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Retrieving impulse response function amplitudes from the ambient seismic field

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SUMMARY

Seismic interferometry is now widely used to retrieve the impulse response function of the Earth between two distant seismometers. The phase information has been the focus of most passive imaging studies, as conventional seismic tomography uses traveltimes measurements. The amplitude information, however, is harder to interpret because it strongly depends on the distribution of ambient seismic field sources and on the multitude of processing methods. Our study focuses on the latter by comparing the amplitudes of the impulse response functions calculated between seismic stations in the Kanto sedimentary basin, Japan, using several processing techniques. This region provides a unique natural laboratory to test the reliability of the amplitudes with complex wave propagation through the basin, and dense observations from the Metropolitan Seismic Observation network. We compute the impulse response functions using the cross correlation, coherency and deconvolution techniques of the raw ambient seismic field and the cross correlation of 1-bit normalized data. To validate the amplitudes of the impulse response functions, we use a shallow M_w 5.8 earthquake that occurred on the eastern edge of Kanto Basin and close to a station that is used as the virtual source. Both S and surface waves are retrieved in the causal part of the impulse response functions computed with all the different techniques. However, the amplitudes obtained from the deconvolution method agree better with those of the earthquake. Despite the expected wave attenuation due to the soft sediments of the Kanto Basin, seismic amplification caused by the basin geometry dominates the amplitudes of S and surface waves and is captured by the ambient seismic field. To test whether or not the anticausal part of the impulse response functions from deconvolution also contains reliable amplitude information, we use another virtual source located on the western edge of the basin. We show that the surface wave amplitudes of the anticausal part agree well with those of a shallow M_w 4.7 event that occurred close to the virtual source. This study demonstrates that the deconvolution technique seems to be the best strategy to retrieve reliable relative amplitudes from the ambient seismic field in the Kanto Basin.

Key words: Body waves; Earthquake ground motions; Seismic interferometry; Seismic noise; Surface waves and free oscillations; Wave propagation.

1 INTRODUCTION

Over the past decade, new opportunities have emerged in seismology through seismic interferometry. Shapiro & Campillo (2004) empirically showed that the impulse response function (IRF) of the Earth can be retrieved by cross correlating ambient seismic field time series recorded by two distant seismometers. Under a certain distribution of ambient seismic field sources, the IRF of the medium retrieved by cross correlation yields the true Green's function (Snieder 2004; Wapenaar 2004). The ambient seismic field

on Earth is, however, mainly excited by the interaction of oceanic waves with shallow seafloor near the coasts (Hasselmann 1963) and by nonlinear interactions between oceanic waves travelling in opposite directions (Longuet-Higgins 1950; Hasselmann 1963). This leads to an uneven distribution of the ambient seismic field sources that potentially biases both phase and amplitude information of the IRFs.

The traveltimes information of the IRF has been, up to now, the primary focus of passive seismology. To reduce the biases caused by the non-uniform distribution of ambient seismic field sources,

Figure 1. Map of the Kanto Basin, Japan, including stations of the MeSO-net (purple triangles), Hi-net (black triangles), JMA (blue triangles) and University of Tokyo (green triangles) networks. Coastlines are represented by the black lines and 500 m spaced basin depth contours derived from the JIVSM are shown by the coloured lines. The names of the TENNOD and E.ZKUM virtual sources and the two receivers for which the waveforms are shown in Figs 2 and 8 are also shown. The epicentres of the M_w 5.8 and 4.7 earthquakes are represented by the red stars and their focal mechanisms are plotted. The two azimuths of 242° and 294° that are used in the following are represented by the two black lines departing from the M_w 5.8 earthquake epicentre. The upper right insert shows the Japanese Islands (black lines), plate boundaries (grey lines) and the zoomed region (red square).

pre-processing of the raw ambient seismic field (1-bit normalization, pre-whitening) is routinely performed before computing the cross correlations (Bensen *et al.* 2007). The traveltimes measurements of the pre-processed IRFs have been used to image Earth structure at a wide range of scales (Shapiro *et al.* 2005; Lin *et al.* 2008; Nishida *et al.* 2008; Boué *et al.* 2013; Lin *et al.* 2013; Mordret *et al.* 2014), to estimate seismic anisotropy (Mordret *et al.* 2013; Takeo *et al.* 2013), to monitor velocity changes in volcanoes (Brenguier *et al.* 2008b) and to study the response of the crust after large earthquakes (Brenguier *et al.* 2008a, 2014).

The amplitude information, in contrast, has attracted less attention because it appears to be more affected by the intensity and directionality of the ambient seismic field sources. Empirical (Stehly & Boué 2017), numerical (Cupillard & Capdeville 2010; Lawrence *et al.* 2013) and theoretical (Tsai 2011; Weaver 2011) studies predict a sensitivity of the amplitudes to the distribution of ambient seismic field sources as well as to the data processing. Despite this, several empirical studies demonstrated that the amplitude information seems to be preserved and can be used to simulate the long-period ground motions of moderate M_w 4–5 (Prieto & Beroza 2008; Denolle *et al.* 2013; Viens *et al.* 2014, 2015) and large M_w > 6.5 (Denolle *et al.* 2014a; Viens *et al.* 2016b) earthquakes, to map

site amplification (Denolle *et al.* 2014b; Bowden *et al.* 2015) and to infer seismic attenuation (Prieto *et al.* 2009; Lawrence & Prieto 2011). However, these studies use different processing techniques that might affect the amplitude of the IRFs, and thus our ability to evaluate their reliability.

