Waveform inversions of SS precursors: An investigation of the northwestern Pacific subduction zones and intraplate volcanoes in China

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Abstract

The arrival time and amplitude of long-period underside reflection from mantle seismic discontinuities (SS precursors) have made major contributions to the understanding of mantle composition and dynamics. In this study, we introduce a nonlinear waveform inversion technique to simultaneously constrain shear velocities and discontinuity depths beneath the northwestern Pacific subduction system. Based exclusively on a large SS precursor waveform dataset, we are able to clearly delineate the morphology of the descending Pacific plate, which flattens at the base of the upper mantle and extends westward by at least ~1500 km toward northern-central China. Our grid search over a range of angles indicates a maximum correlation between shear velocity and transition zone thickness at ~30 degrees, consistent with the reported average slab dip beneath the study region. The strongly positive correlation suggests predominantly thermal, rather than compositional, variations along the descending Pacific plate. Our joint depth-velocity solutions also suggest a 5-10 km depression of the 410 km discontinuity and an average decrease of 1.2% in upper mantle shear velocity beneath the intraplate volcanic fields in northeastern China. This anomaly, which reaches the middle of the upper mantle transition zone beneath the Changbai hotspot, initiates at a significantly shallower (~320 km) depth beneath the Wudalianchi regions. The high amplitude reflection at depths greater than 410 km suggests a water-poor melt layer in possible association with 1) decompression melting from passive upwelling and/or 2) active upwelling through a slab window.

Keywords: SS precursor, Mantle transition zone, Northwestern Pacific, Intraplate volcanism, Waveform inversion

1. Introduction

Intraplate volcanic activities have been well documented in both continental and oceanic regions at distances hundreds to thousands of kilometers away from plate boundary zones. The origin and mechanism of intraplate volcanism vary broadly (Niu, 2005; Zhao, 2007; Tang et al., 2014) and often require the presence of deep mantle plumes (Campbell, 2007; Chen et al., 2007; Zhao, 2007). An ideal laboratory for the study of intraplate volcanism is northeastern Asia, where Cenozoic magmatic centers are densely distributed along the north-south oriented Changbai Mountain range and Wudalianchi volcanic field in northeastern (NE) China. The former is a stratovolcano located approximately 1200 km west of the Japan trench, while the latter consists of cinder volcanoes covering an area of 500 km² toward the north. The origin of these volcanic fields has been linked to mantle plumes as well as subduction-related back-arc spreading and thinning of the lithosphere (Basu et al., 1991; Niu, 2005), though compatible helium isotopic compositions between the Cenozoic basalts from the same region and mid-ocean ridge basalt favor an upper mantle origin (Chen et al., 2007). Further insight was offered by seismic tomography where a horizontally deflected and stagnant Pacific plate at the base of the upper mantle could play a key role in melt generation (Gorbatov and Kennett, 2003; Zhao et al., 2004; Obayashi et al., 2006; Lebedev and van der Hilst, 2008; Li and van der Hilst, 2010).

Models of seismic velocities are complemented by observations of mantle transition zone (MTZ) discontinuities. For the upper mantle assemblage of olivine composition, phase transitions from olivine to wadsleyite and ringwoodite dissociation are widely accepted origins of the 410 km discontinuity (from here on, 410) and 660 km discontinuity (from here on, 660), respectively, at the top and bottom of the MTZ (Anderson, 1967; Ito and Takahashi, 1989). These two mineralogical phase boundaries exhibit opposite Clapeyron slopes (Navrotsky, 1980; Ito and Takahashi, 1989; Katsura and Ito, 1989; Weidner and Wang, 1998), and their sensitivity to temperature and composition have been frequently explored in global seismic imaging (Shearer, 1993; Gu et al., 1998; Gu and Dziewonski, 2002; Lawrence and Shearer, 2006a; Deuss, 2009). Based on mantle reflections (Heit et al., 2010; Gu et al., 2012) and conversions (Li and Yuan, 2003; Liu et al., 2015), a depression in excess of 30 km has been observed at the base of the MTZ beneath the intraplate volcanic fields in NE China. This topography anomaly coincides with a distinctive low velocity asthenosphere, which has been interpreted as the potential source of melting beneath the volcanic centers in NE China (Zhao et al., 2004; Lei and Zhao, 2005; Niu, 2005; Li and van der Hilst, 2010; Tang et al., 2014).

A known source of error in the independent analyses of seismic velocity and discontinuity topography is the trade-off between them (Flanagan and Shearer, 1998; Gu and Dziewonski, 2002; Obayashi et al., 2006; Li et al., 2008; Zhao et

al., 2004; Li and van der Hilst, 2010). Time corrections are typically adopted to minimize the excess topography caused by heterogeneous mantle structures, whereas models of seismic velocities are mostly derived under the assumption of unperturbed mantle interfacial depths. This trade-off was reduced by Gu et al. (2003) and Lawrence and Shearer (2006a) through joint inversions of seismic velocity and discontinuity topography, though much of the information embedded in the waveforms of the secondary reflections were underutilized. In this study, we characterize the upper mantle and MTZ beneath the northwestern Pacific region (Figure 1) using waveform inversions of stacked *SS* precursors (Figure 1d). Our full waveform nonlinear inversion approach enables a simultaneous solution for the travel times of *SS* precursors, which are sensitive to mantle temperatures surrounding olivine phase boundaries (Ohtani et al., 2004; Deuss, 2009; Lessing et al., 2014), and the impedance contrasts imprinted onto the *SS* precursors amplitudes (Shearer, 1991; Chambers et al., 2005; Gu and Sacchi, 2009; Lessing et al., 2015). We will demonstrate that a dense precursor dataset alone is sufficient to resolve major upper mantle seismic anomalies in the northwestern Pacific subduction system.

