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The Effect of Sedimentary Basins on Surface Waves That Pass Through Them

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Abstract

Surface waves propagating through sedimentary basins undergo elastic wavefield complications that include multiple scattering, amplification, the formation of secondary wavefronts, and subsequent wavefront healing. Unless accounted for accurately, such effects may introduce systematic bias to estimates of source characteristics, the inference of the anelastic structure of the Earth, and ground motion predictions for hazard assessment. Most studies of the effects of basins on surface waves have been for waves inside the basins. The purpose of this paper is to investigate wavefield effects downstream from sedimentary basins, with particular focus on continental basins and propagation paths, elastic structural heterogeneity, and Rayleigh waves at 10 s period. Based on wavefield simulations through a recent 3D crustal model of East Asia, we demonstrate significant Rayleigh wave amplification downstream from sedimentary basins in eastern China; Ms measurements made on the simulated wavefield vary by more than a magnitude unit. We show that surface wave amplification caused by basins results predominantly from elastic focusing and that amplification effects produced through 3D basin models are reproduced using 2D membrane wave simulations through an appropriately defined phase velocity map. These effects include the magnitude and pattern of amplification and de-amplification, the formation of a secondary wavefront, the distance to the maximum amplification, and the decay of the amplification with distance downstream from the basin. We discuss how the size and geometry of the velocity anomaly affect focusing and illustrate this with several numerical examples. Finally, by comparing the impact of elastic focusing with anelastic attenuation, we conclude that on-continent sedimentary basins will typically affect surface wave amplitudes much more strongly through elastic focusing than through the anelastic attenuation of the basins.
1. Introduction

The Earth is stratified such that greater structural heterogeneity exists near its principal boundaries and discontinuities. This is particularly true near the free surface, where the presence or absence of sedimentary basins is a principal contributor to lateral heterogeneity in the shallow Earth. The existence of basins makes it more difficult to recover information about structures below them in the deeper crust and mantle, particularly if the structure of the basin is only poorly known. For this reason and others, seismologists often attempt to site seismic stations outside of basins, which, together with the fact that earthquakes commonly occur either deep within basins or below them, means that seismic body waves may often be recorded with a minimal imprint of the effect of basins. This is not true for surface waves, which are trapped near the Earth’s surface and propagate through sedimentary basins even to stations that may be situated outside of them. The purpose of this paper is to investigate how significant sedimentary basins affect the surface waves that propagate through them, with a particular focus on amplitude effects for waves that propagate over regional (rather than teleseismic) distances in a continental setting.

Because of the relevance to seismic hazard, there have been many studies of the effects of sedimentary basins on the amplification of surface waves that are recorded within a basin (e.g., Aki & Larner, 1970; Bard & Bouchon, 1980a,b; Bard et al., 1988; Kawase, 1996; Olsen et al., 1995, 2006, 2009; Olsen, 2000; Alex & Olsen, 1998; Graves et al., 2011; Day et al., 2012; Denolle et al., 2014). Many of these studies interpret such amplitude effects as a “site response” of the basin, which is seen to terminate at the boundary of the basin. In addition, site effects are commonly explained as a 2D vertical cross-section effect whereby lateral heterogeneities play a less important role. There has been appreciably less study of the residual effects on surface waves after they have passed through a basin; thus, the effect of sedimentary basins on through-passing surface waves is more poorly understood. This is particularly true regarding the role of lateral heterogeneity on the surface waveforms at short to intermediate periods that propagate over regional distances. Surface wave focusing/defocusing caused by lateral heterogeneity has predominantly been studied at intermediate to long periods for waves
that propagate over teleseismic distances (e.g., Lay & Kanamori, 1985; Woodhouse & Wong, 1986; Wang et al., 1993; Wang & Dahlen, 1995; Selby & Woodhouse, 2000; Yang & Forsyth, 2006).

There are a number of reasons that motivate the study of the effects of sedimentary basins on through-passing surface waves at short and intermediate periods at regional distances. (1) With the development of ambient noise tomography, surface wave measurements derived from ambient noise cross-correlations have become a common tool to infer crustal and uppermost mantle structure, either alone (e.g., Shapiro et al., 2005; Lin et al., 2007; Yang et al., 2007) or in concert with other observations (e.g., Lin et al., 2012; Shen et al., 2013a,b; Shen & Ritzwoller, 2016; Kang et al., 2016). The shorter periods of the surface wave observations derived from ambient noise, in contrast with earthquake-based measurements, make the surface wave measurements much more sensitive to sedimentary structures. Indeed, the inference of the shear velocity structure of sedimentary basins is now common practice in ambient noise tomography (e.g., Lin et al., 2007; Yang et al., 2007; Shen & Ritzwoller, 2016; Kang et al., 2016). (2) The approximations and assumptions that form the basis for some tomographic methods break down in the presence of the strong lateral heterogeneity that may exist in or near sedimentary basins, particularly near their boundaries. In addition, significant distortions to surface waveforms may be inconsistent with the assumptions and potentially may propagate downstream far from the basin (e.g. Feng & Ritzwoller, 2017, in preparation). (3) Source characterization generally, and moment or magnitude estimation specifically, depend in part on interpreting the amplitude of surface waves that propagate at regional distances. Strong amplitude effects imparted to surface waves by elastic heterogeneities may bias magnitude estimates and other source characteristics. This may be particularly important in the context of nuclear discrimination (e.g., Bowers & Selby, 2009). (4) The inference of the anelastic structure of the Earth may be based on amplitude information from seismic waves, which could be biased by amplitude effects caused by elastic structures. Some studies of surface wave attenuation tomography have taken focusing/defocusing into account (e.g., Dalton & Ekstrom, 2006; Dalton et al., 2008; Bao et al., 2016), but have applied corrections only at long periods. At shorter periods, surface wave Q tomography has been performed by extracting amplitude information from
ambient noise (e.g., Prieto et al., 2009; Lawrence & Prieto, 2011). Such studies typically assume, however, that focusing and defocusing average out in the data processing. The purpose of this paper is to attempt to improve understanding of the nature of elastic propagation effects on surface waves, particularly their amplitudes downstream from sedimentary basins. Our hypothesis is that a significant fraction of the observed amplitude variability is caused by elastic focusing/defocusing due to lateral wave propagation effects. The focus of this paper is on lateral wavefield effects on Rayleigh waves at 10 sec period, which is typically well excited by small earthquakes and nuclear explosions and is also well represented in ambient noise cross-correlations that are commonly used in tomographic studies. The existence and nature of sedimentary basins strongly affect regionally propagating Rayleigh waves at this period.

