We discuss the use of reverse time migration (RTM) with dense seismic networks for the detection and location of sources of atmospheric infrasound. Seismometers measure the response of the Earth’s surface to infrasound through acoustic-to-seismic coupling. RTM has recently been applied to data from the USArray network to create a catalogue of infrasonic sources in the western US. Specifically, several hundred sources were detected in 2007–2008, many of which were not observed by regional infrasonic arrays. The influence of the east–west stratospheric zonal winds is clearly seen in the seismic data with most detections made downwind of the source. We study this large-scale anisotropy of infrasonic propagation, using a winter and summer source in Idaho. The bandpass-filtered (1–5 Hz) seismic waveforms reveal in detail the two-dimensional spread of the infrasonic wavefield across the Earth’s surface within approximately 800 km of the source. Using three-dimensional ray tracing, we find that the stratospheric winds above 30 km altitude in the ground-to-space (G2S) atmospheric model explain well the observed anisotropy pattern. We also analyse infrasound from well-constrained explosions in northern Utah with a denser IRIS PASSCAL seismic network. The standard G2S model correctly predicts the anisotropy of the stratospheric duct, but it incorrectly predicts the dimensions of the shadow zones in the downwind direction. We show that the inclusion of finer-scale structure owing to internal gravity waves infills the shadow zones and predicts the observed time durations of the signals. From the success of this method in predicting...
the observations, we propose that multipathing owing to fine scale, layer-cake structure is the primary mechanism governing propagation for frequencies above approximately 1 Hz and infer that stochastic approaches incorporating internal gravity waves are a useful improvement to the standard G2S model for infrasonic propagation modelling.

1. Introduction

The cataclysmic eruption of Krakatoa in 1883 is considered the loudest sound in recorded human history. It also signalled the beginning of the study of sources of long-period sound in the atmosphere and the use of this sound to probe atmospheric structure. This study details the rich history of this field that is both a branch of acoustics and a branch of geophysics. We discuss what infrasound is, how it propagates through Earth’s atmosphere and the close connection between infrasonics and seismology. Then, we discuss the signal processing methods used to detect and locate infrasound with sensor arrays. We show how these methods led to the discovery of hundreds of infrasonic events in the western US, and the inference from the detailed analysis of several of the high-quality signals that fine-scale structure is required in atmospheric temperature and wind models.

2. The acoustic spectrum

The acoustic spectrum is divided into three bands based on the nominal limits of human hearing. The audible band in the middle of the acoustic spectrum ranges from 20 to 20 000 Hz. Although humans cannot hear sound at higher, ultrasonic, frequencies there are well-known applications of this sound in medicine and industry. Some animals use ultrasound for communication, navigation or for location of prey. Intrinsic energy loss scales with the square of frequency, and as a result, ultrasound decays rapidly and can be used only at close range. Conversely, sound waves in the infrasonic band below the audible range decay very slowly and can be detected instrumentally at ranges of hundreds to thousands of kilometres from the source. Infrasound is analogous to infrared energy in the electromagnetic spectrum. Infrared can be felt as heat but cannot be seen. Although humans cannot hear infrasound, high-frequency infrasound just below 20 Hz can be felt as vibrations or sensed indirectly by its effect on structures (e.g. windows rattled). Some large animals, such as elephants, use infrasound for communication [1]. Below the deepest range of infrasound—acoustic waves with periods of approximately 60 s, corresponding to wavelengths that are greater than approximately 20 km—buoyancy forces are important, and gravity becomes a key factor governing propagation [2]. Gravity waves, which result when a large parcel of air is displaced vertically (e.g. by air flow over topography or convective uplift), are akin to tsunami waves and propagate at speeds of tens of metres per second. Gravity waves are valued for research because they attenuate slowly and can be detected at ranges of hundreds to thousands of kilometres. As we discuss in this study, gravity waves also appear to play an important role in determining the extent of the infrasonic wavefield that is ducted back to the Earth’s surface and in dispersing and shaping infrasonic wavetrains.

3. Infrasonic sources

As mentioned earlier, we became aware of infrasound in 1883 as acoustic-gravity waves from Krakatoa circled the Earth several times over several days following the eruption [3]. Other notable infrasonic sources in the pre-nuclear age include the Tunguska event above Siberia in 1909 and the explosion of the Baden Aniline and Soda Factory plant in Germany in 1921. Phenomena in the atmosphere or in the Earth’s crust that compress a large volume of air can potentially generate infrasound [4]. These include natural sources (e.g. earthquakes, avalanches, storms and
bolides) and man-made sources, such as rocket launches, supersonic aircraft and explosions [5–9]. Although prior to the nuclear age infrasound was generally regarded as a scientific curiosity, its value for monitoring nuclear tests was recognized shortly after the Trinity test in New Mexico in 1945. Infrasound played an important role in nuclear monitoring until the Limited Test Ban Treaty in 1963 banned nuclear tests above ground. With the signing of the Comprehensive Nuclear Test-Ban Treaty (CTBT) in 1996 infrasound was re-adopted by the nuclear monitoring community as a key tool for monitoring the atmosphere. A key advantage of infrasound over satellites for monitoring nuclear explosions is that clouds are transparent to sound but potentially a significant impediment to satellites.

