Acoustic-to-seismic ground coupling: coupling efficiency and inferring near-surface properties

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SUMMARY
A fraction of the acoustic wave energy (from the atmosphere) may couple into the ground, and it can thus be recorded as ground motion using seismometers. We have investigated this coupling, with two questions in mind, (i) how strong it is for small explosive sources and offsets up to a few tens of meters and (ii) what we can learn about the shallow subsurface from this coupling. 25 firecracker explosions and five rocket explosions were analysed using colocated seismic and infrasound sensors; we find that around 2 per cent of the acoustic energy is admitted into the ground (converted to seismic energy). Transfer coefficients are in the range of 2.85–4.06 nm Pa−1 for displacement, 1.99–2.74 μm s−1 Pa−1 for velocity, and 2.2–2.86 mm s−2 Pa−1 for acceleration. Recording dynamic air pressure together with ground motion at the same site allows identification of different waves propagating in the shallow underground, notably the seismic expression of the direct airwave, and the later air-coupled Rayleigh wave. We can reliably infer shallow ground properties from the direct airwave, in particular the two Lamé constants (λ and μ) and the Poisson ratio. Firecrackers as pressure sources allow constraining elastic parameters in the top-most layer. In this study, they provide frequency-dependent values of λ decreasing from 119 MPa for low frequencies (48 Hz) to 4.2 MPa for high frequencies (341 Hz), and μ values decreasing from 33 to 1.8 MPa. Frequency-dependent Poisson ratios ν are in the range of 0.336–0.366.

Key words: Elasticity and anelasticity; Acoustic-gravity waves; Acoustic properties; Earthquake monitoring and test-ban treaty verification; Seismic noise.

1 INTRODUCTION
During most of its history, seismology has regarded elastic and pressure waves, propagating below and above ground, separately, calling the former waves ‘seismic’ and the latter ‘acoustic’. This was in part due to an intellectual boundary, but it was also for convenience: the assumption of a traction-free surface (and of seismic displacements that are discontinuous across it) provided a simple and convenient upper boundary condition for seismic modelling.

Nevertheless, a coupling of seismic and acoustic waves has been noticed occasionally: audible signals have been reported from earthquakes, and this has been established on a sound scientific basis (Mutschlecner & Whitaker 2005). Indeed, large earthquakes set off seismic waves that can be observed up at heights of more than 200 km, for example, Lognonné et al. (2016) and Shani-Kadmiel et al. (2017), passing through the atmosphere as acoustic waves. Similarly, large volcanic eruptions have regularly produced atmospheric pressure perturbations that were recorded by seismic stations (e.g. Neuberg et al. 1994). Seismic observations have also been made of meteorites, for example, for the Great Siberian meteorite that exploded near Tunguska on 30 June 1908 (Whipple 1930; Ben-Menahem 1975; Wheeler & Mathias 2019) and in several others (Brown et al. 2003; Kumar et al. 2017; Varypaev et al. 2019). The most spectacular observations were made perhaps for the Chelyabinsk meteor, with the help of infrasound and seismic networks (e.g. Tauzin et al. 2013). Recent studies point out that seismic sensors are valuable also for studying anthropogenic and other acoustic phenomena, as well as the propagation behaviour of acoustic waves through the atmosphere, in particular after explosive sources (Schneider et al. 2018; Blixt et al. 2019; Fuchs et al. 2019b). This is particularly promising since there are many more seismic stations around the globe than infrasound arrays.

A prominent example of putting the information from different networks (and physical quantities) together is nuclear verification research, greatly guided by CTBTO efforts (e.g. Pilger et al. 2017), where seismic, infrasound and hydroacoustic data are analysed together, to better understand the nature of suspicious events, and especially for estimating the yield of explosions. Extracting the maximum amount of information from the data requires calibrating the information from the various measurement types, and especially understanding acoustic/seismic coupling.
The acoustic-seismic coupling has been known for several decades: Sabatier et al. (1986) modelled the acoustic-seismic coupling, to predict the seismic transfer function. Albert & Orcutt (1989) used gunshots as the source, to compare seismic recordings with those recorded by microphones. Edwards et al. (2007) analysed shockwave data from a space capsule re-entry on coocated seismic and pressure sensors and suggested an energy admittance of up to 2 per cent. In the same year Lin & Langston (2007) experimented with and pressure sensors and suggested an energy admittance of up to 2 shockwave data from a space capsule re-entry on coocated seismic (1989) used gunshots as the source, to compare seismic recordingspling, to predict the seismic transfer function. Albert & Orcutt (2016) studied how frequency and incidence angle impact seismo-acoustic coupling using noise from jet overflights. Averbuch et al. (2020) conducted a seismo-acoustic modelling of infrasound propagation from underwater and underground sources and showed that evanescent wave coupling and leaky surface waves are the main energy contributors to long-range infrasound propagation.

Acoustic/elastic coupling has gained large importance also in material sciences (e.g. Hess 2002), for determining structure and fracture behaviour, even at extremely small spatial scales, and for structural health monitoring (damage imaging) in civil engineering (see ‘ultrasonic guided wave tomography’ in Yan et al. 2010; Rose 2011). The seismo-acoustic coupling may also play an important role in many applications of environmental seismology; the near-surface region of the Earth is an interesting target to be studied by seismic waves (Park et al. 2019), for example for agricultural purposes such as locating the fragipan horizon (Howard & Hickey 2009).

