A simple method to evaluate the seismic/infrasonic energy partitioning during an eruption using data contaminated by air-to-ground signals

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Research Article

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Abstract

A volcanic eruption transmits both seismic waves and infrasound signals. The seismo-acoustic power ratio is widely used to investigate the eruption behaviors and the source dynamics. It is often the case that seismic data during an eruption are significantly contaminated or even dominated by ground shaking due to infrasound (air-to-ground signals). To evaluate the contribution of infrasound-originated power in the seismic data, we need a response function of the seismic station to infrasound. It is rare to obtain a seismo-acoustic data-set containing only infrasound signals, though it is ideal for calculating the response function. This study proposes a simple way to calculate the response function using seismo-acoustic data containing infrasound and independent seismic waves. The method requires data recorded at a single station and mainly uses the cross-correlation function between the infrasound data and the Hilbert transform of the seismic data. It is tested with data recorded by a station at Kirishima volcano, Japan, of which response function has been constrained. It is shown that the method calculates a proper response function even when the seismic data contain more significant seismic power (or noise) than the air-to-ground signals. The proposed method will be useful in monitoring and understanding eruption behaviors using seismo-acoustic observations.

Keywords

Infrasound, Seismo-acoustic observation, Volcano, Ground-coupled air wave.

Introduction

A propagating acoustic wave in the atmosphere induces local ground oscillation [Ben-Menahem and Singh, 1981, Sabatier et al., 1986]. Therefore, it is recorded not only by infrasound sensors but also by seismic sensors. Such signals in seismometer data are called 'ground-coupled air waves' in recent papers in volcanology [e.g., Johnson and Malone, 2007; De Angelis et al., 2012; Fee et al., 2016; Smith et al., 2016; McKee et al., 2018; Matoza et al., 2019]. However, the term 'ground-coupled airwaves' was originally used to indicate the pressure oscillation generated by propagating seismic waves [Donn and Posmentier, 1964]. The bidirectional usage of the term has confused the discussion. Matoza and Fee [2014] distinguished air-ground coupling and ground-air coupling. In this manuscript, we call the former

in seismometer records as an air-to-ground signal and the latter in acoustic records as a ground-to-air signal to clarify the coupling direction.

The generation efficiency of the air-to-ground signal is $\sim 0.1 - 10 \mu\text{m/s/Pa}$ [Ichihara 2016; Novoselov et al., 2020], which is much larger than that of the ground-to-air signal ($\sim 0.0003 \text{ Pa}/(\mu \text{ m/s})$) [e.g., Kim et al., 2004; Watada et al., 2006; Ichihara et al., 2012; Kurokawa and Ichihara, 2020], the former is usually more significant in the seismo-acoustic observation during volcanic eruptions. The air-to-ground signals recorded by seismometers are used to investigate infrasound when few infrasound sensors are available [Johnson and Malone, 2007; Ichihara et al., 2012; De Angelis et al., 2012; McKee et al., 2018]. On the other hand, they can make a significant power in the seismometer records and disturb the analyses of seismic signals associated with eruptions [Nakamichi et al., 2013; Ichihara, 2016] and other surface phenomena like snow avalanches [Heck et al., 2019; Marchetti et al., 2020].

The Volcanic Acoustic-Seismic Ratio (VASR) that is the infrasonic power over the seismic power [Johnson and Aster, 2005], has been used to investigate the energy partitioning between seismic waves and infrasonic waves. The energy partitioning allows us to infer the change of eruptive behaviors [Sciotto et al., 2011; Johnson and Aster, 2005; Ichihara, 2016; Palacios et al., 2016]. When the same eruption mechanism is kept, the seismic and infrasonic eruption tremors tend to vary their amplitudes with the eruption parameter, like plume height, keeping their ratio constant [Ichihara, 2016; Fee et al., 2016; Haney et al., 2018; Sciotto et al., 2019]. In such situations, it is particularly important to distinguish whether the power ratio of seismic and infrasonic data represents the seismic-infrasonic energy partitioning or the ground response to the incident infrasound [Ichihara, 2016]. More essentially, we need to distinguish if the recorded signal is seismic or infrasonic waves. It is achieved by the propagation velocity when seismic arrays are available [Nakamichi et al., 2013, Heck et al., 2019; Marchetti et al., 2020].

