

Seismological Reconstruction of Meteoroids: Trajectories, Physical Parameters, and Strewn Fields in Wind-Perturbed Atmospheres

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Abstract

Every day, the Earth’s atmosphere is struck by meteoroids, and the recovery of fresh meteorite material is the main way of improving our understanding of how the Solar System and the Earth formed. However, meteorite recovery is often limited because not all meteorite-producing meteoroids are observed. Meteoroid detection is currently dominated by optical fireball networks, whose individual cameras are affected by clouds, daylight and spatial coverage gaps. Dense, globally distributed seismic networks offer a time- and weather-independent alternative. Since meteoroids continuously generate acoustic waves during their supersonic flight, the resulting air-to-ground coupled seismic waves can be used to reconstruct their atmospheric path. This provides a distinct physical method of constraining meteoroid parameters. However, exploiting this approach requires a new framework to systematically detect meteoroid signals in seismic data and reconstruct trajectories in wind-perturbed atmospheres. Only then can the seismic method reliably complement the current approaches to estimating meteoroid parameters, trajectories, and meteorite ground locations.

In this work, we outline the development and application of this end-to-end framework, which covers meteoroid detection to meteorite strewn-field reconstruction. First, we compile the Seismo-Acoustic Meteoroid Observation Recording Database (SMORD), which contains 307 meteoroid events and 3,295 manually labelled arrivals across a global set of stations. We then use this database to train a PhaseNet-based deep-learning phase picker in SeisBench. The model achieves 91% precision and 94% recall (AUC 0.89), with a median absolute timing error of 0.02 s. We demonstrate how this model can automatically detect meteoroid onsets and reconstruct the trajectory of a meteoroid over the Adriatic Sea.

We then improve the trajectory reconstruction method by using the uncontrolled re-entry of the Starlink-2382 satellite as a meteoroid equivalent. By incorporating a 3D wind model and a lift-like parameter into a ray-based direct-search optimisation, we significantly improve the travel-time fit to recording data from 226 stations. This allows us to recover an entry velocity of $8.9 \pm 0.7 \text{ km s}^{-1}$ and an effective ablation coefficient consistent with near-complete ablation of the $\sim 100 \text{ kg}$ main satellite fragment. Using collocated acoustic and seismic sensors, we found that the acoustic-to-seismic ground coupling coefficients vary by up to three orders of magnitude (from $4.31 \times 10^{-10} \text{ m s}^{-1} \text{ Pa}^{-1}$ to $5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$), mainly due to differences in near-surface rock stiffness.

Finally, we use probabilistic Markov chain Monte Carlo ensemble sampling to perform a statistically robust reconstruction of the 8 March 2026 Koblenz meteoroid trajectory, which was recorded at 95 stations. Replacing direct ray computation with a precomputed lookup table reduces new ray calculations to simple interpolation, enabling this full Bayesian inversion of the meteoroid trajectory. With this trajectory and the associated uncertainties that can be propagated, we recover the heliocentric orbit and initial mass of $305 +180/-90 \text{ kg}$ of the

meteoroid. Additionally, we identify five distinct fragmentation events along the atmospheric trajectory and calculate a strewn field, which is validated against fragments that struck a building in Koblenz-Güls. Together, these results demonstrate the successful application of the end-to-end acoustic-seismic framework which can shape future meteoroid trajectory reconstructions and enables the rapid, automated calculation of meteorite strewn fields.

Contents

- Abstract i**
- List of Figures vii**
- List of Tables ix**
- List of Acronyms xi**
- List of Symbols xiii**
- 1. Introduction 1**
 - 1.1. Research Aims and Motivation 1
 - 1.2. Outline of the thesis 3
- 2. Meteoroids: An introduction to terminology, origins, flux and physical properties 5**
 - 2.1. Terminology: Meteors, Meteoroids, and Meteorites 5
 - 2.2. Origins of Meteoroids 6
 - 2.3. Meteoroid Flux 8
 - 2.4. Physical Properties of Meteorites 9
- 3. Theoretical background 11**
 - 3.1. Observational Methods and Detection Capabilities 11
 - 3.1.1. Optical Detection 11
 - 3.1.2. Radar Detection 12
 - 3.1.3. Human Visual Observation 13
 - 3.1.4. Infrasound Detection 14
 - 3.1.5. Seismic Detection 15
 - 3.2. Acoustic-Seismic Wave Emission by Meteoroids 16
 - 3.2.1. Ballistic Mach Wave 16
 - 3.2.2. Explosive Fragmentation Events 18
 - 3.2.3. Impact 19
 - 3.2.4. Acoustic-to-seismic Coupling 20
 - 3.2.5. Meteoroid Blast Radius 26
 - 3.3. The Meteor Light Curve 27
 - 3.4. Luminous Efficiency 28
 - 3.4.1. Definition and the Luminous Equation 28
 - 3.4.2. Dependence on Entry Velocity and Physical State 29
 - 3.4.3. Practical Determination of Luminous Efficiency 30

4. Data and Methods	31
4.1. Data	31
4.1.1. Atmospheric Data and Models	31
4.1.2. Waveform Data	32
4.2. Methods	34
4.2.1. Acoustic-Seismic Waveform Processing	34
4.2.2. Trajectory Reconstruction	35
4.2.3. Markov chain Monte Carlo and Bayesian Inference	41
4.2.4. Extending the inferred Meteoroid Trajectory	43
4.2.5. Deep-Learning Phase Identification with PhaseNet and SeisBench	46
5. Seismo-Acoustic Meteoroid Observation Recording Database	49
5.1. Abstract	50
5.2. Non-technical summary	50
5.3. Introduction	51
5.4. Data	53
5.4.1. Meteoroid event catalogs	53
5.4.2. Waveform retrieval and preprocessing	53
5.5. Methods	54
5.5.1. Manual phase picking and labeling protocol	54
5.5.2. Model training with SeisBench PhaseNet	55
5.6. Results	57
5.7. Application	61
5.8. Discussion	63
5.9. Conclusions	67
6. Re-entry and Burn Up of Starlink-2382 Satellite	69
6.1. Abstract	70
6.2. Introduction	70
6.3. Methods	75
6.4. Trajectory and Ablation Modelling	78
6.5. Results	80
6.5.1. Trajectory	80
6.5.2. Finite difference modelling of wave propagation	83
6.5.3. A-S Ground Coupling	85
6.5.4. Ablation estimation	87
6.6. Discussion	90
6.7. Conclusions	91
7. The Koblenz Meteoroid	95
7.1. Abstract	95
7.2. Introduction	96
7.3. Trajectory reconstruction and fragmentation localisation	97
7.4. Strewn Field	100
7.5. Pre-atmospheric heliocentric orbit	100

7.6. Data	102
7.7. Methods	104
8. Summary and Outlook	107
8.1. Summary	107
8.2. Outlook	111
9. Software References	117
9.1. Software references to chapter 5	117
9.2. Software references to chapter 6	117
9.3. Software references to chapter 7	118
10. Network References	119
10.1. Network references to chapter 5	119
10.2. Network references to chapter 6	131
10.3. Network references to chapter 7	133
Bibliography	135
A. Supplement to chapter 6	155
B. Supplement to chapter 7	161