The amplitude of seismic waves can be affected by several effects: the elastic effects, which are the geometrical spreading, multipathing (focusing and defocusing) and scattering effects, and the anelastic effects, which are also called intrinsic attenuation. These processes are particularly relevant in complex shallow crustal structures such as sedimentary basins. The low seismic-wave velocity of the soft sediments that compose the basin and the shape of the basement both contribute to the trapping of seismic waves and thus to their amplification. The Kanto Basin, Japan, is a sedimentary basin that is locally deeper than 4 km and is located beneath the Tokyo Metropolitan area. The Japan Integrated Velocity Structure Model (JIVSM; Koketsu *et al.* 2008, 2012) shows that the basin is constituted of sediments that have S -wave velocities ranging between 0.5 and 1.5 km s⁻¹. The sediment layers lie over a stiff bedrock of S -wave velocity inferred to be 3.2 km s⁻¹. The 500 m spaced contours of the basement are shown in Fig. 1. This complex and deep structure has a resonance period of 6–10 s (Kudo 1978, 1980; Furumura & Hayakawa 2007; Denolle *et al.* 2014b) and seismic

waves amplified by the basin are a potential threat to the large-scale urban structures, such as high-rise buildings or long-span bridges, of the Tokyo Metropolitan area.

The Tokyo Metropolitan area is extremely well instrumented. The dense Metropolitan Seismic Observation network (MeSO-net), which is composed of 296 accelerometer stations, was deployed in shallow 20 m deep boreholes (Kasahara *et al.* 2009; Sakai & Hirata 2009) to assess the seismic hazard in the region. There are also dozens of instruments of the High sensitivity seismograph network (Hi-net) of the National Research Institute for Earth Science and Disaster Resilience (NIED; Okada *et al.* 2004; Obara *et al.* 2005), University of Tokyo, and Japan Meteorological Agency (JMA) networks (Fig. 1). All stations record continuous waveforms that contain signals from several moderate M_w 4–5 earthquakes that occurred close to seismic stations (epicentral distances <10 km). The ambient seismic field recorded by the stations has been used in several seismic interferometry studies to infer the complexity of the wave propagation and seismic amplification in the basin (Denolle *et al.* 2014b; Boué *et al.* 2016; Viens *et al.* 2016a). Therefore, the combination of the large number of stations, the complexity of wave propagation and the numerous shallow earthquakes makes the area the ideal natural laboratory to investigate the recovery of reliable amplitudes from the ambient seismic field.

The goal of this study is to determine the best processing strategy to retrieve reliable IRF amplitudes from the ambient seismic field. After introducing the different processing techniques, we compare the waveforms of the vertical-to-vertical IRFs computed using one virtual source station that is located on the eastern edge of the basin. To validate the surface wave amplitude of the IRFs, we compare their long-period peak ground velocities (PGVs) with those of a shallow earthquake that occurred in the vicinity of the virtual source. Finally, we discuss the effect of seasonal variations on the amplitudes, the retrieval of realistic *S*-wave amplitudes, and the possibility of using the anticausal part of the IRFs using one virtual source located on the western edge of the basin.

2 METHODS: CROSS CORRELATION, COHERENCY, AND DECONVOLUTION

We use the ambient seismic field data recorded in October 2014 by the seismic stations located close to the surface of the Kanto Basin. After instrumental response correction, the acceleration records of the MeSO-net stations are integrated once in time to retrieve the corresponding velocity waveforms and the data set is divided into 1 hr time windows. Waveforms are bandpass filtered between 0.05 and 2 Hz (0.5–20 s) using a 4-pole, 2-pass Butterworth filter and down-sampled to 4 Hz to speed up the computation process. To reduce the bias that localized earthquakes might cause (Bensen *et al.* 2007), we remove windows that have spikes larger than 10 times their standard deviation. We finally compute the Fourier transform of each 1 hr non-overlapping time series after using a zero-padding of 10 times the length of the window to increase the resolution when performing the computation in the frequency domain.

For each pair of virtual source–receiver stations, we compute the IRFs using three different techniques. The first method is cross correlation and can be written as

$$C_{ZZ}(x_r, x_s, t) = \langle \mathfrak{F}^{-1} (\hat{v}_Z(x_r, \omega) \hat{v}_Z^*(x_s, \omega)) \rangle, \quad (1)$$

where $\hat{v}_Z(x_r, \omega)$ and $\hat{v}_Z^*(x_s, \omega)$ are the Fourier transformed time series recorded by the vertical component *Z* at the receiver station x_r and the virtual source x_s , respectively. The * symbol is the complex

conjugate and ω is the frequency. The inverse Fourier transform, denoted by $\mathfrak{F}^{-1}$, is computed to retrieve the cross-correlations in the time domain (*t*). We finally stack the cross correlations over 1 month, represented by the $\langle \rangle$ brackets, to increase the signal to noise ratio (i.e. equivalent to spatially averaging the effect of the ambient seismic field sources). For the cross correlation, we use both raw and 1-bit normalized ambient seismic field data. The 1-bit normalization method, which is widely used in passive seismology, only retains the sign of the data, setting the positive values to 1 and the negative ones to -1 (Campillo & Paul 2003; Shapiro & Campillo 2004; Bensen *et al.* 2007).