Our study begins with a 3D wavefield simulation in East Asia (Fig. 1), a region chosen because of the presence of significant sedimentary basins, identified by cross-hatching in Figure 1 and named in Table 1. The model of the earth is a recent model of crustal and uppermost mantle structure constructed by Shen et al. (2016) for eastern China and surrounding regions. These authors call their model the “China Reference Model”, and take particular pains to represent sedimentary structures and present the model specifically for use for other studies. We concentrate our simulation on the Rayleigh waves at 10 s period, which propagate long distances on continents but are very sensitive to near surface structures.

Another reason for our concentration on East Asia is the interest in monitoring nuclear explosions that occur on the North Korea nuclear test site. In nuclear monitoring, the ability to discriminate nuclear explosions from naturally occurring seismic events such as earthquakes rests in part on the ability to measure reliably and interpret the amplitude of body waves and surface waves that are generated by these sources. Body and surface wave amplitude measurements are commonly converted into the magnitude estimates mb and Ms, respectively. Although many nuclear explosions are characterized by a small mb:Ms ratio relative to most earthquakes, there are exceptions (e.g. Bowers & Selby, 2009; Selby et al., 2012). Strong spatial variations in the amplitudes of surface wave have been observed for nuclear tests in North Korea (e.g., Bonner et al., 2008; Koper et al.,
It is commonly believed that the substantial scatter in Ms estimates results from the impact of sedimentary basins in East Asia, although the dominant physical causes of these amplitude anomalies have been unclear.

The paper is organized into approximately two equal parts. The first part of the paper comprises sections 2 – 4 where we present the results of the 3D simulation across East Asia. We describe the Earth model used in the 3D simulations and the numerical scheme (section 2), present simulated observations of surface wave travel times and amplitudes (section 3) across East Asia at 10 s period, and then present measurements of Ms. The final section illustrates how elastic structure would bias surface wave magnitude (Ms) measurements when the amplitude effects of elastic heterogeneity are ignored. The 3D wavefield simulations across East Asia are based on the code SES3D, which is a 3D spectral element solver in spherical coordinates (Gokhberg & Fichtner, 2016). One of the features of this code is that it is straightforward to implement arbitrary 3D models.

The second part of the paper comprises sections 5 – 8, which aim to illuminate the physical cause(s) of the surface wave amplitude anomalies across East Asia simulated in the first part of the paper. The full waveform numerical examples presented in this part of the paper are carried out by SPECFEM2D (e.g., Komatitch et al., 2001) and SW4 (Petersson & Sjogreen, 2014), which are designed for 2D and 3D wavefield simulations in Cartesian coordinates, respectively. In Section 5 we show that most of the wavefield effects observed in the 3D simulations can be reproduced quite accurately with horizontal 2D membrane wave simulations. Section 6 presents a primary hypothesis of this paper, that elastic amplitude anomalies downstream from sedimentary basins result primarily from lateral focusing. We provide several lines of evidence to support this hypothesis including the coincidence of amplitude anomalies with lateral deflections in wavepaths and that surface wave amplitude anomalies are very similar if simulated in two horizontal dimensions through a phase speed map or through a 3D structural model. In section 7, we discuss how surface wave focusing depends on the scale and geometry of elastic heterogeneity. Finally, in section 8, we compare the effects of anelastic attenuation and elastic focusing and show that on a regional scale elastic focusing through sedimentary
basins is more likely to cause significant surface wave amplitude anomalies than anelastic attenuation produced by the basins.

2. 3D Model and Wavefield Simulation

We begin the paper with a depiction of wavefield effects that emerge from recent generation of 3D crustal models. We employ a recently produced 3D isotropic model of the crust and uppermost mantle beneath China (Shen et al., 2016) developed using Rayleigh wave dispersion measurements (8 – 50 s period) obtained from ambient noise and earthquake tomography. The model is intended as a reference for studies like ours, and the authors refer to it as “China Reference Model”. The model is presented on a 0.5°x0.5° grid and extends from the surface to a depth of 150 km. Surface topography is rendered by changing crustal thickness.

Here we describe only a few salient characteristics of the China Reference Model. Importantly, the authors took care in representing sedimentary basins in which sedimentary structure is summarized with three unknowns at each grid node: sedimentary thickness and the top and bottom shear wave speeds (Vs) in the basin. Vs grows monotonically and linearly with depth in the sediments. The density of the sediments and the crystalline crust are computed from Vs using the scaling relationship of Brocher (2005), whereas Vp/Vs is 2.0 in the sediments and 1.79 in the crystalline crust.

For anelasticity, we replace the 3D Q model of Shen et al. with the 1D Q model of Durek & Ekström (1996) because we are primarily interested in the impact of lateral variations in elastic structures on surface waveforms. The visco-elastic relationship is implemented in the simulation using a series of Standard Linear Solids (SLS). We find a set of relaxation times and the corresponding weights using a simulated annealing algorithm (Kirkpatrick et al., 1983), which results in an almost constant Q from about 10 to 20 s period. The result of using a 1D Q model is that all lateral variations in the wavefield will derive exclusively from elastic structures.

The geographical region of the model is shown in Figure 1 by the blue box near whose boundary we place a perfectly matched layer (PML, Berenger, 1994). Similarly, there is a perfectly matched layer at 200 km depth. We replace water layers with sediments but the sedimentary structure at each location is designed to fit the observed local surface wave...
speeds. The largest impact of this replacement occurs in the Sea of Japan, where water depth is the greatest. Because this results in a physically unrealistic model off the coast, we confine our simulation to the continent as much as possible. The original China Reference Model has an irregular shape; therefore, we smoothly extrapolate the edges of the original model to get a “rectangular” model that is acceptable for our simulations. The resulting model is an oversimplification of sedimentary structure, although it fits surface wave phase speeds quite well across our study region. Shen et al. (2016) show that the standard deviation of the misfit to interstation phase time measurements is less than 1 s at most periods. As is common with tomographic models, the amplitude and sharpness of the structural anomalies are probably underestimated, a problem that grows in significance as structures reduce in spatial size. We believe, however, that the model is the best alternative to represent structural effects on surface waves across the study region.