For this study, we investigated whether the subtle change in impedance of the ground caused by explosions with small net explosive mass (NEM) and small offsets (of up to a few tens of metres) can be measured by nodal geophones. We focused on the acoustic/seismic coupling with two questions in mind (i) how strong it is and (ii) what can be learned about the shallow subsurface from this coupling. Our field experiment tested ground coupling in a controlled yet natural setting. Compared with earlier experiments, our setting has particular advantages: (i) it is set off under controlled conditions and we used several sources in and above the ground (hammer blows, small explosions and rockets); (ii) the use of nearly 100 seismic sensors allowed us to track the wavefield and distinguishing different wave types. We also investigate how seismo-acoustic coupling can be used for retrieval of near-surfaces ground properties and propose a method to estimate the Lamé constants and the Poisson ratio for sites equipped with co-located infrasound and seismic instrumentation.

The paper is organized as follows: we introduce the layout of the experiment, as well as geological properties of the site and the instrumentation used in Section 2. In Section 3, we develop the necessary mathematical apparatus and in Section 4 we show examples of seismic and pressure records, calculate coupling transfer coefficients as well as the energy coupling efficiency and show how to infer elastic parameters of shallow subsurface from such measurements. We discuss our observations in Section 5 and finally draw conclusions.

2 ACOUSTIC TO SEISMIC COUPLING EXPERIMENT

2.1 Description of the experiment

The experiment was performed on 14 May 2019, near the Conrad Observatory on the Trafelberg mountain in Austria (coordinates: 47.9270, 15.8582, elevation 1088 m), on a flat 40 m × 80 m field site made of water-saturated limestone-rich thumb-sized breccia surrounded by trees and forest (see Fig. 1). The core of the experiment is a seismic array of 97 geophone nodes (Fairfield ZLand Gen2, 3-components, 5 Hz corner period) arranged in a concentric ring layout of 20 m diameter (see Figs 1 and 2). The in-ring spacing of the nodes is 2 m and neighbouring rings are separated by 2.5 m. All ring nodes and the central node were deployed at the ground surface using metal spikes that are attached to them (see Fig. 3). 16 of the sensors were installed and slightly buried along N–S and E–W lines in between the rings. All nodes were GPS-synchronized and recorded ground velocity at 2000 samples per second. Additionally, four nodes at the north, south, east and west edge of the ring array were each colocated with a Hyperion IFS-5111 infrasound sensor recording air pressure changes at 1000 samples per second. A particularly useful property of the infrasound sensors is that they are seismically-decoupled (the nominal seismic sensitivity of the infrasound sensors is 0.08 Pa m s−2, at 10 Hz). The spacing between colocated nodes and Hyperions is approximately 20 cm. Please refer to the data set description document (Fuchs et al. 2020) for a more comprehensive explanation of the instrumentation and setup.

We used regular firecrackers and effect rockets as acoustic sources (see Fig. 4). The net size of each firecracker charge is given by the NEM (‘Net explosive mass’) specified by the manufacturer. For our experiment we shot firecrackers of three different charges: L – ‘large’ – 7.5 g NEM, M – ‘medium’ – 6.75 g NEM and S – ‘small’ – 0.8 g NEM. All crackers were fired at 1.6–2.1 m above the ground surface. Flying effect rockets with 75 g NEM (including fuel and effect) were used to create explosive sources at height. Please refer to the data set description document (Fuchs et al. 2020) for more detailed documentation of all active sources that were shot during the experiment.

Next to the experiment site there is a meteorological station called MetLift (see Fig. 1)—a meteorological measurement platform for snowy areas (Dorninger 2012). This station measures air temperature, humidity and pressure as well as wind speed at multiple heights (1, 2, 3, 4, 5 and 7 m), and wind direction. Data were recorded at 1 sample per minute. During the experiment, the temperature was varying in the range of +1 to +3 °C. Relative humidity was varying in the range of 45–70 per cent. Air pressure was steadily dropping from 903.7 to 903.0 hPa and the wind speed was varying in the range of 0–4 m s−1, blowing irregularly from varying directions (see Fig. A1).

2.2 Air/soil properties and wave velocities

Since we had access to meteorological measurements at 1-min intervals we were able to calculate the time-dependent air density using the method from Picard et al. (2008):

\[ \rho_{\text{air}} = \frac{p M_a}{Z RT} \left[ 1 - x_v \left( 1 - \frac{M_v}{M_a} \right) \right]. \]  

(1)

where \( \rho_{\text{air}} \) is the density of air [kg m−3], \( p \) is pressure [Pa], \( T \) is air temperature [K], \( x_v \) is the mole fraction of water vapor, \( M_a \) is the molar mass of dry air [g mol−1], \( M_v \) is the molar mass of water [g mol−1]. \( Z \) is the compressibility factor and \( R \) is the molar gas constant [J mol−1 K−1].

The time-dependent acoustic velocity in the air was calculated as described in Cramer (1993):

\[ v_{\text{air}}^2 = \gamma \times \frac{RT}{M_a} \left( 1 + \frac{2 p B}{RT} \right), \]  

(2)
where \( V_{\text{air}} \) is the speed of sound in \( \text{m s}^{-1} \), \( \gamma \) the specific heat ratio, \( M_m \) the molecular mass of air and vapour mixture (determined via relative humidity) \( \text{g mol}^{-1} \) and \( B \) is the second Virial coefficient.