The second method is coherency. It is similar to pre-whitening in the frequency domain (Bensen *et al.* 2007), and consists of using the amplitude spectrum of the raw data from both stations as a denominator term when performing the cross correlation. This process can be summarized as

$$H_{ZZ}(x_r, x_s, t) = \left\langle \mathfrak{F}^{-1} \left(\frac{\hat{v}_Z(x_r, \omega) \hat{v}_Z^*(x_s, \omega)}{\{|\hat{v}_Z(x_s, \omega)|\} \{|\hat{v}_Z(x_r, \omega)|\}} \right) \right\rangle, \quad (2)$$

where $||$ represents the absolute value and $\{\}$ denotes a smoothing of the spectra using a moving average over 20 points to stabilize the denominator terms. All the other symbols are the same as in eq. (1).

The third method is deconvolution, where only the amplitude spectrum of the data recorded by the virtual source is used in the denominator, and can be written as

$$D_{ZZ}(x_r, x_s, t) = \left\langle \mathfrak{F}^{-1} \left(\frac{\hat{v}_Z(x_r, \omega) \hat{v}_Z^*(x_s, \omega)}{\{|\hat{v}_Z(x_s, \omega)|\}^2} \right) \right\rangle, \quad (3)$$

where all symbols are the same as in eqs (1) and (2).

The traveltime (or phase) information is preserved through all three techniques as the denominator term in Eqs (2) and (3) only affects the amplitude information. The regularization of the amplitude with the denominator term, however, prevents us from directly comparing the amplitudes from the three techniques without any scaling. In the following, we either normalize the amplitudes or calibrate them with earthquake records before comparing the waveforms.

3 EFFECTS OF THE PROCESSING ON THE IMPULSE RESPONSE FUNCTIONS

We use the techniques described in Section 2 to retrieve IRFs between the TENNOD station, which is the virtual source, and receivers in the basin. The virtual source station has been chosen because of its location close to the epicentre of a shallow M_w 5.8 earthquake. This event occurred on 14 March 2012 at a depth of approximately 11 km (F-net/NIED catalogue) and is used to validate the amplitude of the IRFs in Section 4. In the Kanto Basin, the primary source of the ambient seismic field is the Pacific Ocean. Because of the virtual source and receiver geometry with respect to the ambient seismic field sources, we expect a strong asymmetry between the causal and anticausal parts of the IRFs. As the signal of the anticausal part is weak (Supporting Information Fig. S1), we only focus on the causal part of the IRFs for the TENNOD virtual source.

Fig. 2 shows the IRF waveforms, which are normalized to their respective peak absolute value, computed between the TENNOD station and the E.SRTM and E.YMMM receivers (locations shown in Fig. 1). In the 1–10 s period range, several groups of waves are observed and can be explained by wave propagation in the Kanto Basin (Boué *et al.* 2016; Viens *et al.* 2016a). The earliest group of waves, which arrives at 15 and 25 s for the E.SRTM and E.YMMM

Figure 2. Impulse response functions retrieved from the raw cross correlation (red), 1-bit cross correlation (orange), coherency (green) and deconvolution (blue) techniques between the TENNOD virtual source and the (a) E.SRTM and (b) E.YMMM stations, which are located at 44 and 67 km from the virtual source, respectively. The location of these two stations is shown in Fig. 1. All the waveforms are shown for the Z-Z component and are bandpass filtered between 1 and 10 s. The name of each group of arrivals is also indicated.

stations, respectively, has been identified as *S* waves by Viens *et al.* (2016a) and is seen in all the waveforms. However, differences in the *S*-wave amplitudes can be observed among the IRFs, with smaller amplitudes retrieved with the deconvolution method. Nevertheless, the recovery of clear body waves from seismic interferometry without any additional computation to the one described in Section 2 remains unusual and has only been observed in a few studies (Roux *et al.* 2005; Poli *et al.* 2012; Boué *et al.* 2013).

The second and third wave packets in Fig. 2 are likely surface waves. The first group is well retrieved by all the methods, but its duration is shorter for the raw cross correlation IRFs. The second group is weaker for the waveforms obtained from raw and 1-bit cross correlations compared to the two other techniques. To explain the low amplitudes, we compute the Fourier spectra of the IRFs retrieved with the cross correlation and deconvolution techniques of raw data for 28 stations and show them in Fig. 3. These receivers are located within the azimuth range of $274^\circ \pm 4^\circ$, between 15 and 100 km from the virtual source. The IRFs obtained by cross correlation have a narrower frequency content (0.15 to 0.3 Hz), which corresponds to the frequency range of the secondary microseism peak, compared to that of the deconvolution (0.1 to 1 Hz). The discrepancy between the two techniques is expected from theoretical analysis (Vasconcelos & Snieder 2008) and is summarized as follows. The ambient seismic field recorded at one station is the convolution of its source excitation with the Green's function from the source to the receiver. For two stations recording the ambient seismic field generated by the same source, the cross correlation of their signals creates a strong dependence on the source excitation. On the other hand, the deconvolution (and coherency) method cancels out this dependence with the denominator term. In the 1–10 s period range, the secondary microseism sources are dominant, which explains the limited frequency content (0.15 to 0.3 Hz) of the cross correlation IRFs. The deconvolution and coherency methods cancel out the dependence to the secondary microseism sources and allow for the retrieval of the last group of surface waves that has a dominant period content between 1 and 3 s (Viens *et al.* 2016a).

