Identification of infrasonic and seismic components of tremors in single-station records: application to the 2013 and 2018 events at an isolated volcanic island, Ioto, Japan

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Abstract

Infrasonic stations are sparse at many volcanoes, especially those on remote islands and those with less frequent eruptions. When only a single infrasound station is available, the seismic-infrasonic cross-correlation method has been used to extract infrasound from wind noise. However, it does not work with intense seismicity and sometimes mistakes ground-to-atmosphere signals as infrasound. This paper proposes a complementary method to identify the seismic component and the infrasonic component using a single microphone and a seismometer. We applied the method to estimate the surface activity on the isolated volcanic island, Ioto. We focused on volcanic tremors during the phreatic eruption on April 11, 2013, and during an unconfirmed event on September 12, 2018. We used the spectral amplitude ratios of the vertical ground motion to the pressure oscillation and compared those for the tremors with those for known signals generated by volcano-tectonic earthquakes and airplanes flying over the station. We were able to identify the infrasound component in the part of the seismic tremor with the 2013 eruption. On the other hand, the tremor with the unconfirmed 2018 event was accompanied by no apparent infrasound. We interpreted the results that the infrasound with the 2013 event was excited by the vent opening or the ejection of ballistic rocks, and the 2018 event was not an explosive eruption either on the ground or in the shallow water. If there was any gas (and ash) emission, it might have occurred gently undersea. As the method uses the relative values of on-site records instead of the absolute values, it is available even if the instrument sensitivity and the station site effects are poorly calibrated.

Keywords

Volcanic tremor, Infrasound, Vertical ground motion, Spectral ratio, Ioto

Introduction

Eruption in isolated volcanic islands are becoming the focus of attention for their significant growth, as in the case of Nishinoshima (Maeno et al. 2016; Kaneko et. al 2019), and for the hazardous nature, as in Anak Krakatau (Williams et al. 2019; Perttu et al. 2020) and White Island more recently (Global Volcanism Program 2020). The existence of abundant water tends to cause hazardous eruptions like
phreatomagmatic and phreatic explosions (Mastin and Witter 2000; Stix and Moor 2018). It is often the case in isolated islands, the occurrences, the times, and the sequences of eruptions are not identified due to the lack of observations. The detection is particularly hard for small but frequent eruptions because signals are not strong enough to reach the global monitoring network.

Infrasound is generated by activity such as opening vent and emission of volcanic gas and rocks so that it is useful to distinguish the volcano’s surface activity from underground processes (e.g. Ripepe et al. 2018). When infrasound data during an eruption is available only from a single station, it is difficult to distinguish the eruption signals from wind noise. To detect infrasound signals, Ichihara et al. (2012) proposed a cross-correlation analysis between the pressure oscillation and ground motion signals, which has been applied successfully with some improvements (e.g. Cannata et al. 2013; Matoza and Fee 2014; Nishida and Ichihara 2015; Ichihara 2016; Yukutake et al. 2018). However, the method is not applicable when the volcano is seismically very active, and the seismic signal dominates infrasound signal in the seismometer record. Moreover, if the ground velocity associated with the seismic wave is significantly large, it generates pressure perturbation that is noticeable in the infrasound data (Kim 2004; Watada et al. 2006). Such a ground-to-atmosphere signal can be mistaken as an infrasound signal when only a single infrasound station exists.

Ioto is an isolated volcanic island of which seismicity is regularly intense (Ueda et al. 2018). At Ioto, phreatic eruptions frequently occur due to the high geothermal activity (Notsu et al. 2005), and the volcanic activity is pronounced not only on the ground but also undersea detected by remote hydrophones (Matsumoto et al. 2019). In this situation where the volcanic activity is very high throughout the island, there is a need to monitor eruptions and its temporal changes.

This study aims to identify volcanic infrasound using a single pair of seismometer and microphone at Ioto. By comparing the data of the tremors associated with the 2013 eruption and the unconfirmed 2018 event with those of volcano-tectonic earthquakes and human-made infrasound, we distinguish tremors including infrasound and purely seismic tremors.

Volcanic Activities at Ioto in 2013 and 2018

Ioto (Iwo-jima) is one of the most active isolated volcanic islands in Japan, located approximately 1200 km south of Tokyo and belongs to the Izu-Bonin-Mariana island arc. The island, about 8 km × 4 km in size with the highest elevation of 170 m, is just the summit part of a stratovolcano rising about 2000 m
from the sea floor. Eruptions sometimes occur under the sea and are detected by remote hydrophones (Matsumoto et al. 2019). The seismic activity is intense and a large-scale uplift has continued for centuries (Kaizuka et al. 1985; Ueda et al. 2018). Although minor phreatic explosions seem to occur frequently at various points in the island (Corwin and Foster 1959; Notsu et al. 2005; Ueda et al. 2018), most of them have not been confirmed in terms of their occurrences, times, and source vents.