The surface layer at the experiment site is made of water-saturated limestone-rich thumb-sized breccia (see Fig. 3). Knowledge about the bulk elastic properties of this near-surface layer is required for the calculation of the seismo-acoustic coupling efficiency. Therefore, we performed a seismic refraction experiment with hammer beats as active sources (see Fig. 6). This allowed us to estimate the velocity of a \( P \) wave in the soil layer as 1000 m s\(^{-1}\). Since we did not have access to lab measurements we had to assume the soil density. Gegenhuber (2015) reports the density of denudated lime-stones in Austria varying from 2.73 to 2.85 g cm\(^{-3}\) (averaging to 2.79 g cm\(^{-3}\)) with porosity in the range of 2.33–4.85 per cent (averaging to 3.59 per cent). Our sediments are loose; therefore we had to account for a presumably higher porosity using the method for wet bulk density of a rock sample calculation provided in Rieke & Chilingarian (1974):

\[
\rho_{\text{bw}} = \rho_{\text{bd}} \left(1 - \phi\right),
\]

(3)

where \( \rho_{\text{bw}} \) is wet bulk density \( \text{g cm}^{-3} \), \( \rho_{\text{bd}} \) is dry bulk density \( \text{g cm}^{-3} \), \( \rho_s \) is matrix (grain-mineral) density \( \text{g cm}^{-3} \), \( \rho_f \) is density of a fluid \( \text{g cm}^{-3} \) and \( \phi \) is porosity [decimal per cent]. Taking \( \rho_{\text{bw}} = 2.89 \text{ g cm}^{-3}, \rho_f = 1.00 \text{ g cm}^{-3} \) and \( \phi = 47 \text{ per cent} \) (sediments are loose) we estimate the soil density at the experiment site to be approximately 2.02 g cm\(^{-3}\).

3 GROUND MOTION ASSOCIATED WITH ACOUSTIC WAVES

Ben-Menahem & Singh (1981) showed how to compute ground displacement originating from acoustic waves in the atmosphere when apparent velocity of the acoustic wave and total interface pressure in the air are known (their eq. 9.187):

\[
U_x = -i V_{\text{air}} \frac{P_0}{2\alpha(\lambda + \mu)} e^{i \omega (t - x c)}
\]

(5)

\[
U_z = -V_{\text{air}} \frac{P_0}{2\alpha(\lambda + \mu)} \left(\frac{\lambda + 2\mu}{\mu}\right) e^{i \omega (t - x c)}
\]

(6)

where \( U_x \) is the horizontal ground displacement \( \text{m} \), \( U_z \) is the vertical ground displacement \( \text{m} \), \( P_0 \) is the amplitude of the excess interface pressure \( \text{Pa} \). \( c \) is the apparent (= horizontal) velocity in air \( \text{m s}^{-1} \), \( V_{\text{air}} \) is the true velocity in air \( \text{m s}^{-1} \), \( \omega \) is a vector of the angular frequencies \( \text{rad s}^{-1} \), \( \lambda \) is the first Lamé constant \( \text{Pa} \), \( \mu \) is the second Lamé constant \( \text{Pa} \), \( x \) is the offset \( \text{m} \) and \( t \) is the time \( \text{s} \). We will, in the following, consider horizontally propagating acoustic waves; then we have \( c = V_{\text{air}} \).

Excess interface pressure \( P_0 \) can be written as:

\[
p_0 = \frac{P_0}{e^{i \omega (t - x c)}}
\]

(7)

where \( P_0 \) is the amplitude of the acoustic pressure pulse \( \text{Pa} \). Thus, we can substitute eq. (7) in eq. (5) and eq. (6) and we obtain:

\[
U_x = -i V_{\text{air}} p_0 \frac{1}{2\alpha(\lambda + \mu)}
\]

(8)

\[
U_z = -V_{\text{air}} p_0 \frac{(\lambda + 2\mu)}{2\alpha(\lambda + \mu)}
\]

(9)

Surface pressure \( p_0 \) is measured with infrasound sensors, \( U_x \) and \( U_z \) are measured with three-component seismometers (we rotate records from ZNE to ZRT, since backazimuth of the signal is always known to us) and apparent velocity \( V_{\text{air}} \) can be inferred from eq. (2). Therefore, we are able to obtain values for Lamé constants by solving eq. (8) and eq. (9) for \( \lambda \) and \( \mu \):

\[
\lambda = A \frac{U_x}{U_z} - \frac{AB}{AU_z - BU_x}
\]

(10)
Acoustic-to-seismic ground coupling

Figure 2. Detailed view of the seismic ring layout (black and blue symbols) and colocated infrasound sensors (pink). Seismic sensors are either buried (blue) or on the surface (black). The red star shows a failed seismic node with no data available.

\[ \mu = \frac{AB}{AU_z - BU_x} \]  
(11)

where \( A = -\frac{V_{air} p_0}{2\omega} \) and \( B = -\frac{V_{air} p_0}{2\omega} \).

It is perhaps remarkable that an exact solution for the Lamé constants can be given. This is possible, if acoustic pressure is available, in addition to ground motion. The knowledge of \( \lambda \) and \( \mu \) allows us to also derive the Poisson ratio \( \nu \):

\[ \nu = \frac{\lambda}{2(\lambda + \mu)} \]  
(12)

and the bulk modulus:

\[ K = \lambda + \frac{2\mu}{3} \]  
(13)

There are some interesting properties of eqs (8) and (9), for example that the frequency content of the pressure signal should be best-recoverable using ground velocity. The pressure waveform should reappear in \( \dot{U}_x \), while \( \dot{U}_z \) is phase-shifted by 90°. The pressure waveform also appears (approximately) in the negative vertical ground motion \( U_z \), and in the (positive) vertical ground acceleration \( \ddot{U}_z \). The frequency content differs though. This explains some of the observations in the literature, for example in fig. 4 of Edwards et al. (2007).
4 RESULTS