To study the effect of the processing method on the amplitudes, we assume that the dominant signal in sedimentary basins comes

Figure 3. Fourier spectra of the impulse response functions extracted using the deconvolution (black) and the cross correlation (red) techniques between the TENNOD virtual source and 28 receiver stations. The thick lines represent the geometric mean of the 28 Fourier spectra. The waveforms are normalized with the peak value of the closest station to the virtual source (18 km) before being Fourier transformed.

from the surface waves. We bandpass filter the IRFs at the same stations as in Fig. 3 in a period range of 3–10 s. The two main reasons to choose a narrower bandpass filter are the following: first, the cross correlation technique can only retrieve clear signals at periods longer than 3 s; and second, Viens *et al.* (2016a) demonstrated that both phases and amplitudes of the Z-Z (vertical-to-vertical) IRFs from deconvolution are highly similar to the waveforms from the M_w 5.8 earthquake in the period range of 3–10 s. Then, we correct their relative amplitudes for the surface wave geometrical spreading by multiplying the waveforms by $\sqrt{x_s}$, where x_s is the virtual source–receiver distance. Finally, we compute the envelope of the signals using the Hilbert transform and taper out the waves travelling faster than 5 km s^{-1} that are probably non-physical. Fig. 4 shows the envelopes, which are normalized by a single calibration

Figure 4. Envelope of the Z-Z impulse response functions retrieved from the raw cross correlation (red), 1-bit cross correlation (orange), coherency (green) and deconvolution (black) techniques as a function of the distance from the virtual source. All the impulse response functions are normalized with the maximum value of the waveform at the closest station from the virtual source, corrected for the geometrical spreading of surface waves (multiplied by $\sqrt{x_s}$, where x_s is the virtual source–receiver distance), and bandpass filtered in the period range of 3–10 s.

factor taken as the peak amplitude at the closest station to the virtual source, against the distance from the TENNOD virtual source. Despite the geometrical spreading correction, the amplitudes of the first group of surface waves generally decay with the increasing distance, a likely signature of attenuation in Kanto Basin. However, local amplifications can be observed at some stations, as for example, at the two stations located between 60 and 70 km from the virtual source (Fig. 4). We can also note that the amplitude of *S* waves retrieved with cross correlation of raw and 1-bit data is higher than that of the two other techniques for most of the stations. Despite the surface wave geometrical spreading correction, the amplitude of *S* waves decreases faster and is weaker than that of the surface waves, which further shows that the long-period surface waves are dominant in the Kanto Basin.

4 VALIDATION AGAINST EARTHQUAKE RECORDS: ATTENUATION AND AMPLIFICATION

We showed in Section 3 that the dominant signal is the first group of surface waves. To validate the amplitude of this group of waves, we compare their long-period PGVs, which are computed for the waves travelling between 0.3 and 1.4 km s^{−1}, with those of the

M_w 5.8 earthquake. The long-period PGV of surface waves is called surface wave PGV in the following. As only the relative amplitude is preserved by seismic interferometry techniques, the IRFs must be calibrated against earthquake velocity records. To scale up the amplitude of the IRFs to that of the earthquake waveforms for a set of stations (see next paragraph), we first calculate the ratio of earthquake to IRF surface wave PGVs at each station. Then, we compute the mean of the ratios of the selected stations and correct the IRFs by this factor. This correction is the same for all the selected stations but is different for the cross correlation, coherency and deconvolution techniques. We also account for the difference between the epicentre and virtual source locations, which leads to a difference in surface wave geometrical spreading. We effectively shift the virtual source to the earthquake epicentre by correcting the amplitudes with the ratio of the geometrical spreading terms $\sqrt{x_s}/\sqrt{x}$, where x_s is the virtual source–receiver distance and x is the epicentre–receiver distance. This correction is relatively minor (mean value of 0.96) given the proximity of the virtual source to the real source epicentre (8.3 km), the period range of interest (3–10 s), and the fact that the virtual source and the earthquake are located outside of the Kanto Basin (higher seismic wave velocity and therefore longer wavelengths). In addition to the amplitude corrections, we apply a time shift to the IRFs to account for the fact that the earthquake epicentre and the virtual source station are not co-located. We first compute the average *S* wave velocity between the epicentre and the hypocentre using the JIVSM. The obtained theoretical wave velocity is 3.3 km s^{−1} and is assumed to be constant between the epicentre and the virtual source. For each station, we use this theoretical wave velocity and the difference between the virtual source–receiver and epicentre–receiver distances to time shift the IRF. The difference between the two distances is around 7 km for all the stations in the basin and leads to a time shift of approximately 2 s.