Kurokawa and Ichihara [2020] used known seismic waves and acoustic waves recorded by an infrasound sensor and a collocated seismometer as references. They calculated the spectral power ratios between them to evaluate the ground-to-air coupling and air-to-ground coupling efficiency. They used airplane sound as the known acoustic signal, though it has frequency components and incident angles much higher than those of volcanic signals. On the other hand, Ichihara [2016] obtained the ground response to infrasound as a function of frequency at a station in Kirishima volcano, Japan (Fig. 1). This case is unique in that a good infrasound source is available: Sakurajima, about 42 km away (Fig. 1c), frequently transmits explosion infrasound (Fig. 1a). Because of the distance, seismic waves from the source do not

reach the stations at Kirishima, or if they do, they are well separated in time from the infrasound signals. The response function, consisting of the amplitude ratio (Fig. 1d) and the phase shift of the vertical ground velocity to the pressure data (Fig. 1e), was obtained using the beginning 10 s containing the strong pulse (including 3 s before the onset), and those from 15 events were stacked [Ichihara, 2016].

In general, such an ideal infrasound source is rare in the field. Although the previous study assumed the reference data, both seismic and infrasonic, include only infrasound signals, the seismic data may always contain seismic waves when volcanoes are active. This paper presents a convenient method to evaluate the response function and examine the contribution of the air-to-ground signals in seismometer records using data containing both seismic and infrasonic signals. The method requires only an infrasound sensor and a collocated seismometer without significant wind noise. If there is an additional infrasound station, we can also correct for the wind noise effect.

Basic Theory

We denote vertical ground velocity recorded by a seismometer as d_v and pressure change recorded by an infrasound sensor as d_p . For simplicity, we assume the incident waves are dominated by a single seismic wave (v_s) and a single infrasonic wave (p_a). We assume that d_v consists of v_s , an air-to-ground signal (v_a) generated by p_a , and wind-induced ground oscillation (v_w), while d_p consists of p_a and wind noise (p_w). The contribution of the ground-to-air signals is assumed negligible, as mentioned above. These assumptions are represented by

$$d_v = v_s + v_a + v_w, \quad d_p = p_a + p_w. \quad (1)$$

A cross-correlation function between two time series, d_1 and d_2 , in a given time window $[t, t + T]$ is represented by

$$CC(d_1, d_2; t; \tau) \equiv \frac{1}{T} \int_t^{t+T} d_1(t') d_2(t' - \tau) dt', \quad (2)$$

where τ is the time delay of d_1 to d_2 . The corresponding cross correlation coefficient $R(d_1, d_2; t; \tau)$ is

$$R(d_1, d_2; t; \tau) \equiv \frac{CC(d_1, d_2; t; \tau)}{\sqrt{E(d_1; t)} \sqrt{E(d_2; t)}}, \quad (3)$$

where $E(d; t)$ represents the power of a time series d in the time window $[t, t + T]$, namely,

$$E(d; t) \equiv CC(d, d; t; 0). \quad (4)$$

Hereafter, we omit t that specifies the time window.

Among v_s , v_a , v_w , p_a , and p_w in equation (1), we assume no pair except v_a and p_a has a correlation. Although v_s and p_a may have a correlation if their source is common, their correlation should be found with a time shift much larger than that of v_a to p_a , considering the velocity difference between the seismic waves and infrasound [Ichihara et al., 2012]. The correlation between v_w and p_w is small if the distance between the infrasound sensor and the seismometer is larger than the correlation length of wind noise. For example, the distance of 5 m is enough to suppress the wind-noise correlation at frequency > 1 Hz in ordinary wind conditions [Shields, 2005, Ichihara et al., 2012]. Therefore,

$$R(d_v, d_p; \tau) = \frac{CC(v_a, p_a; \tau)}{\sqrt{E(d_v)}\sqrt{E(d_p)}}, \quad (5)$$

$$E(d_v) = E(v_s) + E(v_a) + E(v_w), \quad (6)$$

$$E(d_p) = E(p_a) + E(p_w) = (1 + W_p)E(p_a), \quad (7)$$

where W_p represent the power ratio of the wind noise to the waves. Under the same wind condition, the infrasound data are usually more significantly affected by wind noise. Therefore, we neglect $E(v_w)$ compared with $E(v_s) + E(v_a)$ in the following discussion.