An eruption occurred at about 16:00 on April 11, 2013, has been observed from the ground and the sky and recorded by time-identified photographs (Japan Meteorological Agency 2013). It occurred at Million Dollar Hole (Figure 1) with dark smoke of 400 m height and large ballistic rocks while the seismic activity and crustal movement were less intensive. It accompanied a volcanic tremor lasting about 9 min from 15:59 on April 11.

Another activity seems to have occurred in shallow water near the Ogigahama coast (Figure 1) on September 12, 2018, though the only evidence is water spouts with heights of 5-10 m observed at around 11:00 on the day (Japan Meteorological Agency 2018). It was preceded by predominant uplift and high-frequency seismicity for a few weeks. Data from local seismometers and remote hydrophones during the period indicates that frequent undersea eruptions occurred and induced the earthquakes (Matsumoto et al. 2019). The seismic activity shifted to low-frequency, and volcanic tremors increased from September 12 to 13. Then, the volcanic activity gradually declined. We investigate the 2018 activity in comparison with the confirmed case of the 2013 eruption.

Data and Methods

Data
We used the records at the three seismic stations in Ioto (Figure 1). IOCD station of the Japan Meteorological Agency (JMA) is equipped with a velocity seismometer (L-4C, 1 Hz, Sercel Inc.) and an infrasonic microphone (TYPE7144, > 0.1 Hz, Aco Co., Ltd.) having a horizontal separation of 7.6 m and a vertical difference of 1.5 m. IJSV and IJTV stations operated by the National Research Institute for Earth Science and Disaster Resilience (NIED) have velocity seismometers (J21-3D, 1 Hz, Mitsutoyo Corporation). The sampling frequency of all the instruments is 100 Hz. Note that the IOCD is the only infrasonic station in Ioto and the nearest seismic station to the vent of 2013 eruption and the location where the water spouts were observed in the 2018 activity. The distances of IOCD to these possible sources are 1.03 and 0.94 km, respectively.
This study focuses on two tremor events (Figure 2); TR1 on April 11, 2013, and TR2 on September 12, 2018 (TR2 included the three sequences). TR1 coincided with the 2013 eruption (Japan Meteorological Agency 2013). On the other hand, the volcanic activity associated with TR2 is unknown, though it is the most prominent tremors during the 2018 activity (Japan Meteorological Agency 2018). We expect to determine whether TR2 accompanied infrasound or not by the combined analyses of the data from the seismometer and the microphone at IOCD.

Figure 2 shows the wave traces and the spectrograms of the whole analyzed periods for TR1 and TR2. In both TR1 and TR2, the seismic velocity amplitude is the largest at IOCD and the smallest at JJTV. The relation is consistent with the relative distances between the stations and the active areas as in Figure 1. Although the seismic site-effects should be considered, strictly speaking, it implies the amplitude attenuation with distance from the source. The spectrograms indicate that the dominant frequencies of TR1 and TR2 range in 1-10 and 1-5 Hz, respectively. TR1 is overlapped by an airplane noise, which has a harmonic feature with dominant frequencies about 18 and 36 Hz (Figure 2c).

**Cross-correlation Analysis**

We performed a cross-correlation analysis between the vertical ground velocity and the microphone data at IOCD. The method distinguishes infrasonic signals from wind noise, and helps detect eruption events in the situation where only one microphone is available (Ichihara et al. 2012). With the distance between the seismometer and the microphone, \( d \), of 7.7 m, the range of frequency, \( f \), should satisfy the relation:

\[
v/3 < f \cdot d \leq C_a
\]

\( v \) is the wind velocity, and \( C_a \) is the sound velocity), which guarantees that \( d \) is smaller than the infrasound wave lengths and larger than correlation lengths of wind noise (Shields 2005). We assumed that \( C_a \) is 340 m/s. We also considered that \( v \) is smaller than the maximum wind speed of 11.8 m/s observed at Chichijima island about 280 km to the north of Ioto in April 2013 and September 2018. Then, the relation is rewritten as \( 0.5 < f \leq 44.2 \) Hz. Therefore, we used the frequency band of 1-10 Hz. The cross-correlation coefficient (CC) was calculated for the delay time of the vertical ground velocity to the microphone data from 0.5 to 0.5 s by use of a 5-s time window sliding every 1 s.