Figure 2 illustrates some of the structural features of the China Reference Model by presenting horizontal slices of Vs at the depths of 3 km, 10 km and 20 km as well as topography on the Moho. The model captures many geological structures of the uppermost crust such as the Songliao Basin, the Bohaiwan Basin, the Sichuan Basin and the Subei Yellow Sea Basin, which are all seen at 3 km depth.

As shown in Figure 1, the seismic source is located on the North Korea nuclear test site. The moment tensor is isotropic to remove the radiation pattern from the wavefield. The source has a moment magnitude $M_w = 4.07$. We apply a fourth order Butterworth bandpass filter to the Heaviside step function with corner frequencies at $f_{min} = 0.05$ Hz and $f_{max} = 0.1$ Hz to produce the input source time function (Figure 3).

3. Wavefield Analysis

Figure 4 presents vertical component wavefield snapshots at times of 120, 460, 700, 1020 s after the initiation of the source. The Rayleigh wave dominates these wavefields. A corresponding wavefield animation can be found at https://www.youtube.com/watch?v=BGRdea7aY4k. The wavefront distortion is mostly caused by the sedimentary basins. There are four principal types of distortion. (1) The
wavefront is retarded and buckles inward during propagation through a basin (e.g., 460 s snapshot, Fig. 4b). (2) After the wavefront emerges from the basins it amplifies. (3) The basins generate a secondary wavefront with a smaller radius of curvature that trails the primary wavefront (e.g., 700 snapshot, Fig 4c). (4) There is attendant de-amplification that brackets the region of strong amplification. De-amplification is harder to observe in the 3D wavefield simulation through the China Reference Model, but we will attempt to clarify it in the simulations through idealized structures presented later in the paper. The basin that has the largest impact on the wavefield is the Bohaiwan Basin.

Given the synthetic seismograms from the simulation, we measure the dispersion curves for phase velocity and spectral amplitudes to determine the travel time (Fig. 5a) and amplitude maps (Fig. 5b). We make these measurements using frequency-time analysis (FTAN, Levshin et al., 1972; Levshin & Ritzwoller, 2001). Although the simulation is for the band between 10 and 20 s period, we concentrate interpretation at 10 s period because the impact from sedimentary basins is stronger at the shorter period end of our bandwidth of study. Large amplitude stripes appear (Fig. 5b) where the travel time level lines cave inward (Fig. 5a).

We can determine the apparent propagation direction at each grid point using the eikonal equation:

$$\frac{k}{c(r)} = \nabla \tau(r)$$  \hspace{1cm} (1)

where the left hand side is the unit wavenumber vector located at position $r$ divided by the corresponding phase speed and the right hand side is the gradient of the local travel time. The local direction of the gradient of the travel time field gives the apparent propagation direction. The angular difference between the propagation direction and the great circle path, which reflects the local deflection of the propagation direction, is presented in Figure 5c. As discussed later in section 6, the propagation deflections bracket the amplification stripes in Figure 5b.

For comparison, we also present the direction of propagation and deviation from the great-circle path using travel times computed by 2D ray tracing on a sphere followed by the application of the eikonal equation in Figure 5d. This is computed using the input
phase speed map in Figure 5e from the China Reference Model. Ray tracing is performed with the fast-marching method (FMM) of Rawlinson & Sambridge (2004) for the 10 s Rayleigh wave. We also discuss this result in section 6.

4. Bias of Surface Wave Magnitudes Due to Elastic Propagation Effects

As discussed earlier, reliable surface wave magnitude (Ms) measurements are needed to help discriminate nuclear explosions from earthquakes. Here, we illustrate how large of a bias can be introduced in the Ms measurement unless 3D propagation effects through sedimentary basins are taken into account.

We apply Russell’s empirical formula (Russell, 2006) to measure Ms in our simulations, which is defined as follows:

$$M_s = \log(a_b) + \frac{1}{2} \log(\sin(\Delta)) + 0.0031 \left( \frac{20}{T} \right)^{1.8} \Delta - 0.66 \log\left( \frac{20}{T} \right) - \log(f_c) - 0.43 \quad (2)$$

where $T$ is period, $f_c$ is the corner frequency controlled by the period and the epicentral distance $\Delta$ (in degree). $a_b$ is the measured amplitude filtered with a bandwidth $[1/T-f_c, 1/T+f_c]$. The formula is designed to correct the amplitude measurements empirically for geometrical spreading, anelastic attenuation and surface wave dispersion. We first use this formula to measure Ms based on simulation through the laterally homogeneous model ak135 (Kennett et al., 1995) using the same source (Mw = 4.07) we applied to the 3D simulation. At an epicentral distance of $\Delta = 20^\circ$, at 10 s period, Ms = 3.2. This value of Ms serves as the reference for the analysis of the 3D synthetic data through the China Reference Model.

Figure 6 shows that the Ms measurements vary from 2.5 to 3.9 across the study region through the China Reference Model at 10 s period. The range of variation is similar to measurements made on real data across the region (e.g., Bonner et al., 2008). Because the reference Ms = 3.2, we consider Ms estimates that fall in the range from 3.1 to 3.3 as unaffected by elastic amplification or de-amplification. More than one third of the region has measurements that are biased outside this range. The Ms map clearly illustrates attendant de-amplification zones that bracket the stripes of strong amplification. We discuss this in section 6 with an idealized example.
Figure 7 illustrates the variation of measurements of surface wave amplitude and Ms with azimuth and distance. The selected grid nodes for Figure 7 are identified with colored lines in Figure 6. Assessment of azimuth variation is based on three groups of grid nodes at different distances (distance ~ 500 km, 1100 km, 1700 km) and is shown in Figure 7a and c. For a distance of about 500 km (blue dots), amplitude measurements display very little azimuthal variation and Ms estimates are approximately constant. However, for a distance of about 1100 km (green dots), the measurements oscillate with azimuth and display several peaks due to amplification and de-amplification. At greater distances (~1700 km, red dots), the peaks corresponding to the amplifications become narrower with larger variations in both amplitude and Ms. To illustrate variation with distance, we also group our measurements in three azimuth ranges in Figures 7b and d at azimuths of 235°-236° (blue dots), 253°-254° (green dots), and 300°-301° (red dots), respectively. The 253°-254° azimuth range has the largest amplification effect, while the 235°-236° range is expected to be de-amplified as it is between two amplification stripes. The 300°-301° range has neither amplification nor de-amplification. Two of these groups of measurements (235°-236° and 300°-301°) display a clear decay of amplitudes with distance while for the azimuth range (235°-236°) amplitudes tend to be systematically smaller than the other groups at distances larger than 1500 km. The Ms measurements in the two groups of measurements shown in Figure 7 illustrate how Russell’s empirical formula cannot account for focusing and defocusing.