It is noted that $R(d_v, d_p; \tau)$, $E(d_v)$, and $E(d_p)$ are calculated from the observed data. Our aim here is to estimate $E(v_a)/E(p_a)$ to obtain the seismo-acoustic power ratio $E(v_s)/E(p_a)$ from the observed data.

It is assumed that p_a is propagating along the ground surface in the direction of the x -axis. The component of angular frequency (ω) of p_a is

$$p_{a\omega} = P_\omega \exp \left[i\omega \left(t - \frac{x}{\alpha} \right) \right], \quad (8)$$

where α is the sound speed in the atmosphere, and ω is the angular frequency. When the ground is a homogeneous elastic half space, and its seismic speeds are much larger than α , the vertical ground velocity ($v_{a\omega}$) induced by $p_{a\omega}$, is

$$v_{a\omega} = H_{ps} e^{-\frac{i\pi}{2}} P_\omega \exp \left[i\omega \left(t - \frac{x}{\alpha} \right) \right], \quad H_{ps} \equiv \frac{\alpha}{2(\lambda + \mu)} \left(\frac{\lambda + 2\mu}{\mu} \right), \quad (9)$$

where (λ, μ) are the Lamé's constants of the ground [Ben-Menahem and Singh, 1981].

Equation (9) indicates that $v_{a\omega}$ has a phase shift of $-\pi/2$ with respect to $p_{a\omega}$, and this phase shift does not depend on the frequency. The actual ground is not homogeneous or flat, and the incident infrasound wave is not necessarily along the ground surface. Nevertheless, the phase shift of $-\pi/2$ is observed in many places [Ichihara et al., 2012; Ichihara, 2016]. Although the sensor distance can cause a time shift of correlation, we can correct it [McKee et al., 2018]. On the other hand, the phase delay

significantly deviates from the theoretical value and fluctuates at high frequency (Fig. 1e), possibly due to the ground structure [Ichihara, 2016]. Besides, H_{ps} depends on the frequency, especially in the low frequency (Fig. 1d). Therefore, it is difficult to estimate the response function theoretically. Here we derive an equation relating H_{ps} and the observed data, assuming the phase shift is $-\pi/2$. The relation is tested with data in the following sections.

First, the phase of d_v is shifted by $\pi/2$, using the Hilbert transform, $h_i(d_v)$, as

$$d_v^h = -\text{Imag}[h_i(d_v)], \quad (10)$$

where $\text{Imag}[\cdot]$ takes the imaginary part. The phase-shifted time series are denoted with a subscript h , namely,

$$d_v^h = v_s^h + v_a^h + v_w^h. \quad (11)$$

Second, we calculate $R(d_v^h, d_p; \tau)$, and search its maximum around $\tau = 0$. We denote τ that gives the maximum as τ_{max} . It should be consistent with the propagation time of infrasound between the infrasound sensor and the seismometer. When any signals in d_v , other than v_a , correlate with p_a , $CC(d_v^h, d_p; \tau)$ tends to have a maximum significantly deviated from the expected time delay [Ichihara et al., 2012]. Such data sets are not adequate to use in the current method.

Third, we approximate

$$CC(d_v^h, d_p; \tau_{max}) = CC(v_a^h, p_a; \tau_{max}) = H_{ps}E(p_a) \sim H_{ps}E(d_p). \quad (12)$$

The last approximation holds with negligible wind noise. We use equations (12) and (5) to obtain

$$H_{ps} \sim \frac{CC(d_v^h, d_p; \tau_{max})}{E(d_p)} = R(d_v^h, d_p; \tau_{max}) \sqrt{\frac{E(d_v^h)}{E(d_p)}}. \quad (13)$$

This equation estimates H_{ps} using only the observed data.

