Probing gravity waves in the middle atmosphere using infrasound from explosions

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Key Points:

- Ground-based infrasound recordings of explosions are used to retrieve effective sound speed fluctuations in the mesosphere
- Vertical wave number spectra of the retrieved fluctuations agree with the "universal" gravity wave saturation spectrum
- Infrasound from 49 explosions and radar data show that remote sensing of the middle atmosphere is possible via ground-based infrasound data

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Abstract

This study uses low-frequency, inaudible acoustic waves (infrasound) to probe wind and temperature fluctuations associated with breaking gravity waves in the middle atmosphere. Building on an approach introduced by Chunchuzov et al., infrasound recordings are used to retrieve effective sound-speed fluctuations in an inhomogeneous atmospheric layer that causes infrasound backscattering. The infrasound was generated by controlled blasts at Hukkakero, Finland and recorded at the IS37 infrasound station, Norway in the late summers 2014 – 2017. Our findings indicate that the analyzed infrasound scattering occurs at mesospheric altitudes of 50 – 75 km, a region where gravity waves interact under non-linearity, forming thin layers of strong wind shear. The retrieved fluctuations were analyzed in terms of vertical wave number spectra, resulting in approximate k_z^{-3} power law that corresponds to the "universal" saturated spectrum of atmospheric gravity waves. The k_z^{-3} power law wavenumber range corresponds to vertical atmospheric scales of 33−625 m. The fluctuation spectra were compared to theoretical gravity wave saturation theories as well as to independent wind measurements by the Saura medium-frequency radar near Andøya Space Center around 100 km west of IS37, yielding a good agreement in terms of vertical wavenumber spectrum amplitudes and slopes. This suggests that the radar and infrasound-based effective sound-speed profiles represent low- and high-wavenumber regimes of the same "universal" gravity wave spectrum. The results illustrate that infrasound allows for probing fine-scale dynamics not well captured by other techniques, suggesting that infrasound can provide a complementary technique to probe atmospheric gravity waves.

Plain Language Summary

This study analyzes inaudible acoustic waves (infrasound) detected in Norway following explosions during disposal of military equipment in Finland. We show that infrasound reflects off small-scale structures in the middle atmosphere (within 50-75 km altitude) and we use signals recorded to retrieve so-called effective sound-speed profiles, a proxy of small-scale variations in temperature and horizontal wind. Spectral analysis of the retrieved altitude profiles reveals a power law associated with gravity waves. Such waves are important in the transfer of energy between atmospheric layers and are generated, for example, by upward air flow over mountain ranges. The vertical scales to which infrasound is sensitive to, are estimated to range from 33 to 625 m. Comparisons between spectra obtained using radar and infrasound show good agreement in terms of amplitudes and slopes. This suggests that the radar and infrasound-based effective sound-speed profiles represent different regimes of the same "universal" gravity wave spectrum. This study uses a large, consistent infrasound dataset and independent radar data to show that remote sensing of fine-scale wind and temperature variations in a region of the middle atmosphere for which very few observations are available, is possible by means of ground-based infrasound measurements.

1 Introduction

This study investigates the use of acoustic waves to probe fine-scale wind and temperature structures of the middle atmosphere (i.e. stratosphere and lower mesosphere). Atmospheric infrasound, i.e. low-frequency sound waves in the inaudible frequency range (< 20 Hz) can be generated by both natural (e.g., volcanoes, earthquakes, thunder) and artificial (e.g., rocket launches, sonic booms, blasts) sources. Once generated, infrasound waves can propagate in the atmosphere over long distances as the energy is ducted by waveguides formed by vertical gradients in temperature and wind (Brekhovskikh, 1960; Diamond, 1963). In addition to the source characteristics, infrasound waves also provide information about the medium through which they propagate, and can therefore serve as a tool for atmospheric remote sensing (e.g., Le Pichon et al., 2005; Assink et al., 2019; Smets & Evers, 2014; Chunchuzov et al., 2022).

Probing the middle atmosphere by means of ground- and space-based remote sensing techniques contributes to a better representation of this region in atmospheric models. The latter allows for improved weather forecasts due to the dynamical coupling between different atmospheric layers (Shaw & Shepherd, 2008). The resolution of the atmospheric model products, and therefore the scales of atmospheric processes resolved, strongly depends on available computational capabilities and the scientific problem. For example, high-resolution limited-area models routinely in use at national meteorological services (e.g., Bengtsson et al., 2017) have high horizontal resolution of several kilometers, however, the model top is typically in the lower stratosphere ($\sim$ 10 hPa, or 30 km). In contrast, global numerical weather prediction models (NWPs) and general circulation models (GCMs) with model tops raised into the mesosphere and above (Stocker et al., 2014) have lower resolution and are unable to resolve atmospheric processes at scales smaller than 10 kilometers in operational NWP (Bauer et al., 2015) and tens of kilometers in GCMs (H.-L. Liu et al., 2014; Becker et al., 2022). While not fully resolvable by models, these subgrid-scale processes can be observed by various observational techniques, including radar, lidar and rocket measurements (Rapp & Lübken, 2004; Le Pichon et al., 2015; Schäfer et al., 2020; Strelnikov et al., 2019).

One such subgrid-scale phenomenon is atmospheric gravity waves (GWs). Generated in the lower atmosphere, GWs propagate into the middle atmosphere with increasing amplitude due to the decrease in air density with altitude, until they ultimately become unstable and break. When breaking, GWs generate small-scale eddies or turbulence which in turn interact with other atmospheric waves (Fritts & Alexander, 2003). The transfer of energy and momentum between different atmospheric layers is an important function of atmospheric waves. For example, the middle atmospheric meridional circulation is primarily GW-driven (Fritts & Alexander, 2003) and breaking mesospheric GWs play an important role in the wintertime polar stratospheric downward motion (Garcia & Boville, 1994; Wicker et al., 2023). Momentum deposited by GWs (or GW drag) can modify atmospheric circulation patterns at lower altitudes, therefore affecting the weather and its prediction (McFarlane, 1987). This highlights the need for GW probing and for improvement of GW representation in NWP and GCMs. Efforts are also being made to develop GW-resolving GCMs stretching up to the edge of the thermosphere (e.g. H.-L. Liu et al., 2014; Becker et al., 2022).

GWs interact with other atmospheric waves in various ways, including wave-wave interaction and wave-breaking (Fritts & Alexander, 2003), and cause the presence of localized, three-dimensional small-scale fluctuations in temperature and wind fields. These have been observed in the middle atmosphere by in-situ, ground- and space-based instruments (e.g., Fritts & Alexander, 2003; Tsuda, 2014; Selvaraj et al., 2014; Bossert et al., 2015; Miller et al., 2015; Podglajen et al., 2022). The vertical scales of these fluctuations are significantly smaller than the horizontal scales, and have characteristic vertical length scales ranging from tens of meters to tens of kilometers (Gardner et al., 1993). The presence of such small-scale atmospheric fluctuations is known to affect propagation and scattering of infrasound waves (Chunchuzov & Kulichkov, 2020). Moreover, it has been demonstrated by Bertin et al. (2014) and Lalande and Waxler (2016) that infrasound waveguides are very sensitive to GW induced small-scale fluctuations in wind and temperature (see also Brissaud et al. (2023)). This implies the importance of accounting for fine-scale atmospheric structures when modelling infrasound propagation (Drob et al., 2013; Hedlin & Drob, 2014; Chunchuzov et al., 2022). On the other hand, this also suggests that infrasound observations can be used to probe small-scale atmospheric fluctuations, thereby addressing the need for an enhanced observations of GWs (Cugnet et al., 2019).

The purpose of the current study is to quantify GW activity using a dataset of infrasound recordings from distant ground-based explosions. These signals have been recorded at a ground-based microbarometer array in Norway, every day during the period of mid-August to mid-September for the years 2014-2017. We apply a method that allows for the retrieval of so-called effective sound speed fluctuations in an inhomogeneous layer in the middle atmosphere. The method was developed over several years by Chunchuzov (2002); Chunchuzov et al. (2013, 2015, 2022); Chunchuzov and Kulichkov (2020). Based on the retrieved effective sound speed fluctuations for each event, we calculate the corresponding vertical wavenumber spectrum, and further interpret this in terms of power spectral density (PSD) slope and amplitude. The retrieved GW spectra are further compared to independent wind radar observations as well as to both linear and non-linear theoretical GW saturation models (Dewan & Good, 1986; S. A. Smith et al., 1987; Chunchuzov et al., 2015).

We exploit an infrasound dataset of signals generated by ground-based blasts in Hukkakero, Finland. These signals are detected at 321 km distance from the source, at microbarometer array IS37 in Northern Norway. This dataset has several attractive features making it suitable for atmospheric probing studies. First, the explosive events take place during August and September which is during the atmospheric transition from summer to winter, when the zonal component of the stratospheric winds reverses from westward to eastward (Waugh & Polvani, 2010; Waugh et al., 2017). Second, the known locations of the source and receiver together with the transient nature of the blasts make it possible to clearly identify arrivals from both stratospheric and from mesospheric – lower thermospheric (MLT) altitudes. Finally, yet importantly, the recurring nature of explosive events allows us to study day-to-day variability of the middle atmosphere dynamics.

The paper is organized as follows. A background on infrasound sensitivity to atmospheric structure, infrasound signal processing terminology, and previous studies exploiting Hukkakero explosion-related data is provided in Sect. 2. Section 3 describes the infrasound dataset, signal pre-processing, the SD-WACCM-X atmospheric model used, and the ray-tracing simulations conducted. Its subsection 3.4 elaborates the effective sound speed retrieval methodology. The obtained results are shown in Sect. 4, also further discussed in Sect. 5 including vertical wavenumber spectrum comparison to independent radar measurements and theoretical models.