List of Figures

2.1.	Overview of meteoroid entry stages.	5
2.2.	Asteroid Belt location	6
2.3.	Size-frequency distribution of meteoroids	9
3.1.	Apparent magnitudes of major planets	12
3.2.	Acoustic-seismic wave radiation mechanisms	16
3.3.	Acoustic N-wave	17
3.4.	Dynamic ram pressure over altitude of the Koblenz meteoroid.	18
3.5.	W-shaped wavelet.	21
3.6.	Critical angle versus Rayleigh wave phase velocity.	22
3.7.	Meteoroid-related surface wave on a seismic waveform recording.	24
3.8.	Average light curve out of 4 meteor cameras	27
4.1.	G2S atmosphere specifications.	31
4.2.	Broadband 3-Component seismic recording data.	33
4.3.	Coordinate system for geometrical trajectory inversion	36
4.4.	InfraGa ray geometry.	37
4.5.	Coordinate system for ray-based trajectory inversion	38
4.6.	Beamforming of a meteoroid event	40
4.7.	Illustration of MCMC ensemble sampling.	42
4.8.	Strewn field comparison between optical and seismic trajectory solutions.	44
4.9.	PhaseNet application to a meteoroid signal.	47
5.1.	Vertical component waveform examples of meteoroid signals	52
5.2.	World map of SMORD events	58
5.3.	Statistics of SMORD metadata	59
5.4.	Training results	62
5.5.	Distribution of PN pick probabilities	63
5.6.	Fireball event on 12 th April 2025 across the Adriatic Sea	64
6.1.	Map of the target region in Central Europe	72
6.2.	Observed recordings at Raspberry S & B station AM.R03EA	73
6.3.	Contour Map of acoustic and seismic arrivals	74
6.4.	Grid of atmospheric models	75
6.5.	Map showing different trial trajectories of the geometrical model reconstruction	77
6.6.	Correlation matrix for the eight inverted parameters	81
6.7.	Map showing different trial trajectories of the ray model reconstruction	82
6.8.	Time evolution of the wavefield	85

6.9.	Comparison between the ray and wave model	86
6.10.	Map of collocated A-S recording sites	88
6.11.	Mass over time evolution for different parameters	89
7.1.	Sensor distribution, trajectory and acoustic wave of the Koblenz Meteoroid	98
7.2.	Fragmentation events of the Koblenz Meteoroid	99
7.3.	Meteorite strewn field	101
7.4.	Orbit views of the Koblenz meteoroid	103
8.1.	End-to-end seismic meteoroid trajectory characterisation framework.	108
A.1.	Map of photograph and video locations	155
A.2.	Map with selected ray paths	157
A.3.	Modelled Mach carpet	158
B.1.	MCMC Corner Plot	162
B.2.	Atmospheric state	163
B.3.	Orbital elements and auxiliary orbit-defining parameters	165

List of Tables

3.1.	Complementarity of meteoroid observation techniques	16
5.1.	Guidelines for decision-making for pick quality	54
5.2.	Metadata of a phase pick stored in SMORD	55
5.3.	PN picker performance metrics (modified after Sokolova and Lapalme 2009) . . .	60
5.4.	Comparison of FRIPON trajectory parameters	65
6.1.	Comparison of trajectory parameters	83
6.2.	Parameters to estimate ground compliance	87
6.3.	Comparison of ablation coefficients	90
A.1.	Table of satellite fragments	156
B.1.	Trajectory MAP estimate parameters	164

List of Acronyms

Acronym	Description
AU	Astronomical Unit
CTBTO	Comprehensive Nuclear-Test-Ban Treaty
ECMWF	European Centre for Medium-Range Weather Forecasts
FRIPON	Fireball Recovery and InterPlanetary Observation Network
G2S	Ground-to-Space
HWM	High wind model
IMO	International Meteor Organisation
IMS	International Monitoring System
LTA	Long term average
MCMC	Markov chain Monte Carlo
PMCC	Progressive multi-channel correlation
PSO	Particle Swarm Optimization
SNR	Signal-to-noise ratio
STA	Short term average

List of Symbols

Symbol	Description
x, y, z	Generic coordinates
N	Generic integer number
a	Semi-major axis
A	Shape factor
C_D	Drag coefficient
c_{eff}	Effective speed of sound in a given azimuth direction
c_s	Ambient speed of sound
$\mathbf{d}$	Set of observations (Bayesian Inversion)
d_m	Meteoroid diameter
e	Eccentricity
E_k	Initial kinetic energy
E_l	Energy per unit path length
E_{opt}	Optical energy
G	Gravity constant
i	Inclination (orbit parameter)
m	Meteoroid mass
M	Mach number
$\hat{n}$	Ray direction
p_0	Ambient pressure
p_{ram}	Dynamic ram pressure
R_0	Blast radius
$R_{x,z}$	Rotation Matrices
t	Time
T_J	Tisserand Parameter
T_P	Pressure transmission coefficient
U_w	Zonal wind component
v	Velocity
v_{obs}	Observation velocity
v_P	P-wave velocity
v_S	S-wave velocity
v_R	Rayleigh-wave velocity
v_g	Group velocity
v_∞	Pre-atmospheric velocity
V_w	Meridional wind component

Symbol	Description
Γ_A	Drag-shape factor
μ	Standard gravitational parameter
ρ_a	Ambient atmospheric density
ρ_m	Meteoroid density
σ	Ablation coefficient
τ	Luminous efficiency
ϑ	Inclination angle
$\boldsymbol{\theta}$	Bayes parameter vector
θ	Elevation angle
θ_c	Critical angle
θ_M	Mach Cone opening angle
ω	Argument of perihelion
Ω	Longitude of ascending node
$\odot$	Sun
$\oplus$	Earth

1. Introduction

1.1. Research Aims and Motivation

This thesis presents research motivated by the growing need for robust, scalable and weather-independent methods to detect, characterise and reconstruct the trajectories of meteoroids entering the Earth's atmosphere. The central aim is to provide a comprehensive framework for recovering meteoroid trajectories, physical parameters and strewn field estimates from acoustic-seismic network data. While these tasks were not completely inaccessible to seismological methods in the past, the lack of physically realistic assumptions for the meteoroid trajectory and the associated acoustic waves led to large uncertainties in the reconstructed meteoroid trajectories and thus also in the strewn field estimates. The work presented here is motivated by six key considerations, each of which highlights the scientific importance and practical necessity of the methods developed and presented in this work.

First, the global land coverage of seismic and infrasound stations provides far greater coverage than the camera and radar networks currently used for meteor detection (Eickhoff et al., *in press*). While optical camera networks are effective, they are limited in spatial coverage and subject to gaps in deployment that leave large portions of Earth's surface unmonitored. Seismic and infrasound stations, by contrast, are distributed across nearly every continent, and their data are continuously archived and mostly openly accessible, while meteor camera networks mostly lack this accessible open data approach. However, to harness this existing infrastructure for meteoroid science, new methods must be developed to reliably identify meteoroid-related signals within seismic waveforms and physically consistently invert those waveforms to constrain meteoroid parameters in a data-driven and also sufficiently fast manner. Developing and validating such methods is one of the primary contributions of this thesis.

Second, seismic and infrasound stations operate continuously and under all atmospheric conditions, providing a fundamental operational advantage over optical systems. Camera networks are only operable during the night and in clear weather conditions, meaning that many meteoroid events, including those occurring during the day or in overcast areas, are not detected or are poorly characterised by existing observational frameworks. However, acoustic-seismic networks are unaffected by these diurnal and weather limitations, making them a reliable means of building a complete and unbiased catalogue of meteoroid activity. Additionally, as the acoustic waves emitted from meteoroids follow a directivity, due to wind and atmospheric stratification, seismic and infrasound data can provide accurate back-azimuth information from the signal alone. This physical property of the radiated wave cannot be detected by optical observation methods because the light produced by meteors is radiated more uniformly.