4. Infrasonic propagation

The propagation of atmospheric infrasound is similar in many regards to what is commonly observed in seismic and hydroacoustic propagation, but there are several key differences (figure 1). As in seismic and hydroacoustic propagation, infrasound compresses the elastic medium in the direction of propagation. It also refracts through large-scale changes in the velocity of the medium, reflects from abrupt velocity contrasts and diffracts or is scattered by small-scale heterogeneity. The speed of sound in m s$^{-1}$ is generally governed by the ambient air temperature and the wind $\vec{c} = 20.05\sqrt{T} + \vec{v}_w \cdot \hat{k}$, where $T$ is temperature in kelvin, $\vec{v}_w$ is the wind velocity and $\hat{k}$ is the unit vector in the propagation direction. It is often useful to relate propagation to the effective sound speed, which is the speed of sound in the horizontal propagation direction $c_e = 20.05\sqrt{T} + v_x$. At 25°C with no wind, the speed is 346 m s$^{-1}$. Sound can be thought of as advected by wind, which leads to faster propagation and higher amplitudes in the direction of the wind. However, in the direction against the wind, infrasound may not return to the surface. When it does return to the surface, the propagation is slower, and the amplitudes are usually lower. The key differences between infrasonic and seismic/hydroacoustic propagation are that infrasonic propagation is much slower, the medium through which infrasound propagates changes significantly with time and the speed of sound is strongly anisotropic because the horizontal winds can often be as fast as 60 m s$^{-1}$ (20% of the static sound speed).
Infrasound is ducted by temperature increases at higher altitudes but mostly by favourable winds in the troposphere and the stratosphere (figure 1) [12–15]. Infrasound can also be ducted within the thermosphere due to its high temperature but is predicted to be heavily attenuated owing to absorption [16]. The atmosphere changes constantly at both fine scales (e.g. gravity waves [17,18]) and large scales [15,19]. Therefore, we would expect noticeable differences in infrasonic waveforms recorded from a series of identical sources at the same location separated by hours or days [20]. Sound is commonly channelled along different paths to the source and arrives in several bundles (i.e. is multipathed). In general, because of the attenuation in the thermosphere, long-range propagation primarily depends on the behaviour of the east–west (zonal) stratospheric wind jet (figure 1). At mid-latitudes, the zonal winds reverse twice per year. In the Northern Hemisphere, the zonal winds are directed from west to east in the winter months, and east to west in the summer months. The stratospheric duct often depends critically on the zonal winds. Little energy is expected to return to the Earth’s surface upwind of the source [12]. Recently, considerable advances have been made in our ability to model this time-varying structure [10,21,22] and in predicting infrasonic propagation through it [14,23,24].

As numerous studies have pointed out, and as we show in more detail later, infrasound has often been observed within ‘shadow zones’ that are predicted by using ray theory and full wave methods with smoothed atmospheric models, such as the standard ground-to-space (G2S) model. A shadow zone is classically defined as an area on the Earth’s surface where no infrasound is detected (figure 1). We redefine the term more broadly for this study to refer to a predicted gap in the extent of an infrasonic arrival branch, which permits a shadow zone to exist in the same location where other arrivals are observed at different times. The atmospheric models infrasonic researchers use result from integration of data from multiple sources (such as radiosondes, numerical weather prediction models, climatologies). As the data are integrated, the atmospheric model is averaged, and only the results at the lowest resolution of all input datasets are preserved. A leading hypothesis is that the missing small-scale structure is critically important in governing the spread of the infrasonic wavefield at the Earth’s surface and the duration of signal packets. Several infrasonic studies have suggested that infrasonic wavefront scattering by internal gravity waves in stratospheric and thermospheric ducts can explain the penetration of sound into these shadow zones [23,25–31]. However, critical testing of this hypothesis requires more observational studies.

5. Infrastructure for studying infrasound

(a) Microbarometers

Infrasound was revived as a field of active study largely due to the CTBT, which was adopted by the United Nations General Assembly in 1996. The CTBT comes with a verification regime (International Monitoring System) that includes networks of sensors to monitor the solid Earth, oceans and atmosphere for nuclear tests. Infrasonics is the key technology for detecting atmospheric tests. The design of the infrasonic network calls for 60 stations, 45 of which have already been constructed (figure 2). Each station includes an array of at least four microbarometers with an aperture of 1–3 km. The advantages of an array over a single sensor are that an array permits the detection of infrasound among higher levels of noise and the characterization of infrasonic signal arrival direction and horizontal propagation speed. Adjacent stations in the infrasonic network are separated by approximately 2000 km. Studies suggest that this design is capable of locating via a cross-bearings approach a 1 kT atmospheric explosion using two or more arrays [32,33]. Additional infrasonic arrays that have been deployed in some regions increase the sampling and increase the total number of arrays operating worldwide to nearly 100. However, even infilled with those regional arrays, the network still severely under-samples the variation in observed arrivals due to even the large-scale structure in the atmosphere such as that which creates shadow zones (figure 1).
Figure 2. The global international monitoring system infrasonic network. As of 31 January 2012, 45 of the stations (black diamonds) are certified and transmitting data to the International Data Centre in Vienna. An additional five stations (white squares) are under construction. The remaining 10 stations (white circles) are planned. Of these planned stations, only nine are shown as the location of one station is not finalized.

(b) Seismometers

Although seismic networks have been used for decades to study earthquakes and probe the structure of the Earth’s interior, they also record atmospheric phenomena, presumably through the acoustic-to-seismic coupling phenomenon [8, 34–39]. The main advantage of infrasonic studies using seismic networks is that the seismic networks are far denser than the global infrasonic array network and are consequently proving very useful for the study of atmospheric phenomena in fine detail [9, 40–42]. With the proliferation of seismic networks, there are over two orders of magnitude more seismic stations worldwide (approx. 16 000 [43]) than infrasonic arrays. Some regions (e.g. Japan, Europe, the continental United States) are particularly well instrumented.