2 Background

2.1 Sensitivity of infrasound to atmospheric structure

Infrasound propagation is sensitive to spatial variations in temperature and wind (e.g., Waxler & Assink, 2019). In the direction of propagation, the wind and temperature related propagation effects can approximately be modelled using the concept of effective sound speed, $C_{\text{eff}}(z)$, defined as:

$$C_{\text{eff}}(z) = \sqrt{\gamma R T} + \mathbf{u} \cdot \hat{n}, \tag{1}$$

where, γ, R, T, $\mathbf{u}$ and $\hat{n}$ correspond to the adiabatic index, the gas constant, the absolute temperature, the horizontal wind speed vector and the direction of propagation, respectively. In the infrasound-related context, it is often appropriate to approximate $\sqrt{\gamma R} \approx 20\,\mathrm{m\,s^{-1}K^{-1/2}}$. For cases where ground-to-ground propagation is of interest, it is convenient to introduce the effective sound speed ratio, which is obtained by normalizing $C_{\text{eff}}(z)$ by its value on the ground and which is analogous to the more familiar refractive index. From classical ray theory, acoustic signals that originate from the ground are expected to traverse in waveguides between the ground and the altitudes for which the C_{eff} ratio exceeds unity.

The celerity is defined as the source-receiver great-circle distance divided by the infrasound travel time (i.e., the difference between the arrival time and origin time). The celerity can hence be considered as the average group speed of a guided acoustic wave. When the origin time and location are known, celerity-based models can be used to provide information about the infrasound waveguide through which an acoustic wave propagated. Infrasonic paths with a substantial vertical component have a group speed that is significantly lower than the speed of sound. Conversely, infrasound guided by tropospheric waveguides (that propagates in the troposphere) has a celerity near the local sound speed. Typical celerities for different waveguides are $310-330$ m/s for tropospheric arrivals, $280-320$ m/s for stratospheric arrivals, and $180-310$ m/s for mesospheric and thermospheric arrivals (e.g., Nippress et al., 2014; Lonzaga, 2015).

2.2 Infrasound array processing

An infrasound array is a group of microbarometers distributed in space but installed close enough so that the received sensor signals are sufficiently coherent to estimate the wavefront parameters of the dominant plane wave arriving at the array. This is done using array signal processing techniques that delay and sum sensor traces according to a model for the inter-element delays. This spatial filtering allows for reducing incoherent noise and for separating acoustic signals from different directions of arrival. Identification of the signals of interest is typically based on the observed back-azimuth, apparent velocity, and average inter-sensor coherence. The back-azimuth represents the direction from which the plane wave arrives at the array and is measured in degrees clockwise from the North. The apparent velocity is the velocity the plane wave appears to travel at horizontally along the array. This parameter is estimated based on the time delays between sensors (as well as back-azimuth) and contains information about the angle of incidence θ of the plane-wave, $v_{\text{app}} = c/\sin\theta$ where c is the local sound speed. There is no unique relationship between apparent velocity and altitude from which signal arrives, however higher values of apparent velocity would normally indicate arrival from higher altitudes. The combination of back-azimuth and travel time allows for signal identification and infrasound source location, while v_{app} helps to identify the incidence angle of the ray-path at the ground.

2.3 The Hukkakero blasts in infrasound studies

The site of Hukkakero, Finland (67.94° N, 25.84° E; Fig. 1), has been of particular interest for infrasound related studies over the past years. At Hukkakero, blasts related to the disposal of military explosives occur yearly since 1988 in August-September, typically once a day with a yield of around 20 tons of TNT equivalent (Gibbons et al., 2015). In addition to generating an atmospheric pressure wave, these explosions produce clear seismic signals which allow for the accurate estimation of origin time and location by means of seismic localization techniques (Gibbons et al., 2020). Blixt et al. (2019) showed that the ARCES seismic array in northern Norway records, besides the seismic waves also the ground-coupled airwaves associated with Hukkakero explosions. These explosions are also well-represented in event bulletins like the comprehensive European Infrasound Bulletin (Pilger et al., 2018, Fig. 10), as well as in the Comprehensive Nuclear-Test-Ban Treaty (CTBT) bulletin products.

Infrasound signals that originated from Hukkakero explosions have been exploited in several atmospheric probing studies. Blixt et al. (2019) analyzed 30 years of Hukkakero explosions detected at the ARCES/ARCI seismo-acoustic array (Norway) in terms of back-azimuth deviation due to cross-wind (the component of wind perpendicular to the direction of propagation) influence along the propagation path. The resulting cross-wind estimates obtained showed a good agreement with the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA)-Interim model. Amezcua et al. (2020) presented a way to implement an off-line assimilation of infrasound data into atmospheric

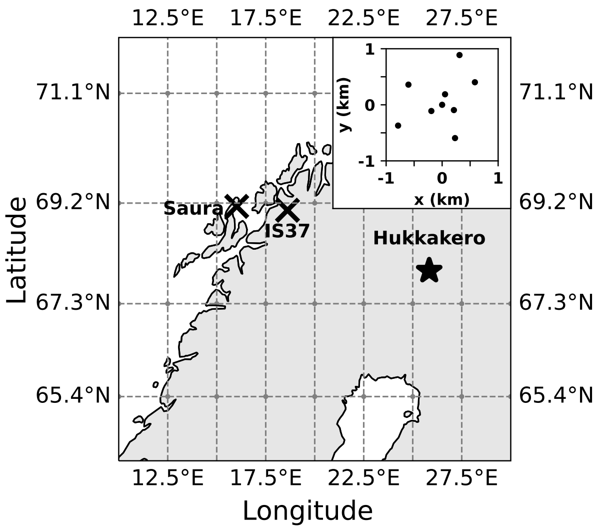

Figure 1. Location of all sources of data used in this study: Hukkakero explosion site, IS37 infrasound array, and Saura medium-frequency radar. The SD-WACCM-X atmospheric model grid is displayed on the map as gray dashed lines. The IS37 array layout is shown in the inset.

models using Ensemble Kalman filters. The study extends the approach by Blixt et al. (2019), demonstrating that assimilation of back-azimuth deviation allows for corrections to atmospheric winds at tropospheric and stratospheric altitudes. Based on the same dataset, Vera Rodriguez et al. (2020) developed an extended inversion methodology that uses infrasound observations to update atmospheric wind and temperature profiles on the basis of the ERA5 re-analysis ensembles.

Still, Hukkakero related infrasound signals have not previously been used to probe small-scale atmospheric inhomogeneities.

3 Materials and Methods

3.1 Infrasound dataset and signal pre-processing

This study exploits Hukkakero explosions and the associated signals recorded at infrasound array IS37 that is located at ~ 320 km distance in Bardufoss, Norway (69.07° N, 18.61° E; Fig. 1). This 10-element array is part of the International Monitoring System (IMS) for the verification of the CTBT (Marty, 2019). The region is also host to a cluster of additional seismo-acoustic monitoring stations (Gibbons et al., 2015). During the years $2014 - 2017$, 57 explosions took place at Hukkakero, however 8 of them (the three last explosions in 2014 and the five last explosions in 2016) were significantly weaker (Gibbons et al., 2015) and are therefore not considered in the current study. Origin times of the analyzed 49 explosions are tabulated in Appendix A.

For each explosion, the back-azimuth and apparent velocity of the dominant wavefront were estimated using a conventional time-domain array processing technique (Melton & Bailey, 1957). The detection of coherent infrasound over the array is based on the evaluation of the so-called Fisher ratio. The Fisher ratio corresponds to a probability of detection of a coherent signal with a specific signal-to-noise ratio (SNR). The associated inter-element time-delays are used to form the so-called best-beam, for which the individual array recordings are time-aligned before summation. Details on the particular im-

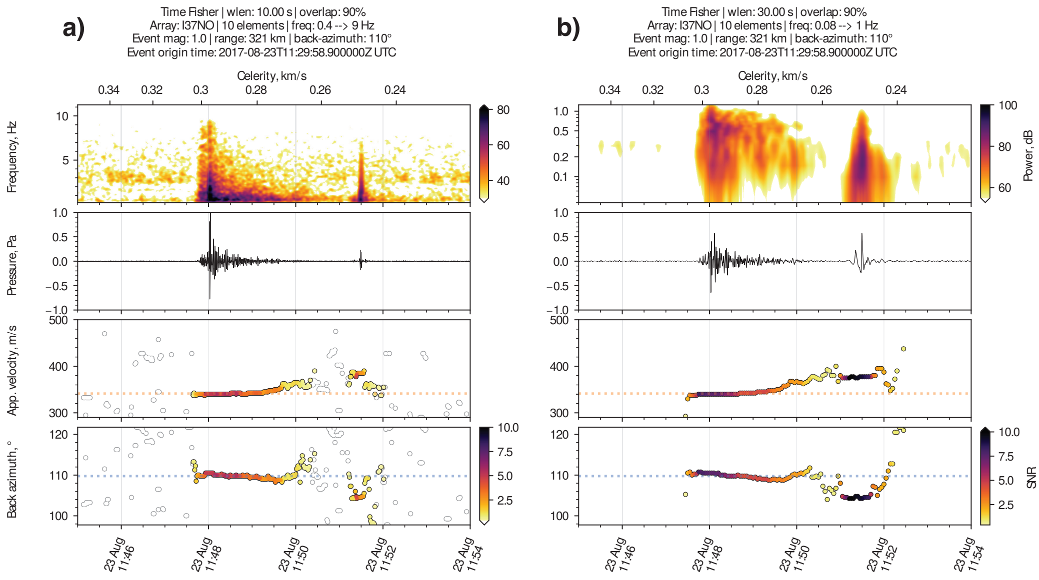

Figure 2. Array processing results for a Hukkakero explosion on 23 August 2017, processed between a) 0.4 – 9 Hz and b) 0.08 – 1.0 Hz. Top panel: spectrogram displayed in decibel. Second panel: the best-beam trace with an orange dashed line indicating the sound speed on the ground ($\approx$ 340 m/s). Third panel: apparent velocity. Bottom panel: the back-azimuth, where the blue dashed line corresponds to the great-circle back-azimuth (110°) towards Hukkakero.

plementation can be found in Evers (2008). The beam waveforms were processed in two partly overlapping frequency bands to highlight the key trace features, 0.4–9 Hz and 0.08 – 1.0 Hz. Figure 2 shows array analysis results for one explosion filtered in both frequency bands. Note, the contribution of ocean ambient noise ("microbaroms") around 0.2 Hz (Vorobeva et al., 2021; De Carlo et al., 2020) and wind noise at low frequencies is negligible compared to the explosion contributions.