To reduce the azimuthal effects related to the non-uniform distribution of the ambient seismic field sources and to the surface wave radiation pattern at the earthquake source (Denolle *et al.* 2014a; Viens *et al.* 2016b), we only calibrate the IRFs within narrow azimuth ranges of 8°. Finally, we select the stations located at distances greater than 30 km from the epicentre to ensure a similar path effect for the earthquake and IRF waveforms. The two corrections applied to the amplitudes and the station selection allow us to compare the IRF surface wave PGVs with the observed, ground truth, earthquake PGVs.

To confirm that the surface wave PGV is a reliable metric to compare the amplitudes, we compare the earthquake waveforms to the amplitude calibrated and time shifted IRFs. We show the earthquake waveforms and the IRFs obtained from raw cross correlation, coherency and deconvolution for 3 stations located in the 264° ± 4° azimuth range from the epicentre in Fig. 5 (map and 3 additional stations are shown in Supporting Information Fig. S2). The waveforms are bandpass filtered between 3 and 10 s and are selected to compare the waveforms at different distances from the epicentre. The wave propagation through the Kanto Basin creates long and complex earthquake strong motions that are generally retrieved by the IRFs, even if some mismatches can be observed. The small phase shifts observed between the IRF and earthquake waveforms are probably related to the fact that we use a constant theoretical velocity to time shift the IRFs rather than frequency dependent values. Nevertheless, the small time delays observed between the IRF and earthquake waveforms do not influence our amplitude measurements in the following. Finally, note that the amplitudes of the IRFs from the three techniques at the E.KDKM station are very different.

Figure 5. Vertical component of the M_w 5.8 earthquake (blue) and Z-Z impulse response functions retrieved from the raw cross correlation (red), coherency (green) and deconvolution (black) techniques at three receiver stations. For each panel, the horizontal arrow starts and ends for waves travelling at 1.4 and 0.3 km s⁻¹, respectively, and corresponds to the range where the surface wave PGVs are selected.

For the stations located within the $264^\circ \pm 4^\circ$ azimuth range from the epicentre, we plot the natural logarithm of the surface wave PGVs as a function of the distance from the epicentre in Fig. 6. We also show the $1/\sqrt{x}$ and $1/\sqrt{x} \times e^{-\alpha x}$ theoretical curves, where x is the epicentre-receiver distance and α is an effective attenuation coefficient. We choose an arbitrary value of $\alpha = 14 \times 10^{-3} \text{ km}^{-1}$, which is close to the values found by Prieto *et al.* (2009) in the Los Angeles sedimentary basin (e.g. $\alpha = 2\text{--}5 \times 10^{-3} \text{ km}^{-1}$ in the 5–10 s period range). In Fig. 6, we can observe that the amplitude decay of the IRF and earthquake surface wave PGVs follows the $1/\sqrt{x} \times e^{-\alpha x}$ theoretical curve relatively well for distances smaller than 80 km and larger than 140 km from the epicentre. However, between 80 and 140 km, there is a large amplification for both predicted and observed values. This amplification correlates well with the deepest

part of basin (Koketsu *et al.* 2008, 2012), which is also shown in Fig. 6. Such amplification at large distance demonstrates that elastic 3-D effects of wave focusing are stronger than attenuation in the Kanto Basin.

To quantify the match between the observed and predicted amplitudes, we compute the root mean square error (RMS error) between the natural logarithm (ln) of the surface wave PGVs. The RMS error can be written as

$$\text{RMS error} = \sqrt{\frac{\sum_{n=1}^N (\ln(\text{PGV}_{\text{eq}}^n) - \ln(\text{PGV}_{\text{IRF}}^n))^2}{N}}, \quad (4)$$

where $\text{PGV}_{\text{IRF}}^n$ and PGV_{eq}^n are the IRF and earthquake surface wave PGVs, respectively. n is the n th station of a total number of stations N located within an azimuth range. For the 68 stations located

Figure 6. Upper panel: map of the Kanto Basin including the basin depth contours from the JIVSM and all the seismic stations. The receiver stations included in the $264^\circ \pm 4^\circ$ azimuth range, which is represented by the two black lines, are used in this figure. Middle panels: natural logarithm of the surface wave PGVs of the earthquake (red) and impulse response functions (IRFs) after amplitude calibration (blue) waveforms for the three different techniques. The black curves are the theoretical curves $1/\sqrt{x}$ and $1/\sqrt{x}e^{-\alpha x}$, where x is the epicentre–receiver distance and α is an attenuation coefficient taken as $14 \times 10^{-3} \text{ km}^{-1}$. The surface wave PGVs are computed from the waves bandpass filtered in the period range of 3–10 s and travelling between 0.3 and 1.4 km s^{-1} . For each panel, the RMS error computed using eq. (4) and the lower and upper bounds of the 95 per cent normal confidence interval are also shown. Bottom panels: velocity profile extracted from the JIVSM for an azimuth of 264° from the epicentre. The S -wave velocity of each layer is indicated and the thick line represents the basin depth.