In summary, elastic structures in sedimentary basins strongly impact both surface wave amplitudes and Ms estimates of regionally propagating surface waves at 10 s period. Amplitudes may be three times larger and the Ms variation can be as large as 1.4 magnitude units across East Asia for events on the North Korea nuclear test site.

5. 2D versus 3D Amplitude Effects

In this section, we investigate whether the wavefield effects observed downstream from sedimentary basins in the 3D simulations presented in sections 3 – 4 fundamentally reflect 3D phenomena or if they can be well approximated with 2D simulations. To do this we compare here results from 2D and 3D simulations.

A complication in comparing 2D and 3D simulations is that overtones are generated in
the 3D simulations that are absent in the 2D simulations. To test this effect, we compare amplitudes and group velocities measured through the laterally homogeneous model ak135 (Kennett et al., 1995) using full waveform seismograms (produced by SW4, Petersson & Sjogreen, 2014) and normal mode synthetics that include only the fundamental mode (Herrmann & Ammon, 2002). The source is an explosion at a depth of 1 km. We use frequency-time analysis to measure the group velocity and the amplitude from the synthetic seismograms as a function of epicentral distance. The amplitude measurements are corrected using a 2D geometrical spreading factor and normalized using:

\[
A = \frac{A_{\text{obs}} \sqrt{D}}{A_{1000} \sqrt{1000}}
\]  

(3)

where \(A_{\text{obs}}\) is the observed amplitude, D is the source-receiver distance along the earth’s surface, \(A_{1000}\) is the amplitude measured at a distance of 1000 km. As Figure 8 shows (red dots), the group velocity and corrected normalized amplitude measurements from the 3D full waveform synthetics display oscillations whose amplitude decays with distance from the source. In contrast, the fundamental mode measurements are range-independent (black dashes).

We conclude that for source-receiver distances greater than about 200 km, overtone interference is weak enough to interpret amplitude measurements made on the fundamental mode. Only at epicentral distances greater than 200 km, therefore, can we compare amplitude measurements obtained on 2D and 3D synthetics.

Figure 9 illustrates the input models for the comparison between the 2D and 3D simulations. The input model for the 3D simulation is a rectangular region with dimensions 3000 km \(\times\) 600 km \(\times\) 200 km, in which an explosive source is located at \(x = 100\) km, \(y = 300\) km, \(z = 1\) km. The velocity structure is ak135 in which a circular sedimentary basin is embedded centered at \(x = 1100\)km, \(y = 300\)km. The basin has a diameter of 200 km and extends to a depth of 5 km. Shear velocity inside the basin increases linearly with depth from 2 km/s at the top (\(z = 0\) km) to 3 km/s at the bottom (\(z = 5\) km) of the basin. Figure 9a illustrates the 3D shear wave speed profiles inside and outside the basin.
The 2D model is constructed from the 3D model. In fact, the 2D phase velocity map for the 10 s Rayleigh wave is the input model for the 2D membrane wave simulation. Because the absorbing boundary conditions (Stacey, 1988) implemented in the 2D membrane wave code (SPECFEM2D) do not perform ideally and artificial reflections from the boundaries are generated, the 2D simulation region is larger than the 3D case. As shown in Figure 9b, the green rectangle represents the size of the 3D simulation region while the black one illustrates the size of the 2D modeling region.

Figure 10 presents wavefield snapshots from the 3D and 2D simulations. The principal difference is that the wavefield in 2D is non-dispersive whereas the wavefront in 3D is dispersed, and is therefore wider. The simulations show several similar qualitative patterns, however. (1) The wavefronts are retarded inside the basin (Fig. 10a,b). (2) Downstream from the basin, there is increased amplitude (Fig. 10c,d). (3) There is the generation of secondary wavefronts where there is amplification (Fig. 10c,d). The secondary wavefront is clearer on the 2D simulation (Fig. 10d).

To perform a quantitative comparison between the 2D and 3D wavefields, we measure the spectral amplitudes at 10 s period for the 2D and 3D synthetics and compare them directly in Figure 11. The amplitude measurements are normalized to the value observed at 400 km. The red rectangular region in the figure indicates the location of the sedimentary basin and the low velocity anomaly. The dominant features of the 2D and 3D results are consistent with each other: they both predict similar amplitude anomalies downstream from the basin, which we conclude is a focusing phenomenon. Both simulations depict amplification inside the basin, although the differences inside the basin are somewhat larger than outside the basin. This may be caused by the interference of multimode surface waves in the 3D modeling, which are not simulated in the 2D computation.

Another characteristic seen in the simulation results presented in Figure 11 is the decay of the surface wave amplification downstream from the anomaly. This decay results from wavefront healing (e.g., Nolet & Dahlen, 2000), which is a diffraction phenomenon consistent with Huygens’ principle that causes wavefront distortion and amplification to decay with distance downstream from the basin. In contrast with body waves
(Marquering et al, 1999) where the travel time perturbation asymptotically approaches zero at large propagation distances, surface waves undergo only incomplete healing of both travel time and amplitude perturbations irrespective of the epicentral distance. This phenomenon is reflected in 2D surface wave sensitivity kernels, which do not go to zero at the center of the kernel (Fig. 12) unlike the body wave travel time kernel.

The similarity of results from 2D and 3D simulations means that phase velocity maps can be used to make amplitude predictions. This is not only computationally much faster, but phase velocity maps are a more direct observable than 3D models, which are inferred from phase velocity maps and perhaps other data. Also, accurate 3D simulations of short period seismic waves through water layers (e.g., Japan Sea) are challenging, but can be circumvented with 2D membrane wave modeling.