Fig. 3 shows a compilation of IS37 infrasound signals from the 49 explosions exploited in the current study. The first arrivals are detected between 17.5–19 minutes (celerity of 281–314 m/s) after the explosion (Fig. 3a) and feature energy in a broad frequency band (Fig. 2a). Typically, the waveform consists of a main arrival with a significantly larger amplitude, followed by a coda ("tail") with progressively increasing apparent velocity with values within the 340–360 m/s. These ranges of celerities and apparent velocities are typical for stratospheric arrivals (Nippress et al., 2014; Lonzaga, 2015) which generally refract or reflect near the stratopause. Similarly extended wave trains have been observed in far-field infrasound recordings following large detonations (Fee et al., 2013; Lalande & Waxler, 2016; Green et al., 2018), and it was assumed that these wave trains originate from interactions with atmospheric perturbations caused by GWs.

After this first wave train, a later arrival can in many cases be observed between approximately 20–23 min after the explosion (a celerity range of 232–267 m/s). Figs. 2b and 3b show the signals in a pass-band between 0.08 – 1.0 Hz. This arrival is characterized by a low-frequency U-shaped waveform, has higher apparent velocity values (i.e., $>$ 360 m/s) and larger back-azimuth deviations compared to the first arrival. All of these characteristics are typical of arrivals returning from the lower thermosphere (Le Pichon et al., 2005; Assink et al., 2012, 2013; Green et al., 2018; Blom & Waxler, 2021).

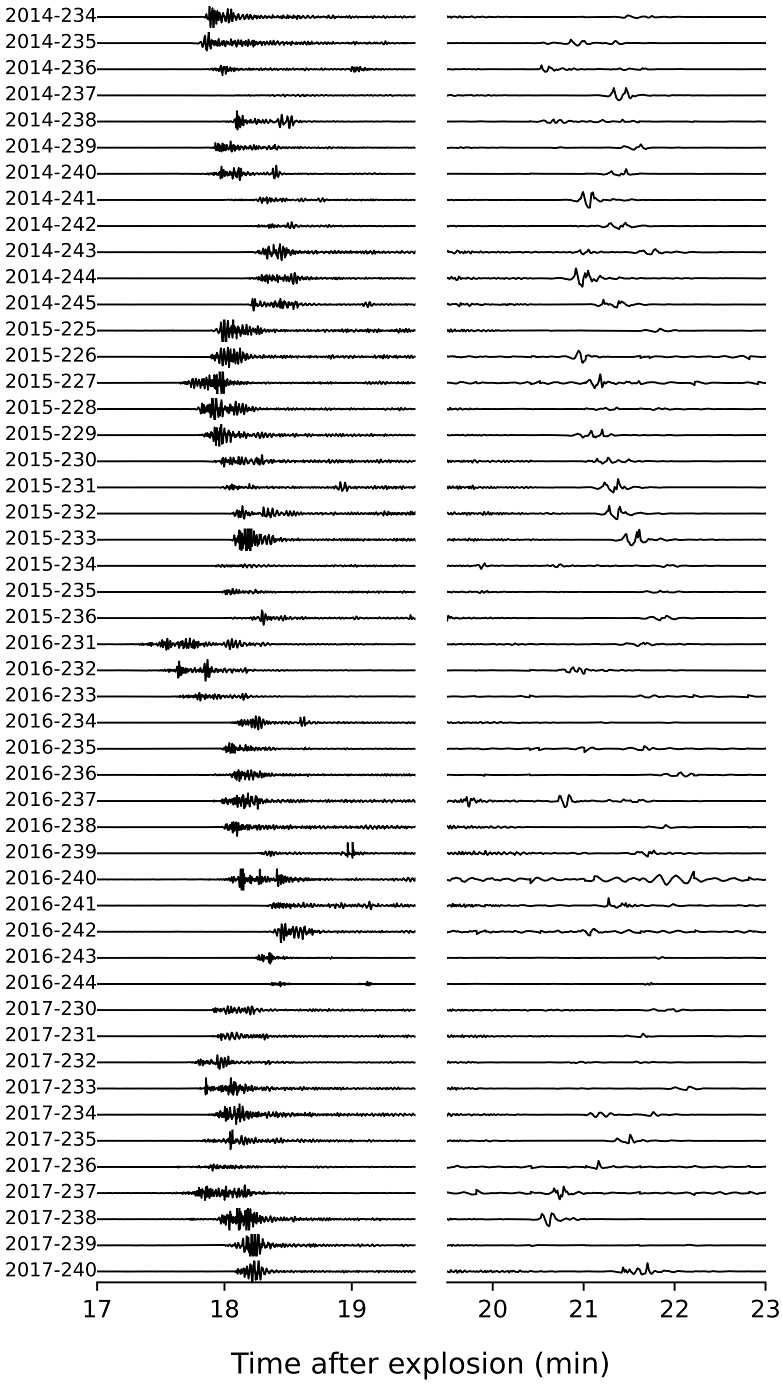

Figure 3. Infrasonic signals from 49 Hukkakero explosions that occurred in the time period 2014-2017. The signals have been recorded at infrasound array IS37 between (left) 17 − 19.5 minutes and (right) 19.5 − 23 minutes. The data are band-pass filtered between (left) 0.4 − 9 Hz and (right) 0.08 − 1 Hz. The y-axis of each trace has ± 1 Pa limit. The left-hand side labels display the year and the day-of-year when events took place.

A closer look at Figure 3 further reveals that several of the events feature an arrival between the stratospheric and thermospheric arrivals, see also Gibbons et al. (2019, Fig. 10.7). Although the current study only exploits the stratospheric arrivals for atmospheric probing, it is worth noting the potential for further analysis and probing based on later arrivals in the wavetrains, for example as demonstrated in Chunchuzov et al. (2011).

3.2 The SD-WACCM-X atmospheric model

In this study, the Whole Atmosphere Community Climate Model with thermosphere and ionosphere extension (WACCM-X; H.-L. Liu et al., 2018) is used as a model atmosphere. The particular version is the *specified dynamics*, SD-WACCM-X, version v2.1 (Sassi et al., 2013), for which the temperature and winds are nudged by the Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2; Gelaro et al., 2017) from the ground up to ~ 50 km. Above that altitude, WACCM-X is free-running. While WACCM-X extends up to about $500-700$ km altitude (145 levels), we only consider the altitude region relevant for infrasound propagation, which is up to 140 km altitude. The model has grid cells of $1.9° \times 2.5°$ in latitude-longitude and a 3-h temporal resolution (see the Data availability Section). For a detailed description of chemical and physical processes and parameterizations included in the model, see the studies by H.-L. Liu et al. (2018); J. Liu et al. (2018).

The WACCM-X model has been validated against observations and empirical models and has shown a good agreement in thermospheric composition, density and tidal amplitudes (H.-L. Liu et al., 2018). The SD-WACCM-X model has been found to be representative of the Earth's atmosphere in studies of different atmospheric phenomena: e.g., elevated-stratopause events (Siskind et al., 2021; Orsolini et al., 2017), dynamics (Kumari et al., 2021), atmospheric tides (Pancheva et al., 2020; Zhang et al., 2021; van Caspel et al., 2022). In contrast to other models routinely used for infrasound propagation, SD-WACCM-X provides a single consistent atmospheric model covering the altitude region relevant for long-range infrasound propagation, with a suitable spatio-temporal resolution. In particular, WACCM should provide a more physical description of the MLT region when compared to atmospheric specifications that are typically used for thermospheric arrival modeling, such as the HWM/MSIS climatological models (Drob, 2019).

Due to the proximity of the source to the receiver, the atmosphere can be approximated as a 1-D layered medium without time dependence. To avoid interpolation in space and time, we extract pressure, temperature, zonal and meridional winds from the grid node closest to the explosion site (Fig. 1) and the time step closest to the explosion origin time. The atmospheric conditions for all 49 Hukkakero events are presented in Fig. 4. Zonal and meridional winds in the stratosphere ($20-50$ km) are weak and have absolute values of up to 18 m/s. Their variation from explosion to explosion is negligible with standard deviation of $1-5$ m/s. This can be explained by the summer-to-winter transition in the stratospheric polar vortex where zonal wind is reversing from the westward summer circulation to the eastward winter circulation (Waugh & Polvani, 2010; Waugh et al., 2017). In contrast, atmospheric winds in the mesosphere - lower thermosphere ($50-120$ km) reach values of up to 100 m/s and vary strongly between explosions (standard deviation of up to 33 m/s) (A. K. Smith, 2012).