Third, acoustic-seismic observations are physically independent of optical and radar measurements, which has important implications for validation and joint analysis. Derived products, such as trajectory estimates or physical parameters obtained from acoustic-seismic waveforms, can be cross-checked against those obtained using optical and radar methods. This provides a means of identifying systematic biases and improving the overall reliability of meteoroid characterisation. Likewise, when multiple data types are available for the same event, acoustic-seismic, optical and radar observations can be combined in a joint inversion framework to more tightly constrain meteoroid parameters than any single method could achieve alone.

Fourth, acoustic-seismic methods are particularly well suited for detecting and characterising meteoroids within a mass range that some existing techniques struggle to cover. Objects with masses between approximately 10^3 g and 10^6 g generally cannot be discovered easily before atmospheric entry using telescopic surveys, yet they are large enough to produce well detectable acoustic-seismic signals upon entry (Waszczak et al., 2017). Improving our understanding of the flux and physical properties of this population would have direct implications for planetary defence and models of the small-body impactor environment (Drury et al., 2026). Acoustic-seismic methods are one of the most practical ways of systematically characterising this class of objects, especially when jointly used with optical and radar techniques.

Fifth, and perhaps most consequential for the broader scientific community, the trajectory constraints derived from acoustic-seismic data can be used to estimate the spatial distribution of meteorite material on the ground: the strewn field. Accurate and rapid strewn field prediction is essential for enabling the timely recovery of meteorites, as weathering of the meteoroid starts immediately after entering the Earth's atmosphere and their scientific value diminishes with exposure time (Bland and Artemieva, 2006). In order to achieve this, the analysis chain from raw waveform data to strewn field estimation must be fast enough to guide field teams within days of an event. A key outcome of this thesis is the development of such an analysis pipeline. Such a pipeline should provide an end-to-end workflow that can be applied to new events with minimal manual intervention (see Chapters 5, 7).

Sixth, collocated acoustic-seismic observations provide a unique insight into the process by which acoustic waves are converted into seismic waves. The energy of an atmospheric pressure wave generated during a meteoroid's entry into the atmosphere is transferred into the solid Earth upon ground incidence (see Chapter 6). A sufficiently detailed understanding of this process would, in principle, allow a seismic meteoroid magnitude scale to be constructed, analogous to those used in earthquake seismology. A magnitude scale for meteoroids could be a valuable tool for estimating the energy of a meteoroid entry directly and consistently from seismic amplitudes alone. Thus it would greatly simplify the rapid characterisation of new events and facilitate systematic comparison across the growing catalogue of acoustically detected meteoroids.

These motivations define the scope and structure of the scientific contributions presented here. The end-to-end pipeline begins with a probabilistic, deep-learning phase identification model that has been trained using data from more than 300 meteoroid events. This model can automatically identify meteoroid-related arrivals in continuous seismic recordings. These phases can then be fed into an ensemble-based Bayesian inversion scheme, which generates probabilistic posteriors for parameters that define the meteoroid. This scheme explicitly

accounts for full atmospheric wind fields through state-of-the-art ground-to-space atmospheric models. Together, these two components enable the determination of meteoroid trajectories, pre-Earth heliocentric orbits and meteorite surface locations within hours to days of a new event. Beyond trajectory reconstruction, the framework provides access to physical parameters such as ablation (i.e., the mass loss rate) properties and meteoroid mass. These quantities were previously unrecoverable from acoustic-seismic data alone. Thus, it establishes acoustic-seismic network analysis as a scientifically robust and operationally feasible addition to optical and radar methods in the broader effort to characterise the meteoroid environment around Earth. Additionally, the methods developed in this thesis lay part of the groundwork for the effort of a future seismic meteoroid magnitude scale.

1.2. Outline of the thesis

This thesis begins with a brief introduction to the topic of meteoroids. It covers their origins within the solar system, the mass flux to which the Earth is continuously subjected across a range of sizes and energies, and the physical and compositional properties of meteorite rocks. This chapter establishes the necessary background knowledge for subsequent technical discussions. I then discuss in detail the theoretical foundation required to detect, measure and analyse meteoroids using various observational methodologies. Particular attention is given to the physical principles underlying each technique, as well as their respective capabilities and limitations. This is followed by a section on the generation of acoustic and seismic waves by meteoroids. This section describes the mechanisms by which energy is coupled into the solid Earth and atmosphere and explains how the resulting signals can be used to characterise meteoroid bodies. A chapter on the data sources and principal methods shared across the three studies is then presented to provide a unified methodological framework that highlights the core scientific contributions of this work.

The subsequent three chapters each present one of the research studies in full, covering its motivation, dataset, discussions and findings. The first study describes the construction of a seismo-acoustic meteoroid observation recording database, which focuses on the training and validation of the deep-learning phase identifier. The second study presents the seismic and acoustic analysis of the re-entry and burn-up of a Starlink satellite. Due to its known orbital and physical properties, the satellite serves as a well-constrained analogue for a natural meteoroid event. The re-entry provides a rare opportunity to benchmark trajectory reconstruction methodology and discuss the acoustic-to-seismic coupling. The third study presents the full Bayesian acoustic-seismic characterisation end-to-end pipeline, which was applied to an especially well-instrumented meteoroid event near Koblenz. The study demonstrates the joint inference of trajectories, physical parameters, and strewn fields from acoustic-seismic network data in wind-stratified atmospheres and shows how quickly trajectory solutions can be generated, and how uncertainties must be carried downstream to subsequent analyses. Finally, a summary and outlook section synthesises the advancements and contributions made across all three studies, reflecting on their broader implications and outlining how future investigations can build on this work.

2. Meteoroids: An introduction to terminology, origins, flux and physical properties

2.1. Terminology: Meteors, Meteoroids, and Meteorites

The study of small solar system bodies that interact with the Earth's atmosphere relies on a set of terms that are unfortunately often used interchangeably. However, it is important to define them sharply to not confuse light phenomena with rock fragments. This distinction is especially important in acoustic-seismic studies, because while optical and radar observation methods observe the meteor (the light phenomenon), acoustic-seismic methods observe the acoustic and acoustic-seismic coupled waves. Both types of waves are produced by the meteoroid body (the rock body in flight).

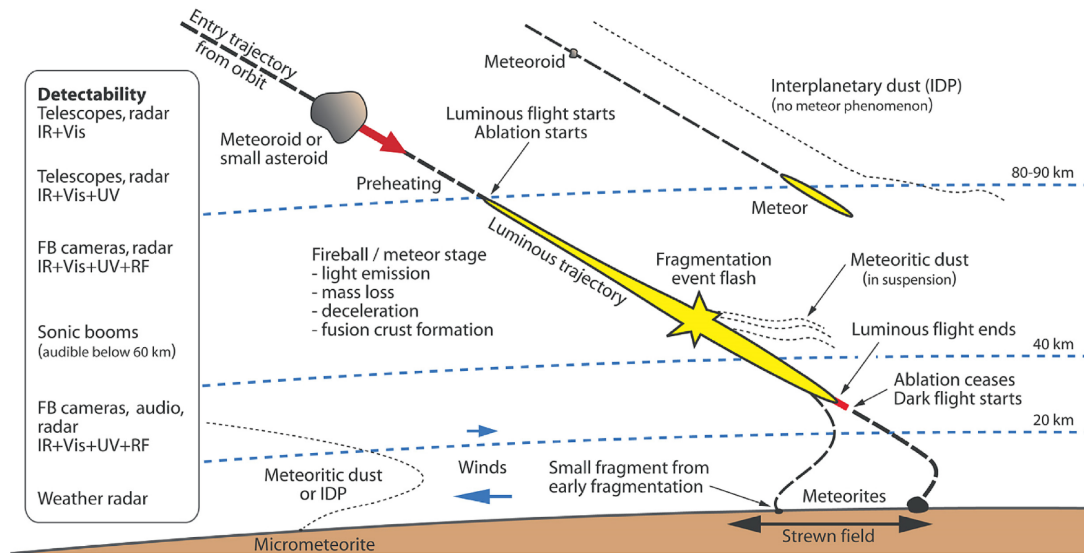


Figure 2.1.: Overview of different stages that meteoroid bodies undergo when entering Earth's atmosphere and meteoroid detectability for different methodologies versus altitude. Taken from Moilanen et al. (2021).