4.1 Waveforms and spectral content

To study acoustic-to-seismic coupling, we have analysed 25 firecracker explosions and five rocket explosions using the ring array and colocated infrasound sensors. Fig. 5 shows an example of waveforms and spectrograms for a firecracker explosion (charge L) at 4.5 m distance from the sensor pair HYP01-R501. This example has been selected for having low wind disturbance.
Figure 5. Explosion experiment: infrasound and seismic measurements of a 'large' charge explosion at colocated seismic (R501) and infrasound (HYP01) sensors (data is not filtered). (a) Spectrograms for corresponding seismic and infrasound records. (b) Normalized and overlaid, on top of each other, seismic and infrasound waveforms. The background shows the time window of the direct airwave used for determining elastic parameters, and that of the air-coupled Rayleigh wave. (c) Section plot (time versus distance). Seismic traces are shown in black (vertical components), infrasound in red. Red solid-line indicates picked acoustic velocity. Purple dashed-line indicates air-coupled Rayleigh wave. The acoustic wave propagates with a velocity of around 0.3 km s$^{-1}$; coherent phases of the subsequent air-coupled Rayleigh wave suggest a similar phase velocity.

The spectrograms and waveforms show that the co-located seismic and infrasound sensors have recorded energy in a similar frequency band for the first arrival, which is followed on the seismic trace (and only there) by a prolonged wave train with a narrow-band spectrum. The pressure signal generated by the firecracker is remarkably short (0.3–0.4 s). The signal onset is characterized by an anticorrelation of seismic and pressure signals, as expected from eq. (6). This is expected intuitively for an elastic subsurface, since the build-up of air pressure above ground causes the ground to move down (Ben-Menahem & Singh 1981; Matson 2018). There is a phase shift between vertical and radial components (see Fig. C1). We will use this relation to extract information about the elastic properties of the subsurface.

The record section in Fig. 5 shows all seismic and acoustic traces recorded along with the entire ring array, for the example explosion. The first onset indicates a constant acoustic velocity of around 333 m s$^{-1}$, independent of distance from the source. There is no indication of seismic phases arriving before the acoustic phase. In addition to the first impulsive arrival at later times, we observe a pronounced low-frequent wave train of up to 0.2 s duration on all seismic traces. The spatial coherence of waveforms in Fig. 5 over the entire ring array indicates a phase speed close to the acoustic wave speed, yet the energy clearly propagates in the ground (it is not on the infrasound traces). We suspect that this is the manifestation of an air-coupled Rayleigh wave (Haskell 1951; Albert & Orcutt 1989). This will be discussed below. Different from those guided
waves, classical Rayleigh waves do not seem to be present. The latter would propagate with around 0.9 times the shear wave velocity, which would correspond to about twice the acoustic velocity in our case: no such waves are visible in the record section.

For comparison, Fig. 6 shows data from a hammer shot at a similar distance from the ring array and the colocated sensor pair. For small offset (<6 m) the first arrival appears to be propagating at acoustic velocities. However, in contrast to the firecracker explosions, a clear seismic arrival is visible at distances larger than 6 m; it is propagating with a velocity of approximately 1000 m s\(^{-1}\). Generally, seismic waveforms show more complexity compared with an acoustic source. The acoustic phase is barely visible, and can only be safely identified on the infrasound sensors.

4.2 Constraining near-surface properties from seismo-acoustic waves

We use the relations from Section 3 to infer subsurface parameters from the seismic expression of the acoustic wave. For this purpose, we transform \(U_x\) and \(U_z\) into the spectral domain using a real fast Fourier transform (see Cooley & Tukey 1965), and solve for \(\lambda\) and \(\mu\). That way we obtain Lamé constants as a function of frequency. We only use frequencies that are actually present in the waveforms (compare Fig. 5) and that can be reasonably resolved given the length of the time window and data sampling rate, into the account. This results in 48–341 Hz as the usable frequency range.

We select a very short time window (0.04 s length) around the first arrival of the acoustic signal for each seismic-infrasound station pair and for each firecracker that was shot at an offset of 4–5 m; this geometry corresponds to an apparent velocity—parallel to the surface—of \(V_{\text{air}} = 333\) m s\(^{-1}\); using equation eq. (2) to calculate acoustic velocities in air. Averaging over all shots and station pairs we obtain the first Lamé constant \(\lambda\) in the range of 119 MPa for low-frequencies (48 Hz) to 4.2 MPa for high-frequencies (341 Hz). The second Lamé constant \(\mu\) is calculated as 33–1.8 MPa for 48–341 Hz frequencies, respectively (see Fig. 7).

The values for both \(\lambda\) and \(\mu\) show a gradual variation with frequency. Maximum values are around 340 MPa for \(\lambda\) and around 100 MPa for \(\mu\). The decrease with increasing frequency is by two orders of magnitude for \(\lambda\) and slightly less for \(\mu\). We also obtain values for the Poisson ratio \(\nu\) (see Fig. 7). These values are in the range of 0.336–0.366 decreasing with the increase of frequencies (and hence towards the shallower part of the profile). A similar effect was also observed in (Liu et al. 1997).

4.3 Acoustic-to-seismic ground coupling

We now turn to the question of which fraction of the acoustic energy couples into the ground, and how this manifests itself in the records. For this, we use the entire wave record, for example in Fig. 5. For each explosion, we analyse the absolute peak vertical ground motion (in units of displacement, velocity and acceleration) recorded at each colocated Node–Hyperion pair (see Table 1). These observations of ground motion, along with absolute peak pressure amplitudes from the associated infrasound sensors, enable us to determine the acoustic-to-seismic energy coupling efficiency (how much percent of energy is coupling from air to the ground) as well as the acoustic-to-seismic coupling transfer coefficient (ratio of ground motion versus pressure at the infrasound station).