When the wind noise is significant, we need another infrasound station close by to evaluate the relation between $E(d_v)$ and $E(p_a)$. The calculation with wind noise is given in Additional File 1. Because H_{ps} is the site property of the station, it should not change significantly with time. Therefore, we evaluate it using equation (13) as an average of the adequately selected time windows.

Considering $E(v_a) = H_{ps}^2 E(p_a)$ and $E(d_p) \sim E(p_a)$, we obtain

$$\frac{E(d_v)}{E(d_p)} \sim \frac{E(v_s)}{E(p_a)} + H_{ps}^2. \quad (14)$$

It is noted that this equation is applicable on the condition that an infrasound signal p_a exists. Equation (14) indicates that we should refer to H_{ps}^2 for interpreting the observed power ratio, $E(d_v)/E(d_p)$. The

seismic power $E(v_s)$ increases the power ratio above H_{ps}^2 . Equation (14) is particularly useful when the air-to-ground signal is comparable with the seismic power. We can also identify the condition of $E(v_s) \ll H_{ps}^2$, in which the observed $E(d_v)/E(d_p)$ cannot provide the meaningful seismic-acoustic energy partitioning.

With wind noise, equation (14) becomes

$$\frac{E(d_v)}{E(d_p)} = \frac{1}{1 + W_p} \left[\frac{E(v_s)}{E(p_a)} + H_{ps}^2 \right], \quad (15)$$

(Additional File 1). When $E(d_v)/E(d_p) < H_{ps}^2$, it indicates that $E(d_p)$ is significantly contaminated by the wind noise power, and thus $E(d_v)/E(d_p)$ is not useful, either.

Method to estimate H_{ps} from data

We use the data presented in Fig. 1a and 1b as d_p and d_v , respectively. They are infrasound from Sakurajima recorded by an infrasound sensor (Hakusan, SI102, 0.1-1000 Hz) and a seismometer (Nanometrics, Trillium 120P) at Kirishima, respectively. The sampling rate is 100 Hz for both data. Ichihara [2016] has shown that the part of d_v containing the air-to-ground signal is well reproduced by d_p with the response function at 1-7 Hz shown by the red dashed line in Fig. 1d and 1e.

As the infrasound signal is larger than the background wind noise (Fig. 1a), we assume $W_p = 0$ in the following analysis. On the other hand, there is noticeable oscillation before the arrival of infrasound in the seismograph (Fig. 1b), indicating that the background seismic signal is comparable with the air-to-ground signal. Specifying the origin of the background signal is out of the scope of this study. Here we treat all signals other than v_a in d_v as v_s .

We use the four frequency bands: 1-3.5 Hz, 3.5-7 Hz, 7-12 Hz, and 12-18 Hz. For the selection of the frequency bands, we refer to the feature of the known response function shown in Fig. 1d and 1e. The phase shift of d_v to d_p is constant around $-\pi/2$ at 1-7 Hz, while the amplitude ratio H_{ps} is relatively stable in 7-18 Hz. In general cases without known response functions, we may arbitrarily select the frequency bands.

We apply a zero-phase-shift band-pass filter to d_p and d_v and calculate the cross-correlation coefficient, $R(d_v, d_p; \tau)$, in the individual bands. Figure 2 shows the results, in which the time origin is taken at the onset of the infrasound signal. In each of Fig. 2a-2d, the lowest panel shows $R(d_v, d_p; \tau)$, calculated in a 2.56-second-long time window sliding every 0.5 s. We can clearly see the typical correlation pattern of

air-to-ground signals, that is a positive peak in $\tau > 0$ and a negative peak in $\tau < 0$ with a node at $\tau = 0$ [Ichihara et al., 2012], continuously after the arrival of the infrasound. It confirms that d_p includes p_a in the period in all the frequency bands.