A meteoroid is defined as a naturally occurring, solid object in interplanetary space with a diameter ranging from approximately $30 \mu\text{m}$ to 1 m Koschny and Borovicka 2017. Objects with a smaller diameter are referred to as interplanetary dust particles, while those above the 1 m diameter threshold are classified as asteroids or cometary nuclei depending on their

origin (Rubin and Grossman, 2010). When a meteoroid enters the Earth's atmosphere at high velocities of $11\text{--}72\text{ km s}^{-1}$ it collides with atmospheric molecules. The velocity range is defined by the escape velocity as the lower limit and the upper limit is given by the heliocentric velocity at 1 astronomical unit (AU) of 42 km s^{-1} combined with Earth's orbital motion at 30 km s^{-1} , totalling 72 km s^{-1} . During the collision with atmospheric molecules and as long as the atmospheric density and meteoroid velocity is more than $2.5\text{--}5\text{ km s}^{-1}$ (Ceplecha et al. 1998), the meteoroid undergoes a process called ablation. This is the combination of mass loss by surface melting, evaporation, and sputtering, which converts kinetic energy into radiation and heat (Rogers et al., 2005). As it travels through the atmosphere, the meteoroid ionises atoms along its path, creating a luminous trail called a meteor. Particularly bright events exceeding an apparent visual magnitude of -4 (approximately the brightness of Venus) are called fireballs or bolides (Ceplecha et al., 1998). Any solid material that survives atmospheric entry and reaches the ground is called a meteorite (see Fig. 2.1). In summary, as long as the body is still in flight, it is called a meteoroid and it can produce the light phenomenon known as a meteor, while any surviving fragments, reaching the planetary surface, are called meteorites.

2.2. Origins of Meteoroids

In order to understand where meteoroids originate, it is necessary to discuss the main gravitational forces and the resulting resonance zones in the inner solar system (Morbidelli et al., 2002). The main asteroid belt, which is approximately torus-shaped and hosts rocky and metallic bodies, lies between the orbits of Mars and Jupiter, approximately $2\text{--}3.3\text{ AU}$ from the Sun (see Fig. 2.2). This is the primary origin of meteoroids that eventually impact Earth (DeMeo and Carry, 2014).

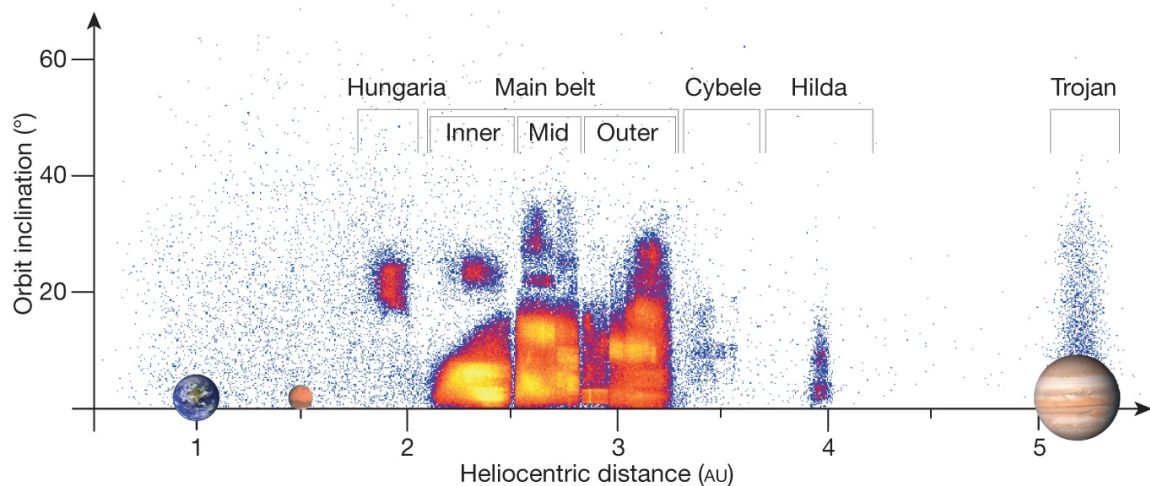


Figure 2.2.: Location of the main asteroid belt with respect to the planets (Earth, Mars and Jupiter) and the Sun. On the x-axis the heliocentric distance in AU is given and on the y-axis the orbit inclination in $^{\circ}$ is given. The colour gradient shows the density of asteroids (yellow represents the highest number density, blue the lowest). Taken from DeMeo and Carry (2014).

The bodies within the asteroid belt are subject to orbital resonances. These are configurations in which the orbital period of a body is in an integer ratio with that of Jupiter, causing repeated gravitational kicks that build up over time and can result in significant orbit alterations. One example of such a resonance is the 3:1 mean-motion resonance, which alters the orbits of asteroids that complete three orbits around the Sun for every one of Jupiter. Resonances like this are the dominant dynamical escape routes through which individual objects can be injected into Earth-crossing orbits (Bottke et al. 2002; Morbidelli and Gladman 1998). Smaller fragments themselves are produced by asteroid-asteroid collisions within the belt. Over timescales of 10^6 - 10^8 yr these smaller fragments with diameters of less than 10 metres are then moved into these resonance zones by the Yarkovsky effect. This effect is a small thermal thrust force due to slowly rotating bodies emitting more heat from their warmer afternoon side than their cooler morning side (Farinella et al. 1998; Vokrouhlicky et al. 2015). Even smaller fragments (in the diameter range of $1\ \mu\text{m}$ to $1\ \text{mm}$) are more prominently moved by the Poynting-Robertson effect, by which the solar radiation causes a dust grain to lose angular momentum relative to the orbit around the star (Burns et al., 1979). However, this is secondary in this study, as the meteoroids discussed here must be bigger to produce measurable acoustic-seismic signals, and are more affected by the orbit alterations due to the Yarkovsky effect.

The secondary meteoroid origin are comets. Depending on their orbital periods, these are classified as Jupiter-family, Halley-type or long-period comets. Jupiter-family comets have an orbital period of less than 20 years and are primarily controlled by Jupiter's gravity. They are also distinguished by a dimensionless orbital parameter called Tisserand's parameter T_J . Jupiter-family comets fall within the range $2 < T_J < 3$, while asteroids have $T_J > 3$. Each time a comet passes close to the Sun, ice is sublimated from its surface due to solar heating, which creates a trail of dust and up-to decimetre-sized particles (Gundlach et al., 2015). These particles spread along the comet's orbit over many revolutions and form a phenomenon known as a meteoroid stream. These streams appear as tube-like concentrations of particles that the Earth passes through once per revolution around the Sun, leading to yearly, predictable meteor showers such as the Perseids (Nesvorný et al. 2010; Jenniskens 2006). Halley-type comets have orbital periods of 20–200 years, while long-period comets have periods of 200 years or more (Levison, 1996). Both of these originate from the Oort Cloud, and they approach on highly eccentric, randomly inclined, and often retrograde orbits. This means that their related meteoroids encounter Earth at higher relative velocities of up to $72\ \text{km s}^{-1}$, compared to the prograde, low-inclination debris of Jupiter-family comets which impact the Earth with a minimum velocity of $11\ \text{km s}^{-1}$ (Ceplecha et al., 1998; Jenniskens, 2006).