6. The Physical Cause of Surface Wave Amplification/De-Amplification

6.1 Elastic Focusing

Wavefield snapshots such as that shown in Figure 4 (e.g., t = 700 s) as well as measurements obtained on the wavefields such as those in Figures 5 – 7 reveal that waves that travel through sedimentary basins (or low velocity anomalies) undergo substantial amplification that persists downstream. We also observe attendant de-amplification that brackets the strong amplification stripes (e.g., Fig 6). The amplification of the wavefields is similar in 2D and 3D, as illustrated in Figure 11.

Some of the principal characteristics of the simulated wavefields are illustrated in the cartoon of Figure 13, which include the formation of a wavefield concavity inside the velocity anomaly, a secondary wavefront with a smaller radius of curvature than the primary wavefront outside the velocity anomaly, a high amplitude stripe directly downstream bracketed by two low amplitude zones, and the decay of the amplitude anomaly with distance from the basin. The primary wavefront diffracts around the low velocity anomaly and outpaces the secondary wavefront, but adheres to it (dashed lines). The arrows in Figure 13 are drawn normal to the wavefronts and represent local wavefield propagation directions. This section and section 7 are devoted to illuminating the cause of surface wave amplification/de-amplification downstream from sedimentary basin.
A 2D lens model is a useful heuristic to illuminate the formation of the secondary wavefront. Figure 14 illustrates the “lens effect” in which a 2D cylindrical wave propagates from left to right, analogous to surface waves propagating in the horizontal plane along Earth’s surface. A lens is a low velocity anomaly, which has a similar impact on the wavefield as a sedimentary basin in the Earth. Assuming that the radius of curvature of the cylindrical wave is much larger than the size of the lens, the cylindrical wave in approximately planar. A wavefront traveling through the lens, shown in red, converges to a focal point, which then serves as a secondary source that generates a secondary wavefront with a smaller radius of curvature compared with the primary wavefront (in blue) that does not travel through the lens.

Wavefield convergence results in amplification. We expect the focal point to have the largest amplification because it has the greatest wavefield convergence. The surface wavefield will remain amplified downstream from the focal point due to the bending of the primary wavefront and its adherence to the secondary wavefront (dashed lines, Fig. 13). This bending or deflection focuses energy and increases amplitudes. By conservation of energy, we also expect two de-amplification stripes to bracket the amplification region (Fig. 13). Indeed, the increase in amplitude is caused by the flow of seismic energy along the wavefront toward the axis of the velocity anomaly, which also causes amplitudes to decrease off-axis along the wavefront.

The 2D lens heuristic is introduced here to illustrate elastic focusing, notably the formation of a secondary wavefront, but we do not expect it to provide accurate quantitative predictions. There are two reasons for this. (1) Most basins are not shaped like a thin lens, which has a thickness that is much smaller than the radius of curvature of the lens surface. Only in this case would the wavefield be expected to converge to a single focal point. Even for a spherical lens, any departure from this ideal shape of thin lens would result in spherical aberration; i.e., the development of multiple points of convergence. (2) Also, the lens equations are ray theoretical, which are inappropriate in the presence of strong diffraction as exists in our applications.
6.2 Wavefront Bending and Amplification

If the sedimentary basins do amplify surface waves by elastic focusing, simulated wavefields will show counter-posing ray deflections and low amplitudes that bracket the amplification of the wavefield downstream from a basin. We now show that these criteria are satisfied, which is evidence that elastic focusing controls the amplification downstream from sedimentary basins.

To explore elastic focusing further, we investigate the distortion of a wavefront propagating downstream from a circular low velocity anomaly, with particular concentration on the direction of propagation of the wavefield. Because the primary wavefield effects are similar in 2D and 3D simulations, we present results here from 2D simulations but the principal conclusions would be unchanged if the experiment were performed in 3D.

Figure 15 shows a horizontal 2D membrane wave simulation using SPECFEM2D (e.g., Komatitch et al., 2001). The source and low velocity anomaly locations are shown in Figure 15a. The distance from the center of the anomaly to the source is a little less than 1000 km. The fractional wave speed perturbation of the low velocity anomaly in the “basin” is given by:

\[
\frac{\delta \beta}{\beta} = \begin{cases} 
\frac{\varepsilon}{2} \left[ 1 + \cos \left( \frac{\pi r}{R} \right) \right] & r \in [0, R] \\
0 & r \in (R, \infty) 
\end{cases}
\]

where \( \varepsilon = -10\% \) is the maximum velocity perturbation in the center of the basin, \( R = 100 \) km is the radius of the low velocity anomaly, and \( r \) is the radial distance from the center of the anomaly. Due to the effect of the low velocity anomaly, the travel time contour is distorted (Fig. 15b) inside and there is an amplification stripe downstream from the low velocity anomaly (Fig. 15c). There is also attendant de-amplification that brackets the amplification stripe (white stripes in Fig. 15c). The emergence of the de-amplification stripes is consistent with the shape of the cross-section of the amplitude sensitivity kernel (Fig. 12b). Using the eikonal equation (eq. 1), the gradient of the travel
time map provides the propagation direction of the wavefield. The angular difference between the propagation and the straight ray directions gives the wavefront deflection, as shown in Figure 15d. The red and blue stripes, which represent positive and negative deflections, come in pairs that bracket the amplification stripe in Figure 15c.

The relation between the propagation directions and the amplification stripe in Figure 15c,d is similar to the relation between the amplification and deflections stripes that appear in the 3D simulation through the China Reference Model shown in Figures 5b,c and in the ray tracing result shown in Figure 5e. Figure 5c has more detailed bifurcations and somewhat larger magnitudes of the off-great circle deflection than Figure 5e, which may be due to wavefield scattering in the full waveform simulation.

We conclude that both the 3D and 2D amplification phenomena shown in Figures 5 and 15 are caused primarily by the elastic focusing of surface waves.

7. Effect of Basin Size on Surface Wave Amplification and Focusing

Surface wave amplification depends on the shape and size of basin (low velocity anomaly). To illustrate this, we present several 2D membrane wave simulations with circular anomalies of different diameters. In the numerical scheme the distance from the source to the center of the circular anomaly is 3000 km. The fractional wave speed perturbation of the anomaly is given by equation (4) with $\varepsilon = -10\%$. We vary the radius of the anomaly, which is $R$ in equation (4), incrementally from 100 km to 400 km in the different simulations.