within the $264^\circ \pm 4^\circ$ azimuth range, the RMS errors are equal to 0.244, 0.461 and 0.255 for the deconvolution, coherency and cross correlation, respectively. The minimum RMS error is obtained for the IRFs retrieved with the deconvolution method. The RMS error computed from the surface wave PGVs of the cross correlation technique is comparable to that from the deconvolution technique

and much smaller than the one from the coherency technique. The large RMS error for the coherency technique is likely to be related to the distribution of ambient seismic field sources the Pacific Ocean. Tsai (2011) theoretically demonstrated that the attenuation of waves retrieved with the coherency method for one-sided far-field surface wave sources is lower than the true attenuation value. Such a

Figure 7. (a) Mean of the RMS error computed from the surface wave PGVs of the earthquake and the impulse response functions retrieved with the raw cross correlation (red squares) and 1-bit cross correlation (orange squares) techniques as a function of the azimuth from the epicentre. For each azimuth, the stations located in the angle $\pm 4^\circ$ are used and the 95 per cent normal confidence interval of the bootstrap is indicated by the error bars. (b) Same as panel (a) for the RMS errors computed from the surface wave PGVs of the earthquake and the impulse response functions computed with the coherency (green) and deconvolution (black) methods.

Figure 8. Impulse response functions retrieved with the deconvolution method between the virtual source (TENNOD) and the (a) E.SRTM and (b) E.YMMM stations from the raw data recorded in April (black), June (blue), August (orange), October (red) and December (green) 2014. The waveforms are normalized with the peak amplitude of the impulse response retrieved for the month of April and the maximum value of each waveform is indicated between parentheses. All the waveforms are shown for the Z–Z component and are bandpass filtered between 3 and 10 s.

Table 1. Mean and standard deviation of the peak values of the waveform for the months of April, June, August, October, December 2014, normalized to peak value of the month of April 2014. For each month, we use the deconvolution method between the TENNOD virtual source and 272 receiver stations located within the 238° – 298° azimuth range from the epicentre.

Normalized peak value	April	June	August	October	December
Mean	1.00	1.04	1.05	1.05	1.11
Standard deviation	0.00	0.19	0.21	0.20	0.21

distribution of ambient seismic field sources is very similar to that in the Kanto Basin where a lower wave attenuation can be observed for the coherency technique (Fig. 6).

To evaluate the consistency of the surface wave PGVs in the RMS estimates, we bootstrap eq. (4) by randomly resampling

1000 times with replacement the original surface wave PGV values. We finally evaluate the 95 per cent normal confidence interval of the RMS errors using the bootstrapped data. The lower and upper bounds of the 95 per cent normal confidence interval are shown in Fig. 6. The bootstrap shows that the difference between

deconvolution and cross correlation is not supported by the data variance.

To study the azimuthal dependence of the RMS error, we compute the RMS error for stations located within 8° azimuth ranges, every 2° between 242° and 294° from the epicentre. These azimuth ranges were chosen because they contain at least 15 stations. Results are shown for raw and 1-bit cross correlations in Fig. 7(a) and for coherency and deconvolution in Fig. 7(b). The largest RMS error is obtained for the coherency technique at most of the azimuths and the smallest RMS errors are found for the raw cross correlation and deconvolution methods. There is an increase of the RMS error for the four techniques with increasing azimuth, with peak values at an azimuth of $284^\circ \pm 4^\circ$. The increase of RMS error might be caused by ambient seismic field sources being less efficient in that direction and causing a poorer recovery of the IRFs. This effect seems to be significant for some azimuths and should be further investigated. However, the shape of the array combined with the earthquake location prevents us from studying this effect over a broader azimuthal range.

To evaluate the bias that can be caused by the non-homogeneous number of stations for each azimuth, we use the bootstrap method previously described. As expected, fewer stations in a given azimuth range leads to a greater uncertainty in the RMS value (Fig. 7). Note that the difference between deconvolution and cross correlation is not significant for any of the azimuths.

5 DISCUSSION

5.1 Seasonal variations

The IRFs studied in the previous sections are computed using 1 month of data recorded in October 2014. The stability of the amplitude measurements throughout the year has been demonstrated in the Los Angeles sedimentary basin in the 4–10 s period range (Meier *et al.* 2010; Prieto *et al.* 2011). To verify that the amplitude measurements are also stable throughout the year in the Kanto Basin, we compute deconvolutions using eq. (3) for the months of April, June, August, and December 2014. The waveforms between the TENNOD virtual source and the E.SRTM and E.YMMM receivers are shown in Figs 8(a) and (b), respectively. For the two receivers, all the waveforms are normalized to the maximum value of the month of April. The shape of the waveform slightly changes over the year, but the different wave packets (*S* and surface waves) are reconstructed for each month. Moreover, the maximum amplitude is almost constant throughout the year and no clear seasonal pattern can be deduced for these two stations, thus supporting the stability of the amplitude measurements.

We compute the peak absolute values of the waveforms normalized with the month of April for the 272 stations located in the 238° – 298° azimuth range from the epicentre, and summarize their mean and standard deviation values in Table 1. The amplitudes are the lowest in April, but weakly vary with the seasons (less than 10 per cent) and the standard deviation is around 0.20 for each month. This shows that the peak value of the waveforms is almost stable throughout the year in the Kanto Basin.