Over large timescales, particles from a meteoroid stream are dispersed across the entire inner solar system due to gravitational perturbation. These particles contribute to the diffuse, sporadic meteoroid background, which is not annually recurring. Sporadic meteoroids have no apparent association with a specific parent body or radiant point in the sky (Campbell-Brown, 2008). It is important to classify meteoroids into these four source families: asteroidal, Jupiter-family comet, stream, and sporadic. This is because different source families produce meteoroids with distinct chemical compositions, and therefore different densities and porosities. These physical properties influence how and where the body ablates in the atmosphere. Associating a meteoroid observation with its source family enables inferences to be made about its parent population. In practice, classification is important for forecasting meteoroid

streams, modelling the impact risk to spacecraft, and studying metal atoms deposited in the atmosphere by meteoroid ablation (Plane 2012; Moorhead et al. 2020). All these applications rely on estimating the flux of meteoroids in order to assess the total amount of material impacting the Earth over a given time period.

2.3. Meteoroid Flux

The Earth's atmosphere is constantly subjected to impacts from meteoroids across a wide range of sizes and masses. Estimates of the total mass accreted per year are around 40 kt per year, with most of this mass carried by particles in the 50-500 μm diameter range (Love and Brownlee 1993; Plane 2012). The cumulative size-frequency distribution (see Fig. 2.3) follows an approximate power law, meaning that smaller objects are exponentially more numerous than larger ones. Earth is struck by approximately 10^5 - 10^6 meteoroids in the c. 1 kg class per year, but only once per year by a meteoroids with a mass of 10^5 - 10^6 kg (Bland and Artemieva 2006; Brown et al. 2002). From an observational perspective, sub-centimetre meteoroids produce faint meteors that can only be detected by sensitive camera or radar measurements and little to no measurable acoustic waves. However, centimetre- to decimetre-scale bodies generate meteors that are visible to the naked eye and can be recorded with wide-field camera networks (e.g., Colas et al. 2020), and the acoustic waves can sometimes be recorded at individual infrasound or seismic stations. Bodies of the decimetre to metre scale produce bright bolide events, releasing energy in the range of 1-10 kt of TNT equivalent, which converts to 10^{12} - 10^{13} J. These bodies can reliably be detected by infrasound and seismic instruments and they occur on timescales of years to decades (Brown et al. 2002; Jenniskens et al. 2009).

Meteoroids can also be a hazard to human life and infrastructure. The relevant threshold for this lies in the 1-50 m diameter range. Bodies smaller than approximately 25 metres are commonly destroyed by ram pressure, which builds up during the descent of a meteoroid due to the increase in ambient pressure. When this pressure exceeds the meteoroid body's structural integrity, it explodes violently. These explosions are also called fragmentations or airbursts and result in quasi-spherical propagating pressure waves at altitude. For example, the 2013 Chelyabinsk event involved a 19-metre-diameter, 12,000-tonne stony meteoroid that released approximately 400-500 kt TNT-equivalent energy in a fragmentation at around 30 km altitude. The resulting shockwave shattered windows within a 90 km radius and injured over 1,500 people (Brown et al. 2013; Popova et al. 2013). Additionally, the main mass impacted into a local lake so that impact cratering could not be observed. Another example is the 1908 Tunguska event, which is estimated to have produced an fragmentation energy equivalent to 10-15 Mt TNT and to have flattened approximately 2,000 km² of Siberian forest without producing a confirmed impact crater (Boslough and Crawford, 2008). At the upper end of this range, as a contemporary example, is the asteroid Apophis, which was initially assigned a 2.7% probability of impacting Earth in 2029 (Chesley, 2005), releasing a potential kinetic impact energy of up to 1000 Mt TNT equivalent ($\approx 4 \times 10^{18}$ J) and measuring approximately 300 m in diameter (Shuvalov et al., 2017). In the context of energies, the 2011

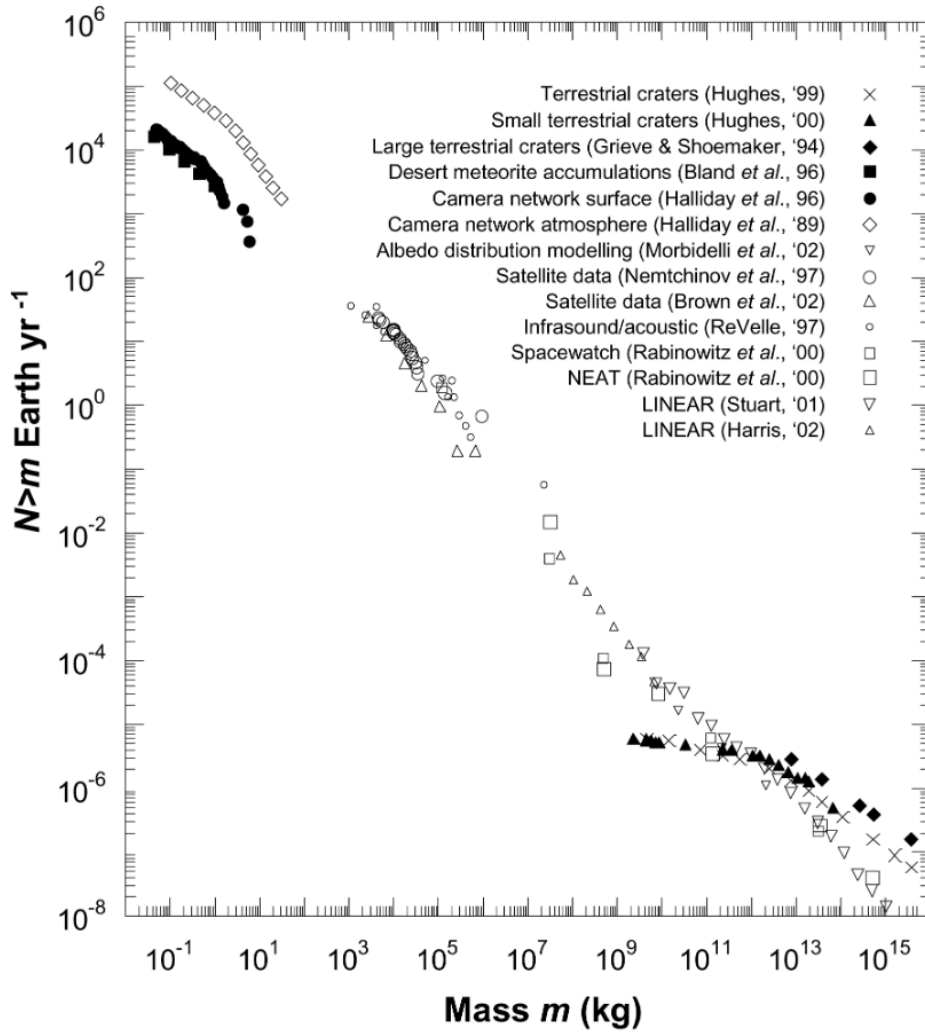


Figure 2.3.: Size-frequency distribution of meteoroids for Earth's upper atmosphere (open symbols) and surface (filled symbols) as number of events larger than a given mass. Taken from (Bland and Artemieva 2006).

Tōhoku earthquake released 1.9×10^{17} J (c. 40 Mt TNT equivalent), which is comparable to the frequency of meteoroid events of this magnitude, occurring approximately once per decade to century (Norio et al., 2011). This demonstrates that, although meteoroid impacts can be locally devastating, large-scale destructive events are rare. Additionally, they are spatially unconstrained, which differentiates them from chronic seismic or volcanic hazards.