The amplitude curves for different simulations are presented in Figure 16a from which we draw two conclusions. (1) The maximum amplitude increases as the size of the basin (velocity anomaly) grows. This also can be explained with the lens model. When the size of a lens increases, the primary wavefield out-distances the secondary wavefield by a larger amount than for a small lens, which increases the curvature of the diffracted part of the wavefield and increases the amplitudes. This is summarized in Figure 16b, which shows both the increase in the maximum amplitude with the size of the velocity anomaly and the increase in the maximum ratio between the observed amplitude and the amplitude of the wavefield without the basin, which we call the maximum amplitude ratio. The
maximum amplitude increases sub-linearly with the size of the velocity anomaly, although the maximum amplitude ratio is more nearly linear with the size of the velocity anomaly.

(2) The location of the maximum amplitude also increases with basin radius. If the incoming wavefront were perfectly planar, then based on the lens heuristic the location of the maximum amplitude would be the focal point. However, the incoming wavefront is only approximately planar. To estimate the focal length of the velocity anomaly for the non-planar incoming wavefront, we can use the thin lens approximation. If we define the image distance \( d_i \) as the distance from the center of the velocity anomaly to the location of the maximum amplitude, we can approximately determine the focal length using the thin lens formula:

\[
\frac{1}{d_i} + \frac{1}{d_o} = \frac{1}{f}
\]  

where \( d_o = 3000 \text{ km} \) is the object distance from the source to the center of the velocity anomaly, and \( f \) is the focal length to be computed. Figure 16c shows that the focal length defined in this way increases approximately linearly with the radius of the anomaly (blue dots: computed focal length from eq. (5); black line: linear regression curve). Note that the analysis above is only approximately valid, because a circular low velocity anomaly is not an ideal thin lens and it is not expected to have a single focus.

There are many other factors that will affect elastic focusing and surface wave amplitudes, including the curvature of the wavefront (or source-to-basin distance), the shape of the basin or velocity anomaly (e.g., the aspect ratio), the wavelength of the wavefield, and the velocity perturbation. We discuss only one of these here, namely the shape of the velocity anomaly and display its potentially profound impact on amplitudes by presenting a somewhat extreme example. Figure 17 shows surface wave amplitudes measured in a 2D membrane wave simulation through a rectangular low velocity anomaly centered 1000 km from the source. The aspect ratio of the velocity anomaly is extremely small; i.e., the long direction of the anomaly (perpendicular to the direction of propagation) is much greater than the short direction (along the direction of propagation). In this case, there is barely any amplification downstream from the basin, as shown in Figure 17b, although
there is amplification in the basin. Amplification is caused by focusing due to the bending of the wavefront, but when a plane wave hits an anomaly that is very long in the direction transverse to the propagation direction, the wavefront will not bend, which is why there is little focusing.

In summary, a rule of thumb is that amplification strengthens and the focal length increases when the size of the low velocity anomaly increases, approximately linearly for a circular anomaly. However, this relation breaks down for velocity anomalies with extreme aspect ratios.

8. Anelastic Attenuation versus Elastic Focusing

The simulations we have presented so far concentrate on the effect of near surface elastic heterogeneity on surface wave amplitudes. Here, for comparison we address the effect of anelastic attenuation by presenting more 3D simulations using the code SW4. (None of the simulations in this section are in 2D.)

The quality factor Q is lower in sedimentary than non-sedimentary rocks, thus strong anelastic attenuation will decrease amplitudes of surface waves that pass through sedimentary basins, somewhat offsetting amplitude gains caused by elastic structures. Here, we show that the amplification effect due to focusing is expected to be much larger than the attenuation effect due to the anelasticity, at least for a circular basin.

We present several different 3D simulations in Cartesian coordinates to investigate this problem. The background input velocity structure in 3D is the 1D model ak135 with a circular basin inserted. The basin has a smoothly varying cross-sectional depth, defined as:

\[
z = \begin{cases} 
\frac{z_{\text{max}}}{2}[1 + \cos(\pi r / R)] & r \in [0,R] \\
0 & r \in (R,\infty) 
\end{cases}
\]

where \(z_{\text{max}} = 4 \text{ km}\) is the maximum depth of the basin, \(R = 200 \text{ km}\) is the basin’s radius, \(r\) is the radial distance from the center of the basin. The depth of the basin varies as a half-cosine with radial distance from the center of the basin. The basin’s center is 1000
km from the source. The shear wave quality factor $Q_s$ and the shear wave speed $\nu_s$ are constant in the basin. Given $Q_s$, $Q_p$ is determined with the empirical relationship of Clouser & Langston (1991) and $\nu_p$ is determined from $\nu_s$ by the relationship of Brocher (2005). The three simulations have input basins that differ in quality factor and shear velocity, but have the same geometry.

The model parameters inside the basin for the three input models are summarized in Table 2 and are described as follows. Model 1 has both elastic and anelastic heterogeneity, being a low Vs, low Q basin. Model 2 has only elastic heterogeneity, being a low Vs but normal Q basin. Model 3 has only anelastic heterogeneity, being a normal Vs but low Q basin. Thus Model 1 contains both a Vs and Q anomaly, Model 2 contains only a Vs anomaly, and Model 3 contains only a Q anomaly. Our main interest is to compare the results of Models 2 and 3, which contain purely elastic and purely anelastic effects, respectively.

We use the amplitude ratio in Figure 18a to illustrate the results, which is the ratio of observed amplitude in the simulations ($A_{obs}$) to the reference amplitude observed with the horizontally homogeneous model ak135 ($A_{ref}$). Strong anelastic attenuation (Model 3) alone only decreases amplitudes to a small extent (red dots). The elastic structural anomaly (Model 2) has a much larger effect on amplitudes downstream from the basin (green dots). A basin that has both low velocity and low quality factors (Model 1) still results in a significant increase in the amplitudes downstream from the basin (blue dots).

Figure 18b shows that the magnitude of the downstream amplitude anomalies in simulations with a purely anelastic heterogeneity varies approximately linearly with the size of the basin, as does the maximum elastic amplitude ratio (Fig. 16b). However, the anelastic effect is much smaller. As shown in Figure 18a, for a basin with a radius of 200 km, the magnitude of anelastic amplitude decay would be ~7% but the maximum elastic amplification would be ~200%. Thus, the expectation is that elastic amplification effects will dominate anelastic attenuation effects for circular basins.