The acoustic-to-seismic coupling transfer coefficient is defined as the ratio of the maximum absolute value of the vertical amplitude of the envelope of the seismic signal compared to the maximum absolute value of the envelope of the pressure signal (Edwards et al. 2007):

\[
C_{\text{AS}} = \frac{\text{max envelope}(A_{\text{seis}})}{\text{max envelope}(A_{\text{pressure}})},
\]

where \(A_{\text{seis}}\) is the recorded vertical seismic amplitude at the moment of incident and \(A_{\text{pressure}}\) the recorded pressure amplitude at the moment of incident.

The acoustic-to-seismic energy coupling efficiency compares kinetic energies in air and soil (Edwards et al. 2007):

\[
E_{\text{AS}} = \frac{1}{2} \rho_{\text{soil}} v_{\text{soil}}^2, \quad \frac{1}{2} \rho_{\text{air}} v_{\text{air}}^2,
\]

where \(\rho_{\text{soil}}\) is the density of soil [kg m\(^{-3}\)], \(\rho_{\text{air}}\) is the density of air [kg m\(^{-3}\)], \(v_{\text{soil}}\) is the particle motion velocity in the soil layer [m s\(^{-1}\)] and \(v_{\text{air}}\) is the particle velocity in the air [m s\(^{-1}\)]. The particle velocity \(v_{\text{air}}\) in the air needs to be calculated from the pressure measurements, again following Edwards et al. (2007):

\[
v_{\text{air}} = \frac{p}{\rho_{\text{air}} V_{\text{air}}},
\]

where \(p\) is measured air pressure change, \(\rho_{\text{air}}\) is density of air and \(V_{\text{air}}\) is the corresponding acoustic velocity.

Fig. 8 shows the ranges of transfer coefficients that we obtained from the 24 explosions, recorded at the different offsets. These are given separately for measures of ground displacement, velocity and acceleration. Fig. 8 also shows the energy coupling efficiency for all individual explosions. Table 2 gives the results in numerical form, providing the average value of the results at different distances (and its uncertainty).

Transfer coefficients are in the range of 2.85–4.06 nm Pa\(^{-1}\) for displacement, 1.99–2.74 \(\mu\)m s\(^{-1}\) Pa\(^{-1}\) for velocity and 2.2–2.86 mm s\(^{-1}\) Pa\(^{-1}\) for acceleration. We do not observe any clear relationship between transfer coefficients and source offset. The energy coupling efficiency varies in the range of 1.42–2.39 percent.

The uncertainty estimates, such as given in Figs 7 and 8, are based on the variability of repeated measurements. It is important to acknowledge that this ignores errors in assumed values used to calculate parameters from the coupling coefficients, which introduce systematic errors. Such errors are especially important for the energy coupling, which depends on a rather uncertain soil density.

5 DISCUSSION

5.1 Transfer coefficients and coupling efficiency

Values for the transfer coefficients and the coupling efficiency that we have obtained are listed in Table 2. These can be compared with earlier results, which either result from observations of sonic booms, or laboratory studies. Table 3 gives this comparison. Due to the rather different frequency content, we restrict the comparison to velocity coupling coefficients, since they physically relate with energy admittance.

We see that the measurements in Table 3 have the same order of magnitude. Given the crude (amplitude-based) measure, one should not overemphasize differences, yet laboratory studies appear to produce higher values than the longer-period natural studies. Our values agree closely with the values from the sonic boom studies. Values of the energy admittance that we find (1.42–2.39 percent) are in excellent agreement with the value of 2.13 percent that Edwards et al. (2007) find.
Figure 6. Refraction experiment: infrasound and seismic measurements of hammering in the vicinity of colocated seismic (R517) and infrasound (HYP03) sensors (data is not filtered). (a) Spectrograms. (b) Normalized and overlaid, on top of each other, seismic and infrasound waveforms. (c) Section plot of stacked hammer shots. Seismic traces are shown in black (vertical components), infrasound traces in red. Different apparent group velocities are indicated on the right. The seismic wave propagates with a velocity of around 1 km s$^{-1}$. The red line indicates the arrival of the acoustic wave, propagating with a velocity of around 0.3 km s$^{-1}$. This acoustic wave is barely visible in the seismic traces. Please note that left y-axis starts with T -0.02 s for the extra space on the right y-axis.

5.2 Nature of the air-guided Rayleigh waves

The later part of the seismic traces shows an entirely different character in the explosion experiment, compared with the refraction experiment. There is a strong and characteristic narrow-band signal following the initial acoustic wave. The apparent velocity of this wave is clearly very close (and probably identical) to that of the acoustic wave. Similar observations have been made by several authors (Haskell 1951; Press & Ewing 1951), and attributed to air-coupled Rayleigh waves.

Langston (2004) has modelled the phenomenon, and he showed that the waveform character of the air-coupled Rayleigh depends

strongly on details of the subsurface velocity model, for example the presence of the non-surface low-velocity layer and its velocity details. A slight change in that velocity profile may invert the (retrograde or prograde) polarization, and it may strongly enhance the wave, or render it completely invisible. There is much information about subsurface structure contained in the air-coupled Rayleigh wave, and this (occasionally very strong) wave may be contributing greatly to the general complexity of the appearance of seismo-acoustic wave trains—which may sometimes be too quickly attributed to other factors such as building resonances, topography, or the type of instrumentation.