2. Data and Method

We utilize a global dataset of broadband and long-period seismograms, recorded between 2006 and 2014, from the Incorporated Research Institutions for Seismology (IRIS). The midpoints of the source-receiver pairs densely sample the structure beneath NE China and the northwestern Pacific subduction zones. We restricted the maximum depth of earthquakes to 75 km to mitigate the interference between *SS* precursors and their depth phases (Schmerr and Garnero, 2006; An et al., 2007), and adopt a minimum magnitude (M_w) cutoff of 5.5 to ensure sufficient reflection amplitudes. We further constrain the distance from 100° to 160° to minimize the interferences from *ScSScS* (Shearer, 1993; Schmerr and Garnero, 2006) and topside reflections from upper mantle discontinuities. After deconvolving the instrument responses, we apply a Butterworth bandpass filter with corner periods at 15 s and 75 s to the transverse component seismograms. We eliminate all traces with signal-to-noise ratio (SNR) less than 4.0 according to the definition of Gu et al. (2012), which is more restrictive than the majority of earlier studies due to a substantially larger data volume. The filtered seismograms are then inspected visually to eliminate duplicate records or overlapping events (Schmerr and Garnero, 2006), and a polarity reversal is performed if necessary. This refined dataset contains 7868 waveforms from 1301 events. The densities of earthquakes and stations are particularly high in the southern and northern hemispheres, respectively (see Figure 1a).

The amplitudes of *SS* precursors, which are comparable to those of noise, are greatly enhanced through stacking after aligning the waveforms on the first major swing of *SS* and normalizing each seismogram with respect to its amplitude. We then deconvolved *SS* from the transverse component seismograms to minimize source effects. To correct for the move-out, a constant time shift is applied to each waveform based on the differential time between the predicted *SS-S410S* (Dziewonski and Anderson, 1981) and the reference time for a source-receiver distance of 130° (Gu and Sacchi, 2009). The move-out corrected seismograms are then time shifted to account for variations in crustal thickness and surface topography using CRUST1.0 (Laske et al., 2013) and ETOPO1 (Amante and Eakins, 2009), respectively. Finally, the time-shifted seismograms are sorted into common midpoint (CMP) circular bins (Shearer, 1991; Gu et al., 2012) along three parallel great-circle transects (profiles A to C) with 3° inline spacing (see Figure 1b). Circular bins with a radius of 2° (approximately 15% overlap between two adjacent bins in the inline direction) are adopted to prevent over-smoothing while providing a sufficient number of traces in each bin for noise suppression. The stacked seismogram in each cap is calculated as the weighted sum of time-corrected seismograms (Gu and Sacchi, 2009). It is worth noting that the majority of the source and receiver locations do not overlap with topographic anomalies, e.g. the Rocky Mountains, Alps and Apennines, that may cause confounding observations within the MTZ (Figure 1c). Moreover, while the

azimuthal distribution is dominated by two perpendicular orientations (NE, NW), the crossing ray segments effectively average out the path heterogeneity and enable a reasonable recovery of the velocity and discontinuity depth in most bins (Zheng and Romanowicz, 2012).

[Figure 1 about here.]

Figure 1: (a) Global distribution of the earthquake (red stars) and station (blue triangles) locations used in this study. (b) Distribution of the *SS* bouncepoints from the earthquake-station pairs. The convergent plate boundaries and slab contours are indicated by the red and blue lines, respectively. The contour lines are taken at a constant interval of 50 km starting at 100 km depth (Hayes et al., 2012). The *SS* precursor waveforms are stacked into 30 bins along three parallel profiles A, B and C. (c) Path coverage of the *SS* precursors that are used in the stacking and inversion procedures. (d) A graphical representation of theoretical ray paths of *SS* and its precursors for a source-receiver distance of 130.7 degrees. The left panel shows an observed seismogram aligned on the maximum amplitude of the *SS* phase. *S410S* and *S660* are marked based on the predicted arrival times of PREM (Dziewonski and Anderson, 1981).

2.1. Nonlinear Waveform Inversion

A potential pitfall during an *SS* precursor analysis is the velocity-discontinuity depth trade-off, as the timing of the secondary reflections is corrected based on relatively smooth shear velocity models obtained from earlier studies of body and/or surface waves. An improvement was proposed by jointly inverting for velocity and discontinuity depth using travel times (Gu et al., 2003) and multiple waveforms (Lawrence and Shearer, 2006a). In this study, we quantitatively investigate the waveform information of an *SS* precursor dataset sampling the northwest Pacific region. We model the full waveforms of *SS* precursors using the Genetic Algorithm (GA), an effective nonlinear inversion technique (Stoffa and Sen, 1991; Haupt and Haupt, 2004). A similar waveform inversion approach was recently utilized to model the waveforms of P-to-S converted phases (Chen et al., 2015), though the emphasis of that study is on the continental crust rather than the MTZ (this study).