There are two disadvantages of seismic networks in the study of infrasonic propagation. The first is that the coherence distance of infrasound (approx. 200–2000 m) is usually much shorter than the distance between seismic stations. This means that one cannot use precisely the same methods with seismic networks that one uses with infrasonic arrays to detect and characterize coherent infrasound. The second disadvantage is that a single sensor is more susceptible to noise than an array comprising several sensors at the same site. Wind noise is a critical problem that can degrade single-sensor recordings of infrasound (see Walker & Hedlin [44] and references therein).

The studies we present in this study are based on data from two broadband seismic networks in the USA. The first network, the USArray Transportable Array, comprises 400 three-component seismic stations deployed in the continental US on a 70 km spaced Cartesian grid spanning an area of approximately 2 000 000 km² (figure 3). As the name suggests, the network is continually reconfigured by removing stations on the trailing (western) edge and redeploying them on the leading (eastern) edge. Each station remains at one location for 2 years. The second deployment is the 90-station High Lava Plains (HLP) network (figure 3). This network was deployed to study the crustal and upper mantle structure beneath southeast Oregon and southwest Idaho and was fortuitously located along an azimuth of 300° between approximately 280–800 km range from the Utah Test and Training Range (UTTR). UTTR is one of the main sources of infrasound in the western US where the routine detonation of Trident rocket motors on the Earth’s surface occurs, creating repeating explosions of approximately 20 ton equivalent yield.
Figure 3. Two broadband seismic networks in the western US on 16 June 2008. The small grey circles are stations in the USAge Array Transportable Array. The darker, larger, grey stations to the northwest of UTTR (black star) are in the High Lava Plains network. The concentric circles are 200 km distance markers from UTTR. Three infrasonic arrays in the region are represented by the black triangles and are plotted to underscore the difference in sampling between the infrasonic arrays and seismic sensors.

As we discuss later, the infrasonic and seismic networks support distinctly different and complementary avenues of research in infrasound. Each comes with advantages and disadvantages. We exploit the advantage that the seismic networks offer—a relatively dense spatial sampling of the infrasonic wavefield. In this study, we illustrate how we use these networks to infer that the recently developed four-dimensional atmospheric temperature and wind models are generally valid within the model resolution and that the approach for the stochastic inclusion of internal gravity waves in the G2S model is required to explain finer scale propagation effects observed in the seismic data.

6. Seismic detection and location of atmospheric phenomena

There are two commonly used techniques for detecting infrasound that was recorded by multiple sensors. The first technique involves cross-correlation of the waveforms recorded by different pairs of sensors. We defer a discussion of this technique to Cansi [45]. The second method called ‘beamforming’ forms the basis for the reverse time migration (RTM) method described in this study and is used for infrasonic source location by Walker et al. [46]. Generally speaking, plane-wave beamforming for an infrasonic array generates a beam function

\[ B(t, \theta, v) = \sum_{i} a_i(t + t_o(\theta, v, r_x, r_y)), \]

where \( a_i \) is the waveform amplitude, \( n \) is the number of array receivers, \( t \) is time, \( t_o \) is the time shift to align the waveforms at the reference point, \( r_x \) and \( r_y \) are the x and y relative position of...
the receivers, $\theta$ is trial signal azimuth of arrival and $v$ is trial apparent speed across the array. The trial time shift is $t_o = \vec{r} \cdot \hat{k} / v$, where $\hat{k}$ is the unit vector in the trial propagation direction across the array that depends on $\theta$. In other words, for each time sample, trial azimuth and apparent velocity, all array waveforms are projected in space and time to a common reference point (usually at the centre of the array) and summed. For infrasonic arrays, the typical range in $v$ is 300–450 m s$^{-1}$ because infrasound usually propagates at grazing angles to the Earth’s surface such that the apparent speed is close to the true sound speed at the array. The maxima in $B(t, \theta, v)$ indicate possible detections of infrasound at the associated arrival time, azimuth and apparent phase velocity.

The RTM approach is physically quite different, but mathematically just a minor modification of the plane wave beamforming equation. RTM works by ‘back projecting’ or ‘stacking’ energy recorded at the Earth’s surface that travels along the surface at a predicted apparent velocity to all possible source locations on a prescribed spatio-temporal grid (figure 4). Possible event locations in the grid are marked by constructively interfering back-projected energy. RTM originates from the exploration seismology community [47–49]. With the advancement of computers and larger sensor networks, this imaging method is now being applied to many academic, engineering and medical problems. Each problem has a specific focus and context, leading to a variety of different names, including ‘back projection’, ‘source scanning’, ‘time reversal’ and ‘stacking’ [50–55]. For example, Shearer [56] used RTM to detect and locate source of earthquakes by reverse time migrating P-wave, S-wave and surface wave energy generated by earthquakes and recorded by the International Deployment of Accelerometers seismic network. More recently, a similar technique was used to detect, locate and characterize the rupture details of earthquakes [57–61]. RTM has also been used to detect and locate sources of infrasound in the western US [46,62].

Mathematically, RTM is ‘beamforming at the source’ and generates a beam function

$$B(t, s_x, s_y) = \sum_{i} a_i(t + t_o(s_x, s_y, r_x, r_y)),$$

where $t$ is source time and $s_x/s_y$ is source position; the only differences between RTM and plane wave beamforming are that the RTM trial beam parameters are the source longitude and latitude and that the definition of $t_o$ is now the predicted travel time between trial source and receiver.

