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Short period ScP phase amplitude calculations for core–mantle boundary with intermediate scale topography

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ABSTRACT

The core–mantle boundary (CMB) topography plays a key role in constraining geodynamic modeling and core–mantle coupling. It's effective to resolve the intermediate lateral scale topography (hundreds of km) with short period core reflected seismic phases (ScP) due to their small Fresnel-zones at short epicentral distances (<3336 km (30°)). We developed a method based on the ray theory and representation theorem to calculate short period ScP synthetics for intermediate lateral scale CMB topography. The CMB topography we introduced here is axisymmetric and specified with two parameters: H (height) and L (diameter, or lateral length scale). Our numerical computation shows that a bump ($H > 0$) and dip ($H < 0$) model would cause defocusing/weakening and focusing/amplifying effects on ScP amplitude. Moreover, the effect of frequency and combination of L and H are quantified with the amplification coefficients. Then we applied this method to estimate a possible CMB topography beneath northeastern Japan, and a CMB model with $L = 140$ km, $H = 1.2$ km overall matches the observed pattern of 2D PcP/ScP amplitude ratios. However, it is difficult to totally rule out other factors that may also affect PcP/ScP pattern because of limitation of ray-based algorithms we used here. A hybrid method combining ray theory and numerical method is promising for studying complicated 3D structure and CMB topography in the future.

1. Introduction

The core–mantle boundary (CMB) topography has been under extensive geophysical investigations since it plays significant roles in studies of mantle convection, geomagnetic field evolution and decadal variation of Earth's rotation (Lay and Garnero, 2004; Calkins et al., 2012; Roberts and Aurnou, 2012). For the past decades, numerous seismic studies have demonstrated the CMB topography features ranged from global size to a scale of a few kilometers with various methods. Normal modes and travel time techniques both reported that the global undulations on the CMB are within several kilometers (Creager and Jordan, 1986; Morelli and Dziewonski, 1987; Doornbos and Hilton, 1989; Ishii and Tromp, 1999; Sze and van der Hilst, 2003; Tanaka, 2010; Soldati et al., 2013), while some authors argued for the trade-offs between the topographic and volumetric structures. (Rodgers and Wahr, 1993; Pulliam and Stark, 1993; Murphy et al., 1997; Obayashi and Fukao, 1997; Garcia and Souriau, 2000; Kárason and van der Hilst, 2001; Soldati et al., 2003; Koelemeijer et al., 2012; Colombi et al., 2014). Based on the scattering theory, a small-scale topography model of 100–350 m' variation along with the correlation length of 7–20 km explained the observed 1 Hz PKKP precursors (Doornbos, 1980; Earle and Shearer, 1997), although some contaminations from volumetric scattering in the lower mantle could make this value overestimated.

On the intermediate lateral scale (hundreds of km), the travel times and amplitudes of short period (1 Hz) core reflected phases (e.g. PcP and ScP) are found to be effective to resolve the topographic variation at the CMB, as their Fresnel-zones are comparable to the intermediate lateral scale. Neuberg and Wahr (1991) elucidated that the CMB topography amplitudes beneath their particular area are at most 2–3 km across 50–400 km after comparing PcP travel time residuals and the PcP/P amplitude ratio pattern with synthetics (Neuberg and Wahr, 1991). Rost and Revenaugh (2004) reported rapid PcP/P amplitude ratio variations beneath Kenai Peninsula of Alaska, and hypothesized the small-scale CMB topography cause after ruling out possibilities such as attenuation, radiation pattern and slab effect (Rost and Revenaugh, 2004). Recently, Wu et al. (2014) developed an algorithm for calculating short period synthetic PcP seismograms with CMB topography and proposed a depressed CMB up to 6 km across several hundreds

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km length to fit the abnormal PcP/P amplitude ratios observed by Rost and Revenaugh (Wu et al., 2014). In these studies, PcP observations are mostly within epicentral distances of 3336 km (30°) and larger to exploit the relatively strong P wave reflection at CMB. Around epicentral distances of 2224 km (20°), the reflection coefficient is reduced and PcP is often contaminated by S wave and its coda, thus not desirable for modeling CMB topography.