2.4. Physical Properties of Meteorites

Meteorites offer a glimpse into the past. The oldest meteorite material found on Earth is older than the planet itself. The Murchison meteorite, for example, which was found in Victoria, Australia, in 1969, and contains presolar silicon carbide grains which are dated to be about 3 ± 2 billion years older than the Solar System (Heck et al., 2020). Studying meteorite material

that has impacted Earth is the only way to directly measure the composition of solid solar system material. Approximately 86% of all observed falls are chondrites. These primitive, chemically undifferentiated stony meteorites have not undergone large-scale melting since their formation 4.56 billion years ago, thus preserving the early solar system's chemistry (Burbine et al., 2002). Undifferentiated refers to the fact that these bodies never separated into a metallic core and silicate mantle, unlike terrestrial planets or large asteroids. Chondrites can be further subdivided into ordinary chondrites, which account for around 80% of found chondritic material. They consist primarily of olivine, pyroxene, and dispersed iron-nickel metal grains, and have bulk densities of 3,200-3,800 kg m⁻³ (Consolmagno et al. 2008; Britt and Consolmagno 2003). The other subdivision of chondrites are carbonaceous chondrites. These are compositionally more primitive and contain hydrated silicates, complex organic compounds and, in some cases pre-solar grains, which is material predating the formation of our solar system entirely (Heck et al., 2020). Their higher porosity and volatile content yields lower bulk densities of 1,500-2,700 kg m⁻³ (Consolmagno et al. 2008; Scott and Krot 2014). Differentiated meteorites originate from the interiors of bodies that have undergone global melting and gravitational separation (Weiss and Elkins-Tanton, 2013). Achondrites are stony meteorites originating from asteroidal crusts and mantles, while iron meteorites are solidified metallic cores. Stony-iron meteorites originate from the core-mantle boundary of differentiated asteroids. These meteorites cover a density range of approximately 3,000-7,900 kg m⁻³ (Consolmagno et al., 2008).

Cometary meteoroids are severely poorly represented in meteorite collections because their fragile, porous structure causes them to completely disintegrate upon atmospheric entry. Inferring their properties requires special in-situ spacecraft measurements, such as those carried out by ESA's Giotto (Reinhard, 1987) and the Rosetta missions (Glassmeier et al., 2007), or ground-based meteor spectroscopy (Cepilecha et al., 1998). Cometary meteoroids mainly consist of anhydrous, magnesium-rich silicates, such as olivine and pyroxene, as well as carbonaceous matter (including carbon, hydrogen, oxygen and nitrogen) and sulfide minerals. Cometary meteoroids have bulk densities as low as 100-300 kg m⁻³ (Blum et al. 2006; Jessberger et al. 2001). This wide range of densities, spanning nearly two orders of magnitude from fragile cometary aggregates to dense iron meteorites, directly affects how they behave when they enter the atmosphere and whether they survive. For these reasons, the source-family classification is an essential foundation for any quantitative treatment of meteoroid observations.

3. Theoretical background

3.1. Observational Methods and Detection Capabilities

Meteoroids passing through Earth's atmosphere are studied using a variety of observational techniques. The following chapters provide a brief introduction to each technique. Each technique probes a different physical aspect of a meteoroid event and is sensitive to a different range of sizes and masses of objects. Because no single method can cover the full spectrum of meteoroid masses, and as the different methods constrain different parameters more effectively, a combination of detection and analysis techniques is required to gain a comprehensive understanding of a meteoroid event. The following subsections describe the five main techniques, how they work and their key limitations.

3.1.1. Optical Detection

Optical methods exploit the luminosity of the meteor produced when a meteoroid ablates in the upper atmosphere. As the body decelerates at an altitude of around 70–110 km, the surrounding gas column is heated, which excites and ionises atmospheric oxygen molecules and evaporated meteoroid atoms. These ionised particles produce broadband continuum emission and discrete atomic spectral lines across the near-ultraviolet to near-infrared range (Borovička 1994; Ceplecha et al. 1998).

Ground-based optical systems range from wide-field all-sky cameras like fish-eye lens instruments with a field of view of 180 ° used in networks such as the European Fireball Network (Oberst et al., 1998), AllSky7 (Hankey et al., 2023), the Fireball Recovery and Interplanetary Observation Network (FRIPON) (Colas et al., 2020), and the Global Meteor Network (GMN) (Vida et al., 2021), to narrow-field intensified video cameras such as those used by the Desert Fireball Network (DFN) (Bland et al., 2012). A single camera station can record meteors occurring within a horizontal radius of roughly 300–500 km based on lines of sight at typical meteor altitudes of 50–100 km. For this reason, the separation between stations is usually 100–200 km, enabling stereoscopic triangulation of the trajectory with an accuracy of a few hundred metres (Ceplecha et al., 1998).

To measure the brightness with which an object appears to an observer on Earth the apparent magnitude is used. The apparent magnitude scale is logarithmic and inversely rated, which means that brighter objects have smaller (more negative) numbers. One magnitude difference converts to a change of brightness by the factor of roughly 2.5. Photographic and low-light systems reach limiting apparent magnitudes for meteors of approximately +6 to +8, which

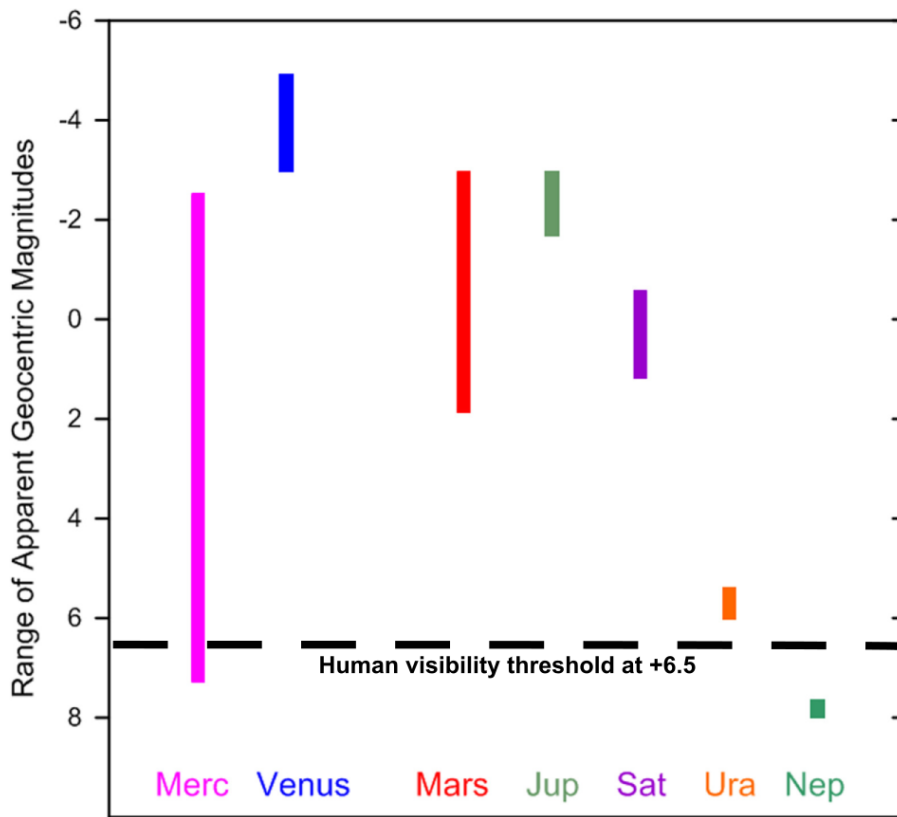


Figure 3.1.: Apparent magnitudes of major planets as seen from Earth. Modified after Mallama and Hilton (2018)

correspond approximately to the brightness of the planets Uranus to Neptune in the night sky (see Figure 3.1). These apparent magnitudes convert to meteoroid masses ranging from 10^{-4} - 10^{-3} g for a typical entry velocity of 30 km s^{-1} (Ceplecha, 1987). Intensified video systems can push this sensitivity to masses slightly below 10^{-4} g (Koten et al., 2004). These systems can be used in tandem with dedicated spectrographs to enable simultaneous compositional analysis through atomic emission spectroscopy (Trigo-Rodríguez, 2019).