We use the band-pass-filtered data in 120 s after time zero as d_p and d_v . First, we shifted the phase of d_v forward by $\pi/2$ to obtain d_v^h . Second, we calculate $R(d_v^h, d_p; \tau)$. Third, we take the maximum of $R(d_v^h, d_p; \tau)$, denoted as R_{max} , in $\tau = \pm 0.01$. This span, corresponding to one data point, is allowed, considering the time delay between the two sensors separated by ~ 5 m and possible phase delay due to noise. Fourth, we calculate $E(d_p)$ and $E(d_v^h)$ by the mean square of the data. Finally, we calculate H_{ps} , using equation (13).

Result and discussion

Figure 3a compares the value of H_{ps} obtained by the current method and by Ichihara [2016]. They are consistent. It is noted that the previous value was obtained using the beginning 10 s of 15 events, while the current value was obtained from the 120-second-long data of a single event. Figure 3b shows the corresponding $R(d_v^h, d_p; \tau)$ for each band. They have the maximum values within $\tau \pm 0.01$ s, confirming that the phase shift of $-\pi/2$ was successfully corrected.

The data used in this analysis are ideal to evaluate H_{ps} because d_p and d_v mainly include p_a and v_a , respectively. We do not always have such data sets. We evaluate the effect of v_s with the assumption that v_s is independent of p_a . We artificially add v_s made in the following way. We generate a random function that is centered at zero and has the same length as d_v . We apply the same band-pass filter to this function and call it as d_{bg} . We normalize d_{bg} to have the same energy as d_v , by multiplying it by $\sqrt{E(d_v)/E(d_{bg})}$. Then, we amplify it by an arbitrary factor Γ to make a hypothetical v_s . Namely,

$$v_s = \Gamma d_{bg} \sqrt{\frac{E(d_v)}{E(d_{bg})}}. \quad (16)$$

Assuming the original d_v purely consists of v_a , that is $E(d_v) = E(v_a)$, Γ is regarded as $\sqrt{E(v_s)/E(v_a)}$. The actual $\sqrt{E(v_s)/E(v_a)}$ should be larger than Γ because d_v contains some background signal other than v_a . We may see it by comparing the amplitudes before and after the infrasound arrival in d_v and d_p (Fig. 2). It is also supported by Fig. 3b that $R_{max} < 1$.

Including v_s into d_v , we follow the same procedure as the previous paragraph to calculate H_{ps} . For each frequency and each Γ , we tested with 100 random functions. The mean and the standard deviation

of H_{ps} , τ_{max} , and R_{max} are presented as functions of $\sqrt{E(v_s)/E(v_a)}$ in Fig. 4a, 4c, and 4d, respectively. We also show the relation of the individual H_{ps} and R_{max} in Fig. 4b. As the background signal power, $E(v_s)$, increases, the cross-correlation coefficient decreases (Fig. 4d). On the other hand, H_{ps} stays around the expected value (Fig. 4a and 4b). The average values are almost the same as those without artificial noise, H_{ps}^0 (the closed circles on the vertical axis in Fig. 4a).

The same test is performed using actual seismic data recording volcanic tremors. Shinmoe-dake, an active cone of Kirishima volcano, had sub-plinian eruptions in 2011 [Nakada et al., 2013], three weeks after the signal in Fig. 1 was recorded. A volcanic tremor continued there from 18 January 2011 to the first sub-plinian eruption on 26 January 2011. Two phreatic eruptions occurred at 1:27 a.m. (JST) on 19 January and 7:31 a.m. (JST) on 26 January 2011. The tremor always had power in the frequency range of 3.5-7 Hz [Ichihara and Matsumoto, 2017]. We randomly selected 100 windows with a length of 120 s between the two phreatic eruptions, apply the band-pass filter at 3.5-7 Hz, and use it for d_{bg} in equation (16). Figure 5a, 5c, and 5d showed the result in the format to be compared with the green lines in Fig. 4a, 4c, and 4d. The normalized spectra of the 100 signals (d_{bg}) are shown in Fig. 5b. It is aligned in the increasing order of $(H_{ps} - H_{ps}^0)/H_{ps}^0$ for $\Gamma = 10$, which is shown in the upper panel of (b). We could not identify the spectral features that deviate H_{ps} from H_{ps}^0 .