Infrasound travels horizontally hundreds to thousands of kilometres in stratospheric ducts that are approximately 50 km in vertical height. As a consequence of this geometry, the groups of stratospheric arrivals that return to the Earth’s surface at these ranges have predicted travel times that are nearly linear functions of arc distance. The speed that describes this propagation is called ‘group velocity’. Temperature and horizontal wind lead to a variation in this group velocity of approximately 280 to approximately 320 m s$^{-1}$. At a range of 1000 km, the difference in predicted arrival times for these two velocities is 375 s. If one assumed the same velocity for RTM imaging of all the events, the energy would not always align coherently at the source, and many events would not be detected. A way around this problem is to perform a grid search over optimum group velocity, redefining the beam function

$$B(t, s_x, s_y, v_g) = \sum_{i} a_i(t + d(s_x, s_y, r_x, r_y)/v_g),$$

where $d$ is the arc distance between source and receiver, and $v_g$ is the trial group velocity. This is only a first-order solution; a more accurate approach may be to jointly solve for an analytical horizontal anisotropy model that provides an azimuthal variation in optimum group velocity and maximizes the amplitude of $B$. For the results presented below, $v_g$ ranged from 280 to 350 m s$^{-1}$.

Stratospherically ducted infrasound that is recorded at one station does not exhibit phase coherence with the corresponding infrasound recorded at a station 5 km away; the frequency-dependent coherence of infrasound ranges from 200 to 2000 m. Because the RTM method requires phase coherence between nearby stations in order to achieve constructive interference at the
source location, we pre-process the waveforms to obtain the envelope functions of 1–5 Hz bandpass-filtered data. The envelopes are then decimated to a sampling rate of 10 mHz to regularize the envelope durations. This often results in just a handful of time samples representing several infrasonic arrivals at each station (figure 5). Automatic gain control with a time window 1000 s long is applied to regularize the running maximum amplitude of the decimated envelopes, which prevents a handful of stations with high noise levels from masking the detection of an event. As a result of this automatic gain control, peaks in $B$ that are interpreted as possible events mathematically indicate high coherence of envelope amplitude rather than the sum of envelope amplitude along a suite of possible travel time curves.

For detecting sources in $B$, a detector function $Q(t)$ is calculated that is the maximum value of $B$ at each time sample looking in the $s_x$, $s_y$ and $v_g$ domains. A high-pass filter with a 6 h cut-off period is subsequently applied to $Q$ to remove long-period ‘noise’. The signal-to-noise ratio (SNR)
of each peak in $Q$ is calculated by converting the peak amplitude to decibels with respect to the daily median value. Peaks with values greater than 15 dB are generally considered significant and may indicate possible events. We manually inspected and picked maxima in $Q$ that were classified as detections. A picking interface was developed for this process (figure 6). The analyst initially searches for maxima in $Q$ above approximately 15 dB (figure 6a). A maximum is selected, which defines the optimum event parameters $s_x^*, s_y^*, t^*$, and $v_g^*$, and the selection updates both a map with a cross-section through $B(s_x, s_y, t^*, v_g^*)$ (figure 6b) and a plot of envelope functions versus range from $(s_x^*, s_y^*)$ (figure 6c). The receivers that were used in the image can be shown on the map. The analyst can isolate stations only in certain source-to-receiver azimuth ranges and distances to investigate what stations contribute the most to the peaks in $Q$. Although the analyst also overlays predictions from known earthquake times and positions, visual discrimination between seismic and acoustic-to-seismic signals is relatively easy due to the greater than 10 times faster seismic apparent velocities. Such seismic signals show as nearly horizontal lines in the record section and do not constructively sum along acoustic travel time curves. If the picked acoustic event is considered significant, the analyst assigns the event a grade of A to C, with ‘C’ representing a probable event with a relatively low SNR. ‘A’ events are routinely detected out to at least 500 km range and have SNR above 25 dB, whereas ‘C’ events often have a range out to 200 km and SNR between 10 and 20 dB.

Because of the subjectivity introduced in manually picking events, the bootstrap method is used to quantify the uncertainty in the optimum event parameters [51]. A significant peak in $B$ defines the $i$th event. For each $i$th event, we resample with replacement 100 times the full set of stations used in the location process that defined $B$. This generates a new $B_i$ and $Q_i$. The maximum value over time in $Q_i$ is found, and the associated optimum parameters for this bootstrap-resampled event location are saved. Uncertainties in $(s_x^*, s_y^*)$ are characterized with the

![Figure 5. Processing steps performed on the USArray seismic data prior to RTM. This time window brackets two arrivals from the 19 February 2008 Oregon bolide explosion. For this event, both low-resolution (initial) and high-resolution (final) results were obtained. The final step is the application of an automatic gain control (AGC). From Walker et al. [46].](http://rsta.royalsocietypublishing.org)
smallest geographical uncertainty ellipse that encloses 67 per cent of the 100 bootstrap locations. The uncertainties in $t^*$ and $v^*_g$ are provided by the width of the probability distribution function that encloses 67 per cent of the bootstrap values.

This method was used by Walker et al. [62] to detect and locate 901 infrasonic events that occurred in 2007–2008, defining version 1 of the Western US Infrasonics Catalogue (WUSIC-1). These events cluster spatially around four primary ‘hotspots’: southern Nevada, central Nevada, northwest Utah and southwest Idaho (figure 7). These areas spatially correlate with the regions of military training activities. Possible explanations for these sources include the detonation of ordinance and brief durations of supersonic flight. The RTM method was also attempted with the 2004–2006 USArray, but the number of stations available during this time period, at the 70 km station spacing, gave rise to unacceptably high uncertainties that would have diluted the value of the WUSIC if included.