**Seismic- and Infrasonic- Spectral Ratio**

Seismic waves propagating in the ground and infrasound waves propagating in the atmosphere individually generate both of ground motion and pressure oscillation. Therefore, it is not apparent whether an oscillation recorded by a single sensor (either a seismometer or a microphone) is a seismic- or infrasonic-wave. When infrasound propagates along the ground surface, the vertical ground velocity, \( w_p(f) \), induced
by the infrasonic pressure wave, \( p_{in}(f) \), is given in the frequency domain by

\[
w_p(f) = \frac{C_a}{2(\lambda + \mu)} \cdot \frac{\lambda + 2\mu}{\mu} \cdot p_{in}(f)e^{-\pi/2},
\]

(1)

where \((\lambda, \mu)\) are Lame’s constants for the ground (Ben-Menahem and Singh 1981). The effective values of Lame’s constants vary with stations and depend on frequency because they are influenced by the shallow structure of the ground and topography (Langston 2004; Nishida and Ichihara 2015). Empirically, \(|w_p(f)/p_{in}(f)|\) ranges 0.1-10 \( \mu \text{m}\cdot\text{s}^{-1}/\text{Pa} \) (e.g. Langston 2004; Matoza and Fee 2014; Nishida and Ichihara 2015; Ichihara 2016). On the other hand, when the vertical ground velocity due to the propagating seismic wave, \( w_{in}(f) \), induces local air pressure perturbations, \( p_w(f) \), in a homogeneous fluid medium assuming that the time scale of the vertical motion is short compared with the acoustic cut-off period in the atmosphere, the relation is formulated as

\[
p_w(f) = \rho C_a \cdot w_{in}(f),
\]

(2)

where \( \rho \) is the density of air (Cook 1971; Donn and Posmentier 1964). There exist records of ground-to-atmosphere signals excited by large earthquakes, which have a good agreement with the theory (Kim 2004; Watada et al. 2006). In these cases, \(|w_{in}(f)/p_w(f)|\) is as large as 3000 \( \mu \text{m}\cdot\text{s}^{-1}/\text{Pa} \). Because the spectral amplitude ratio of seismometer data to microphone data is significantly different depending on whether the wave is seismic or infrasonic, we use it to distinguish the waves. For convenience, the observed spectral amplitude ratio will be referred to as \((w/p)_{obs}\).

We calculated \((w/p)_{obs}\) for each of TR1 and TR2 in the following steps. 1⃝ Power spectral densities (PSDs) of the seismic data (the vertical component) and infrasonic record were individually calculated in a 10-s window sliding with 5-s overlapping for the periods shown in Figure 2. 2⃝ The wind is the most critical noise in the infrasound record, which has significant power in low frequencies below 1 Hz (Fee and Garces 2007). Therefore, we focused on the frequency range above 1 Hz in searching for volcanic signals. For each time window, we calculated the powers of the infrasound data in high- and low-frequency bands, \( E_h = \int_{f_1}^{f_2} PSD(f)df \) and \( E_l = \int_{0.5}^{f_1} PSD(f)df \), respectively. If \( \sqrt{E_h} > 3\sqrt{E_l} \), we employed the time window. 3⃝ For each of TR1 and TR2, we averaged the PSDs over the time windows selected in 2⃝ to obtain the mean PSDs, \( P_W(f) \) and \( P_P(f) \), for the seismic and infrasonic data, respectively. Then, we obtained the spectral amplitude ratio, \((w/p)_{obs} = \sqrt{P_W(f)/P_P(f)}\). 4⃝ We also evaluated mean PSDs
for the background noise spectra, $P_{PW}^b(f)$ and $P_{PP}^b(f)$, for seismic- and infrasonic-data, respectively. We searched the background time windows from 14:00-17:00 of April 11, 2013, for TR1 and from 0:00-24:00 of September, 8, 2018, for TR2. When the infrasonic PSDs above 1 Hz are smaller than $10^{-3}\ \text{Pa}^2/\text{Hz}$ and the seismic PSDs are below 5 ($\mu\text{m} \cdot \text{s}^{-1})^2/\text{Hz}$ in a time window, we regarded it as a background noise window. We took 100 time windows for each of TR1 and TR2, and averaged the PSDs to obtain $P_{PW}^b(f)$ and $P_{PP}^b(f)$. The mean PSDs for the signals ($P_{W}(f)$ and $P_{P}(f)$ obtained in (3)) and those for background ($P_{PW}^b(f)$ and $P_{PP}^b(f)$ evaluated in (4)) were compared. The meaningful frequency bands were defined by the following condition:

$$\frac{P_{W}(f) - c_1 \cdot P_{PW}^b(f)}{c_1 \cdot P_{W}^b(f)} > 10, \quad \frac{P_{P}(f) - c_2 \cdot P_{PP}^b(f)}{c_2 \cdot P_{PP}^b(f)} > 10,$$

