2018). A joint gravity and full-wave seismic inversion greatly improves the density structure in southern California (Bao et al., 2018b; Guo et al., 2018). We simulated wave propagation between the North Korea nuclear test site and the station MDJ for several stochastic models and showed that traveltimes and amplitudes are significantly affected by small-scale heterogeneities. We studied the distribution of velocity heterogeneities in southern California. We observed that the power spectral densities of the resolved spectra of the western U.S., the local tomographic model and the stochastic model from modeling of scattered waves follow a linear trend consistent with the power-law distribution (Bao and Shen, 2017a; Bao et al., 2018a), providing strong justification for extrapolating the power-law distribution of the resolved tomographic model to shorter wavelengths.

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LIST OF SYMBOLS, ABBREVIATIONS, AND ACRONYMS

AFRL	Air Force Research Laboratory
AFSPC	Air Force Space Command
AFWA	Air Force Weather Agency
AGU	American Geophysical Union
AW	Adjoint Wavefield method
CCP	Common Conversion Point
EOF	Empirical Orthogonal Function
GCMT	Global Centroid Moment Tensor
GSN	Global Seismic Network
GEOFON	Global Seismic Network operated by Helmholtz-Zentrum Potsdam
HPC	High-Performance Computing
IRIS	Incorporated Research Institutions for Seismology
IRIS DMC	IRIS Data Management Center
LAB	Lithosphere-Asthenosphere Boundary
MTZ	Mantle Transition Zone
NSF	National Science Foundation
PEP _T	P-wave Energy Partitioning on the Transverse component
PML	Perfectly Matched Layer
PSD	Power Spectral Density
PS-RTM	Passive-Source Reverse Time Migration
RTM	Reverse Time Migration
SEM	Spectral Element Method
SI	Scattering Integral method
SSA	Seismological Society of America
SSPRK	Strong Stability Preserving Runge-Kutta method

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