7. Seismic analyses of atmospheric structure

In the following analyses, we rely on seismic SNR for different tasks. It is important to understand that seismic SNR is not equivalent to infrasonic amplitude. The SNR of the seismic motion imparted by incipient infrasound is well known to be controlled by the infrasonic frequency, elevation angle and complex geology beneath each station (acoustic-to-seismic coupling). Therefore, the absence of a seismic signal does not necessarily preclude the existence of an infrasonic signal (false negatives probably exist). However, the presence of a seismic signal travelling at infrasonic speeds indicates a companion infrasonic signal without question (no false positives), and seismic SNR can be used to map the minimum spatial extent of detected infrasound. Nonetheless, it is worth noting that the stations used in this analysis have all been deployed in vaults above sediments to minimize variations in site effects.
The authors have also compared seismic and infrasonic waveforms for acoustic-to-seismic coupled signals that were recorded by the NVIAR array in western Nevada, which is operated by Southern Methodist University. The waveform structures were often similar, with nearly identical signal durations; seismic coda lengthening owing to reverberation between geological strata under the seismometers was not observed.

The WUSIC-1 represents a new source of infrasonic events for both large-scale and fine-scale propagation studies. Below, we select events from this catalogue to show that large-scale structure observed in the spatial distribution of detected infrasound is consistent with that predicted by the large-scale structure in the G2S atmospheric models. We will also show that the finer details of propagation that are better observed by a very dense IRIS PASSCAL seismic network are consistent with predictions that incorporate the effects of internal gravity waves in the G2S atmospheric model.

(a) Large-scale structure: anisotropy

Two random events from near the southwest Idaho corner were selected to study the stratospheric infrasonic propagation characteristics during the summertime and wintertime (figure 8). Both of these events have unconstrained source location uncertainties. The SNR in decibels is determined for each station by comparing the maximum absolute value of the 1–5 Hz bandpass-filtered data in the signal window to the corresponding median value of the envelope function immediately prior to the signal window. Once again SNR values higher than 15 dB are considered significant;
Figure 8. An example of (a) eastward and (b) westward ducting demonstrated by map views of USArray stations colour-coded by detected signal-to-noise ratio for two WUSIC events. The star indicates the source location determined by RTM. Although the source location uncertainty for these events is undetermined, the dotted ellipse is the largest confidence ellipse found for the WUSIC and provides some insight. The thick black lines indicate the azimuth range used to select envelope waveforms for plotting in figure 9. The concentric circles expand out in 200 km increments. Results from ray tracing using the standard G2S atmospheric model are compared with the observations in (c) and (d). Colour here represents the density of points where stratospherically ducted rays returned to the Earth’s surface (red dots in figures 1 and 9). The classic first shadow zone between 0 and 200 km is clearly shown. Black vectors indicate the mean wind velocity in the G2S stratosphere between 40 and 60 km. Figure modified from Walker et al. [62].

However, because this is an automated process, seismic stations that often experience ‘noise bursts’ will yield artificially high SNR values. Therefore, we manually inspected the energy in the signal and noise windows to cull out such problem stations. As the final step, 15 dB is subtracted from each SNR value so that 0 dB represents the detection threshold.

For the winter event, SNR values up to 45 dB exist to the east southeast of the source location (figure 8a). As expected for amplitude decay owing to geometrical spreading, the SNR decays with increasing range, with a maximum value of SNR at the station closest to the source. Because the source location procedure is not biased by amplitude, we infer from this consistency that the source location is accurate.

The most important feature is that two lines radiating from the station bracket the majority of stations that detected the source (figure 8a). The envelope and bandpass-filtered waveforms for these stations show the alignment of energy along the optimum group velocity (figure 9).
This event is an ‘A’ event, and a coherent envelope signal is observed out to 800 km range. Upon inspection of the 1–5 Hz band-pass-filtered data, and guided by the optimum celerity of 290 m s\(^{-1}\), we interpret individual sets of arrivals as stratospherically ducted phases (\(uS\), \(uS_2\), etc.; figure 1). Three things must be noted regarding the arrival structure. First, the \(uS\) arrival branch begins...
around 100 km and extends to 400 km. There are clear observed arrivals within the shadow zone. Second, the \( uS_2 \) arrival branch extends from 300 to 600 km. Lastly, there is approximately 100 km of overlap between the \( uS \) and \( uS_2 \) branches between 300 and 400 km range. The intrusion into the shadow zone and the latter observation of overlap between \( uS \) and \( uS_2 \) are relevant in the next section on modelling fine-scale structure.

We performed atmospheric ray tracing to predict the spatial distribution of infrasound observable at the Earth’s surface (figures 8c and 9a). The three-dimensional range-dependent environment for this modelling was provided by the G2S-MERRA model [10,21], a product defined by the MERRA model with 0.5° × 0.67° spatial resolution and 72 vertical levels up to approximately 75 km (approx. 1 km vertical resolution) and the HWM07/MSISE00 empirical climatologies above 75 km [22,63]. Rays were launched in all directions away from the source. A two-dimensional histogram of the locations where the rays returned to the Earth’s surface was generated and subsequently averaged using a 50 km wide boxcar filter. Although there is quite a bit of scatter in the SNR values among the predicted shadow zones, the predominant feature to note is that the ray modelling predicts infrasound in the same region as the majority of stations that detected a signal. Vectors indicating the average G2S wind velocity between 40 and 60 km are also shown. The observations and predictions suggest that the infrasound was ducted to the east southeast by stratospheric winds.