These results support that we can evaluate H_{ps} properly, even if the seismic data are significantly contaminated by the background signal (seismic waves or other noise). When we have persistent infrasound and can average a long time window or many windows, we may estimate H_{ps} more accurately.

Conclusions

We developed equation (13) that relates the ground response to infrasound (H_{ps}) and the observed seismo-acoustic data (d_v and d_p). We have confirmed that it provides a proper value of H_{ps} , even when the observed seismic data contain significant seismic signals as well as the air-to-ground signals. Constraining H_{ps} is essential for interpreting the observed seismic-acoustic power ratio to the energy partitioning between seismic waves and infrasound at the source.

Availability of data and materials

The datasets used and/or analysed during the current study are available from the corresponding author on reasonable request.

Competing interests

The authors declare that they have no competing interests.

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Authors' contributions

MI developed the methods, made the analyses, and drafted the manuscript. KY confirmed the equations. DM raised the needs in volcanology to motivate this study. All authors read and approved the final manuscript.

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Figure Captions

Figure 1

Infrasound from an explosion, at 22:38 (hh:mm JST) on 13 January 2011, of Sakurajima volcano recorded by an infrasound sensor (a) and a broad-band seismometer (b) at the SMN station near Kirishima volcano, 42 km to the north of Sakurajima. (c) The locations of Sakurajima and Kirishima, southern Kyushu, Japan. The amplitude (d) and phase (e) response of the ground velocity to incident infrasound at the SMN station. The red dots represent the response function at 1-7 Hz used by Ichihara [2016]. All the plots are reproduced from Ichihara [2016], but the phase delay in (e) has been converted from the range of $[-\pi, \pi]$ to $[-2\pi, 0]$.

Figure 2

The cross-correlation analyses of the data in Fig. 1 at (a) 1-3.5 Hz, (b) 3.5-7 Hz, (c) 7-12 Hz, and (d) 12-18 Hz. In each panel, the top and middle panels show the band-passed infrasonic data (d_p) and seismic data (d_v), respectively. The bottom panel shows the cross-correlation coefficient between the seismic data, d_v , and infrasonic data, d_p , with the vertical axis representing the time delay, τ , of d_v to d_p . The horizontal dashed line indicates $\tau = 0$.

Figure 3

(a) The amplitude response of the ground velocity to infrasound, H_{ps} , obtained by this study is shown for 1-3.5 Hz (red line), 3.5-7 Hz (green line), 7-12 Hz (blue dots), and 12-18 Hz (purple dots). The results are compared with the response function obtained by Ichihara [2016] (Fig. 1d and 1e). (b) The corresponding cross-correlation coefficient between the phase-shifted seismic data, d_v^h , and the infrasonic data, d_p . The colors indicate the frequency bands as in (a).

Figure 4

The effect of background seismic signals, v_s , in the seismic data, d_v , on the estimation of H_{ps} . The colors indicate the frequency bands: 1-3.5 Hz (red), 3.5-7 Hz (green), 7-12 Hz (blue), and 12-18 Hz (purple). A hundred random functions are band-pass filtered and used as v_s . The horizontal axis is the square-root of power ratio of v_s to the original d_v that is approximated as v_a . It is noted that the actual power ratio $E(v_s)/E(v_a)$ is larger than $E(v_s)/E(d_v)$ as d_v itself contains a significant power of background signals. The lines and error bars show the means and the standard deviations of H_{ps} (a), the maximum correlation coefficient, R_{max} (d), and the time delay, τ_{max} , that gives R_{max} (d) for the hundred random functions. The closed circles on the vertical axes are the values without v_s . The

relation between the individual R_{max} and H_{ps} for all the cases are shown with circles in (b).