3.1.2. Radar Detection

When a meteoroid ablates, it leaves behind a cylindrical column of plasma consisting of free electrons and ions, which is called the meteor trail. Radio waves in the HF to VHF range (approximately 2-100 MHz) reflect from this plasma column, forming the basis of radar meteor detection (Ceplecha et al., 1998). Radar meteor analysis can be divided into two echo types. These are trail and head echoes. The more common type is the trail echo, produced by the elongated column of plasma that the meteoroid leaves in its wake. Trail echoes are predominantly specular. There, the radar receives a strong return from the short segment of the trail oriented perpendicular to the line of sight, where the contributions from successive electrons add in phase, producing a sharp, brief echo (Ceplecha et al., 1998). Trail echoes are further classified by the electron line density of the column. Underdense trails, with line

densities below approximately 10^{14} electrons m^{-1} , are transparent to the incident wave, which scatters from individual electrons to give a short-lived echo whose amplitude decays as the trail diffuses. Overdense trails, above this threshold, have sufficient line density to reflect the wave collectively, behaving as a conducting cylinder and producing longer-duration echoes from which deceleration and ablation profiles can be extracted (Mathews 2004; Holdsworth et al. 2004).

The second type is the head echo, a non-specular return scattered from the dense, approximately spherical plasma region immediately surrounding the meteoroid itself. Unlike trail echoes, head echoes move across the sky at the meteoroid velocity and are observed largely independently of the trail-perpendicularity condition, making them well suited to high-power large-aperture radars (Close et al., 2004). Because the head echo tracks the body directly, it yields precise velocity and deceleration measurements along the observed path.

Operational and historical meteor orbit radars, such as the Canadian Meteor Orbit Radar (CMOR at 6 kW), the Advanced Meteor Orbit Radar (AMOR), and the higher-power Southern Argentina Agile Meteor Radar (SAAMER at 60 kW), operate at frequencies of approximately 26–40 MHz (Jones et al., 2005; Baggaley et al., 1994; Fritts et al., 2010). They detect meteor trails over an altitude range of 70–110 km (Ceplecha et al., 1998), with an effective detection radius extending to c. 200–250 km from the antenna (Hocking et al., 2001). Dedicated regional networks, such as the Belgian RADio Meteor Stations (BRAMS) network, complement these wide-area systems by deploying arrays of receivers optimised for forward-scatter detection. This increases spatial coverage and detection rates within a defined geographic region (Lamy et al. 2022). Most importantly, radar is sensitive to far smaller meteoroids than any other methods. The minimum detectable electron line density for well-calibrated systems corresponds to meteoroid masses of approximately 10^{-6} g, which extends the minimal detectable mass by several orders of magnitude compared to optical techniques (Galligan and Baggaley 2004; Webster et al. 2004). However, a key practical limitation is that specular reflection geometry restricts radar observations to only a small part of the meteoroid’s total trajectory. Within this portion, radar systems provide the most accurate pre-atmospheric velocity estimates of any observational technique, making them particularly valuable in complementing optical and acoustic-seismic methods in multi-instrument analyses.

3.1.3. Human Visual Observation

Observing meteors with the naked eye is a low-cost addition to instrumental networks. Under dark-sky conditions, the human eye has a limiting visual magnitude of around +6.5, depending on the dark-sky quality (Crumey, 2014). At a meteor altitude of 90 km, this limiting magnitude corresponds to meteoroid masses of 0.01–1 g for an entry velocity of 30 km s^{-1} , meaning the human eye is approximately comparable to wide-field photographic systems in terms of mass sensitivity. The effective observability radius is determined by the geometry of the meteor trail and the minimum detectable luminosity at the observer’s position. It amounts approximately to 300–400 km for a zenith-pointing meteor (Rendtel, 2026).

Organised visual programmes, such as the Fireball Program coordinated by the International Meteor Organisation (Hankey and Perlerin 2014; International Meteor Organization (IMO) 2025b), collect zenithal hourly rate (ZHR) data. The ZHR is a normalised count, corrected for the observer's limiting magnitude and radiant elevation, which corresponds to the average number of meteors that a single observer can see in an hour (Rendtel, 2026). It is an approximate way to estimate the intensity of a given meteor shower. Additionally, the IMO also provides meteoroid trajectory estimations based on human observations, which can be used as a first-order trajectory guess for other trajectory reconstruction methods.

Although human observations are less sensitive and precise than those of instrumental networks, the large number of observers (thousands worldwide, centred around North America and Central Europe) provides semi-global sky coverage and shower-profile data, especially for major showers (Rendtel 2014; Molau and Rendtel 2009).

3.1.4. Infrasound Detection

Meteoroids large enough to generate infrasound have frequencies between 0.01 Hz and 20 Hz. The lower limit is the Brunt–Väisälä frequency, below which the disturbance transitions into the acoustic-gravity wave regime and the upper limit is near the lower threshold of human hearing. The primary mechanisms of acoustic wave generation are discussed in detail in section 3.2. Fundamentally, a meteoroid generates acoustic waves through its supersonic motion through the atmosphere and during explosive fragmentation events. These acoustic infrasound waves then propagate over long distances due to atmospheric waveguide effects. The most important waveguides are temperature gradients, and wind gradients at the tropopause (c. 10 km) and in the thermosphere (c. 120 km). These waveguides direct acoustic energy around the globe with minimal attenuation (Drob et al. 2003; Le Pichon et al. 2010).

The International Monitoring System (IMS; Christie and Campus 2009) is the most important global network for infrasound meteoroid detection. Originally established to verify compliance with the Comprehensive Nuclear-Test-Ban Treaty, the IMS currently consists of 53 certified arrays distributed worldwide (Hupe et al., 2022). Each of these arrays is made up of a cluster of microbarometers sensitive to pressure changes in the infrasound range, with instrument sensitivity in the order of 1 mPa (Christie and Campus, 2009). The network can detect meteoroid fragmentations with energies above 1–10 t of TNT equivalent anywhere on Earth, corresponding to meteoroid masses of 0.1–1 kg, depending on entry velocity and trajectory geometry (ReVelle 1976; Brown et al. 2002). For large-scale events, such as the 2013 Chelyabinsk meteoroid, infrasound signals were recorded at distances of up to 15,000 km (Le Pichon et al., 2013). A key limitation of infrasound is that long atmospheric travel times and path complexity make accurate source localisation challenging without complementary optical or seismic data. Another limiting factor is the low number of infrasound arrays distributed around the globe. However, single-station infrasound observations can be used as well for meteoroid detections.

3.1.5. Seismic Detection

Large meteoroids transfer acoustic energy into the Earth's solid surface through two pathways: 1) direct ground coupling of atmospheric pressure waves (air-to-ground coupling), and 2) direct seismic wave excitation at the impact point for terminal meteorite-dropping and impact events (Edwards et al. 2008; Borovička and Kalenda 2003). The resulting seismic signals are impulsive, with most of the energy concentrated in the 0.5–40 Hz range, and they can be recorded by short-period and broadband seismometers at distances of up to approximately 350 km, depending on instrument placement, atmospheric conditions, and the velocity and trajectory geometry of the meteoroid (see 5.5). If arrivals from multiple stations are available, they can be used to reconstruct the meteoroid trajectory. These methods range from purely geometrical approaches (Pujol et al. 2005; Vida 2023) to ray-based reconstruction methods (see chapters 5; 6; 7).