As a proof of concept we first apply nonlinear inversions to a synthetic seismogram computed using GEMINI 2.2 (Friederich and Dalkolmo, 1995). The Green's function calculation assumes a spherically symmetric Earth model and uses a frequency domain method to numerically solve a system of ordinary first-order differential equations (Friederich and Dalkolmo, 1995; Friederich, 1999; Fichtner and Igel, 2008). We then utilize GA, a global optimization approach simulating the process of natural evolution (Haupt and Haupt, 2004), to explore solutions of mantle velocity and discontinuity depths that match the simulated

waveform with 10% noise. We apply a binary-coded genetic algorithm within
 which variables are defined as indices to a regular discrete search space. Every
 model parameter (discontinuity depth and velocity perturbations) is represented
 by a binary string, while the inversion attempts to minimize the following cost
 function by adjusting the model parameters:

$$J = \|W(d_{obs} - d(I\hat{m}))\|_2^2 + \mu \|DI\hat{m}\|_2^2 \quad (1)$$

where I is the interpolation operator sampling the model $\hat{m}$, consisting of shear
 wave velocities at discontinuities, at a constant depth interval and D is a discrete
 approximation to the first-order derivative. The symbols d_{obs} and $d(I\hat{m})$ represent
 the observed and synthetic seismograms, respectively, and W is the weighting
 matrix. Finally, μ is a regularization parameter controlling the trade-off between
 the first and second terms (i.e., misfit and model norm, respectively), and is
 determined through trial and error. Our model solution consists of 6 layers,
 ranging from 220 km to 720 km along the depth axis, and the average spacing is
 83.3 km. The S velocity is allowed to vary by $\pm 4\%$ from PREM, which is larger
 than the frequently reported range below the lithosphere - the primary target of
 this study (Grand et al., 1997; Mégnin and Romanowicz, 2000; Fukao et al.,
 2001; Gu and Dziewonski, 2002; Gu et al., 2003; Zhao et al., 2004; Obayashi et
 al., 2006; Li and van der Hilst, 2010; Ritsema et al., 2011). Due to the
 interference of the crustal phases and the side-lobes of the main SS arrival, we

mainly target the waveforms corresponding to the structure below 200 km depth. Our experiments suggest that an average of 1.5% shear velocity perturbation in the shallow mantle (above 220 km), which is comparable to the reported value beneath Northeast China (see Simmons et al. (2010) and Tang et al. (2014)), can cause a 2 km movement on the MTZ discontinuity depth and a perturbation of 0.01 km/s (or less) on the MTZ velocity. Neither artifacts are significant enough to alter the main observations of this study.

We fix the ratio between density and shear velocity based on PREM, and perturbations in density are continuously updated as shear velocities vary during the inversions; a subjective choice that scales between the velocity and density is often adopted to reduce computational cost (Marone et al., 2007; Yuan and Romanowicz, 2010; Liu and Gu, 2012). In general, the arrival times are highly sensitive to velocity and discontinuity depth, whereas the amplitude information embedded in the waveform offers constraints both on velocity and density. While density plays a minor role due to the strong dependence of the objective function on the phase/timing information, a reasonable input density model (PREM in this study) helps stabilize the solutions, especially the shear velocities.

Figure 2 shows the results of the nonlinear inversion based on synthetic data with 10% Gaussian noise. Our input model is modified from PREM, which consists of a low velocity layer atop the 410 and a high velocity layer at the base of the MTZ. We allowed 120 solutions in each generation to ensure diversity at a

manageable computation cost and to steer clear of local extrema. The input and recovered models (see Figure 2b) exhibit only minimal deviations, which are mainly associated with discretized, relatively coarse, model space. The model evolution history suggests almost immediate convergence to the final model from an initial suite of random models (Figure 2d).

[Figure 2 about here.]

Figure 2: (a) An input transverse component seismogram with 10% noise (gray) and the inverted trace after 15 generations (black). The precursors are magnified by a factor of 3 and the difference (data-synthetic) seismogram has been offset by -0.08 for clarity. (b) The input (gray) and the recovered (black) model velocities. The red lines indicate the lower and upper boundaries of the model space. (c) The best and mean values of the objective function in each generation. (d) Color-coded velocity output from nonlinear inversion of the noise-added synthetic seismogram. Random input velocity structures are assumed at the first generation.

3. Results

3.1. SS Precursor Amplitude

Stacks of SS precursor waveforms show robust reflections off the 410 and 660 (*S410S* and *S660S*, respectively) within a laterally varying upper mantle beneath

the northwestern Pacific region. To evaluate standard errors and the robustness of the observed precursors, we adopt a bootstrapping resampling algorithm (Efron and Tibshirani, 1991; Shearer, 1993; Sacchi, 1998; Deuss, 2009) that takes automatic measurements from 200 randomly selected subsets of the data. During each trial, the amplitudes of the *S410S* and *S660S* are directly determined from the positive peaks associated with the precursors. The amplitude uncertainties of the secondary arrivals are subsequently estimated from the standard deviation of bootstrapped measurements (Deuss, 2009). Overall, the maximum standard error in amplitude is 0.2% based on bootstrap resampling.

Interpolated reflectivity maps are constructed from the *S410S* and *S660S* amplitudes (Figures 3a and 3b), which are defined as maximum amplitudes within a 30 s time window centered at the respective PREM predicted arrival times of the *S410S* and *S660S*. Large-scale low and high reflectivity zones are detected, both at the top and the bottom of the MTZ. Two distinctive high amplitude zones are observed on the *S410S* reflectivity map (see Figure 3a). A SW-NE trending high reflectivity zone is present landward from the Wadati-Benioff zone on top of the transition zone, extending from the North China Craton to the Wudalianchi volcanic center and reaching a maximum value of ~7% (amplitudes are normalized with respect to the *SS* phase amplitudes) beneath the Changbai hotspot (see Figure 3a). A second reflectivity zone is observed to the oceanic side of the subducting Pacific plate, approximately 500 km away

from the Japan trench. The centers of these two anomalies are approximately 1500 km apart.