(3)

The background noise levels were adjusted by $c_1$ and $c_2$ so that the mean PSDs for the signal and the noise were equal at 0.5 Hz. Namely, $c_1 = P_{W}(0.5)/P_{PW}^b(0.5)$, and $c_2 = P_{P}(0.5)/P_{PP}^b(0.5)$. The shifting was applied to remove the effect of temporal change in wind noise. The spectral characteristics of the tremors and the background noises obtained by the method are compared in Figure 3.

For reference, we evaluated $(w/p)_{\text{obs}}$ for known infrasonic- and seismic-signals, which are airplane sound propagating in the atmosphere (PN) and seismic waves generated by tectonic earthquakes (EQ). The method was similar to the above. For PN, we analyzed the data from 16:00-17:00 of April 11, 2013 and from 10:00-11:00 of September 12, 2018, in which we found clear airplane signals in the spectrograms. As PN signals had powers in high frequency, we changed the frequency range in step (1) to 10-40 Hz, and the threshold of equation (3) to 30 in step (5). In step (2), 6 and 54 time windows met the requirement in 2013 and 2018 data, respectively. As regards EQ, we used 51 and 118 tectonic earthquakes that occurred in Ioto in March-April 2013 and September 2018, respectively. The timetables are shown in Supplementary materials referring to the NIED catalog. Because many earthquakes occurred in Ioto, signals that had good cross-correlation between the seismometer and the microphone (the CC larger than 0.6) and peak seismic amplitudes larger than 50 $\mu\text{m}/\text{s}$ at IOCD were selected. Twenty-second records from 10 s before the peaks were used for calculating $P_{W}(f)$ and $P_{P}(f)$. Then, we performed step (5) to select the meaningful frequency band.
Results

The seismic- and infrasonic- records and the seismic-to-infrasonic CCs in the analyzed periods for TR1 and TR2 are shown in Figure 4. Figure 5 shows results of the same analysis for the reference signals (PN and EQ). If both of the seismometer and the microphone record infrasound propagating along the ground surface, the CC would have a positive peak near $\tau = 1/(4f_0)$, a negative peak near $\tau = -1/(4f_0)$, and a node at $\tau = 0$ (Ichihara et al. 2012; Yukutake et al. 2018). We observe some change in the pattern for a few minutes from 16:00 in TR1 (Figure 4c). However, the pattern is not so clear as that for PN (Figure 5c). It suggests that the major contribution to the TR1 signal recorded by the seismometer is not infrasonic origin but seismic waves. Nevertheless, the subtle pattern change may be due to the coexistence of infrasound with the seismic tremor of TR1. On the other hand, CCs of TR2 and EQ share a feature with a positive peak around $\tau = 0$ and a negative peak in $\tau > 0$ as shown in Figures 4d and 5d. It suggests a seismic origin for TR2 signal in both seismic- and infrasonic- data.

Figure 6 compares the mean power spectra and $(w/p)_{obs}$ of TR1 and TR2 against PN and EQ. Figure 6c shows that $(w/p)_{obs}$ of TR1 is closer to that of PN than EQ. The infrasonic amplitude is too large to be generated by the observed ground velocity. On the other hand, both powers of the seismic and infrasonic data during TR2 are comparable to the ground-to-atmosphere signal of EQ (Figures 4b, 5b, and 6b). These results support the inference from CC that a pressure wave accompanied TR1 but not TR2. Although the existence of infrasound for TR2 cannot be completely ruled out, it would have been very weak ($< 0.1$ Pa), if it happened.