Figure 4 also shows $C_{\text{eff}}(z)$ ratio profiles (see Sect. 3.1) that have been computed using the SD-WACCM-X model (see Sect. 2). It can be seen that around 50 km altitude the ratio is close but does not exceed unity for most profiles, except for the events on 13 and 14 Aug 2015 (days 225 and 226). This indicates that the presence of a strong stratospheric waveguide for the Hukkakero-IS37 configuration in late summer is rather rare and therefore (strong) stratospheric returns would not be expected at IS37. In contrast, the effective sound speed ratio exceeds unity around lower thermosphere in all cases. This

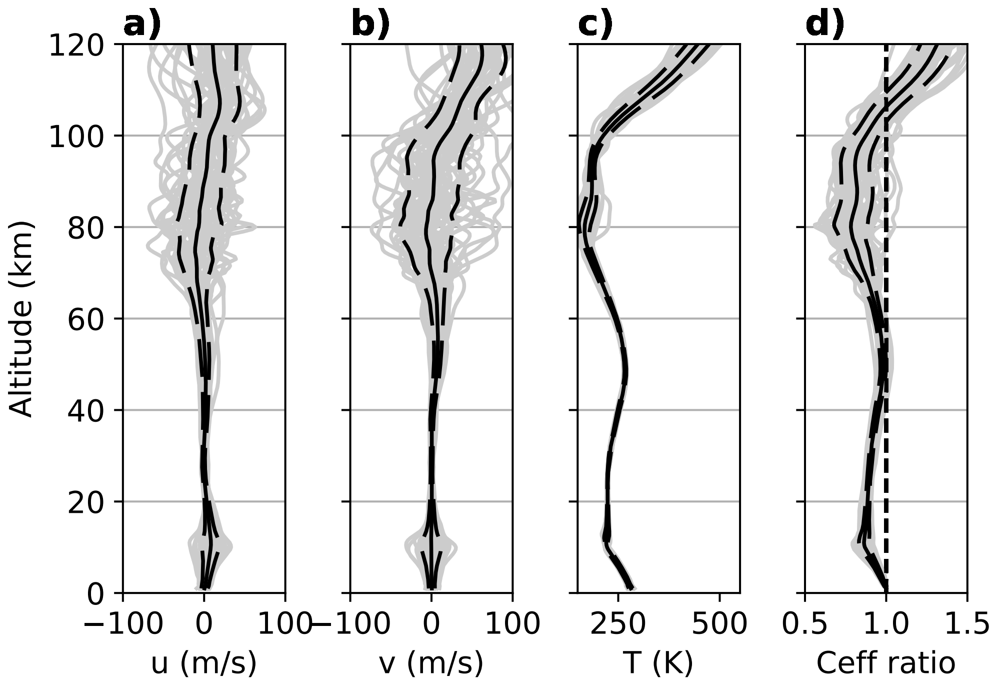

Figure 4. SD-WACCM-X atmospheric specifications for the 49 analyzed Hukkakero explosions, extracted at the grid point closest to the site around the time of the explosion. a) zonal wind, b) meridional wind, c) temperature, d) effective sound speed ratio.

can be attributed to the strong temperature gradient, which guarantees the presence of a thermospheric waveguide.

The effects of small-scale atmospheric fluctuations on stratospheric arrivals is particularly enhanced during periods of the year when the C_{eff} ratio near the stratopause is close to unity (Assink et al., 2014). Under these conditions, the small perturbations (e.g., gravity waves induced wind and temperature perturbations) can cause conditions favorable for i) refraction or ii) reflection. The propagation effects (refraction or reflection) strongly depend on the vertical scale of the atmospheric fluctuations in comparison to the infrasonic wavelength. For relatively large vertical scales, refraction of infrasonic waves can be simulated with ray theory, showing variations in travel time and backazimuth (Kulichkov, 2010). In contrast, infrasound scattering (or partial reflection) on vertical scales comparable to the infrasonic wavelength is a full-wave effect that cannot be simulated using ray theory. However, several studies (Chunchuzov & Kulichkov, 2020; Green et al., 2018; Blixt et al., 2019) have reported observations of partial reflections from stratospheric altitudes in the region where no stratospheric rays are predicted (i.e., the shadow zone).

3.3 Ray-tracing using the SD-WACCM-X model

For each of the analyzed Hukkakero events, we simulated infrasound propagation through its SD-WACCM-X atmospheric profile using the InfraGA ray tracer in 2-D Cartesian mode (see the Data availability Section for links and references). Rays were launched from the location of Hukkakero in the direction of IS37 with inclination angles ranging from 0 to 60 degrees measured from the horizontal.

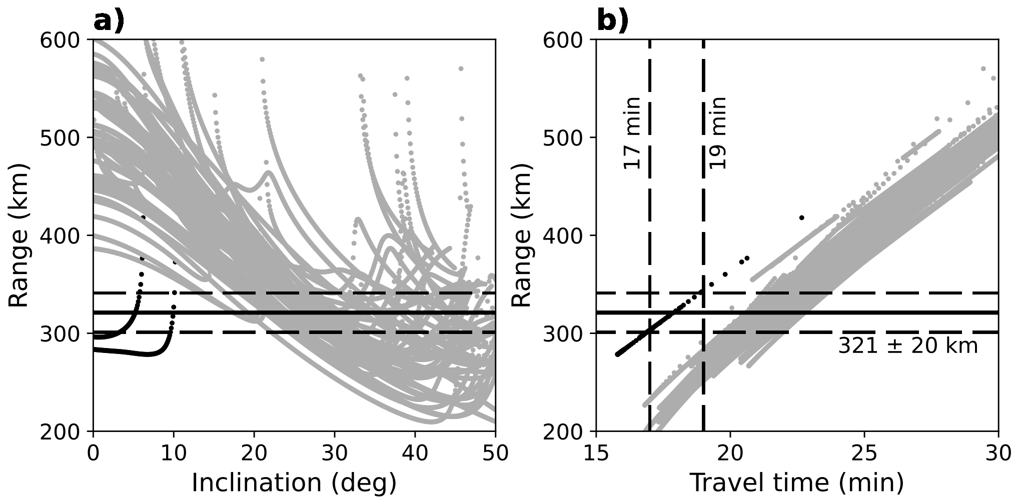

Figure 5. The first ground intercept information predicted by InfraGA for all explosive events. a) Eigenray departure inclination versus the distance from the source, b) travel time versus distance from the source. The eigenray turning height is color coded ($<$ 60 km - black dots, $\geq$ 60 km - gray dots). The Hukkakero-IS37 great-circle distance and the tolerance distance interval considered for ground intercept are indicated as a solid black line and dashed black lines, respectively. Observed travel time of the first arrival at IS37 is between 17 and 19 min (dashed black lines).

Fig. 5a shows ray departure inclination angle against distance from Hukkakero for refracted paths predicted by ray theory. Almost all of the predictions correspond to thermospheric refracted paths with turning heights in the lower thermosphere, near ~ 100 km (gray dots). As was mentioned before, these thermospheric arrivals are often observed at IS37 station Fig. 3. Fig. 5b shows the corresponding travel time (in min) for these rays. Stratospheric arrivals with arrival times between 17-19 min that correspond to our observations (Fig. 3) are only predicted for two events that occurred on 13 and 14 August 2015 (days 225 and 226). It follows from analysis of the SD-WACCM-X profiles (Fig. 4), that for these two days the $C_{\text{eff}}(z)$ ratio exceeds unity in the stratosphere.

From the ray-tracing simulations, it can be concluded that i) IS37 is located in a stratospheric shadow zone (i.e. there is no refraction-supported stratospheric duct) for the vast majority of cases and ii) refracted infrasound reaches the station via thermospheric ducts. Therefore, it is presumed that the stratospheric signals arrive at IS37 station after being partially reflected in the middle atmosphere (Kulichkov, 2010; Chunchuzov et al., 2011).

Fig. 6 illustrates the raypaths of a stratospheric and a thermospheric arrival at IS37 for the analyzed Hukkakero events. The $C_{\text{eff}}(z)$-ratio profile shown in the figure is computed based on the SD-WACCM-X model for 22 August 2017 at 12:00 UTC. The only arrival predicted by ray tracing is a thermospheric refracted ray that propagates up to 113 km and is predicted to arrive at IS37 after $\sim$ 22 minutes, which matches the observations (see Fig. 3).

The reflected rays are not predicted by the classical ray theory but are instead constructed using a mirroring procedure akin to the approach in, e.g., Blixt et al. (2019). We trace all rays until they reach the midpoint between Hukkakero and IS37 and then

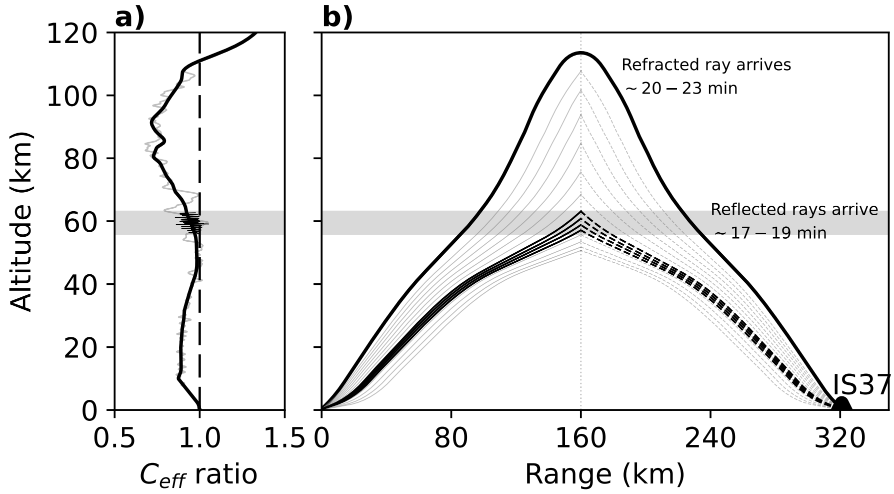

Figure 6. A schematic representation of infrasound raypaths from Hukkakero to IS37 relevant to this study. a) Effective sound speed ratio in direction of IS37 with a conceptual gravity wave perturbations (gray) and inhomogeneous layer of $C_{\text{eff}}(z)$ fluctuations (black). b) Thermospheric ducting simulated by ray theory and explaining later arrivals (20-23 min) with U-shape (thick black line). Earlier arrivals (17-19 min) that are not predicted by ray theory can be explained by infrasound being scattered by small-scale $C_{\text{eff}}(z)$ fluctuations in an atmospheric layer (dashed black lines).

mirror them to continue the path back to the surface. Due to acoustic reciprocity, this is a valid approach in a range-independent medium. It is hypothesized that these rays have scattered from an atmospheric layer with small-scale fluctuations in wind and temperature. The travel time is then estimated as twice the propagation time to the midpoint. The altitude range of the reflective layer is defined from the two rays that match best the observed beginning and ending of the processed infrasound signal. In case of a large discrepancy between the predicted and observed travel time for the lower boundary, we calculate the lower layer altitude as $z_j = z_{\text{top}} - C_{\text{eff}}(t_{\text{end}} - t_{\text{obs},j})$, assuming a constant effective sound speed in the layer. Here z_{top} is the upper boundary of the reflective layer obtained from ray tracing calculations, $t_{\text{obs},j}$ is a set of discrete times describing the observed travel time of the arrival, t_{end} is the end of the analyzed signal window.