Strong *S660S* reflections are observed in the vicinity of the Wadati-Benioff zone (see Figure 3b). The largest *S660S* amplitude, which attains ~8% of that of *SS*, is identified within the reported slab contours (Hayes et al., 2012). While some of the amplitude variations may be affected by the data uncertainties (Figures 3c and 3d), especially in the southern part of the study region due to low data density, the largest amplitudes of the *S410S* and *S660S* are resolved to 95% confidence levels.

In comparison with PREM, which suggests respective reflection amplitudes of 3.9% and 6.1% relative to *SS* at a source-receiver distance of 130° for *S400S* and *S670S*, the observed values are 1.6-2.3 times higher beneath China (the former) and the Japan Basin (the latter). Part of the amplitude may be affected by the differential attenuation (*Q* factor) between *SS* and its precursors. However, our experiment based on conservative choices of *Q* values suggests a contribution no greater than 0.15%, well within the measurement uncertainties (see supplementary materials).

[Figure 3 about here.]

Figure 3: The interpolated reflectivity maps of *S410S* (a) and *S660S* (b) amplitude variations. (c) and (d) show the standard errors of the reflection amplitudes based on a bootstrapping analysis. The solid circles denote the centers of the averaging bins and the magenta lines show the slab contours (Hayes et al., 2012). The numbers at the center locations of the averaging bins in panel (d) show the total number of waveforms in each cap used in stacking.

3.2. *Shear Velocity Variation*

We perform nonlinear waveform inversions to simultaneously recover shear wave velocities and the depths of the upper mantle discontinuities. Following the steps of iterative forward modeling procedure described in Section 2.1, the *SS* precursor waveforms are inverted based on 15 generations, and 120 simulated waveforms are generated in each generation through parallel programming. For consistency, the synthetic waveforms are filtered between the same corner periods as the observed data and the source mechanism is removed by isolating and deconvolving *SS* (Figure 4). In general, the waveforms of *S410S* and *S660S* are accurately modeled by the synthetic seismograms, especially in view that 1) PREM contains no velocity and density jumps at mid MTZ depths (Dziewonski and Anderson, 1981), causing slight mismatches of the waveforms between

S410S and *S660S*, and 2) the expected misfit between the data and synthetics is nonzero due to the uncertainties (e.g., random noise and imperfect structural corrections) in precursor amplitude and arrival time. The 520 km discontinuity (for short, the 520) is locally enhanced within the Wadati-Benioff zone (Gu et al., 2012), though its lateral variations are complex in all three cross-sections (see Figure 4).

[Figure 4 about here.]

Figure 4: *SS* precursor waveform inversion results for profiles A to C. In all panels the black and red lines represent the observed and synthetic stacks, respectively. The key features in the waveform data are highlighted using the dashed boxes. The symbols D and E represent areas with depressed and elevated discontinuity topography, respectively. The letters H and L indicate the high and low amplitude precursory arrivals, respectively.

We determine a local 1D velocity model for each cap based on the stacked *SS* precursor waveform. A 3D model of the study region is then constructed from the resultant 1D models using a linear interpolation algorithm. The inverted shear velocities under NE China and the northwestern Pacific subduction regions (Figure 5) highlight the morphology of the Japan-Kuril trench, showing an average velocity in excess of 1.5% relative to PREM. This high velocity structure intersects the base of the upper mantle near the Sea of Japan and extends

westward toward Central Asia. Its presence and the contrasting discontinuity topography of the 410 (elevated by 10-25 km) and 660 (depressed by 10-27 km) are required to match the low amplitude precursory arrivals (see Figures 3 and 4). The presence of a high velocity anomaly (cold subducting oceanic lithosphere) within the MTZ can reduce the impedance contrast across the 660 km boundary and produce low *S660S* amplitudes (see Figures 4 and 5). The eastern side of this dipping high velocity structure shows a distinct low velocity zone, reaching its minimum value (approximately 1.5% slower than that of PREM) within the MTZ beneath the northeastern Honshu arc. The presence of this anomaly is required by a high amplitude, early-arriving *S410S* oceanward of the subducting Pacific plate (see Figures 3 and 4), which is partially accommodated by a relative depression (~15 km) of the 410. This low velocity structure appears to continue to the base of the MTZ under the Japan and Ryukyu trenches.

The most intriguing low velocity zone is observed in the upper mantle beneath NE China, potentially extending 1000 km further inland from the eastern end of the Wudalianchi volcanic belt to the northwestern facade of the Greater Hinggan Mountains (see Figures 5a and 5b). The center, and the deepest part, of this broad anomaly resides beneath the Changbai Mountain Range, initiating at shallow depths and persisting down to a depth of ~500 km (see Figures 5a-c). This low velocity structure is accompanied by a 5-10 km depression of the 410, which reflects an increased *SS-S410S* differential time in this region (see Figure 4). The

observed low velocity structure under the Wudalianchi volcanic field is limited to depths above 350 km, which is significantly shallower in comparison with that beneath the Changbai hotspot.

[Figure 5 about here.]

Figure 5: Lateral variations in *S*-wave velocity relative to PREM (Dziewonski and Anderson, 1981) at 300 km (a), 400 km (b), 500 km (c), and 600 km (d) depths beneath NE China and the northwestern Pacific region. The blue and red colors represent fast and slow anomalies, respectively. The solid circles mark the centers of the averaging caps and the white triangles denote the locations of major intraplate volcanoes.