Figure 5

A similar background-signal test as in Fig. 4, but volcanic tremors recorded by the same seismic station are used as v_s . We performed the analysis only at 3.5-7 Hz. We randomly selected 100-time windows between 1:27 on 19 January 2011 and 7:31 on 26 January 2011, of which the normalized power-spectral densities of the band-pass-filtered data are displayed in the lower panel of (b). The horizontal axis shows the window number of v_s aligned in the increasing order the deviation of H_{ps} for the case $\sqrt{E(v_s)/E(v_a)} = 10$. The upper frame shows the deviation: $\Delta H_{ps} \equiv (H_{ps} - H_{ps}^0)/H_{ps}^0$, where H_{ps}^0 is the value obtained without v_s , which is the closed circle on the vertical axis in (a). In (a), (c), and (d), the lines and error bars show the means and the standard deviations of H_{ps} , τ_{max} , and R_{max} , respectively.

Figures

Figure 1

Infrasound from an explosion, at 22:38 (hh:mm JST) on 13 January 2011, of Sakurajima volcano recorded by an infrasound sensor (a) and a broad-band seismometer (b) at the SMN station near Kirishima volcano, 42 km to the north of Sakurajima. (c) The locations of Sakurajima and Kirishima, southern Kyushu, Japan. The amplitude (d) and phase (e) response of the ground velocity to incident

infrasound at the SMN station. The red dots represent the response function at 1-7 Hz used by Ichihara [2016]. All the plots are reproduced from Ichihara [2016], but the phase delay in (e) has been converted from the range of $[-\pi, \pi]$ to $[-2\pi, 0]$.

Figure 2

The cross-correlation analyses of the data in Fig. 1 at (a) 1-3.5 Hz, (b) 3.5-7 Hz, (c) 7-12 Hz, and (d) 12-18 Hz. In each panel, the top and middle panels show the band-passed infrasonic data (dp) and seismic data (dv), respectively. The bottom panel shows the cross-correlation coefficient between the seismic data, dv, and infrasonic data, dp, with the vertical axis representing the time delay, τ , of dv to dp. The horizontal dashed line indicates $\tau = 0$.

Figure 3

(a) The amplitude response of the ground velocity to infrasound, H_{ps} , obtained by this study is shown 314 for 1-3.5 Hz (red line), 3.5-7 Hz (green line), 7-12 Hz (blue dots), and 12-18 Hz (purple dots). The 315 results are compared with the response function obtained by Ichihara [2016] (Fig. 1d and 1e). (b) The corresponding cross-correlation coefficient between the phase-shifted seismic data, d_{hv} 316, and the infrasonic 317 data, d_p . The colors indicate the frequency bands as in (a).

Figure 4

The effect of background seismic signals, vs, in the seismic data, dv, on the estimation of Hps. The colors indicate the frequency bands: 1-3.5 Hz (red), 3.5-7 Hz (green), 7-12 Hz (blue), and 12-18 Hz (purple). A hundred random functions are band-pass filtered and used as vs. The horizontal axis is the square-root of power ratio of vs to the original dv that is approximated as va. It is noted that the actual power ratio $E(v_s)/E(v_a)$ is larger than $E(v_s)/E(dv)$ as dv itself contains a significant power of background signals. The lines and error bars show the means and the standard deviations of Hps (a), the maximum correlation coefficient, Rmax (d), and the time delay, $\tau_{\max}$, that gives Rmax (d) for the hundred random functions. The closed circles on the vertical axes are the values without vs. The relation between 327 the individual Rmax and Hps for all the cases are shown with circles in (b).

Figure 5

A similar background-signal test as in Fig. 4, but volcanic tremors recorded by the same seismic station are used as vs. We performed the analysis only at 3.5-7 Hz. We randomly selected 100-time windows between 1:27 on 19 January 2011 and 7:31 on 26 January 2011, of which the normalized power-spectral densities of the band-pass-filtered data are displayed in the lower panel of (b). The horizontal axis shows the window number of vs aligned in the increasing order the deviation of H_{ps} for the case $\sqrt{E(v_s)/E(v_a)} = 10$. The upper frame shows the deviation: $\Delta H_{ps} \equiv (H_{ps} - H_{0ps})/H_{0ps}$, where H_{0ps} is the value obtained without vs, which is the closed circle on the vertical axis in (a). In (a), (c), and (d), the lines and error bars show the means and the standard deviations of H_{ps} , τ_{max} , and R_{max} , respectively.

Supplementary Files

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