3.4 Effective sound speed retrieval

We applied the approach of Chunchuzov et al. (2015) to retrieve fine-scale effective sound-speed variations in the middle atmosphere. This method was designed to be applied to stratospheric and thermospheric arrivals in the shadow zone, assuming that infrasound was scattered from inhomogeneous atmospheric layers with fine-scale $C_{\text{eff}}(z)$ fluctuations. It was demonstrated in (Chunchuzov et al., 2013) that temperature variations contribute relatively little to the effective sound-speed fluctuations ($\sim 20\%$) compared to wind variations ($\sim 80\%$). Therefore, we associate $C_{\text{eff}}(z)$ fluctuations with variations in horizontal wind.

This section presents the salient details behind the algorithm for the retrieval procedure, and provides a description of the main underlying assumptions. For a more detailed derivation of the equations and discussion of the method, we refer to (Chunchuzov et al., 2015; Chunchuzov & Kulichkov, 2020; Chunchuzov et al., 2022). For convenience, most nomenclature and designations used in the current study are the same as in these original studies.

The fine-scale effective sound-speed inversion approach is based on:

1) The assumption that infrasound is scattered or partially reflected at the midpoint between the source and receiver in a moving atmospheric layer with vertical fluctuations in the effective refractive index,
$$\varepsilon(z) = -2(\Delta c + \Delta u \sin\theta_0)/(c_1 \cos^2\theta_0), \tag{2}$$
where Δc are the sound speed fluctuations; Δu is the projection of wind fluctuations on the source-receiver radius vector; c_1 is the average sound speed in the layer; and θ_0 is the angle of incidence on the layer at altitude z. The effective refractive index, $\varepsilon(z)$, is assumed to be non-zero only inside the moving layer. A detailed derivation of Eq. 2 is provided in Appendix B.
2) The relationship between the vertical profile of the effective refractive index fluctuations, $\varepsilon(z)$, and the scattered signal waveform, $p'(t)$ is:
$$p'(t) = -\frac{p'_m r_0}{4R_1}\int_{-\infty}^{\infty} f(t - R_1/c_1 - z/a)\,\frac{d\varepsilon(z')}{dz'}\,\mathrm{d}z\ , \tag{3}$$
where p'_m is the peak signal amplitude recorded at distance r_0 close to the source; R_1 is the total distance along the propagation path; $f(t)$ is the normalized acoustic pressure waveform at r_0; $a = c_1/(2\cos\theta_0)$ is a coefficient representing the speed of the infrasound in the refractive layer; and $d\varepsilon(z)/dz$ is the spatial derivative of $\varepsilon(z)$. The dimensionless waveform of the scattered signal is defined as $I_0(t) = p'(t)R_1/(p'_m r_0)$.
3) The assumption that the initial signal waveform, $f(t)$, has an N-wave shape (Lonzaga et al., 2015) near the source and a duration T_0 at the reflective layer altitude.

After integrating Eq. 3 and solving the resulting equation (more details in Chunchuzov and Kulichkov (2020)), the relation between the effective refractive index profile and the dimensionless waveform of the scattered signal becomes

$$I_0(t) = -\frac{\varepsilon(a[t - R_1/c_1]) + \varepsilon(a[t - R_1/c_1 - T_0])}{4}. \tag{4}$$

Equation 4 can be solved numerically for a set of discrete time samples with respect to $\varepsilon(z)$ using the method of least squares (see Appendix A for details). Next, the effective sound speed fluctuations, $\Delta C_{\text{eff}}(z)$, can be estimated from the $\varepsilon(z)$ profile using Eq. 2 (Appendix B). However, several parameters need to be specified before solving Eq. 4:

- The average sound speed c_1 is obtained by matching the travel time predicted by ray-tracing simulations to the observed travel time, and thereby determining the altitude range of the reflective layer and averaging the sound speed within it, as well as angle θ_0.
- An estimate of the peak overpressure close to the source, p'_m, is obtained using the model by Kinney and Graham (1985) based on the blast yield. The typical yield of Hukkakero explosions is presumed to be approximately 20 ton of TNT equivalent (Gibbons et al., 2015). According to the Kinney and Graham (1985) model with the initial conditions $W = 20$ ton TNT, $P_{\text{ref}} = 1.01325{\cdot}10^5$ Pa, and $\rho_{\text{ref}} = 1.225$ kg/m^3 (Atmosphere, 1976), the peak overpressure at $r_0 = 1$ km from the source becomes $p'_m = 2320$ Pa.
- As the initially generated shock wave propagates, it experiences attenuation and becomes distorted due to non-linear propagation effects, which become more prominent with increasing height due to decreasing atmospheric density with altitude

(Lonzaga et al., 2015; Blom & Waxler, 2021). One of the distortion features associated with non-linear propagation is period lengthening, which occurs since positive and negative phases of the pressure wave travel at slightly different speed (Hamilton & Blackstock, 2008). This contributes to decreasing the amplitude of the acoustic pulse as its duration increases following the acoustic-pulse conservation law (Kulichkov et al., 2017). To get an estimate of the N-wave duration at the reflective layer altitude, weakly non-linear propagation simulations were performed using InfraGA. Properties of the initially generated shock wave (peak overpressure of 2320 Pa and positive pressure phase of 0.11 s) were calculated based on the Kinney and Graham (1985) model described above. Values of T_0 in the range of $1-2$ s were found to correspond to altitudes in the range of $50-80$ km. This is the region from where we expect rays to reflect from, following the travel-time based mirroring simulations as described in Sect. 3.3.

4 Results

This study analyzes the first (stratospheric) Hukkakero arrivals in the infrasound recordings described in Sect. 3.1 and illustrated in Fig. 3. For the 49 Hukkakero blasts investigated, we processed a 30 second segment of the infrasound best-beam signal traces using the recipe provided in Sect. 3.4. Figure 7 displays the $\Delta C_{\text{eff}}(z)$ profiles retrieved. There is a day-to-day variability in the reflective layer altitude, with all $\Delta C_{\text{eff}}(z)$ profiles being located within stratopause–lower mesosphere altitudes of $50 - 75$ km with the average depth of 7.75±0.38 km. Previous studies demonstrate that infrasound signal characteristics observed for events with similar strength and source-receiver geometry are highly sensitive to varying middle atmospheric winds and temperatures (Le Pichon et al., 2002; Drob, 2019; Averbuch et al., 2022). Therefore, the difference in the arrival time between events, as displayed in Fig. 3, can be related to the variation in the infrasound probing altitude. This is confirmed by the overall agreement in the arrival time variations for the explosions studied and the associated altitude variation of the retrieved fluctuation profiles, see Fig. 7. It should be noted that the same $\Delta C_{\text{eff}}(z)$ retrieval procedure can also be applied to later arrivals, which correspond to higher altitudes, as demonstrated in Chunchuzov et al. (2022).

The majority of the effective sound-speed fluctuations retrieved, $\Delta C_{\text{eff}}(z)$, have amplitudes of up to 5 m/s. However, for some cases, the amplitudes reach up to 15 m/s. Exceptionally high $\Delta C_{\text{eff}}(z)$ amplitudes of up to 25 m/s are estimated from the waveform recorded on 27 August 2016 (day 240 shown as the gray profile in Fig. 7). There are two reasons behind it. First, the signal amplitude reaches 2 Pa which is larger than for any other event. Second, rapid changes in the waveform amplitude make it difficult for the fitting procedure to find an appropriate solution (see Appendix B). We consider this event as an outlier and suggest that it should be interpreted as a refracted rather than reflected arrival, and therefore remove it from the analysis.

The root-mean-square error (RMSE) of $\Delta C_{\text{eff}}(z)$ retrieved varies within $6-18\%$ (see Appendix A). This RMSE is calculated based on the difference between the left- and right- hand sides of Eq. 4 (see Appendix B for details).

Next, we perform a vertical wavenumber spectral analysis of the retrieved $\Delta C_{\text{eff}}(z)$ profiles by estimating the PSD using Welch's method (Welch, 1967) with a Hamming window (window length of 750 m or 50 samples and 50% overlap). Figure 8 displays the vertical wavenumber power spectral density of the retrieved effective sound-speed fluctuation profiles, as well as their mean. It can be seen that negative PSD slope is present for all events. The vertical wavenumber, k_z, that corresponds to the beginning of the negative slope is denoted the dominant wavenumber, m_*. Based on the analyzed events, $m_* = 2.15 \cdot 10^{-3} \pm 4.4 \cdot 10^{-4}$ cycles/m (see Appendix A). Fitting the k_z^p power-law within

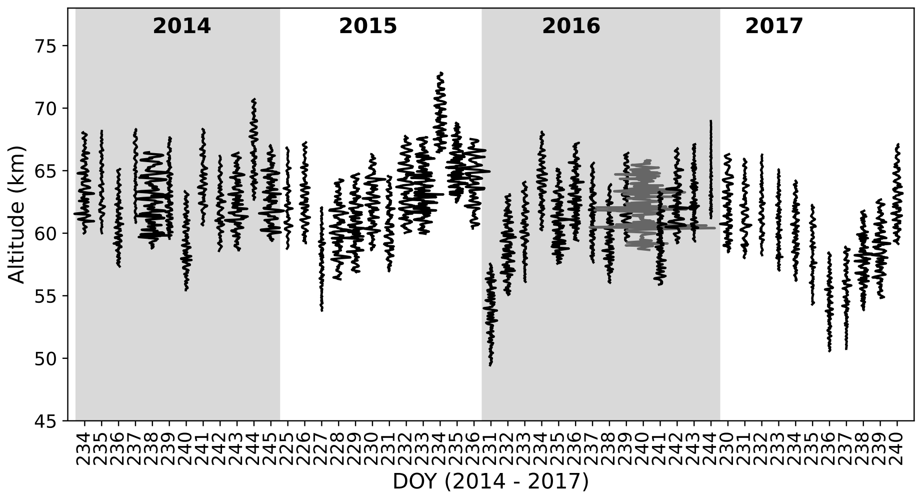

Figure 7. Retrieved fluctuations of the effective sound speed $C_{\text{eff}}(z)$. The $C_{\text{eff}}(z)$ profile on 27 August 2016 (day 240) with exceptionally high values (more details in the text) is displayed in gray to avoid overlapping with other profiles.

