Effects of crust and mantle heterogeneity on PP/P and SS/S amplitude ratios

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[1] Long-period (T > 16 s) PP/P and SS/S amplitude ratios have coherent geographic variations. On average, PP/P is ~10% higher than predicted by the Preliminary Reference Earth Model (PREM) when PP surface-reflection points are within continental regions, and ~10% lower than PREM predictions for oceanic reflection points. Spectral-element synthetics show that this variation can be attributed mostly to the effect of crustal thickness on the long-period PP reflection coefficient. The anomalies of SS/S are attributed mostly to the effect of crustal thickness on the long-period PP reflection coefficient. The anomalies of SS/S are similar determined from spectral-element waveforms of S and SS for 3-D models of the crust and mantle is similar to the observed variation of SS/S. This suggests that wave propagation effects are largely responsible for the observed SS/S variation, not only intrinsic attenuation.

INDEX TERMS: 7203 Seismology: Body wave propagation; 7205 Seismology: Continental crust (1242); 7260 Seismology: Theory and modeling; 5144 Physical Properties of Rocks: Wave attenuation

1. Introduction

[2] The high level of correlation between global models of shear-velocity heterogeneity indicates that, at least at long wavelengths, the shear-velocity structure of the mantle is well known [e.g., Dziewonski, 2000]. Tomographic models to date are constrained primarily by body-wave travel-time and surface-wave phase-delay measurements. Seismic wave amplitudes can provide complementary constraints on velocity heterogeneity as well as attenuation. However, modeling of amplitudes is difficult. Surface-wave amplitude measurements display a large scatter. Hence, models based on these data are necessarily in spatial resolution compared to travel-time and phase-velocity models [Woodhouse and Wong, 1986; Romanowicz, 1990, 1995; Selby and Inghthouse, 2000; Billien et al., 2000].

[3] In this paper, we analyze a global data set of PP/P and SS/S amplitude ratios following recent analyses of the body-wave phases S and SS by Bhattacharya et al. [1996] and Reid et al. [2001]. Contrary to amplitudes of individual body-wave phases, the precision of PP/P and SS/S amplitude ratios is not compromised by uncertainties in earthquake source parameters (e.g., epicenter, mechanism, and seismic moment). Using the spectral-element method (SEM) developed by Komatitsch and Tromp [2002a, 2002b], we demonstrate that the effects of seismic wave focusing and defocusing cause long-period body-wave amplitude variations in addition to surface-wave amplitude variations with patterns similar to those seen in the data.

2. Measurement of Amplitude Ratios

2.1. Observations

[4] Using broadband waveform data from the GSN and GEOSCOPE networks (1980–2000), we measure PP/P and SS/S amplitude ratios with respect to the Preliminary Reference Earth Model (PREM) [Dziewonski and Anderson, 1981]. We make these measurements by cross-correlating low-pass filtered (T > 16 s) observed and synthetic waveforms. The synthetics are computed by normal-mode summation using the PREM velocity and Q structure and Harvard CMT source parameters.

[5] We define the correlation function between the observed, d(t), and synthetic, s(t), waveforms within a window W by:

\[ \Psi_{ds}(\tau) = \int_w d(t)s(t-\tau)dt. \]  \hspace{1cm} (1)

First, we determine the time shift \( \tau_m \) of the synthetic waveform for which \( \Psi_{ds}(\tau) \) has its maximum value. This time shift is regarded as the body-wave travel-time delay. Given \( \tau_m \), we define two quantities that characterize the amplitude ratio between \( d(t) \) and \( s(t-\tau_m) \):

\[ A_1 = \frac{\Psi_{ds}(\tau_m)}{\Psi_{ss}(\tau_m)} \]  \hspace{1cm} and \hspace{1cm} \[ A_2 = \frac{\Psi_{ds}(\tau_m)}{\Psi_{pp}(\tau_m)} \]  \hspace{1cm} (2)

\( A_1 \) and \( A_2 \) minimize, respectively,

\[ \int_w [d(t) - A_1s(t-\tau_m)]^2 dt \]  \hspace{1cm} and \hspace{1cm} \[ \int_w [A_2s(t-\tau_m) - s(t-\tau_m)]^2 dt. \]  \hspace{1cm} (3)