Discussion

The values of $(w/p)_{obs}$

The spectral ratios of seismic data to infrasonic data, $(w/p)_{obs}$, were calculated to discuss the volcanic activities with TR1 and TR2, as presented in Figure 6c and 6d. Here, we consider if the values of $(w/p)_{obs}$ are reasonable, focusing on the reference signals of EQ and PN. EQ that is seismic wave has $(w/p)_{obs}$ in agreement with the theoretical value of ground-to-atmosphere signals (Kim 2004; Watada et al., 2006), which is given in equation (2) and indicated as the dotted line in the figure. PN is acoustic wave and has $(w/p)_{obs}$ ranging 1-10 $\mu$m-s$^{-1}$/Pa. The range is included in the observed values for atmosphere-to-ground signals, 0.1-10 $\mu$m-s$^{-1}$/Pa, from various sources like volcanoes (Nishida and Ichihara 2015; Ichihara 2016; Matoza and Fee 2014) and thunder (Lin and Langston 2007, 2009a).
The theoretical amplitude ratio for atmosphere-to-ground signals is given by equation (1). It assumes
that the atmospheric wave is propagating along the ground surface at speed much lower than the seismic
waves, and the ground and the atmosphere are homogeneous half-spaces (Ben-Menahem and Singh 1981).
Equation (1) with \( \frac{w}{p}_{\text{obs}} \sim 10 \mu \text{m/s}^{-1}/\text{Pa} \) yields the shear modulus of \( \sim 20 \text{ MPa} \) using \( \lambda > \mu \). The
shear modulus is equivalent to that of loose-packing sand or clay (Lo Presti et al. 1997), and Chidorigahara
area in Ioto, including IOCD station, is composed almost entirely of poorly-consolidated volcanic sands
and gravels (Corwin and Foster 1959). It has also been reported that the near-surface P wave velocity of
the area is \( \sim 500 \text{ m/s} \) and close to the sound velocity (Kumagai and Takahashi 1985). Then, the velocities
of the S wave and the surface waves should be smaller than the sound velocity. The low seismic velocities
violate the assumption of equation (1). Besides, the incident angle of the airplane noise is not horizontal.
Lin and Langston (2009a,b) analyzed seismic and infrasonic data of thunder-induced signals and showed
that the ground motion is controlled by the average thickness and velocities of the near-surface layers
including the topmost soft and thin layer. We avoid further interpretation of the \( \frac{w}{p}_{\text{obs}} \) for infrasound
because of many unknown factors. Nevertheless, it is certain that \( \frac{w}{p}_{\text{obs}} \) for atmosphere-to-ground
waves is much smaller than that for ground-to-atmosphere waves.

The proposed method has an advantage that it does not use the absolute values of the record.
Instruments at permanent monitoring stations are not necessarily well calibrated. The field calibration
of an infrasonic station is an issue. Yukutake et al. (2018) made an on-site calibration for the single
microphone that recorded the 2015 Hakone eruption. They conducted the calibration after the volcanic
activity declined, and found a significant deviation of the microphone response from its specification.
On-site calibrations would be more difficult at isolated islands. The seismic stations can be tested using
distant earthquakes recorded simultaneously by multiple stations. The same technique is not useful
for infrasound, even if there is a good source and enough stations. The spatial amplitude distribution
depends significantly on the atmospheric structure (Lacanna et al. 2014), which is also difficult to monitor
especially at isolated islands.

**Volcanic activity associated with TR1 and TR2**

We found that TR1 accompanied infrasound. The infrasound might have coincided with the vent
opening or the ejection of large ballistic rocks that were observed from the ground and the sky (Japan
Meteorological Agency 2013). The seismic-infrasonic cross-correlation (Figure 4c) shows a subtle pattern
change for a few minutes with TR1. In the power-spectral analysis, we selected the time windows in which
the signal might dominate wind noise (step ②). According to the span of the selected time windows, we infer that the explosive activity of the 2013 eruption lasted at least 60 s from 16:01 on April 11.

TR2 did not accompany apparent infrasound signals. Matsumoto et al. (2019) reported that remote hydrophones detected no relevant signal on September 12, either, even though a small splash was observed from the sky. Explosions under shallow water and violent water jets into the air generate detectable infrasound signals (Ichihara et al. 2009; Lyons et al. 2019, 2020). On the other hand, gas emission into the atmosphere by a buoyant plume does not efficiently emanate infrasound (Ichihara et al. 2009). Therefore, we conclude that the unconfirmed 2018 event that generated TR2 was not an explosive eruption either on the ground or undersea.

Conclusions

We have analyzed two volcanic tremor events of Ioto, which were TR1 with the 2013 eruption and TR2 with the unconfirmed 2018 activity. The aim was to determine whether the events accompanied infrasound indicating the volcano’s surface activity, by a single microphone recorded the events with a co-located seismometer. With a pair of microphone and seismometer, we can sometimes detect infrasound by the seismic-infrasonic cross-correlation method. However, the method was not applicable in the studied case of Ioto because seismic activity was intense. Even in such a case, comparing the spectral amplitude ratios \( (w/p)_{obs} \) of the events with those of known seismic- and infrasonic- signals gave information.