$k_z > m_*$ provides an estimate of $p = -3.35$ for the mean PSD and $p = -3.50 \pm 0.39$ for all profiles (see Appendix A).

The power-law exponents obtained in this study are close to the k_z^{-3} power-law which is known to correspond to the “universal” spectrum of horizontal wind fluctuations induced by gravity waves or gravity wave saturation spectrum (Fritts & Alexander, 2003). Various theories were proposed to explain the dynamics behind gravity wave saturation, i.e., instability and wave-wave interaction. The saturation spectrum amplitude was shown to correspond to $CN^2k_z^{-3}$ with C typically varying within $0.1 - 0.4$ (Hines, 1991) depending on the theory and assumptions made. The first attempt to describe universality in measured wind spectra (e.g., Endlich et al., 1969; Dewan et al., 1984) was made by Dewan and Good (1986) who assumed saturation via convective instabilities at each vertical wave number independently and yielded $C = 1$. Later, this theory was extended by S. A. Smith et al. (1987) to account also for amplitude limiting instabilities arising from the whole wave spectrum instead, and value of $C = 1/6$ was obtained. These traditional linear saturation theories were criticized in Hines (1991) and Chunchuzov (2002), where it was shown that small-scale anisotropic inhomogeneities with k_z^{-3} vertical wavenumber spectrum are shaped due to non-resonant internal wave-wave interactions. Chunchuzov et al. (2015) compared vertical wavenumber spectra of effective sound-speed fluctuations retrieved from infrasound detections of five volcanic eruptions and one explosion. Based on this analysis, a value of $C = 0.2$ for the upper stratosphere was proposed.

The power-laws corresponding to linear (Dewan & Good, 1986) and non-linear (Chunchuzov et al., 2015) theoretical models are displayed in Fig. 8 together with error bars indicating possible variability in theoretical PSD amplitude ($C = 0.1 - 0.4$). In both theoretical models, the altitude regime is controlled via the Brunt-Väisälä frequency, N. We use $N = 1.66{\cdot}10^{-2}$ rad/s in our calculations, which is typical for the lower mesosphere (Dewan & Good, 1986). Theoretical models show a good agreement with the mean spectrum of the retrieved $\Delta C_{\text{eff}}(z)$ profiles. This allows us to conclude that the infrasound-based vertical wavenumber spectra that are obtained in this study are consistent with previously obtained theoretical spectra, taking into account the confidence intervals of those measurements (Fritts & Alexander, 2003).

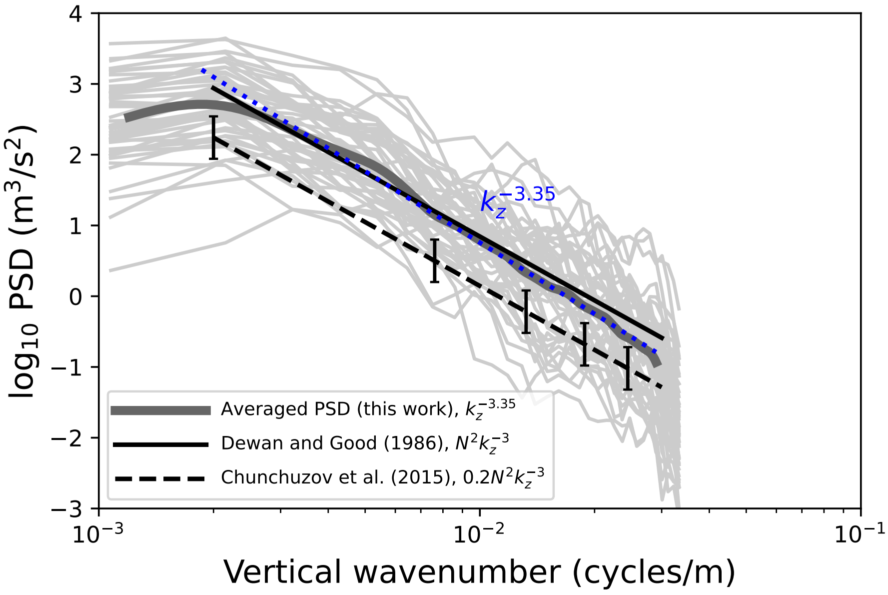

Figure 8. Vertical wavenumber power spectral density (PSD) of the retrieved $\Delta C_{\text{eff}}(z)$ fluctuations (light gray lines) and their mean (dark gray line) versus theoretical models by Dewan and Good (1986) (black solid) and Chunchuzov et al. (2015) (black dashed). Black error bars indicate variability in theoretical PSD amplitude based on other theories mentioned in the text. The blue dotted line indicates the power-law fitting region for the mean PSD.

From the spectral analysis, we can estimate the outer and inner vertical scale of atmospheric inhomogeneities that infrasound is sensitive to, based on the vertical wavenumber limits within which the k_z^{-3} power-law establishes. Denoting the highest vertical wavenumber as m_b, we obtain $L_{\text{inner}} = 1/m_b = 33 - 37$ m and $L_{\text{outer}} = 1/m_* = 386 - 585$ m. Note that the limited altitude range of the $\Delta C_{\text{eff}}(z)$ profiles retrieved restricts the sensitivity to motions with smaller vertical wavenumbers (larger vertical scales). This could be improved by processing longer segments of infrasound waveforms as was demonstrated in e.g., Chunchuzov et al. (2013, 2015).

5 Discussion

The current study applies the effective sound-speed retrieval procedure by Chunchuzov et al. (2015) to infrasound recordings in the shadow zone. This is the first time the aforementioned approach is applied to a large and consistent dataset. Because we are retrieving $\Delta C_{\text{eff}}(z)$ profiles along a fixed source-receiver path and because the explosion yields are similar for each event, we can consider the variability in the infrasound recordings as being related to atmospheric dynamics.

The results show that vertical wavenumber PSDs obtained from the $\Delta C_{\text{eff}}(z)$ profiles are close to the "universal" gravity wave saturation spectrum of k_z^{-3}. The very end of the vertical wavenumber spectra in Fig. 8 corresponds to motions at scales of tens of meters. This is on the edge of transition from the gravity wave saturation regime to the turbulence regime where the theory predicts a transition from a k_z^{-3} power-law to $k_z^{-5/3}$ (e.g., Gardner et al., 1993). The vertical wavenumber where this transition occurs may have different values based on the latitude and altitude of interest, for example, the value of $2{\cdot}10^{-3}$ cycles/m was proposed in (Gardner et al., 1993) for mid-latitude mesopause region. In contrast, Endlich et al. (1969) analyzed vertical wind profiles measured during different seasons and found that their PSDs follow the k_z^{-3} power-law up to the vertical wavenumber of 10^{-2} cycles/m. However, the turbulence regime is outside of the scope of this study, and we leave this question open for further research.

As $C_{\mathrm{eff}}(z)$ fluctuations are mostly associated with variations in horizontal wind (Sect. 3.4), it would be interesting to compare the vertical wavenumber spectra obtained in this study to spectra of wind measured near the IS37-Hukkakero region (Fig. 1). For this purpose, the spectral characteristics of 11 infrasound-based $\Delta C_{\mathrm{eff}}(z)$ profile retrievals from 2017 were compared against independent wind measurements available from the Saura medium-frequency (MF) radar near Andøya, Norway (69.14° N, 16.02° E; Fig. 1). This radar is located $\sim$ 100 km west of the IS37 infrasound station and $\sim$ 420 km north-west from Hukkakero (Fig. 1), and operates on 3.17 MHz with 58 2kW pulsed transceiver modules. Its observation capabilities include wind measurements, estimates of turbulent kinetic energy dissipation rates, and electron density, as well as meteor observations. The observations typically provide measurements within the $\sim 50-100$ km altitude range with a vertical resolution of $1-1.5$ km (Singer et al., 2008). Hence, the system can observe vertical variations at wavenumbers below approximately $k_z = 10^{-3}$ cycles/m.

The wind data used for the validation has been derived from Doppler-Beam-Swinging experiments measuring the radial velocity for one vertical and four oblique soundings including statistical interferometric Angle of Arrival correction (see Renkwitz et al., 2018).

First, we directly compare the Saura radar winds to the SD-WACCM-X model winds. As the effective sound speed $\Delta C_{\mathrm{eff}}(z)$ is taken along the horizontal infrasound propagation direction (Eq. 1), we project the Saura radar wind on the same unit vector pointing from Hukkakero towards IS37: $\mathbf{u}{\cdot}\hat{n} = u\sin(\phi){+}v\cos(\phi)$, where ϕ is the Hukkakero-IS37 azimuth. The same projection was applied to the SD-WACCM-X wind profiles, extracted at the grid node located between the Saura radar and IS37 (Fig. 1). This comparison between Saura radar and SD-WACCM-X winds is displayed in Fig. 9a,b. Although the radar measurements do not fully cover the altitude region where the infrasound-based $\Delta C_{\mathrm{eff}}(z)$ profiles are retrieved (highlighted in Fig. 9a,b), it can still be seen that the Saura wind measurement features a pattern similar to the SD-WACCM-X model. There is a weak wind pattern ($<$ 50 m/s) that alternates between positive and negative values, mostly modulated by tidal waves. Above 70 km, a noticeable discrepancy between measured and modeled winds is observed. This may be related to a lower temporal resolution of the model compared to the radar, the distance between the sampling locations, or to inaccuracies in the parametrization of gravity wave breaking used in the SD-WACCM-X model. Moreover, note that above $\sim$ 50 km SD-WACCM-X is not supported by any observational dataset and is, therefore, expected to deviate more from the measurements. This discrepancy between the radar measured winds and SD-WACCM was shown in (de Wit et al., 2014), and is not unique to our measurements.