At longer periods, the intersection of Rayleigh–Lamb wave branches probably plays a role (Lognonné et al. 2016), causing mode
coupling between acoustic (Lamb) and surface wave (Rayleigh) modes. At shorter periods, acoustic signals are sufficiently broadband to match resonance frequencies of subsurface structure. Air-coupled Rayleigh waves probably consist of narrow higher-order Airy phases (Edwards et al. 2007; Haskell 1951). To explain these phenomena Kanamori et al. (1992) had suggested reverberations in thick sedimentary basins, but (Langston 2004) explains convincingly that the resonances occur within shallow low-velocity layers.
frequencies (for $\nu$ is 12.46 MPa at the lowest frequencies and 5.30 MPa at the highest soills they obtain values in the range 0.3–0.4. The Young modulus the obtained values of

Now we turn to the question: which seismic velocity corresponds to

5.3 Inferences on the subsurface structure

Remarkably, one obtains an exact solution for the Lamé constants, by recording the pressure, as well as vertical and horizontal ground motions. The values that we have inferred show a gradual variation with frequency, which indicates that these values are not random. Towards high frequencies, they decrease to smaller values, which corresponds to a ‘softening’ of the material towards the surface. It is also worth noting that our technique is approximate (e.g. by ignoring the air-coupled Rayleigh wave), but it leads to a simple procedure that does require extensive synthetic calculations, which would require much knowledge of the subsurface, geometry, etc. At the same time, the results agree with synthetic modelling presented in Langston (2004), where they can be compared (their fig. 16c). It is clear from his modelling that the detailed seismic expression can be complex, especially in the latter part of the wave (‘air-coupled Rayleigh wave’).

Maximum values of $\lambda$ are around 340 MPa, and they decrease with increasing frequency, by two orders of magnitude. $\mu$ has a similar gradual variation with frequency; maximum values are around 100 MPa. Toward high frequencies, there is also a strong decrease of $\mu$, but slightly weaker than for $\lambda$. Values for the Poisson ratio would in principle be possible in the entire range between 0 and 0.5. The values which we obtain are in a much narrower range between 0.33 and 0.36 although. This range corresponds well to the type of soil at the experiment site. Indeed, our values are comparable with those stated in Bowles et al. (1996): for cohesionless, medium and dense soils they obtain values in the range 0.3–0.4. The Young modulus is 12.46 MPa at the lowest frequencies and 5.30 MPa at the highest frequencies (for $\nu = 0.336$). The first value is a typical value for clay soil and the latter for clay or loose sand. These values seem quite reasonable for our geological setting of breccia that is a bit more compacted with depth. The bulk modulus is at 141 MPa at 47 Hz and 5.2 MPa at 341 Hz. These values are much lower than those which one would get for solid rock.

5.4 Near-surface layer and depth resolution

Now we turn to the question: which seismic velocity corresponds to the obtained values of $\mu$ and $\lambda$. For this purpose, we use the density value which we have assumed before and use

This gives us values for $V_p$ of 303 m s$^{-1}$ at 47 Hz and 62 m s$^{-1}$ at 341 Hz. Those values are much lower than of 1000 m s$^{-1}$ that we have obtained from the refraction experiment. This suggests that the surface layer has a lower velocity than the layer below, due to perhaps looseness of sediments and/or saturation with water, and our refraction experiment has overlooked it.

If the solid half-space has a faster shear wave velocity than the air, that wave will leak into the air. So in that case it would technically be a leaky wave, not a guided wave. On the other hand, if the shear wave velocity of the halfspace is slower than air, it will not leak into the air. This may explain why the Rayleigh wave appears after the Stoneley wave in Fig. 5b). This later arrivals may as well be scattered waves.

Indeed, one can observe in Fig. 6 that the onset for the first 5 m offset has a different apparent velocity, but the waves did not separate into isolated phases yet. Also, seismic and acoustic waves overlap, which makes it difficult to determine a seismic velocity for the near-surface layer from the refraction experiment alone. The result from the low-frequency Lamé constants suggests a seismic velocity that is quite close to the acoustic velocity. That similarity may explain why the air-coupled Rayleigh wave is present so strongly in this study (by a strong acoustic-seismic coupling). The higher frequency Lamé constants indicate lower seismic velocities, probably corresponding to a near-surface layer of loose unconsolidated sediments.

As the Lamé constants are frequency-dependent, one may in principle determine the depth to which they correspond to from the depth range which the waves are sensitive to. The decay of the surface wave amplitude with depth can be written approximately (see Ben-Menahem & Singh 1981; Tanimoto & Wang 2018) as

$$A = A_0 e^{-2zf/\rho}.$$  \hspace{1cm} (19)

where $A_0$ is initial amplitude, $A$ is amplitude, $f$ is frequency [Hz] and $z$ is depth [m]. We may thus assume that the character of a wave, at a given frequency $f$, depends on properties of the material down to a depth, where $A/A_0$ has decayed to 1/e. This suggests depth ranges down to 3 m for $f = 48$ Hz and down to 50 cm for $f = 348$ Hz. These values are to be understood as maximum depth though, and most of the sensitivity is to shallower depths. It is quite likely that the material in the topmost meter dominates the character of the elastic parameters at all frequencies.

5.5 Wider context and implications

Windows of strong seismic noise (or other wave phases such as air-guided Rayleigh waves) can be avoided, therefore. An analog to our problem exists in marine seismics, where water-guided waves can appear; they are called Scholte waves (Scholte 1947). These waves have found some use for inverting for subsurface structure (e.g. Boiero et al. 2013).