Using vertical component recordings, we measure \( A_{1p}, A_{2p}, A_{1pp}, \) and \( A_{2pp} \) for 80-s wide time windows centered on the theoretical arrival times of P and PP, respectively. \( A_1 \) is equal to \( A_2 \) only when the waveforms of \( d(t) \) and \( s(t) \) in the cross-correlation window \( W \) are identical. Therefore we obtain

\[ A_{1pp}^p = \min(A_1^{pp}, A_2^{pp}) \] \hspace{1cm} and \hspace{1cm} \[ A_{2pp}^p = \max(A_1^{pp}, A_2^{pp}) \] \hspace{1cm} (4)
which are minimum and maximum estimates of the PP/P amplitude ratio, respectively. The best estimate of PP/P and its uncertainty is

\[
PP/P = \frac{A_{pp}^{p/p} + A_{pp}^{p/p}}{2} \pm \frac{A_{pp}^{p/p} - A_{pp}^{p/p}}{2}. 
\]

(5)

We retain measurements of PP/P when

\[
\left| \frac{A_{pp}^{p/p} - A_{pp}^{p/p}}{A_{pp}^{p/p} + A_{pp}^{p/p}} \right| < 0.1. 
\]

(6)

This corresponds to data with a least-squares waveform fit between \(d(t)\) and \(s(t - \tau_m)\) in the P and PP windows, i.e., after the time shift \(\tau_m\) has been applied, of at least 80%. Our data set includes 10,550 PP/P measurements at epicentral distances larger than 55°. We apply the same measurement procedure and selection criteria to transverse component waveform data to obtain 7,250 SS/S amplitude ratio measurements.

### 2.2. 3-D Model Simulations

We simulate PP/P and SS/S amplitude ratios by applying equations (5) and (6) to synthetic vertical and transverse component seismograms, respectively. These synthetics are computed using the Spectral-Element Method (SEM) [Komatitsch and Tromp, 2002a, 2002b] for crustal model CRUST2.0 [Bassin et al., 2000], model ETOP05 of ocean bathymetry and topography [NOAA, 1988], and 3-D models of S and P wave velocity heterogeneity in the mantle. Model S20RTS [Ritsema et al., 1999] is used to describe S velocity variations in the mantle. The P velocity heterogeneity in the mantle is assumed to be identical to that of S20RTS, except for a depth-dependent scaling factor \(R = \frac{\text{d} \log V_p}{\text{d} \log V_p}\), which increases linearly from 1.3 at the surface to 3.0 at the core-mantle boundary [Ritsema and van Heijst, 2002]. The structure is identical to that in PREM.

Since the SEM simulations are time consuming (~40 hours per simulation on a PC cluster with 151 733 MHz CPUs), we limit ourselves to calculating synthetic seismograms for 50 worldwide earthquakes at about 400 global and regional network stations. The number of synthetic seismograms is similar to the number of recordings, ensuring that the gross characteristics of the data are simulated accurately.

### 3. Global Distribution of PP/P

The global distribution of PP/P amplitude ratios has a strong ocean/continent correlation (Figure 1a). PP/P is, on average, ~0.85–0.90 (i.e., the PP amplitude is smaller than predicted by PREM) when PP reflects off oceanic crust while PP/P is evenly distributed about 1 for continental PP reflection points. A relatively high average value of ~1.10–1.15, compared to the PREM predicted value, is obtained for PP/P data with PP reflections points within continents.

Figure 1. (a) Histograms of PP/P (left) and SS/S (right). The data are grouped according to whether the associated PP (or SS) surface reflection point is underneath the oceans (red) or continents (blue). (b) PP/P (left) and SS/S (right) amplitude ratios with respect to PREM and Harvard CMT source parameters, plotted at the PP surface reflection point. The top panel shows the observed amplitude ratios, while the lower panel shows amplitude ratios determined from the SEM synthetics. The large-scale trends in PP/P and SS/S are emphasized by expanding their values in spherical harmonics up to degree and order 6. The correlations between the observed and SEM predicted PP/P and SS/S maps are, respectively, 77% and 54%.