Next, we interpolate the SD-WACCM-X profiles to the radar vertical grid and perform a spectral comparison between the SD-WACCM-X and Saura radar wind profiles closest in time to the explosion onset. The obtained vertical wavenumber spectra are displayed in Fig. 9c together with gravity wave saturation theories from Fig. 8. One can see a good agreement in PSD amplitudes between the radar, atmospheric model and GW saturation theories. However, it's clear that SD-WACCM-X wind spectra have steeper slope and seem to underestimate amplitudes at ranges $10^{-4}-10^{-3}$ cycles/m. A more detailed look into SD-WACCM-X and Saura radar horizontal winds over long time periods is needed to fully understand the nature of such discrepancy. We leave this question open for further research suggesting that parametrization of subgrid-scale processes in SD-WACCM-X can probably be improved.

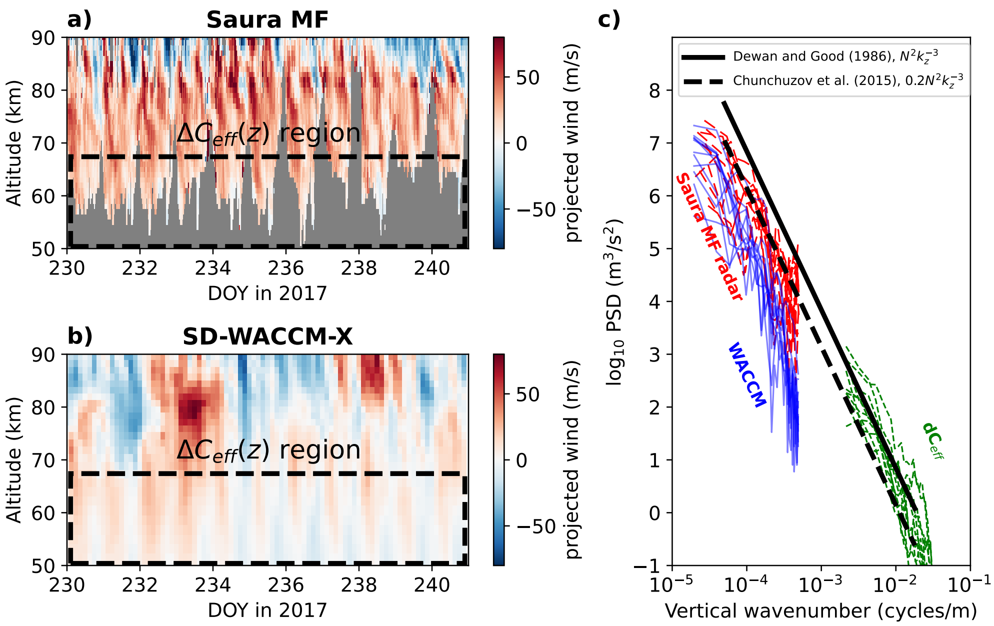

Figure 9. a) Projection of wind measured by Saura MF radar and b) predicted by SD-WACCM-X on the vector connecting Hukkakero and IS37. c) Vertical wavenumber spectra of the Saura radar winds (red dashed), SD-WACCM-X winds (blue solid) and retrieved $\Delta C_{\mathrm{eff}}(z)$ fluctuations (green dotted) for the explosions in 2017, versus theoretical models by Dewan and Good (1986) (black solid) and Chunchuzov et al. (2015) (black dashed).

To resolve the high-wavenumber part of the spectrum that the Saura radar and SD-WACCM-X are insensitive to due to their vertical resolution, the infrasound-retrieved $\Delta C_{\mathrm{eff}}(z)$ profiles retrieved are used. The vertical wavenumber spectra for the 2017 $\Delta C_{\mathrm{eff}}(z)$ profiles are presented in Fig. 9c. As was shown earlier (Fig. 8), the high-wavenumber part of the spectrum follows the k_z^{-3} power-law and agrees well in amplitude with linear and non-linear gravity wave saturation theories. The overall agreement found allows us to suggest that Saura radar and infrasound-based $\Delta C_{\mathrm{eff}}(z)$ profiles represent low- and high-wavenumber parts of the same “universal” GW spectrum.

Possible avenues for future research can include application of the same effective sound-speed retrieval approach to later mesospheric and thermospheric arrivals observed at IS37 (Fig. 3). This would provide an opportunity to study thicker atmospheric layers and to possibly look at other physical phenomena that could be responsible for infrasound scattering (e.g., polar mesospheric summer echoes). Another possible direction of research could be comparing the effective sound-speed fluctuations obtained in this study to other measurement techniques with high vertical resolution, e.g., lidar. Moreover, studying the 3D wind field and temperature fluctuations caused by gravity wave could be performed by applying the retrieval approach to several infrasound stations around the Hukkakero explosion site e.g., ARCES/ARCI (Karasjok, Norway), KRIS (Kiruna, Norway) and APA/APAI (Apatity, Russia) (Gibbons et al., 2015).

6 Summary

In this study, infrasound waves from 49 blasts between 2014 and 2017 are used to retrieve effective sound speed fluctuations, $\Delta C_{\mathrm{eff}}(z)$, in the middle atmosphere. The applied retrieval recipe is based on approaches previously developed by Chunchuzov et al. (2013, 2015). It is based on a relation between the waveform of the scattered infrasound signal recorded on the surface in the shadow zone and the $C_{\mathrm{eff}}(z)$ fluctuation profile in an inhomogeneous atmospheric layer. The results obtained demonstrate that the infrasound scattering occurs in the lower mesosphere between 50 and 75 km altitude. This atmospheric region is also known to be altitudes where gravity waves start to break (Garcia & Solomon, 1985). Therefore, information about the $\Delta C_{\mathrm{eff}}(z)$ retrieved from ground-based infrasound measurements is of direct interest for studying the GW activity and for potential improvement of GW parameterization schemes used in numerical weather prediction models. The spectral analysis of retrieved effective sound speed fluctuations in terms of vertical wavenumber spectra revealed that the tail of the mean spectrum follows a k_z^{-3} power law. This law corresponds to the "universal" spectrum of horizontal wind fluctuations induced by gravity waves (Fritts & Alexander, 2003). The spectral characteristics of the 11 infrasound-based $\Delta C_{\mathrm{eff}}(z)$ profiles retrieved for 2017 were compared against independent wind measurements by the Saura MF radar. Good agreement in amplitudes and slopes of the spectra was demonstrated, indicating that the infrasound and the radar measurements represent the high- and low-wavenumber sections of the "universal" gravity-wave spectrum, respectively. Therefore, the current study opens the way for remote sensing of GW activity by means of ground-based infrasound measurements and to improve the representation of small-scale wind inhomogeneities in upper atmospheric model products. The latter would be beneficial for the infrasound scientific field since advanced simulations of infrasound propagation require atmospheric specifications with high vertical resolution (Hedlin & Drob, 2014; Chunchuzov et al., 2015; Lalande & Waxler, 2016; Sabatini et al., 2019). Moreover, the prospects of using explosive event sequences as *datasets of opportunity* for middle atmospheric remote sensing can pave the way for an enhanced GW representation in atmospheric models.

Appendix A Retrieved parameters and comparisons

Table A1 provides details about the spectral analysis performed in Sect. 4.

Table A1. Explosion origin time, dominant wavenumber and the slope for the corresponding spectrum.

Origin time (yyyy-mm-dd HH:MM:SS, UTC)	DOY	m_* [cycl/m]	exponent in k_z^p	RMSE relative to max amplitude
2014-08-22 11:59:59	234	2.15e-3	-3.79	0.06
2014-08-23 10:29:59	235	1.07e-3	-3.43	0.08
2014-08-24 11:59:59	236	2.15e-3	-3.29	0.13
2014-08-25 10:29:59	237	1.07e-3	-3.23	0.10
2014-08-26 10:59:59	238	2.15e-3	-3.04	0.07
2014-08-27 10:59:59	239	2.15e-3	-2.95	0.08
2014-08-28 10:59:59	240	2.15e-3	-3.30	0.08
2014-08-29 10:29:59	241	2.15e-3	-3.83	0.13
2014-08-30 10:29:59	242	2.15e-3	-3.95	0.10
2014-08-31 10:59:59	243	2.15e-3	-3.63	0.08
2014-09-01 09:59:59	244	2.15e-3	-3.67	0.13
2014-09-02 09:29:59	245	2.15e-3	-3.25	0.09
2015-08-13 10:59:59	225	2.15e-3	-3.71	0.08
2015-08-14 10:04:59	226	2.15e-3	-3.54	0.14
2015-08-15 10:59:59	227	2.15e-3	-3.87	0.09
2015-08-16 10:59:59	228	2.15e-3	-3.56	0.09
2015-08-17 11:59:59	229	2.15e-3	-3.02	0.13
2015-08-18 09:59:59	230	2.15e-3	-3.86	0.06
2015-08-19 09:29:59	231	2.15e-3	-2.90	0.08
2015-08-20 09:29:59	232	2.15e-3	-3.57	0.13
2015-08-21 09:29:59	233	2.15e-3	-3.19	0.08
2015-08-22 11:29:59	234	2.15e-3	-2.84	0.11
2015-08-23 11:29:59	235	2.15e-3	-2.65	0.09
2015-08-24 12:00:00	236	2.15e-3	-3.52	0.06
2016-08-18 12:29:59	231	2.15e-3	-3.18	0.10
2016-08-19 11:29:59	232	2.15e-3	-4.00	0.12
2016-08-20 13:29:59	233	2.15e-3	-3.76	0.07
2016-08-21 13:00:00	234	2.15e-3	-3.71	0.12
2016-08-22 11:59:59	235	2.15e-3	-3.60	0.09
2016-08-23 12:59:59	236	1.07e-3	-2.78	0.18
2016-08-24 11:59:59	237	2.15e-3	-3.06	0.12
2016-08-25 11:29:59	238	3.23e-3	-4.11	0.10
2016-08-26 11:29:59	239	2.15e-3	-3.36	0.10
2016-08-27 12:59:59	240	3.23e-3	-4.07	0.06
2016-08-28 10:59:59	241	2.15e-3	-3.13	0.13
2016-08-29 09:59:59	242	2.15e-3	-3.46	0.10
2016-08-30 07:54:59	243	3.22e-3	-3.13	0.07
2016-08-31 08:49:59	244	3.23e-3	-3.80	0.06
2017-08-18 11:59:59	230	2.15e-3	-4.25	0.18
2017-08-19 11:00:00	231	1.08e-3	-3.46	0.16
2017-08-20 12:00:00	232	2.15e-3	-3.70	0.08
2017-08-21 12:59:59	233	3.22e-3	-4.23	0.07
2017-08-22 11:59:59	234	2.15e-3	-3.47	0.10
2017-08-23 11:29:59	235	2.15e-3	-4.11	0.07
2017-08-24 11:29:59	236	2.15e-3	-4.06	0.14
2017-08-25 09:59:59	237	2.15e-3	-3.75	0.10
2017-08-26 10:59:59	238	2.15e-3	-3.59	0.07
2017-08-27 11:29:59	239	2.15e-3	-3.34	0.08
2017-08-28 10:29:59	240	2.15e-3	-3.40	0.11
	Mean:	2.15e-3	-3.50	
	STD:	4.40e-4	-0.39	