In contrast to studies of PcP, limited number of surveys have used the ScP phase to explore the intermediate CMB topography variation. Benefiting from the relatively large reflection coefficient at CMB, ScP phases are easily detected at large distances (>3336 km (30°), but less than 6670 km (60°)) due to S to Pdiff conversion at the CMB). The ScP data at these distances are widely used in studying lowermost mantle anomalies (Castle and van der Hilst, 2003), especially the D″ and ULVZ (Garnero and Vidale, 1999; Reasoner and Revenaugh, 2000; Persh et al., 2001; Rost and Revenaugh, 2003; Rost et al., 2005). Vidale and Benz (1992)suggested a flat CMB under northeastern Pacific with an amplitude of less than 500 m due to the coherence of P, PcP and ScP waveforms (Vidale and Benz, 1992). Castle and van der Hilst (2000) observed large ScP/P amplitude ratios and conjectured that the combination of high attenuation along P path, low attenuation along ScP path and focusing effect from CMB topography may create such particular amplitude ratios (Castle and van der Hilst, 2000). For the relatively large epicentral distances (3336–6670 km (30–60°)), the effect of heterogeneities along relatively long ScP paths through D″ and ULVZ probably mask the topography's.

Instead, CMB topography variation could be better constrained with the traveltime and amplitude of steep-angle ScP (epicentral distance around 2224 km (20°) and less) due to the smaller Fresnel zone on CMB. Since most earthquakes are double couple source and generate strong shear wave energy, the short period steep-angle ScP wave usually has high signal noise ratio for moderate strong earthquakes. In addition, the ScP at short distances arrives much later than PcP, and is much less contaminated by coda of the direct S wave. Schweitzer (2002) combined PcP-P, ScP-P and PcP–ScP differential traveltime residuals to invert CMB undulations beneath Europe and argued that uncertainties of CMB topography maps that had been derived so far are fairly large partly due to errors of ScP arrival and trade-off between topography and tomography, even if source depths and site effects were corrected (Schweitzer, 2002). On the other hand, the absolute arrival time measurements could be biased by hand-pick errors though the onsets' accuracy could be improved by advanced phase-matching tools and different band-pass filtering techniques. In order to increase resolution of the CMB topography variation, the amplitude information of steep-angle ScP could be utilized.

In this study, we implemented an algorithm for computing short period ScP synthetics based on ray theory and representation theorem, similar to Wu et al. (2014). We modeled two types of CMB topography (bump and depression or dip), and explore the effects on ScP amplitude for different frequencies, lateral length scale (L) and height (H, $H > 0$ for CMB bump and $H < 0$ for a CMB dip) of the topography. Then, we applied this algorithm to assess intermediate scale CMB topography beneath northeast Japan by modeling ScP observation from a Mw6.7 aftershock of the 2013 Sea of Okhotsk earthquake. Lastly, we discuss the limitations of the method and possible improvement that it may work in more complicated mantle structures.

2. Method

2.1. Theory

For computing 3D global synthetic seismograms, a variety of numerical approaches have been implemented, including spectral element, finite difference, finite element and pseudo-spectral element (Igel et al., 2002; Furumura et al., 1998; Komatitsch and Tromp, 1999). These methods can handle strong lateral variation or 3D topography variation of interfaces, but with limitations of long computational time or tremendous memory storage for short period seismic wave simulation (around 1 Hz). To reduce the computational time and memory cost, some authors recently developed 2.5D algorithms, which are schemes of 2D computational domain with out-plane spreading corrections (Cormier, 2000; Nissen-Meyer et al., 2007; Jahnke et al., 2008; Li et al., 2014a). However, these approaches are still time-consuming and are not straightforward to handle 3D topography variations of CMB. In contrast, the ray theory based methods are very attractive because of the high computational efficiency, straightforward implementation of intrinsic attenuation and flexible CMB topography settings. Kampfmann and Müller (1989) calculated short-period (1 Hz) PcP synthetics with sinusoidal topography based on Kirchhoff method and their results are consistent with the observation of pronounced scatter in PcP amplitudes at epicentral distances less than about 60°and very low amplitudes beyond 70° (Kampfmann and Müller, 1989). Wu et al. (2014) applied a method based on ray theory and representation theorem to study the CMB topography features beneath the Kenai Peninsula of Alaska, and they successfully explained rapid variation of short period PcP amplitudes observed with dense seismic array (Rost and Revenaugh, 2004). Here we extend this algorithm for modeling ScP waveforms, and demonstrate some differences in implementing the algorithm as compared to PcP.