Changes in MTZ thickness across the study region, which are calculated from the inverted transition zone discontinuities depths (Figure 6a), are consistent with those of Gu and Dziewonski (2002) using a correlation based method (Figure 6b). Both studies show increased MTZ thickness north of the Ryukyu trench and beneath the Songliao Basin in northeastern China, especially in the present study where values exceed the global average of 242 km (Gu et al., 1998; Gu and Dziewonski, 2002; Lawrence and Shearer, 2006b). The largest difference, which reflects improvements in methodology and vastly improved spatial resolution, is an anomalously thick MTZ in this study beneath Northeast China due to a 25-30 km depression of the 660 (see Z1 in Figure 6a). This anomaly is consistent with the recent findings of Liu et al. (2015) based on migrated receiver functions (see

the dashed-line box, Figure 6a). To a lesser extent, the MTZ beneath the Ryukyu trench is approximately 15 km thicker in this study than Gu and Dziewonski (2002), despite similar shapes and signs (see Z2 in Figure 6a).

As will be discussed in Section 4, the combination of a depressed and amplified 410, a severely deformed 660 and a low velocity asthenosphere beneath the Changbai-Wudalianchi hotspot has major implications for the intraplate volcanism in NE China.

[Figure 6 about here.]

Figure 6: (a) MTZ thickness variations inferred from the inverted depths of the 410 and 660 by this study. (b) MTZ thickness variations reported by Gu and Dziewonski (2002). The red and blue colors represent regions of narrow and thick MTZ, respectively. Z1 and Z2 indicate anomalously thick MTZ beneath NE China and the Ryukyu trench, respectively. The dashed-line box in panel (a) outlines a relatively thick MTZ (>270 km) in Liu et al. (2015), obtained from depth migrated P-to-S converted waves.

4. Discussion

4.1. General Assessment

This study presents a nonlinear optimization method that accurately determines mantle large-scale shear velocity and discontinuity topography based on *SS* waveform information alone. In comparison with a recent study of *SS* precursors for the same region (Gu et al., 2012), the current study benefits from denser data coverage, especially in the northern flank of the study region where data density increased by ~50% due to four extra years of recordings. The resolution improved further from the *SS* precursors amplitude information, which is essential for resolving the impedance contrast at major seismic discontinuities, as well as from the reduced trade-off between the velocity and discontinuity depth. While the scaling of the reported features (>300 km) may be smaller than the nominal resolution (>1200 km), the use of multiple stations (Rost and Thomas, 2009; Schmerr and Thomas, 2011) and dense sampling (Cao et al., 2011; Gu et al., 2012; Zheng and Romanowicz, 2012) can significantly reduce the effective Fresnel zone and smearing along an isochron (Rost and Thomas, 2009; Cao et al., 2010).

The westward dipping high velocity structure beneath the back-arc region (Figures 7a-c), which is consistent with earlier findings from *P* and *S*-wave travel times (van der Hilst et al., 1991; Fukao et al., 1992; Gorbatov and Kennett, 2003; Huang and Zhao, 2006; Obayashi et al., 2006; Li and van der Hilst, 2010; Zhao et al., 2012) and waveform (Méglin and Romanowicz, 2000; Friederich, 2003) inversions, has been widely associated with subducted Pacific lithosphere beneath

Japan and Kuril islands. The general morphologies of the high velocity structures are concordant with those of Obayashi et al. (2006) (see Figures 7d-f), both of which are well supported by the locations of deep focus earthquakes in the study region.

The fate of the subducted Pacific plate has been a source of considerable debates (van der Hilst et al., 1993; Grand, 2002; Huang and Zhao, 2006; Schmerr and Thomas, 2011). While relatively unimpeded penetration into the lower mantle has been documented along Mariana (Creager and Jordan, 1986; Bijwaard et al., 1998; Zhao et al., 2004; Huang and Zhao, 2006; Li et al., 2008) and Kuril (Jordan, 1977; Li et al., 2008; Schmerr and Thomas, 2011; Gu et al., 2012) islands, substantial deformation and changes in slab dip have been frequently documented at the base of the upper mantle as evidence of slab deflection or bending (Zhao et al., 2004; Fukao et al., 2001; Huang and Zhao, 2006; Gu et al., 2012). Beneath the Sea of Japan, the seismic structure in the transition zone is dominated by a 1.5-2.0% increase in P velocity (Fukao et al., 1992; Huang and Zhao, 2006; Obayashi et al., 2006), but the deflected slab segment is less well imaged by shear velocities across the study region (Fukao et al., 2009; Gorbatov and Kennett, 2003). The result of our waveform inversion of SS precursor provides further evidence for slab deflection below the Sea of Japan and stagnation beneath eastern China (see Figures 7a-c).

[Figure 7 about here.]

Figure 7: (a)-(c) West-east vertical cross-sections of the inverted shear wave velocity anomalies along profiles A to C from 200 to 720 km depth beneath the northwestern Pacific subduction zones and NE China. The magenta circles indicate the locations of deep earthquakes within a 2° distance from each profile. The depths of the 410 and 660 km discontinuities are indicated by the dashed lines. A1 and A2 mark the downgoing and stagnant parts of the Pacific slab, respectively, while A3 and A4 denote low shear velocity anomalies beneath the Changbai-Wudalianchi hotspot. A5 indicates low seismic velocities on the ocean side of the Wadati-Benioff zone. (d)-(f) Similar to (a)-(c), but for an earlier model of *P*-wave velocities (Obayashi et al., 2006).