Appendix B Derivation of the inversion equations

B1 Derivation of Eq. 2

Consider a stationary atmosphere consisting of an inhomogeneous moving layer within $z_0 \leq z \leq z_H$ and a homogeneous half-space below and above it. The sound speed $c(z)$, wind velocity $\boldsymbol{v}(z)$ and density $\rho(z)$ have continuous first and second order derivatives, and are constant outside the inhomogeneous layer with values of c_1, $\boldsymbol{v}_1$ and ρ_1. The layer is filled with stratified sound speed, wind velocity and density fluctuations $\Delta c(z)$, $\Delta \boldsymbol{v}(z)$ and $\Delta\rho(z)$ on top of the background atmosphere. Therefore, sound speed, atmospheric wind and density within the inhomogeneous layer are defined as: $c_{1+\Delta}(z) = c_1 + \Delta c(z), \boldsymbol{v}_{1+\Delta}(z) = \boldsymbol{v}_1 + \Delta\boldsymbol{v}(z), \rho_{1+\Delta}(z) = \rho_1 + \Delta\rho(z)$. In terms of the relative fluctuations, it's assumed that $\Delta c/c_1$, $\Delta v/c_1$ and $\Delta\rho/\rho_1$ are of the same order of smallness, namely $M = |\Delta c/c_1| \ll 1$.

A plane monochromatic acoustic wave $A\exp(i(\xi_x x + \xi_y y + \mu z - \omega t))$ propagates from the source to the receiver upward through the homogeneous atmosphere and incident on a moving inhomogeneous layer at an angle θ measured from the vertical. Here A is complex wave amplitude, ω is wave frequency, $\boldsymbol{\xi} = (\xi_x, \xi_y)$ is the horizontal propagation vector, $\mu = (k_0^2 - |\boldsymbol{\xi}|^2)^{1/2}$ is the vertical wavenumber, and $k_0 = \omega/c_1$ is the wavenumber in the homogeneous atmosphere. The projection of the wind velocity $\boldsymbol{v}(z)$ on the source-receiver radius vector $\boldsymbol{\xi}$ is defined as $u(z) = \boldsymbol{v}(z)\boldsymbol{\xi}/|\boldsymbol{\xi}|$.

We introduce the squared effective refractive index following Chunchuzov et al. (2013) as:

$$N^2(z) = \left(n^2\beta^2 - \frac{\xi^2}{k_0^2}\right)\left(\frac{\rho_0}{\rho\beta^2}\right)^2, \tag{B1}$$

where $n = c_1/c$ is a refractive index in a stationary medium, $\beta = 1 - \boldsymbol{\xi}\boldsymbol{v}(z)/\omega$, ρ_0 is a density dimension coefficient, $\xi = k_0 \sin\theta(1 + u_1\sin\theta/c_1)^{-1}$.

Small relative fluctuations of the effective refractive index in an inhomogeneous layer are defined as:

$$\varepsilon(z) = \ln\frac{N_{1+\Delta}^2}{N_1^2} = \ln\frac{n_{1+\Delta}^2\beta_{1+\Delta}^2 - \xi^2/k_0^2}{n_1^2\beta_1^2 - \xi^2/k_0^2} + 2\ln\frac{\rho_1}{\rho_{1+\Delta}} + 4\ln\frac{\beta_1}{\beta_{1+\Delta}}, \tag{B2}$$

where

$$n_1 = 1, \quad n_{1+\Delta} = \frac{c_1}{c_1 + \Delta c}, \quad \text{and} \quad \beta_1 = 1 - \xi u_1/\omega = \left(1 + \frac{u_1\sin\theta}{c_1}\right)^{-1}, \tag{B3}$$

and

$$\beta_{1+\Delta} = 1 - \frac{\xi(u_1 + \Delta u(z))}{\omega} = \beta_1\left(1 - \frac{\Delta u(z)\sin\theta}{c_1}\right). \tag{B4}$$

Substituting parameters from Eq. B3 into Eq. B2 and assuming the first-order of smallness for the natural logarithm, $\ln(x/y) \sim (x-y)/y$, yields

$$\varepsilon(z) = \frac{-2[\Delta c/c_1 + \Delta u(z)\sin\theta/c_1] + \mathcal{O}(M^2)}{\cos^2\theta} + 4\frac{\Delta u(z)\sin\theta}{c_1} - 2\frac{\Delta\rho}{\rho_1}. \tag{B5}$$

As θ approaches $\pi/2$ the last two terms can be neglected and Eq. 2 is obtained.

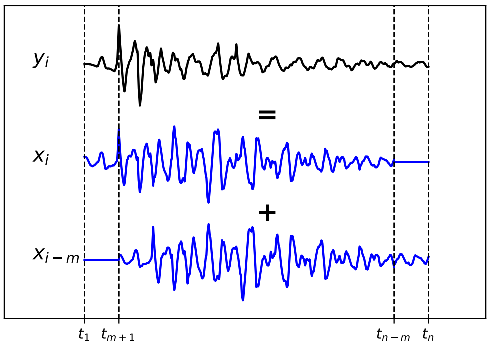

Figure B1. A synthetic example of the Eq. 4.

B2 System of equations to solve Eq. 4

In this section, we provide the same explanation on how to numerically solve Eq. 4 as presented in (Chunchuzov & Kulichkov, 2020), but complemented with more detail.

Eq. 4 represents the dimensionless waveform of scattered signal as a sum of two effective refractive index profiles shifted in time by the time interval T_0. Let us denote values of the scattered signal at discrete times t_i as $y_i = I_0(t_i)$ where $i = 1, 2, .., n$ (n is the number of samples), and effective refractive index values as $x_i = -\varepsilon(a[t_j - R_1/c_1])/4$ with non-zero values at $1, 2, .., n - m$ and $x_{i-m} = -\varepsilon(a[t_j - R_1/c_1 - T_0])/4$ with non-zero values at $m + 1, m + 2, .., n$, where m is the number of t_i values within the time interval T_0. Fig. B1 demonstrates Eq. 4 with the notation introduced.

Thus, the following system of linear algebraic equations with respect to x_i can be obtained from Eq. 4:

$$\begin{cases} y_i = x_i, & \text{for } 1 \le i \le m \\ y_i = x_i + x_{i-m}, & \text{for } m+1 \le i \le n-m \\ y_i = x_{i-m}, & \text{for } n-m+1 \le i \le n. \end{cases} \tag{B6}$$

The number of unknowns in the system B6, $n-m$, is less than number of equations, n, and the system is therefore overdetermined. In this case, the least squares method can be used to find an approximate solution by minimizing the difference $|\alpha X - Y|$ where $X = x_j, \quad j = 1, 2, .., n - m$, $Y = y_i, \quad i = 1, 2, .., n$, and α is the matrix of coefficients.

After the solution $X = x_j$ has been found, the profile of the effective refractive index can be retrieved as $\varepsilon(a[t_j - R_1/c_1]) = -4x_j$. Next, the effective sound fluctuation profile is obtained from $\varepsilon(z_j)$ values using Eq. 2 as:

$$\Delta C_{\text{eff}}(z_j) \approx \Delta c(z_j) + \Delta u(z_j)\sin\theta_0 = -\frac{\varepsilon(z_j)c_1\cos^2\theta_0}{2} = 2x_j c_1 \cos^2\theta_0. \tag{B7}$$

Open Research Section

The 3-hourly SD-WACCM-X model product data are available via `https://www.earthsystemgrid.org/dataset/ucar.cgd.ccsm4.SD-WACCM-X_v2.1.atm.hist.3hourly_inst.html` (last access June 2022). The InfraGA infrasound propagation code (e.g., Blom & Waxler, 2017, 2021) is provided under open access by Los Alamos National Laboratory at `https://github.com/LANL-Seismoacoustics/infraGA` (last access June 2022).

The IS37 infrasound station is part of the International Monitoring System (IMS) of the Preparatory Commission for the Comprehensive Nuclear-Test-Ban Treaty Organization (CTBTO). Data access can be granted to third parties and researchers through the virtual Data Exploitation Centre (vDEC) of the International Data Center: `https://www.ctbto.org/specials/vdec/`. The dataset of Saura wind measurements used in this study is available via `https://www.radar-service.eu/radar/en/dataset/mzuBmhtrDxSGIBNd?token=leArdOpgjcsMPpeNSFyO`. More data can be obtained by contacting Toralf Renkwitz.

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