For the summer event, SNR values up to 40 dB are found to the west of the source location (figure 8b). A general decay with range is also observed, but the highest SNR values are not closest to the source. Furthermore, if we assume the RTM location is correct, the envelope and bandpass-filtered waveforms suggest that the US arrival starts at 40 km and extends to 270 km range (not shown). Previous ray-tracing results have shown that it is difficult, if not impossible, to observe stratospherically ducted infrasound returning to the surface at ranges as short as 40 km (figure 1). One hypothesis to explain this is that the source that gave rise to this ‘C’ event is mislocated. We therefore move the source 120 km to the east while compensating for this move by adjusting the source time and optimum celerity to realign the envelope energy. We find that the new location is within the southwest Idaho hotspot, is close to a station with a higher SNR, and is now only approximately 35 km from two military bombing ranges: Saylor Creek and Juniper Butte. Furthermore, the new celerity changes from 280 to 300 m s\(^{-1}\), which is now consistent with the summertime average [62].

Ray-tracing results for the summer event predict an observed distribution of stratospherically ducted infrasound to the west of the source (figure 8d). Although there is also a lot of scatter in the SNR values among the shadow zones, the highest SNR values are located at a range of 225 km where the most rays return to the surface. We interpret the arrival branches in the bandpass-filtered waveforms as stratospherically ducted arrivals (figure 9d). The exact start of the \( uS \) arrival branch is difficult to interpret, but it may start by approximately 150 km (also within the shadow zone) and extend to 400 km. The \( uS_2 \) branch is better defined extending approximately from 300 to 600 km. Lastly, 100 km of overlap between \( uS \) and \( uS_2 \) exists approximately between 300 and 400 km range. The observations and predictions suggest that the infrasound was ducted towards the west by stratospheric winds.

In summary, based on the consistency between the observations and predictions for the two randomly selected summer and winter events, we infer that the large-scale features in the G2S atmospheric model are accurate enough to correctly predict the large-scale characteristics of infrasonic propagation, in the 1–5 Hz frequency band, through ground-to-stratosphere ducts.

(b) Fine-scale structure

It is well known that impulsive sources typically give rise to infrasonic arrivals that have signal durations that increase with range up to several tens of seconds long. In the absence of additional information, these durations could be interpreted as a result of forward scattering within the Fresnel zone surrounding the ray path or multipathing due to slightly different take-off angles that have a significant influence on the ray path geometry.
Over the past decade, since the advent of the time-varying, three-dimensional G2S atmospheric velocity model [21], it has routinely been observed that shadow zones predicted by propagation modelling of impulsive wavefronts through G2S models are not observed in reality. In other words, sensors sometimes observe infrasonic energy between 1 and 5 Hz in these regions where the G2S model predicts no energy to exist. As with signal durations, the observed ‘infilling’ of G2S-predicted shadow zones can be interpreted to be a result of either diffraction or scattering of the infrasonic wavefront or inaccuracies in the G2S model.

Although few of the source mechanisms in the WUSIC events catalogue have been identified, we know from their spatial clustering and tendency to occur during local working hours that most are man-made. A subset of these events, if not most of them, are surface explosions associated with routine military training activities. Therefore, impulsive WUSIC events may be useful for studies of infrasonic arrival durations and shadow zones.

In this section, we test the hypothesis that the dispersion in time of infrasound and the predicted shadow zones at the edges of arrival branches can be explained due to the lack of fine-scale, layer-cake structure in the G2S model that ray-trace modelling is very sensitive to (because propagation is nearly horizontal; figure 1). For this, we analyse a WUSIC event that is known to be an explosion with a precisely known source location and time. The explosion occurred at the UTTR site in western Utah, where Trident rocket motors with explosive yields of approximately 20 tons are routinely detonated as required by a US–Russia treaty. All Trident rocket motor explosions at UTTR occur during the summer so that infrasonic energy will be carried to the west, away from Salt Lake City, by the summertime zonal winds. For this analysis, we take advantage of the very dense spatial sampling afforded by the HLP seismic network as well as that provided by the USArray (figure 3) to look in fine detail at the seismic definition with range of the arrival branches (figure 10). As mentioned earlier, because we are relying on the seismic detection of this infrasound, we observe only the minimum extent of the arrival branches. The HLP network was deployed from 200 to 800 km along an azimuth of 300° from UTTR. A handful of USArray stations provide additional observations from 0 to 200 km and at greater ranges along this azimuth.

The explosion occurred on 16 June 2008 at 20.32.27.38 UT. Seismic energy in the 0.8–3.0 Hz band that exhibits a group velocity of about 300 m s\(^{-1}\) can be observed out to approximately 800 km range (figure 10). We observe four main branches of infrasonic arrivals. Ray tracing through the G2S model along the azimuth of the network provides travel time predictions (green dots) that match the majority of the observed arrivals, suggesting propagation in a stratospheric duct. The earliest to latest arriving branches are \(uS_1\), \(uS_2\), \(uS_3\) and \(uS_4\), respectively. Looking at branch \(uS\), we observe a significant, and rapid, decrease in SNR away from the source and the splitting of the branch into two sub-branches (\(uS'\) and \(uS''\)) near a range of 400 km. Branches \(uS_2\) and \(uS_3\) gradually increase in SNR from onset near the source and then decay below noise at the distant edge. Branch \(uS_4\) emerges from noise near a range of 700 km but is not observed in the seismic data beyond 800 km. We also observe particularly long signal packets (up to 80 s) at the leading edge of branch \(uS_2\).

The agreement between the G2S arrival times and the signal onset times is excellent. However, the G2S arrivals do not explain the duration of the signal packets. Also, the G2S arrivals do not explain the observed extended ranges of the branches; there is significant penetration of observed infrasound at the edges of all four branches into the adjacent G2S-predicted shadow zones.