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Table 1. Peak envelope of ground motion measures: displacement (DISP), velocity (VEL), acceleration (ACC) and pressure (PRES) for various offsets and source types. Rockets are fired at 10 m offset and ≈40 m height.

| Ground measure | Offset (m) | Explosives | Rockets |
|----------------|-----------|------------|---------|
| DISP (µm)      | 1.37      | 0.413E     | 0.143   |
| VEL (mm s$^{-1}$) | 1.15      | 0.264      | 0.125   |
| ACC (m s$^{-2}$) | 1.71      | 0.446      | 0.174   |
| PRES (Pa)      | 240       | 143        | 50.1    |
| Signal duration (s) | 0.025 ± 0.001 | 0.05 ± 0.001 |

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as the signal gets ‘trapped’ in the layer and undergoes P–SV con-
versions multiple times.

\[ f(z) = \frac{1}{\sqrt{\rho}} \left[ \frac{1}{\lambda + 2\mu} \right] \]

\[ V_p = \sqrt{\frac{\lambda + 2\mu}{\rho}} \]

\[ V_s = \frac{1}{C} \sqrt{\frac{\mu}{\rho}} \]
One may hope to dig more deeply into the complex seismo-acoustic waves and infer useful properties from them. The ground motion directly associated with the acoustic wave seems to be a useful point of entry into this topic. So far, the waveform character of such waves has rarely been studied, probably because the details of the waveform have been deemed to be too complex. Inferences have so far been relatively crude generally, for example specifying coupling constants, etc. What we have presented in this paper may
be as close as one gets to ‘a simple conversion formula on the back of an envelope’, without doing extensive calculations of synthetic seismograms.

Besides, the acoustic wave can be interpolated easily between a set of infrasound sensors, since its velocity is well-known. With this, the analysis can be done also at sites that are equipped with a 3-component seismometer only, but not an infrasound sensor. Our method can thus be extended to 3-D, even if the number of collocated station pairs are limited.

6 CONCLUSIONS

We have observed that acoustic and air-coupled Rayleigh waves dominate the seismic data of an explosion experiment. The wave character of the seismic wavefield depends strongly on which kind of source is used: hammering creates $P$ waves as well as Rayleigh waves (ground roll), while an explosion produces acoustic waves, followed by air-coupled Rayleigh waves.

It turns out that one can constrain near-surface properties by comparing collocated pressure and seismic sensors, and we have used the acoustic wave to directly infer elastic parameters of the near-surface material. The techniques described in this paper are probably of value for engineering purposes.

We have also provided information on how much energy couples from the acoustic wave into the ground. We have calculated transfer coefficients and the energy coupling efficiency for small explosions (gram-scale NEM) and very small offsets (1–30 m for fixed height firecracker explosions on the ground level and 10 m horizontal offset for rocket explosions). Data are available at frequencies up to 500 Hz, which allowed us to investigate higher frequency modes probably of value for engineering purposes.

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7 CODE AND DATA AVAILABILITY, RESOURCES

All waveform data used in this study is freely available for download at the European Integrated Data Archive (EIDA) at https://www.efdsn.org/networks/detail/6A_2019/ (last accessed June 2020) using the network code 6A (Fuchs et al. 2019a).

Data processing and analysis was done using Python 3.7.3 (van Rossum 1995), Pandas 0.25.1 ( McKinney 2010), NumPy 1.15.14 ( van der Walt et al. 2011), ObsPy Toolbox 1.1.1 (Krischer et al. 2015), PyProj 1.9.5.1 (Snow et al. 2019), IPhython 7.8.0 (Pérez & Granger 2007) and SciPy 1.0 (Virtanen 2020). Additional seismic processing was done in Seismic Unix 43R1 (Stockwell & John 1999). Maps and figures were produced with Plotly 4.2.1 and Matplotlib 3.1.1 (Hunter 2007).

Meteorological data from the MetLift and all code to reproduce the results of this work is available at https://github.com/IMGW-univie/ground-coupling (last accessed June 2020).

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Table 2. Mean coupling transfer coefficients (CTC) for displacement (DISP), velocity (VEL), and acceleration (ACC) as well as mean energy coupling efficiency (ESE) calculated for different offsets. Error is a standard error of the mean described in Kokoska & Zwillinger (2000).

| Ground measure | Offset (m) | 0–10 | 10–20 | 20–30 |
|----------------|-----------|------|-------|-------|
| DISP CTC       | 2.85e-09 ± 1.24e-10 | 3.08e-09 ± 1.08e-10 | 4.06e-09 ± 5.00e-10 |
| VEL CTC        | 1.99e-06 ± 7.64e-08 | 2.26e-06 ± 5.58e-08 | 2.74e-06 ± 1.72e-07 |
| ACC CTC        | 2.20e-03 ± 7.74e-05 | 2.57e-03 ± 7.48e-05 | 2.86e-03 ± 1.83e-04 |
| ESE            | 2.39 ± 0.39 per cent | 1.96 ± 0.28 per cent | 1.42 ± 0.17 per cent |

Table 3. Comparison of coupling coefficients for various studies.

| Measurement type | Frequency | Velocity coupling coefficient (in μm s⁻¹ Pa⁻¹) | Soil type |
|------------------|-----------|-----------------------------------------------|-----------|
| Sonic booms from aircrafts (McDonald & Goforth 1969) | 0.2 Hz | 0.4–3.5 | Basement rock |
| Sonic boom from Stardust reentry (Edwards et al. 2007) | 2 Hz | 7.3 | Clay-rich playa |
| Sonic boom from Stardust reentry (Edwards et al. 2007) | 2 Hz | 1.99–2.74 | Volcanic deposits |
| 8 March 2005 Mount St.Helens (Matoza & Fee 2014) | 7–12 Hz | 6–7 | Silty loam |
| Laboratory (Bass et al. 1980; Sabatier et al. 1986) | 20–25 Hz | 7–8 | Loess |
| Laboratory (Bass et al. 1980; Sabatier et al. 1986) | 50 Hz | 14 | Sandy soil |
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**APPENDIX A: WEATHER DATA**