The representation theorem (Aki and Richards, 2002) states that the displacement $\vec{u}(\vec{x},t)$ consists of contributions from the body force $\mathbf{f}$ throughout volume V and the tractions on the boundary $\sum$. For seismic wave propagation inside the Earth, the body force $\mathbf{f}$ is considered as zero. Therefore, the nth ($n = 1, 2, 3$) component of ScP displacement $u_n(\vec{x},t)$ recorded at surface location $\vec{x}$ is the integral of contributions from each sub-source on the core mantle boundary $\sum(\vec{\xi})$ (Eq. (1)).

$$u_n(\vec{x},t) = \int_{-\infty}^{\infty} d\tau \iint \left\{G_{in}\left(\vec{\xi},t-\tau;\vec{x},0\right)T_i\left(\vec{u}(\vec{\xi},\tau),\vec{n}\right) - u_i(\vec{\xi},\tau)c_{ijkl}n_jG_{kn,l}\left(\vec{\xi},t-\tau;\vec{x},0\right)\right\}d\Sigma(\vec{\xi}) \quad (1)$$

where G is the Green function between CMB and the surface, T and $\vec{u}(\vec{\xi},t)$ are the traction and displacement on the CMB respectively, $\vec{x}$ and $\vec{\xi}$ refer to the receiver and the reflection point on the CMB respectively. On the other hand, the boundary conditions (T and $\vec{u}(\vec{\xi},t)$) and Green functions can be derived with geometric ray theory for a point shear dislocation source (Aki and Richards, 2002). Thus, the ScP displacement waveform follows Eqs. (2)–(4).

$$u_n(\vec{x},t) = -\int_{-\infty}^{\infty} d\tau \iint U(\vec{S},\vec{\xi})U(\vec{\xi},\vec{x})\frac{D_n}{\alpha(\vec{\xi})}\{[\lambda(\vec{E}\cdot\vec{N}) + 2\mu(\vec{E}\cdot\vec{C}) \times (\vec{N}\cdot\vec{C})]\dot{f}(\tau - T^{SV}_{S-\xi})\}\delta(t-\tau-T^{P}_{x-\xi}) - [\lambda(\vec{C}\cdot\vec{N}) + 2\mu(\vec{C}\cdot\vec{E})(\vec{N}\cdot\vec{E})]f\left(\tau - T^{SV}_{S-\xi}\right)\dot{\delta}(t-\tau-T^{P}_{x-\xi})\}d\Sigma(\vec{\xi}) \quad (2)$$

with

$$U(\vec{S},\vec{\xi}) = \frac{(\vec{F}\cdot M\cdot\vec{A})\cdot \mathrm{Re}\, f}{4\pi\rho^{\frac{1}{2}}(\vec{S})\rho^{\frac{1}{2}}(\vec{\xi})\beta^{\frac{5}{2}}(\vec{S})\beta^{\frac{1}{2}}(\vec{\xi})\Re^{SV}(\vec{\xi},\vec{S})} \quad (3)$$

$$U(\vec{\xi},\vec{x}) = \frac{1}{4\pi\rho^{\frac{1}{2}}(\vec{\xi})\rho^{\frac{1}{2}}(\vec{x})\alpha^{\frac{1}{2}}(\vec{\xi})\alpha^{\frac{3}{2}}(\vec{x})\Re^{P}(\vec{\xi},\vec{x})} \quad (4)$$

where λ, μ are Lame constants, ρ is the density, α and β are P-wave and S-wave velocity respectively, $\Re^{P}(\vec{\xi},\vec{x})$ and $\Re^{SV}(\vec{\xi},\vec{S})$ are P-wave