Drob et al. [26] presented an approach based on the Naval Research Laboratory’s physics-based Maslov method to populate background G2S models of wind and temperature with physics-based realizations of small-scale internal gravity waves (IGWs). In their method, an empirically derived gravity wave source spectrum based on Warner & McIntyre [64] is used to initialize a random realization of the gravity wave structure in the troposphere. Multiple realizations of the gravity wave structure share the same statistical characteristics. A spectral ray-tracing technique adapted from Broutman et al. [65,66] is then used to propagate the gravity waves upward through the rest of the atmosphere. The result is a single realization of a space- and time-varying perturbation model, with stable statistical characteristics, that is then added to the lower-resolution G2S model.
Figure 10. Vertical component recordings of an explosion that occurred at the UTR site on 16 June 2008. The recordings were made by stations in the HLP network and the USArray that were located within a 150 km wide corridor from UTR along an azimuth of 300°. The waveforms have been bandpass-filtered from 0.8 to 3.0 Hz and normalized by pre-signal noise levels; the waveform amplitude is the relative signal-to-noise rather than true amplitude. The green dots are predicted travel times of rays that were shot from UTR through the standard G2S model. The red dots represent rays shot through one realization of gravity waves added to the same G2S background model. Just rays that landed in the 150 km wide corridor at 300° were included. Recordings from the HLP network made near 800 km from UTR were not included in this figure due to low SNR.

A second set of rays shot through the perturbed (G2S-IGW) model are shown in figure 10 (red dots). The G2S-IGW and G2S arrivals differ in several ways. First, the G2S-IGW arrivals are observed at closer ranges than the G2S arrivals, filling in the classic first shadow zone (figures 1 and 8). Similarly, the G2S-IGW arrivals extend each edge of the arrival branches several tens of kilometres in range, penetrating the adjacent shadow zones predicted by G2S. In all these cases, the range extent of the arrival branches defined by G2S-IGW overlap the observed arrivals much more completely than those defined by G2S.

The majority of the G2S-IGW arrivals are also delayed and spread out in time relative to the G2S arrivals. In all of these cases, the precise timing and duration of these G2S-IGW arrivals match the arrivals of the time-dispersed energy that originated from the explosion.

In summary, ray-trace modelling through the relatively low-resolution G2S model predicts the observed, large-scale anisotropy of propagation in a stratospheric duct. However, the modelling does not predict the finer details of the spatio-temporal extent of the infrasonic
wavefield. Ray-trace modelling through a G2S model that has been stochastically modified by the statistical effects of internal gravity waves explains more completely the observed spatio-temporal distribution of infrasound in the 0.8–3 Hz band.

8. Discussion

(a) Large-scale ground-to-space structure predicted observed anisotropy

We show that rather elementary processing methods allow us to uncover a distinctly different, atmospheric type of source in the dense seismic network data. The large catalogue of atmospheric events (WUSIC) produced by Walker et al. [62] provides the foundation for large-scale, statistical studies of propagation of infrasound in the atmosphere.

Our above analyses of a subset of WUSIC events indicate the validity of the large-scale structure in G2S atmospheric model. This observation is supported by a larger, systematic study of a suite of explosions at the UTTR site [39]. These G2S models appear to accurately register variations in the atmospheric temperature and wind fields—in particular, the seasonally varying stratospheric winds that are known to be key in the ducting of infrasonic energy back to the Earth’s surface. This is significant because the stratospheric winds in the G2S model above approximately 30 km altitude are not directly measured, but are predicted only from direct measurements of pressure and temperature. Although it is generally thought that the large-scale stratospheric winds in the MERRA model between 35 and 65 km are correctly predicted at mid latitudes, because the general air circulation is uncoupled from the effects of topography/clouds and because gravity wave effects are less important at such large scales (D. Drob 2012, personal communication), our results at two snapshots in space–time provide an empirical validation of the inferred winds in the G2S model.

(b) Fine-scale internal gravity wave structure predicts observations via multipathing

Our study provides further evidence, however, that these models are insufficient for predicting the finer structure of the infrasonic wavefield and the duration of infrasonic arrivals. The dense networks that we have used allow us to examine in detail the emergence of stratospherically ducted infrasonic branches with increasing range from the source and their eventual decay below noise. The seismic data, which can show only the minimum range extent of infrasonic arrival branches, clearly show infrasonic arrivals at ranges predicted to be shadowed by ray tracing through the standard G2S model. In a related study [39], full-wave synthetics in the 0.8–3 Hz band computed using the same G2S model also failed to match the observations. The range definitions of the shadow zones predicted by the full-wave synthetics and rays were similar.

The seismic stations provide many independent observations of the arrival time and signal duration. The signal durations are considerably longer than those predicted by ray tracing through the standard G2S model. Other studies also show that full-wave synthetics generated using these basic models do not reproduce the duration of the observed signals [31,39].