We concluded that TR1 included infrasound, while TR2 did not. The infrasound in the part of TR1 might have been excited by the vent opening or the ejection of ballistic rocks. TR2 was not an explosive eruption either on the ground or in the shallow water. If there was any gas (and ash) emission, it might have occurred gently undersea.

Infrasonic observation is useful for the detection of eruptions. However, available infrasonic stations are limited at volcanoes on isolated volcanic islands or with less frequent eruption. Using \( (w/p)_{obs} \) with a pair of seismic and infrasonic sensors would provide a possibility of extracting infrasound signals covered by seismic signals and wind noise. Because the method refers to \( (w/p)_{obs} \) of known signals, it is available without perfect calibrations for the instruments. Concerning to the infrasonic reference signals, we could use various artificial and natural sources like airplane noise, bolide shockwaves, and thunder (Langston 2004; Lin and Langston 2007, 2009a).
List of abbreviations

TR1: Tremor associated with an eruption in April 11, 2013 at Ioto, TR2: Tremor associated with an unconfirmed activity in September 12, 2018 at Ioto, PN: Referenced plane sound, EQ: Referenced tectonic earthquake, CC: Cross-correlation coefficient, \((w/p)_{obs}\): Spectral amplitude ratio between seismic record and infrasonic record, PSD: Power spectral density.

Availability of data and materials

The data set analyzed in this study is not officially available at the request of JMA and JSDF.

Competing interests

The authors declare that they have no competing interest.

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Authors’ contributions

AK performed the analysis and drafted the manuscript. MI offered technical support for the present study, and helped with discussion and revision of the manuscript. All authors read and approved the final manuscript.

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Preparing illustrations and figures
Figure 1. Map of Ioto island with the station locations shown by open circles. The red triangle in the inset marks the location of Ioto in Japan. IOCD has a seismometer and a microphone, of which data are mainly used in this study. IJSV and IJTV are seismic stations. The crosses indicate the 2013 eruption vent (Japan Meteorological Agency 2013) and the location where the water spouts were observed in the 2018 activity (Japan Meteorological Agency 2018).

Figure 2. Vertical ground velocity of TR1 on April 11, 2013, and TR2 on September 12, 2018. (a, b) The raw records at IOCD (black), IJSV (yellow), and IJTV (blue). (c, d) The spectrograms for the data at IOCD. The color bars indicate the power in dB = 10 log_{10}(PSD/PSD_{ref}) with PSD_{ref} = 1 (m·s^{-1})^2/Hz.

Figure 3. Power spectral features of tremors (magenta) and background noise (gray). The symbols show the values of the meaningful frequency bands selected by step ⃣ of the text. (a, b) The upper magenta line with the upward triangles is $P_W$, and the associated gray line is $P^b_W$, on the left axis. The lower magenta line with inverted triangles is $P_P$, and the associated gray line is $P^b_P$, on the right axis. (c, d) The spectral amplitude ratio between seismic record and infrasonic record, $(w/p)_{obs}$, as a function of frequency. The horizontal dashed line indicates the theoretical value for the ground-to-atmosphere wave calculated by equation (2) with $C_a = 340$ m/s and $\rho = 1.16$ kg/m$^3$.

Figure 4. (a, b) TR1 and TR2 waveforms recorded at IOCD. The black shows the vertical component of the seismometer, and the magenta shows the microphone record. The data were filtered within the frequency band of 1-10 Hz. (c, d) The cross-correlation of the seismic and infrasonic data in (a, b) in a 5-s time window sliding every 1 s. The vertical axis, $\tau$, is the time delay of the seismic to infrasonic data.

Figure 5. a, b) PN and EQ waveforms recorded at IOCD on September 12 and 18, 2018, respectively. The black shows the vertical component of the seismometer, and the magenta shows the microphone record. PN and EQ data are filtered within the frequency band of 10-30 Hz and 1-10 Hz, respectively. (c, d) The cross-correlation of the seismic and infrasonic data in (a, b) in a 5-s time window sliding every 1 s. The vertical axis, $\tau$, is the time delay of the seismic to infrasonic data.
Figure 6. The same plots of tremors (magenta) as in Figure 3, compared with the corresponding values of EQ (black) and PN (blue).