There are three primary exceptions to the dominance of elastic amplification over anelastic attenuation. (1) Anelastic attenuation begins to set on as soon as the wavefield
enters the basin and persists approximately constant downstream after the wavefield exists the basin. Elastic amplification does not set on immediately, but maximizes near the focal point which will occur several hundred kilometers downstream from the basin and then decays slowly (Fig. 16a). This caveat is more important for large basins whose focal point lies at a greater distance from the basin. For a basin with a 400 km radius, anelastic attenuation will be stronger than elastic amplification within about 250 km from the basin edge in our simulations, but will be weaker outside this distance. (2) Elastic amplification reduces downstream from the focal point, whereas anelastic attenuation remains constant with distance. Thus, for waves that propagate far enough from a basin the elastic amplification may decay to become commensurate with anelastic attenuation. This will be more likely for small basins where the amplification decays more rapidly with distance. However, elastic amplification does not asymptotically approach zero (as discussed earlier). Even for a basin that is only 100 km in radius, the elastic focusing is expected to be much stronger than anelastic attenuation at continental scales (distances less than 10000 km). (3) For basins with extreme aspect ratios, such as shown in Figure 17, elastic amplification may be minimal and anelastic attenuation may have a larger impact on amplitudes simply because there is barely any focusing.

In conclusion, we expect elastic amplification downstream of on-continent sedimentary basins to dominate anelastic attenuation except very near the basin edge and for basins with extreme aspect ratios.

9. Conclusions

This paper explores the nature of elastic propagation effects on short period surface waves, particularly their amplitudes downstream from sedimentary basins. Our hypothesis is that a significant fraction of amplitude variability observed in regionally propagating surface waves (e.g., Bonner et al., 208) is caused by elastic focusing/defocusing due to lateral wave propagation effects through shallow structures. The focus of this paper is to understand elastic focusing effects on Rayleigh waves at 10 sec period, which is typically well excited by small earthquakes and nuclear explosions and is also well represented in ambient noise cross-correlations that are commonly used
in tomographic studies. The existence and nature of sedimentary basins strongly affect regionally propagating Rayleigh waves at this period.

The paper is organized into approximately two equal parts. The first part of the paper presents the result of a 3D simulation across East Asia based on the code SES3D (Gokhberg & Fichtner, 2016) through the 3D model refereed to as the “China Reference Model” by Shen et al. (2016). An isotropic source is placed on the Korea Nuclear test site and the simulated waveforms display significant Rayleigh wave amplitude anomalies across East Asia caused by the on-continent sedimentary basins near the test site. Sedimentary basins amplify surface waves that propagate downstream from them, they generate de-amplification regions that brackets the amplification stripes, and produce a secondary wavefront with a smaller radius of curvature than the primary wavefront. The amplitude anomalies decay slowly as distance from the basins increases. Ms measurements obtained on these waveforms vary from 2.5 to 3.9 across the region, similar to measurements on real data (e.g., Bonner et al., 2008).

The second part of the paper aims to illuminate the nature and physical cause of the surface wave amplitude anomalies simulated in the first part of the paper. We posit a 2D lens heuristic that explains the formation of the secondary wavefront and understand amplification/de-amplification to result from lateral focusing due to the bending of the primary wavefront. We provide several lines of evidence to support this hypothesis. This evidence includes the coincidence of amplitude anomalies with lateral deflections in wavepaths, which is consistent with focusing, and the fact that surface wave amplitude anomalies are very similar if simulated in two horizontal spatial dimensions through a phase speed map or through a 3D structural model. We discuss how surface wave amplification depends on the size and shape of velocity anomalies such that both the maximum amplification and the focal length increase approximately linearly with the size of a circular basin.

The fact that 2D phase speed maps can be used as proxy for 3D structural models to predict surface wave amplitude anomalies has several practical implications. First, 2D membrane wave simulations are much faster. Second, the existence of water layers can complicate simulations through a 3D model that does not happen in 2D simulations.
Third, phase speed maps are closer to data than 3D models and are therefore more accurate representations of heterogeneity. Thus, phase speed maps present several advantages in computing amplitude anomalies, including that they may be more accurate than those computed through a 3D model.

Finally, we compare the effects of anelastic attenuation to elastic focusing and show that on a regional scale elastic focusing through sedimentary basins is more likely to cause significant surface wave amplitude anomalies than anelastic attenuation produced by sedimentary basins except very near the basin edge or for basins with extreme aspect ratios.

In the future, it is important to test the principal conclusions of this paper with real data. This will include tests to observe strong amplification stripes downstream from sedimentary basins, and perhaps also the de-amplification and propagation deflection stripes that bracket the amplification. In addition, it is also important to test whether the observed features are predicted well with high quality velocity models. To achieve this, there are three major requirements that need to be satisfied. (1) A dense array with high quality seismometers is needed to record accurate spatially resolved amplitude information. The array should be located near to a large sedimentary basin. (2) Seismic events upstream from the basin are also needed with magnitudes large enough to be recorded by the array. Ideally they would also be small enough and far enough to be considered as point sources. (3) A high-resolution 3D model (or 2D phase velocity map) also should be available for the study region.