4.2. MTZ Structure and Dynamics

The simultaneous solutions of velocity and discontinuity depth enable a detailed examination of the effective mantle temperatures within the MTZ (Schmerr and Garnero, 2006). On the global scale, the depths of the 410 and 660 are either uncorrelated (Flanagan and Shearer, 1998; Gu et al., 1998) or slightly anticorrelated (Revenaugh and Jordan, 1991; Gossler and Kind, 1996; Li et al., 2003; Gu et al., 2003). The lack of correlation is largely attributable to smaller-scale heterogeneities (Schmandt and Humphreys, 2010), compositional variations (Weidner and Wang, 1998; Schmerr and Garnero, 2007), or non-vertical

structural geometry (Gu et al., 2012). Considerations for localized dipping thermal-chemical structures are crucial in these correlation analyses, especially in the vicinity of major subduction zones. To properly account for slab dip, we compute the correlation coefficients between MTZ thickness and the average MTZ shear velocity for slab dip angles ranging from 15 to 90 degrees. The maximum correlation coefficient (0.72) from all three profiles is determined at 30 degrees, which is 22% higher than the uncorrected (90 degrees dip) value (Figure 8). This effective slab dip is in excellent agreement with previously reported values (van der Hilst et al., 1993; Gudmundsson and Sambridge, 1998; Huang and Zhao, 2006; Hayes et al., 2012), which highlights 1) the need to properly consider inclined slab morphology, and 2) the dominance of thermal, rather than compositional, variations within the MTZ. A closer examination of the correlation coefficient at the optimal slab dip (see Figures 8c and 8d) suggests substantial improvements over the conventional vertical structure assumption in global analyses (Flanagan and Shearer, 1998; Gu et al., 1998; Tauzin et al., 2008; Houser and Williams, 2010). The strength of the correlation (0.72) largely reflects the severity of thermal perturbations, relative to the ambient mantle, within the subducting oceanic lithosphere. It is enhanced further by a dipping low velocity structure east of the Wadati-Benioff zone (see A5 in Figures 7a and 7b), which has been previously reported by Obayashi et al. (2006) and Gu et al. (2012) in possible connection with 1) convective return flow in connection with slab

dynamics (Bercovici and Karato, 2003; Ohtani et al., 2004), and 2) partial subduction of a residual Mesozoic hot thermal plume (Larson, 1991; Tatsumi et al., 1998; Honda et al., 2007).

[Figure 8 about here.]

Figure 8: (a) A cross-section of our shear velocity model along profile C. The dashed and solid straight lines indicate how the MTZ thickness and average MTZ velocity are measured along slab dips of 30 and 45 degrees, respectively. (b) The correlation between MTZ thickness and the average MTZ shear velocity perturbations at a range of dip angles. (c) Observations of apparent MTZ thickness versus the apparent average MTZ shear velocity along slab dip of 30 degrees. The least-square fit and correlation coefficient are also shown. (d) Similar to (c) but for slab dip of 90 degrees.

4.3. Mantle Beneath the Changbai-Wudalianchi Hotspot

A notable seismic anomaly that reduces the anticorrelation between the depths of 410 and 660 is the low velocity zone beneath the Changbai-Wudalianchi hotspot. The geometry of this low velocity structure (see A3 and A4 in Figures 7b and 7c) is in excellent agreement with those from earlier studies of this proposed back-arc region (Tatsumi et al., 1990; Lei and Zhao, 2005; Huang and Zhao,

2006; Li and van der Hilst, 2010; Tang et al., 2014). This anomaly has been linked to the horizontally deflected Pacific slab and the subsequent upwelling of hot asthenospheric material (Tatsumi et al., 1990; Li and van der Hilst, 2010; Zhao and Liu, 2010), though more recent studies (Tang et al., 2014; Liu et al., 2015) have suggested a slab gap west of the stagnant slab.

In this study, depressed 410 and 660 beneath the Changbai and Wudalianchi volcanic fields imply discontinuous seismic velocities within the MTZ (see A2 in Figures 7b and 7c), above which a low velocity structure (see A3 and A4 in Figures 7b and 7c) continues to the shallow mantle (Figure 11). Anomaly A3 appears to terminate at mid-MTZ depths (~500 km), which is moderately deeper than earlier estimates (Zhao and Ohtani, 2009; Li and van der Hilst, 2010). Judging from the MTZ velocity characteristics, the transition from low to high velocity appears to be gradational rather than abrupt. The presence of the low velocity zone does not significantly impact the amplitude of the 520, which is diffuse above the stagnant part of the Pacific slab (see Figures 4b and 7b).

To examine the robustness and resolution of the velocity and discontinuity depth solutions we conduct a hypothesis test, a frequently adopted technique for linearized inversions. In this test, we first introduce a low velocity structure resembling that beneath the Changbai hotspot (Figure 9). This hypothetical anomaly terminates near 410 km, overlying a high velocity structure similar to the stagnant oceanic lithosphere. The input depths of the 410 and 660 km

discontinuities are both depressed by more than 10 km near the largest shear velocity perturbations. After the synthetic seismograms are computed for each cap location based on these input parameters, we add up to 0.2% Gaussian noise depending on the observed standard errors from bootstrap resampling (Section 3.1). The recovered models through the same inversion procedure show two well-defined seismic velocity anomalies with the same signs and shapes as the input (see Figure 9). The maximum amplitudes of the recovered anomalies are about 80% and 95% of the assumed values of the respective low and high velocities; the recovery of the latter anomaly is superior due to a larger, more laterally coherent input structure. The amplitudes near the edges of the major anomalies are preferentially reduced due to the lateral averaging with a substantially weaker structure to the west. The depressions of the phase boundaries near the center of the profile are recovered to about 90-95% relative to the input values. The performance of the inversion is equally effective at recovering a relatively shallow low velocity zone beneath the Wudalianchi hotspot. In this case, the inversion recovers over 75% of the maximum input velocity and discontinuity topography, as well as a minor lateral gradient east of the volcanic center (see Figure 9). Our hypothesis test demonstrates that 1) a restoring resolution analysis is equally effective in nonlinear inversions as their linear counterpart, and 2) the morphologies of the low velocity zones beneath the volcanic centers and the underlying stagnant slab can be sufficiently resolved in our study region.