*Figure A1.* Representation of the METLIFT meteorological data for the duration of the experiment. Given are temperature (in deg C), relative humidity, air pressure (in hPa), wind speed (in m s⁻¹) and direction. Time is in UTC.
APPENDIX B: ROCKETS

Fig. B1 shows a pronounced decrease in frequency for the signal that is generated during the lift-off phase of the rocket. Within 0.5 s the dominant frequency drops from above 200 Hz to approximately 100 Hz. The source of this signal is the sound emitted by the firing of rocket fuel during the lift-off phase of the rocket. The rockets are heavily accelerating during this phase. Thus, we expect to measure a frequency shift caused by the Doppler effect, as the source of sound is moving away from the receiver at accelerating velocities. However, assuming a constant source frequency of 200 Hz, taking the speed of sound as 330 m s\(^{-1}\) and assuming a final rocket velocity of 40 m s\(^{-1}\) (= 144 km hr\(^{-1}\)) the expected Doppler shift is less than 30 Hz. Consequently, a drop in the frequency of more than 100 Hz cannot solely be explained by a moving source. We suspect that the source frequency emitted by the burning of fuel inside the small fuel chamber changes as fuel is consumed. A gradually lowered source frequency in combination with the Doppler effect may explain the frequency drop we observe.

**Figure B1.** Rocket experiment: Infrasound and seismic measurements of rocket explosion at colocated seismic (R525) and infrasound (HYP04) station (as in Fig. 5). Data were filtered between 50 and 499 Hz with minimum-phase bandpass filter. Signal is separated for two stages: lift-off on the left and explosion on the right.
Figure C1. (a) and (b) Particle motion for explosion 1–2 recorded at the station R509. The azimuth of arriving signal: $90 \pm 0.6^\circ$, third-order Butterworth minimum-phase band-pass filter applied between 10 and 400 Hz. View (a) demonstrates the 0.03 s interval of particle motion during the passage of the acoustic wave. One can observe an elliptical prograde particle motion indicated by the black arrow. View (b) presents the 0.17 s interval immediately following passage of the acoustic wave. One can observe retrograde elliptical motion, indicated with the black arrow, polarized along the direction of the incident wave or an air-coupled Rayleigh wave (as observed in e.g. Ewing 1957). Colours and opacity represent time: from opaque blue at $\Delta T_0$ to saturated red at $\Delta T_{end}$. (c) Comparison of the responses to the explosion in terms of displacement (top and middle view) and pressure changes (bottom view). Data is not filtered. Radial displacement is more impulsive then vertical, which agrees with the suggestion of (Langston 2004).
# APPENDIX D: ABBREVIATIONS USED

Table D1. List of abbreviations used in the text.

| Abbreviation | Name                          | Units       |
|--------------|-------------------------------|-------------|
| $p$          | pressure                      | Pa          |
| $T$          | air temperature               | K           |
| $Z$          | compressibility factor        |             |
| $R$          | molar gas constant            |             |
| $\gamma$     | specific heat ratio           |             |
| $B$          | second virial coefficient     |             |
| $x_v$        | mole fraction of water vapor  |             |
| $M_d$        | molar mass of dry air         | g mol$^{-1}$|
| $M_w$        | molar mass of water           | g mol$^{-1}$|
| $M_{mix}$    | the molecular mass of air and vapor mixture | g mol$^{-1}$|
| $\rho_{air}$ | density of air                | kg m$^{-3}$ |
| $\rho_{dry}$ | wet bulk density              | g cm$^{-3}$ |
| $\rho_{wd}$  | dry bulk density              | g cm$^{-3}$ |
| $\rho_g$     | matrix (grain-mineral) density| g cm$^{-3}$ |
| $\rho_f$     | density of a fluid            | g cm$^{-3}$ |
| $\rho_{soil}$| density of soil               | kg m$^{-3}$ |
| $\phi$       | the porosity                  | Decimal per cent |
| $U_x$        | horizontal ground displacement| m           |
| $U_z$        | vertical ground displacement  | m           |
| $P_0$        | total interface pressure      | Pa          |
| $\omega$     | angular frequency             | rad s$^{-1}$|
| $\lambda$    | first Lamé constant          | Pa          |
| $\mu$        | second Lamé constant         | Pa          |
| $\nu$        | Poisson ratio                 |             |
| $x$          | offset                        | m           |
| $t$          | time                          | s           |
| $z$          | depth                         | m           |
| $V_{air}$    | apparent velocity in the air  | m s$^{-1}$  |
| $V_p$        | P-wave velocity               | m s$^{-1}$  |
| $V_s$        | S-wave velocity               | m s$^{-1}$  |
| $v_{soil}$   | particle motion velocity in the soil layer | m s$^{-1}$ |
| $V_{mair}$   | molecular velocity in the air | m s$^{-1}$  |
| $A_{vair}$   | recorded vertical seismic amplitude at the moment of incident |             |
| $A_{pressure}$| the recorded acoustic amplitude at the moment of incident |             |
| $A$          | amplitude                     |             |
| $A_{final}$  | amplitude at the amplitude threshold |             |