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References


Figure Captions

Figure 1. The blue box encloses the study region in which stippled areas identify the sedimentary basins, which are named in Table 1. The red star is the location of the source for the 3D simulation near the North Korea nuclear test site.
Figure 2. Horizontal cross-sections from the China Reference Model (Shen et al., 2016) of Vs are presented at depths of (a) 3 km, (b) 10 km, and (c) 20 km as well as (d) crustal thickness. The grey polygons in (a)-(c) indicate major basins in the study region (Table 1).
Figure 3. Source time-function. The upper panel is the time series while the lower one is the time derivative of the source time function in the frequency domain. (Far field displacement is proportional to the derivative of the source time function).
Figure 4. Wavefield snapshots at times $t = 120, 460, 700, 1020$ s. (a) At $t = 120$ s, the wavefront is still circular. (b) At $t = 460$ s, the wavefront is traveling through the Bohaiwan and Subei basins and has been greatly distorted. (c) At $t = 700$ s, a secondary wavefront has been generated where amplification occurs. (d) At $t = 1020$ s, the Sichuan basin’s impact on the wavefront is apparent.
**Figure 5.** Plots of 10 s Rayleigh wave measurements of (a) travel time, (b) amplitude, (c) angular difference between the propagation direction and the local great circle direction. (The angle is defined in polar coordinates with the axis pointing to the East, counterclockwise direction is positive.) (d) Angular difference between 2D ray tracing propagation direction and the local great circle direction. (e) Input phase velocity map (10 s Rayleigh wave) for the ray tracing computation in (d).
Figure 6. Surface wave magnitude (Ms) map (10 s), computed using Russell’s formula (Russell, 2006) from the amplitude map (Fig. 5b) taken from the 3D simulation through the China Reference Model. The colored solid and dashed lines mark locations for the azimuthal and distance dependent curves shown in Fig. 7. Ms ~ 3.2 for a laterally homogeneous model.
**Figure 7.** Detailed plots of the 10 s Rayleigh wave amplitude and Ms as a function of distance and azimuth from the 3D wavefield simulation through the China Reference Model. Locations of the points on these plots are presented in Fig. 6 where those lines are colored coded as the symbols in this figure. (a) and (b) are the amplitude measurements while (c) and (d) are the Ms estimates. The full amplitude field across the study region is shown in Fig. 5b and the Ms across the region is shown in Fig. 6. Ms ~ 3.2 for a laterally homogeneous model.
Figure 8. Synthetic experiment to test overtone interference on measurements of the fundamental mode in a 3D simulation. Red dots result from a 3D simulation through a laterally homogeneous model whereas the black dashed lines represent fundamental mode synthetic results through the same model. (a) Measured group velocity and (b) corrected normalized amplitude measurements are presented as a function of epicentral distance. Oscillations in the measurements made on the 3D synthetics are caused by overtone interference near the source where the overtones and fundamental mode are not yet separated in time. Amplitude interference decays below about 1% at distances beyond 200 km, whereas group velocity perturbations extend to greater distances.
Figure 9. Specification of the input 2D and 3D models to test the similarity of wavefield measurements obtained in 2D and 3D simulations. (a) Vertical Vs profile for the 3D input model (red line with dots: inside the basin, black line: outside the basin). (b) Input model geometry for the simulations. The green rectangle is the size of 3D simulation region whereas the black rectangle is the 2D simulation region. The red star is the location of the source. Receivers are aligned on the horizontal blue dashed line.
Figure 10. Wavefield snapshots extracted from the 3D and 2D simulations through a circular low velocity anomaly for the simulation geometries shown in Fig. 9. (a, b) Distortion of wavefront inside the anomaly for the 3D and 2D cases, respectively. (c, d) Amplification (focusing) and the generation of a secondary wavefront for the 3D and 2D cases. The 3D wavefield is more protracted in time due to surface wave dispersion.
Figure 11. Normalized amplitudes obtained at 10 s period from the 2D and 3D simulations (green dots: 2D amplitudes, blue dots with dashed line: 3D amplitudes, red line: 1D (horizontally homogeneous) amplitudes). The simulation geometries for the 2D and 3D cases are shown in Fig. 9 and wavefield snapshots in Fig. 10. The colored box indicates the distance range of the low velocity “basin”.
Figure 12. Cross-sections of 2D surface wave sensitivity kernels, shown to illuminate wavefront healing. (a) Cross-section of the travel time kernel, (b) cross-section of the amplitude kernel. (Sensitivity kernels are not used in any of the quantitative analysis presented here).
Figure 13. Cartoon summarizing some of the principal wavefront characteristics in our simulations including the formation of a concavity inside the low velocity anomaly, the formation of a secondary wavefront with a smaller radius of curvature than the primary wavefront, the bending of the primary wavefront as it adheres to the secondary wavefront (dashed lines), and amplification and de-amplification downstream from the velocity anomaly. Wavefronts are shown with solid lines and normal to them are shown with arrows, which represent local wavefield propagation directions.
Figure 14. The 2D optical lens heuristic to explain the generation of a secondary wavefront (red line) downstream from the focal point. The radius of the cylindrical wave is assumed to be much larger than the size of the lens in this figure, so that the wavefield approximates a plane wave.
Figure 15. A 2D membrane wave simulation to illustrate the impact of a low velocity anomaly on the deflection of the propagation direction. (a) Input velocity model and source location, (b) travel time map, (c) amplitude map, (d) angular difference between the propagation direction and the straight ray path. (The angle is defined in polar coordinates where the axis points to the right and the counterclockwise direction is positive). The impact of a 2D low velocity anomaly is to produce a high amplitude streak downstream from the anomaly in (c) which is bracketed by two de-amplification stripes. The amplification stripe is also bracketed by lines of wavefield deflections clockwise (blue) and counterclockwise (red) in (d). This signature of wavefield amplification/de-amplification and deflection is characteristic of surface wave focusing and is also seen in the 3D wavefield simulation (e.g., Fig. 5b,c).
Figure 16. (a) Amplitude measurements from 2D simulations through circular low velocity anomalies with radii ranging from 100 to 400 km. (b) The maximum amplitude and maximum amplitude ratio plotted as a function of the radius of the velocity anomaly. Amplitude ratio is the amplitude in the simulation with the velocity anomaly divided by the amplitude in the laterally homogeneous model. (c) The focal length is computed using equation (5). The blue dots are the computed values and the black curve is the linear regression curve.
**Figure 17.** 2D membrane wave simulation to contrast with Fig. 16 to illustrate the importance of the shape of the velocity anomaly. (a) Source location (red star) and position of the low velocity anomaly (red rectangle with the long axis in the vertical direction, normal to the long axis in the wave propagation direction). The receivers are aligned along the blue dashed line. (b) The amplitude measurements obtained through the low velocity anomaly in (a).
Figure 18. (a) Amplitude ratio from three 3D simulations with different input models with varying amounts of elastic and anelastic heterogeneity. Model 1: both elastic and anelastic heterogeneity, Model 2: only elastic heterogeneity, Model 3: only anelastic heterogeneity. A full description of the models is summarized in Table 2. (b) Anelastic amplitude ratio is plotted as a function of the radius of the basin, where the anomaly is purely anelastic (i.e., no elastic heterogeneity).
# Tables

Table 1. Major basins in the study region. Numbers are found in Fig. 1

<table>
<thead>
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<th>Zones</th>
<th>Major Basins</th>
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Table 2. Velocity and quality factor inside the basin for different simulations in section 8

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