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[Figure 9 about here.]

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Figure 9: Input models (panels a and b) and the inverted results (panels c and d) of the synthetic tests along profiles B and C. The depths of the 410 and 660 are indicated by the thick dashed lines.

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To further validate the recovered images from the 1D waveform inversions of SS precursors (see Figure 7), we implement a 2D waveform tomography using the same nonlinear inversion technique (Figure 10). The mantle is discretized into cells of 3° (in the horizontal direction) by 50 km (in the vertical direction). To simplify the problem, we assume known discontinuity depths from 1D inversions (see Figure 7) and solve for perturbations of shear velocities within each cell; it is worth noting that we define a finer vertical cell size of 10 km around the discontinuity depths (220, 300, 400, 520 and 670 km depths). In this approach, the travel times of the precursors and their corresponding reflection coefficients are calculated using the 2D ray tracing method (Lebedev and Nolet, 2003; Zhao et al., 2004; Lawrence and Shearer, 2006a; Priestley et al., 2006). Then, the reflectivity profiles are convolved with the source time functions extracted from the stacked waveforms (Shearer et al., 1999). We generate over 2000 solutions at each generation, while 5% of the initial random solutions are replaced by shear velocities from GyPSuM (Simmons et al., 2010). The optimized synthetic waveforms from 2D nonlinear inversions closely match the observations along

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profile B (see Figure 10b). The velocity model shows a westward dipping structure similar to that recovered by 1D waveform inversions, which appears to flatten toward NE China. The maximum velocity of the slab (1.5%) is comparable to that of 1D inversion, whereas the low velocity zone above the 410 (-1.1%) is slightly weaker than the 1D counterpart. Part of the amplitude difference of the low velocity zone between 1D and 2D inversion outcomes could be explained by the inclusion of a more realistic, low velocity lithosphere in the latter experiment.

[Figure 10 about here.]

Figure 10: 2D waveform tomography results using the stacked waveforms observed along profile B (see Figure 4). The results clearly show a high velocity structure, associated with the subducting plate, and reduced shear wave velocities beneath NE China. (a) Schematic view of the *SS* ray paths used in waveform tomography. The sources and receivers are indicated by stars and triangles, respectively. The thick black lines outline the area covered by the waveform tomography approach. The outlined area is zoomed on from 200 km to 750 km depth. (b) The visual comparison between the observed (black) and synthetic (red) waveforms shows a good agreement between the input and modeled waveforms. (c). The normalized fitness values of the objective function in each generation. (d) Color-coded cross-section (outlined area in panel a) of the shear wave velocity perturbations. The blue and red colors represent the fast and slow velocities, respectively. The key features of the recovered image are also labeled (LVZ stands for low velocity zone).

Our models consistently suggest a relatively warm mantle in the vicinity of the 410 km discontinuity. The observed 5-10 km depression of the 410 km seismic discontinuity beneath the intraplate volcanic fields in mainland China can be attributed to the response of the olivine to wadsleyite phase transition, an exothermic reaction with a positive Clapeyron slope of 1.5-4.0 MPaK⁻¹ (Akaogi et al., 1989; Houser and Williams, 2010; Lessing et al., 2014), to the presence of a high temperature anomaly above or across this boundary. By assuming a temperature dependence of the shear wave velocity ($\frac{\partial \ln V_S}{\partial T}$) of $-7 \times 10^{-5}/\text{K}$ for a dry upper mantle (Houser and Williams, 2010), a *S*-wave velocity reduction of 1.5% results in an approximate temperature increase of 215 K relative to the average value of 1694 K for the olivine phase change (Houser and Williams, 2010). The phase boundary moves toward the higher pressures (by 320-860 MPa) and greater depths, which are consistent with recent reports of 20-30 km of depression in the same region (Li et al., 2000; Li and Yuan, 2003; Liu et al., 2015). Different mechanisms have been proposed to explain the origin of widely distributed low velocity regime in the upper mantle beneath northeastern China, most of which are associated with the subducting Pacific plate. An incomplete list of these mechanisms include 1) partial melting induced by the deep dehydration of the slab (Zhao and Ohtani, 2009), 2) passive upwelling of asthenospheric materials in response to slab accumulation within the transition zone (Bercovici and Karato,

2003; Kuritani et al., 2009; Faccenna et al., 2010), and 3) upwelling from a slab gap (Tang et al., 2014; Liu et al., 2015).

The first mechanism infers significant amount of water near the 410 km discontinuity, while up to 3 wt% water could be present in the mantle transition zone (Bercovici and Karato, 2003). A relatively wet mantle around the 410 would elevate the olivine to wadsleyite phase boundary. While the extent of the elevation in presence of water remains questionable (Frost and Dolejš, 2007; Schmerr and Garnero, 2007), an elevated 410 is simply inconsistent with either the observed depression from this (see Figure 7) and earlier studies (Li and Yuan, 2003; Liu et al., 2015), or the reported water content in this region (Ichiki et al., 2006; Chen et al., 2007; Fukao et al., 2009; Ye et al., 2011). In a case of water saturated mantle, a hydrous melt lens may form atop the 410 and produce multiple phase transitions. This was reported in South America where a relatively deep, modest wet to dry wadsleyite reflection was detected at 415 km (Schmerr and Garnero, 2007). Based on numerical experiment from the same study, the expected amplitude (which increases with water content) is significantly lower than the predicted value from PREM and far below our observations under the Changbai hotspot.