To verify whether or not the small amplitude variations observed over the year impact the retrieval of earthquake surface wave PGVs, we calibrate the IRF amplitudes for each month following the same procedure as in Section 4. Fig. 9 shows the RMS errors computed using eq. (4) for each month against the azimuth from the epicentre. Small variations can be observed over the different months, but the

Figure 9. Mean of the RMS error computed from the surface wave PGVs of the earthquake and impulse response functions extracted with the deconvolution technique using the data recorded in April (black), June (blue), August (orange), October (red) and December (green) 2014 as a function of the azimuth from the epicentre. For each azimuth, the stations located in the angle $\pm 4^\circ$ are used.

general trend remains the same throughout the year. The month of October allows us to retrieve minimum RMS errors in the central part of the basin where strong and complex 3-D wave propagation effects are observed. One possible explanation for the minimum RMS error in October 2014 is that two typhoons passed over the Kanto Basin during this month and might have excited the ambient seismic field more efficiently. However, further work is needed to understand the potential effect of strong meteorological events on our results.

5.2 *S*-wave amplitudes

Within the $274^\circ \pm 4^\circ$ azimuth range, clear *S* waves can be retrieved. To investigate the reliability of their amplitudes, we perform the same analysis as done for the surface waves. We first bandpass filter the data between 4 and 10 s and select the phases travelling between 1.7 and 5 km s⁻¹. We calibrate the amplitude of the *S* waves following the same procedure as in Section 4, but with a ratio that accounts for the geometrical spreading of body waves (x_n/x_h , where x_n is the virtual source–receiver distance and x_h is the hypocentre–receiver distance). The long-period *S*-wave PGVs of the earthquake and IRF waveforms are shown in Fig. 10. For the coherency and deconvolution methods, the *S* wave PGVs agree well with those of the earthquake for distances larger than 60 km from the hypocentre. Similarly to surface waves, *S* waves are clearly amplified in the deepest part of the basin. The overestimation of the IRF *S*-wave PGVs for distances smaller than 60 km from the hypocentre might be explained by the fact that body waves travel directly from the hypocentre to the receivers through the basement of the Kanto Basin. Therefore, body waves from the surface-to-surface instruments likely undertake a different ray path than those from the deep source to the surface instruments. Nevertheless, reliable *S*-wave amplitudes can be extracted from the ambient seismic field at sufficient distance from the hypocentre using the deconvolution technique.

Figure 10. Upper panel: map of the Kanto Basin including the basin depth contours from the JIVSM and all the seismic stations. The receiver stations included in the $274^\circ \pm 4^\circ$ azimuth range, which is represented by the two black lines, are used in this figure. Middle panels: natural logarithm of the S -wave long-period PGVs of the earthquake (red) and impulse response functions (IRFs) after amplitude calibration (blue) waveforms for the three different techniques as a function of the distance from the hypocentre. The S -wave PGVs are computed from the waves bandpass filtered in the period range of 4–10 s and travelling faster than 3.5 km s^{-1} . For each panel, the RMS error computed using eq. (4) and the lower and upper bounds of the 95 per cent normal confidence interval are also shown. Bottom panels: velocity profile extracted from the JIVSM for an azimuth of 274° from the hypocentre. The S -wave velocity of each layer is indicated and the thick line represents the basin depth.

5.3 Can the anticausal part of the deconvolution be used?

So far, we have only studied the amplitude information of the causal part of the IRFs for seismic waves propagating westward, away from the coast. For earthquakes located to the West of the basin, however, seismic waves propagate towards the East and most of the energy of IRFs is contained in the anticausal part (Supporting Information Fig. S3). Such a configuration is, for example, similar to Southern California where earthquakes are expected along the San Andreas Fault and their seismic waves will radiate towards the coast where the Los Angeles Metropolitan area is located. Therefore, evaluating the

reliability of the amplitude information contained in the anticausal part of IRFs is critical to the generalization of the ground motion prediction using the ambient seismic field.

We retrieved the IRFs with the deconvolution method from the ambient seismic field recorded in October 2014 by regarding the E.ZKUM station as the virtual source. This station is located close to a M_w 4.7 earthquake (Fig. 1), which occurred on 2012 January 28 at a depth of 20 km (F-net/NIED catalogue). We calibrate the amplitude of the IRFs using the same procedure as in Section 4 and compute the surface wave PGVs for the waves propagating at

Figure 11. Upper panel: map of the Kanto Basin including the basin depth contours from the JIVSM and all the seismic stations. The receiver stations included in the $70^\circ \pm 4^\circ$ and $90^\circ \pm 4^\circ$ azimuth ranges, which are represented by the black lines, are used in this figure. Middle panels: natural logarithm of the surface wave PGVs of the surface waves of the $M_w 4.7$ earthquake (red) and that of the anticausal IRF from deconvolution after amplitude calibration (blue) waveforms as a function of the distance from the epicentre. For each panel, the RMS error computed using eq. (4) and the lower and upper bounds of the 95 per cent normal confidence interval are also shown. The surface wave PGVs are computed from the waves bandpass filtered in the period range of 3–10 s and travelling between 0.3 and 3.5 km s^{-1} . Bottom panels: velocity profile extracted from the JIVSM for the 70° and 90° azimuths from the epicentre. The S-wave velocity of each layer is indicated and the thick line represents the basin depth.