In the spirit of Occam’s razor, the simplest explanation of the penetration of sound into the shadow zones and the prolonged infrasonic arrival durations (pulse lengthening) is that both are due to the interaction of the infrasonic wavefield with small-scale, internal gravity waves in the atmosphere. This conclusion is also obtained by earlier studies ([31] and references therein). If this is correct, the key unanswered question is how exactly are the shadow zones infilled and pulses lengthened? Previous studies have interpreted scattering from small-scale structure to be the dominant mechanism [29,31]. Our ray-tracing analysis suggests that simple multipathing may be the dominant mechanism, at least for the 0.8–3 Hz wavefield where the wavelengths range from 70 to 410 m. The three-dimensional ray tracing used in this study captures the full range-dependent physics of travel time prediction for wavelengths that are smaller than approximately 10 per cent of the size of obstacles encountered by the wavefront. Because propagation is nearly horizontal, and because the gravity wave structure is elongated in the horizontal relative to...
the vertical, we consider the vertical variation to be most important. The dominant vertical wavelength of the gravity wave spectrum added to the G2S model, which is quite strongly amplified in the stratosphere, is approximately 3 km. Using half a wavelength, 1.5 km, as the size of the perturbations experienced by nearly vertical wavefronts, ray tracing can adequately describe propagation for wavelengths less than approximately 150 m (frequencies above 2 Hz). As frequency decreases from 2 Hz, the importance of scattering theory increases for propagation with the specific G2S-IGW model used above. It should be noted that the largest analysed infrasonic wavelength is still only 27 per cent of the size of the perturbations. Rayleigh scattering for the above model, which begins when the ratio of the infrasonic wavelength to the perturbation size is close to unity, begins when frequencies reach approximately 0.2 Hz.

The fact that most observed G2S-IGW arrivals were delayed with respect to the G2S arrivals provides some insight in the propagation. The path taken by a ray through the heterogeneous atmosphere is critically dependent on the takeoff angle from the source. In the above study, a fan of 6001 rays were shot from the source at each of 360 azimuths. As each of these rays propagate along different paths through heterogeneity, they will tend towards slow regions and away from fast regions. Because the gravity waves are a random perturbation to the background G2S model, the additional slower anomalies created by the IGWs should produce a significant amount of pulse lengthening.

Lastly, the UTTR example analysed above, as well as those analysed in Hedlin et al. [39], may also provide insight into observations of propagation for the Idaho events. The overlap of $uS$ and $uS_2$ between 300 and 400 km that is explained by G2S-IGW may also explain the 300–400 km overlap for both Idaho events (figure 9).

(c) Implications for operational monitoring

If our above analyses hold up upon further scrutiny, then operational centres tasked with monitoring infrasonic recordings for signals of interest can use fast ray-tracing algorithms with multiple realizations of G2S-IGW to predict infrasonic signal envelope functions. Such functions not only provide empirical probability distribution functions, which can be used to improve source location programs, but they can also be used to improve detection capability by defining matched filters that can be implemented with multiple arrays simultaneously to search for signals of interest. In addition, if the large-scale anisotropy of the G2S model is further validated, especially at other latitudes, the enhanced G2S-IGW model can provide more stable predictions of back azimuth deflection due to winds that blow across the source–receiver plane. Such corrections are of critical importance to the accurate estimation of infrasonic source location via a cross-bearing approach using a handful of globally spaced arrays.

9. Conclusions

Our analyses of dense seismic network data for several events in the WUSIC show good agreement between observations and predictions. Randomly picked winter and summer infrasonic events that occurred in southwest Idaho were observed to a range of approximately 800 km in the southeast and west directions, respectively. Ray-trace propagation modelling with the standard G2S global atmospheric model predicts the spatial distributions of both sets of seismic observations fairly well, suggesting that the stratospheric winds between 30 and 50 km in the G2S model, which are inferred only from pressure and temperature measurements, are generally correct at the latitudes spanned by the Pacific northwest states. Although the large-scale G2S structure appears to be correct, ray tracing through this structure often predicts shadow zones, or gaps in the infrasonic arrival branches, that are not observed in reality. An explosion from a site in northern Utah generated infrasound that was observed by a dense seismic network out to approximately 800 km to the west northwest. Indeed, G2S ray-trace modelling predicts shadow zones that are not observed. However, when one traces rays through a G2S model that includes the effects of IGWs, shadow zones are not predicted, and the G2S-IGW arrivals explain
much more completely the observed spatial and temporal distribution of infrasound. The success of rays in explaining these observations with the G2S-IGW model suggests that multipathing may be the dominant physical mechanism governing propagation, at least for frequencies above approximately 1 Hz and for the particular G2S-IGW model used above.

The apparent ease with which seismometers in the USArray and HLP networks have detected signals from atmospheric phenomena serves as a reminder that the Earth is an interconnected system. Air pressure variations owing to a wide range of atmospheric phenomena, including wind, acoustic signals, gravity waves and meteorological phenomena, interact with the surface of the Earth and readily transmit to seismic variations. Consequently, dense seismic networks provide an additional tool for the study of atmospheric phenomena, the nature of infrasonic wave propagation from these sources and the structure of the atmosphere. Although seismic recordings require an added layer of care in their interpretation, seismic networks routinely provide considerably greater sampling of the infrasonic wavefield than that achievable by the global infrasonic array network. The infrasonic network directly records infrasound with higher fidelity, but the sparseness of the network makes it difficult to study important aspects of infrasonic propagation. Indeed, it seems clear that each source of data helps to address weaknesses of the other, and the two can be used jointly to further our understanding of the nature of our atmosphere and phenomena occurring within it.

Ray-tracing results for figure 10 and the G2S models used throughout this study were provided by Dr Doug Drob (Naval Research Laboratory, Washington, DC). Relu Burlacu (at the University of Utah) provided ground-truth information about the UTTR explosions. Dr Matt Fouch (Arizona State University) and Dr David James (Carnegie Institution of Washington) provided access to data from their High Lava Plains Seismic Experiment. Earthscope and IRIS provided data from the USArray. Discussions with Drs Doug Drob and Roger Waxler benefitted this study. This research was supported by the US National Science Foundation grant no. EAR-1053576.

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