![Figure 1](image1.png)

![Figure 2](image2.png)
residuals correlate with positive (negative) values of
with a radius of 5
residuals with SS reflection points that fall within a circular cap
data in both (a) and (b) have been smoothed by averaging at least 3
compared to SS waveforms simulated using the PREM model. The
/C1
blue '+' symbols represent positive SS/S anomalies. (b)
represent negative SS/S residuals with respect to PREM, while
parameters plotted at SS surface reflection points. Red circles
amplitude ratios with respect to PREM and Harvard CMT source
(a) Observed (top) and SEM predicted (bottom) SS/S
Figure 3. (a) Observed (top) and SEM predicted (bottom) SS/S
amplitude ratios with respect to PREM and Harvard CMT source
parameters plotted at SS surface reflection points. Red circles
represent negative SS/S residuals with respect to PREM, while blue '+'
symbols represent positive SS/S anomalies. (b) ΔSS→S distribution from Reid et al. [2001]. SS waveforms associated with
positive (negative) ΔSS→S values are relatively broad (narrow)
compared to SS waveforms simulated using the PREM model. The
data in both (a) and (b) have been smoothed by averaging at least 3
residuals with SS reflection points that fall within a circular cap
with a radius of 5°. It is expected that negative (positive) SS/S residuals correlate with positive (negative) values of ΔSS→S.

[9] The ocean/continent distribution of PP/P is sensitive to
crustal thickness variations. Figure 2 shows that values of PP/P
between 0.9 (oceans) and 1.15 (cratons) can be explained by the
effect of crustal thickness on the long-period PP reflection coeffi-
cient. The PP reflection coefficient relative to the reflection coefficient for PREM, in which the crust is 21.4 km thick, varies from
0.9 to 1.1 when the crust ranges in thickness from 0 to 45 km,
in agreement with the observed range of PP/P values. The ocean/

4. Global Distribution of SS/S

demonstrates that the large-scale variation of PP/P can be deter-
dined accurately. Since SS/S is measured in the same manner and
with similar accuracy as PP/P, the large-scale, but complex, variation seen in SS/S is robust.

[12] SS/S amplitude ratios do not clearly correlate with surface
geology. The high geometric mean value of 1.09 ± 0.05 of SS/S
measurements (Figure 1a) and their large-scale geographic varia-
tion (Figure 1b) are similar to the SEM predicted geometric mean value (1.07 ± 0.02) and the SEM predicted geographic variation.
Maps of the observed and SEM-predicted SS/S, expanded in
spherical harmonics up to degree and order 6, both show minima
of SS/S in the central and southern Pacific and northwestern
Europe and a maximum beneath South America. The correlation
coefficient between these maps is 54%.

[13] The geographic variation of SS/S is similar to the varia-
tion of ΔSS→S measurements presented by Reid et al. [2001] (Figure 3), especially in eastern Asia and the Pacific, where overlapping coverage is best. ΔSS→S quantifies SS waveform
broadening. It is expected that high values of SS/S correlate with
low values of ΔSS→S and, vice versa. The good correlation further
underscores that SS/S and ΔSS→S are meaningful data, albeit that,
in disagreement with Reid et al. [2001], we prefer not to attribute the
data patterns entirely to attenuation.

5. Conclusions

[14] While differential body-wave traveltimes have often been
used to constrain seismic velocity variations in the mantle [e.g.,
Woodward and Masters, 1991; Kuo et al., 2000], we have
demonstrated that coherent patterns of PP/P and SS/S amplitude
ratios can also be determined from a large set of high-quality
digital waveform data.

that incorporate the effects of wave propagation through a 3-D
crust and mantle model reproduce the gross characteristics of the
data. The PP/P amplitude ratio appears to be determined mostly by
the reflection coefficient of PP and yields a clear ocean/continent
variation. Like surface-wave amplitudes, SS/S amplitude variations
point to the significant effects of velocity gradients in the mantle on
wave propagation.

[16] SS/S data are valuable for refining shear-velocity models of
the mantle. Moreover, constraining attenuation in the (deep) mantle
using body-wave amplitudes must go hand-in-hand with analysis of
the effects of focusing and defocusing on body-wave ampli-
tudes.

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