velocities between 0.3 and 3.5 km s^{-1} . Results are shown for two azimuths in Fig. 11. We choose the $70^\circ \pm 4^\circ$ azimuth range because of the large number of stations and the $90^\circ \pm 4^\circ$ azimuth range to sample the deepest part of the Kanto Basin. For these two azimuth ranges, the RMS errors (e.g. 0.409 and 0.458) are comparable to

those of the causal part (Section 4). Observed and simulated surface wave PGVs for the stations located above the deepest part of the basin, which reaches a maximum depth of 4.1 km, are clearly amplified compared to the stations located above a shallower basin depth of approximately 3 km. This demonstrates that the anticausal

part of the IRFs obtained from deconvolution also contains reliable amplitude information and can be used to predict earthquake ground motions in the Kanto Basin.

6 CONCLUSIONS

We retrieved IRFs from the ambient seismic field recorded in the Kanto Basin using three seismic interferometry techniques. We first showed that the IRFs constructed from the cross correlation technique have a narrower frequency content than the IRFs extracted with the coherency and deconvolution methods. To validate the surface wave amplitudes of the calibrated IRFs, we compared their long-period PGVs to those observed during a shallow M_w 5.8 earthquake that occurred nearby the TENNOD virtual source. For the $264^\circ \pm 4^\circ$ azimuth range, the surface wave PGVs of the waveforms retrieved with the cross correlation and deconvolution techniques agree well with those of the earthquake. Moreover, complex 3-D wave-propagation effects, likely caused by the basin structure, are well captured by the ambient seismic field. However, the agreement between the surface wave PGVs decreases for epicentre-receiver angles close to a direction parallel to the coast. This is likely to be related to the location and the strength of the ambient seismic field sources in the Kanto region that affect the quality of the IRFs. This feature needs to be further investigated to retrieve better IRFs.

We studied the sensitivity of the amplitudes to seasonal variations and showed that despite small variations, they remain relatively constant throughout the year for the deconvolution method. We also showed that the relative amplitudes of S waves are preserved for distances greater than 60 km from the hypocentre with the deconvolution and coherency techniques. Finally, we conducted the same analysis on the surface wave amplitude of the anticausal part of the IRFs from the deconvolution using an earthquake that occurred on the western part of the basin. The surface wave amplification caused by the deepest part of the Kanto Basin is also well retrieved.

This analysis demonstrates that IRFs from the cross correlation and deconvolution techniques reproduce relatively well the surface wave amplitudes of earthquakes in the Kanto Basin compared to the coherency technique. Moreover, IRFs from the deconvolution technique have a broader frequency content compared to those from the cross correlation technique. Finally, IRFs from deconvolution better retrieve S -wave amplitudes. Therefore, the deconvolution technique seems to be the most appropriate method to predict earthquake ground motions in the Kanto Basin from the ambient seismic field. In the Tokyo metropolitan area, this technique could be used to predict the long-period ground motions of large earthquakes that could occur in the near future along the Nankai Through and the Itoigawa-Shizuoka Tectonic Line.

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SUPPORTING INFORMATION

Supplementary data are available at [GJI](#) online.

Figure S1. Anticausal and causal parts of the IRFs retrieved with the (a) raw cross correlation, (b) 1-bit cross correlation, (c) coherency and (d) deconvolution methods as a function of the distance from the TENNOD virtual source. All the waveforms are bandpass filtered between 3 and 10 s and are normalized with the maximum value at the closest station from the virtual source. All the receivers are located in the $264^\circ \pm 4^\circ$ azimuth range from the virtual source station. The dashed and dotted lines represent the 1.4 and 0.3 km s⁻¹ moveouts, respectively.

Figure S2. Upper panel: map of the Kanto Basin including the basin depth contours from the JIVSM and the locations of the earthquake, the TENNOD virtual source and the six receivers used in Fig. 5 and in this figure. Lower three panels: vertical component of the $M_w 5.8$ earthquake (blue) and Z–Z IRFs retrieved from the raw cross correlation (red), coherency (green) and deconvolution (black) techniques at three receiver stations. The waveforms are bandpass filtered between 3 and 10 s. For each panel, the horizontal arrow starts and ends for waves travelling at 1.4 and 0.3 km s⁻¹, respectively, and corresponds to the range where the surface wave PGVs are selected. Note that the velocity scale of the bottom panel is different from the other panels.

Figure S3. Anticausal and causal parts of the IRFs retrieved with the deconvolution method for the stations located in the (a) $70^\circ \pm 4^\circ$ and (b) $90^\circ \pm 4^\circ$ azimuth ranges from the E.ZKUM virtual source. All the waveforms are bandpass filtered between 3 and 10 s and are normalized with the peak absolute value at the closest station from the E.ZKUM virtual source. The dashed and dotted lines represent the 3.5 and 0.3 km s⁻¹ moveouts, respectively.

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