A favorable solution for the observed 410 amplitude and depth will require significantly increased impedance contrast across a depressed 410, which could be satisfied by the presence of a melt layer in a relatively dry, warm mantle.

Increasing Mg by up to 4% within the melt layer can effectively depress the 410 km discontinuity by about 10 km (Fei and Bertka, 1999; Schmerr and Garnero, 2007), which corroborates our observations in NE China. Both decompression melting associated with passive upwelling (Bercovici and Karato, 2003; Faccenna et al., 2010) and a significant sub-slab component from the slab window (Tang et al., 2014) can induce partial melt atop the 410. This would depress the 410 phase boundary, enhance the impedance contrast and substantially increase the *S*-wave reflection amplitude, which are supported by our observations beneath the Changbai hotspot. Insufficient resolution around the suggested slab window (see Figures 7 and 11) makes it difficult to clearly differentiate between these two candidate mechanisms, however.

Further inferences could be made regarding the mantle beneath the two volcanic centers examined in this study. Despite the considerable depth difference between the Changbai (about 500 km) and Wudalianchi (about 350 km) hotspots, trace element analyses (Basu et al., 1991; Kuritani et al., 2009) suggest a comagmatic upper mantle origin with EM1 (Enriched Mantle 1) composition (Jackson and Dasgupta, 2008). Hence, it is convincible that the mantle upwelling beneath the Changbai hotspot is responsible for the source of partial melt for both volcanic centers in view of its greater vertical extent and strength (see Figures 5a and 11). Further work will be needed to quantify the accuracies of the aforementioned mechanisms and interpretations. Finally, increased seismic velocities at the base

of the MTZ under northeast China and a deep 660 (>680 km) relative to the regional average of 655-665 km (Shearer, 1993; Gu et al., 2003; Houser and Williams, 2010) are consistent with the respective measurements based on analyses of receiver functions (Li et al., 2000; Li and Yuan, 2003; Liu et al., 2015) and migration of long period SS precursors (Gu et al., 2012). A maximum depression of 20-30 km is reported on the 660, possibly in association with a decrease of 300-400 K in MTZ temperature (Li and Yuan, 2003; Ye et al., 2011). A negative temperature anomaly of ~300 K and a greater depth of the 660 km discontinuity jointly suggest increased *S*-wave velocities of approximately 2% in the presence of a negative Clapeyron slope from the post-spinel phase transition (Houser and Williams, 2010; Lessing et al., 2014). The low temperature anomaly and locally depressed 660 could be explained by the existence of water within the MTZ, the amount of which depends on the volume entrained within the slab as well as the residence time of the slab (Fukao et al., 2009). *P* and *S* waveform modeling from northeast China detects a 130 km thick high velocity zone at the base of the MTZ, which may contain up to 0.3 wt% water based on estimates from Ye et al. (2011).

[Figure 11 about here.]

Figure 11: (a) Isosurfaces showing the major seismic anomalies beneath Northeast China and the northwestern Pacific subduction zones in the depth range of 200 to 700 km. The blue and red colors represent fast and slow shear velocities, respectively, at the respective isosurfaces of +0.65% and -1.0%. The definitions of major seismic anomalies (A1-A5) are the same as in Figure 7. The black lines show the slab contours, which are taken at 50 km depth intervals starting at 100 km (Hayes et al., 2012), and the magenta circles indicate the locations of deep-focus earthquakes. (b) A schematic illustration of a NW-SE vertical section (see Figure 11a) showing the possible origins of the intraplate volcanism in Northeast China. Three possible mechanisms for the intraplate volcanism (discussed in Section 4.3) are also illustrated: (1) deep dehydration process of the slab, (2) passive upwelling in response to the subduction of the Pacific plate, and (3) escaping of the hot sub-slab material through a slab window.

5. Conclusions

Our study demonstrates that shear wave velocity and the topography of MTZ discontinuities beneath northeastern China and the northwestern Pacific region can be simultaneously recovered using the waveforms of *SS* precursors alone. A statistically significant positive correlation is observed between MTZ thickness and the average MTZ velocity along an effective slab dip of ~30 degrees, which

reflects the dominant effect of thermal heterogeneity in the upper mantle and MTZ. A strong low velocity zone is found beneath active volcanoes in northeastern China, persisting to a depth of >500 km, which implies that the active volcanoes in eastern China (Changbai and Wudalianchi hotspots) may be fueled by a hot thermal anomaly from MTZ depths.

Passive upwelling and/or a slab window could be responsible for a relatively water-poor melt layer atop the 410, which cause 5-10 km depression and substantially enhance reflection amplitude on the olivine to wadsleyite transition. A distinctive low velocity structure is also observed east of the Wadati-Benioff zone, which narrows MTZ by 30 km relative to its regional average. This low velocity zone is possibly associated with (i) a fossil superplume (~140 Ma) or (ii) upwelling hot mantle material in response to slab and lower mantle interaction. Overall, our findings from nonlinear *SS* precursor waveform inversions offer a new window into the process and fate of the subducting oceanic plate in the northwestern Pacific region. From a technical stand point, the nonlinear joint inversion technique introduced in this study represents a new phase in the retrieval of seismic properties pertaining to large scale thermal/chemical structures in the mantle. This is not aimed to be a replacement of the existing travel time and waveform tomographic approaches, especially in view of the relatively coarse spatial resolution of the *SS* precursor dataset, but an

improvement that effectively incorporates the amplitudes of secondary reflections
or conversions from mantle interfaces.

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