The minimum detectable meteoroid mass is strongly dependent on local noise conditions and site characteristics, such as station placement and local geology. In typical conditions, seismic instruments can consistently record signals from meteoroids with masses of 1–2 kg travelling at 20 km s^{-1} . For individual events and recording stations even 1–10 g meteoroids have produced a measurable arrival signal (Neidhart et al., 2021). Seismic data are particularly valuable for constraining the terminal altitude of energy deposition, as well as part of the dark flight that larger meteoroids undergo, when decelerating to velocities below $2.5\text{--}5 \text{ km s}^{-1}$, predominantly at altitudes in the range of 20–25 km (Ceplecha and McCrosky, 1976). At this point, the meteoroid continues to produce acoustic waves, that can be measured at seismic stations, but it is invisible, or dark, to optical methods as no meteor is produced anymore. Although ground-impact events, in which a meteoroid hits the Earth's surface at speeds much greater than its terminal velocity, are exceedingly rare, seismic data can be used to constrain the impact location based on recorded P- and S-wave arrivals (Moilanen et al. 2026; Daubar et al. 2024; Brown et al. 2008).

To summarize this chapter the different methodologies are presented as Table 3.1 to directly infer the advantages and limitations of each method:

Technique	Min. mass	Detection range	Key constrained parameter(s)	Main limitation
Radar	$\sim 10^{-6} \text{ g}$	200–250 km	Flux; velocity	Limited coverage; samples short trajectory segment
Optical	$10^{-4}\text{--}10^{-3} \text{ g}$	300–500 km	Light curve; composition; trajectory	Night / clear-sky only; limited coverage
Human visual	0.01–1 g	300–400 km	Shower activity (ZHR)	Low precision; observer-dependency
Infrasound	0.1–1 kg	global	Source energy; trajectory, fragmentation timing	sparse network; far range dominated by path complexities
Seismic	<1–2 kg	~350 km	Terminal energy deposition; trajectory, fragmentation timing	Site-dependent coupling

Table 3.1.: Comparison of the five discussed meteoroid observation techniques by mass sensitivity, spatial reach, and the parameters each constrains most effectively. Techniques are ordered by ascending minimum detectable mass, illustrating the complementary coverage that motivates a multi-instrument approach: radar and optical methods sample the abundant low-mass population, whereas infrasound and seismic methods access the rarer, high-energy bodies relevant to planetary defence and meteorite recovery.

3.2. Acoustic-Seismic Wave Emission by Meteoroids

When a meteoroid enters Earth’s atmosphere at supersonic velocity, it can radiate waves in three distinct physical ways (see Fig. 3.2). These have characteristic directivity patterns that depend on the meteoroid’s entry parameters (ReVelle 1976; Edwards et al. 2008).

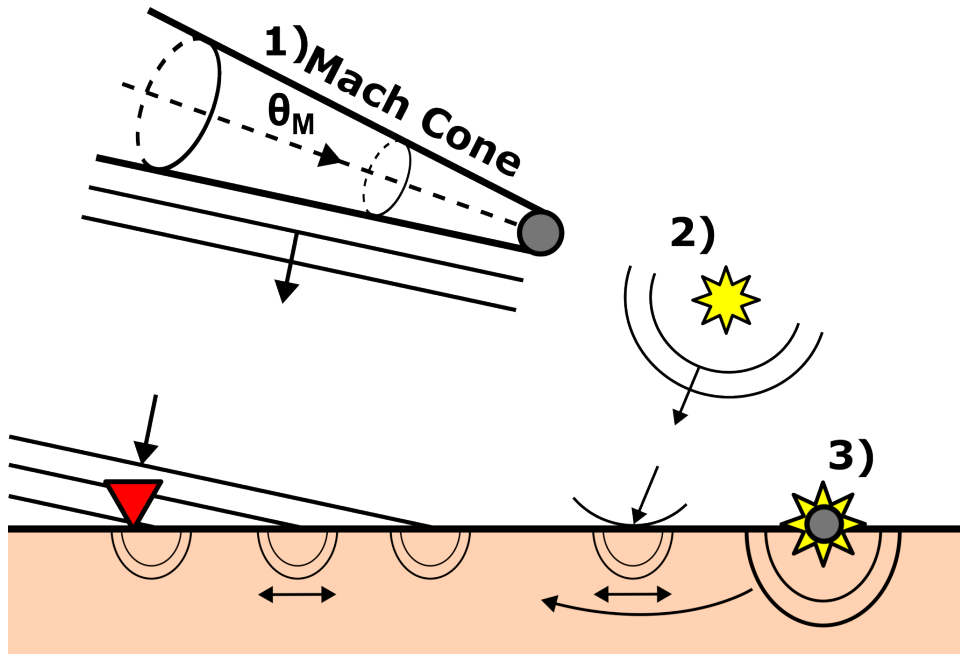


Figure 3.2.: The three wave radiation mechanisms: 1) Ballistic Mach waves generate quasi-cylindrical acoustic waves. The opening of the Mach Cone is defined by the half-angle θ_M . 2) Explosive fragmentations generate quasi-spherical acoustic waves. 3) A meteoroid impact generates seismic waves. The meteoroid is represented by the grey circle, the recording station by the red inverted triangle and the subsurface by the orange half-space. Modified after Edwards et al. 2008.

3.2.1. Ballistic Mach Wave

Any object moving faster than the local speed of sound generates a conical shock wave, called the Mach cone, through the constructive interference of the continuous pressure disturbances it emits. The half-angle θ_M of the cone is given by the Mach relation:

$$\sin(\theta_M) = \frac{c_s}{v} = M^{-1}, \quad (3.1)$$

where c_s is the local speed of sound (e.g., c. $280\text{--}340\text{ m s}^{-1}$ at an altitude of $70\text{--}100\text{ km}$), v is the meteoroid velocity, and M is the Mach number. For a typical meteoroid entry velocity of 30 km s^{-1} , $M \approx 90\text{--}110$ and $\theta_M < 1^\circ$, this results in an extremely narrow cone (ReVelle 1974). The Mach wave is a quasi-cylindrical blast wave that propagates radially outward along the entire flight path.

The physical nature of the Mach wave is well described by line-source blast wave theory (ReVelle, 1974). In this theory, the meteoroid's supersonic flight is modelled as a continuous cylindrical energy deposition along its trajectory. A key parameter that emerges from this theory is the blast radius R_0 , which is defined as the radial distance from the trajectory at which the overpressure of the shock wave decays to the ambient atmospheric pressure and the wave transitions from a non-linear shock to a linear acoustic wave (ReVelle, 1974; Silber and Brown, 2014). In the near field, within R_0 , the wave is strongly non-linear and causes significant heating and compression of the surrounding air. Beyond R_0 , in the far field, the wave evolves into an N-wave: it is a bipolar pressure pulse consisting of a leading compressional phase followed by a trailing decompressional phase, which is the waveform shape most often observed at ground-based infrasound stations (Kinney and Graham, 2013). The time period of the N-wave, and therefore the dominant frequency of the recorded signal, scales with the energy deposited per unit path length and therefore with the size of the meteoroid, with larger bodies producing longer-period signals. The frequency spectrum of the acoustic Mach wave peaks in the range $0.1\text{--}10\text{ Hz}$ (ReVelle 1976; Silber et al. 2015).

The geometry with which the Mach cone intersects the ground has important consequences

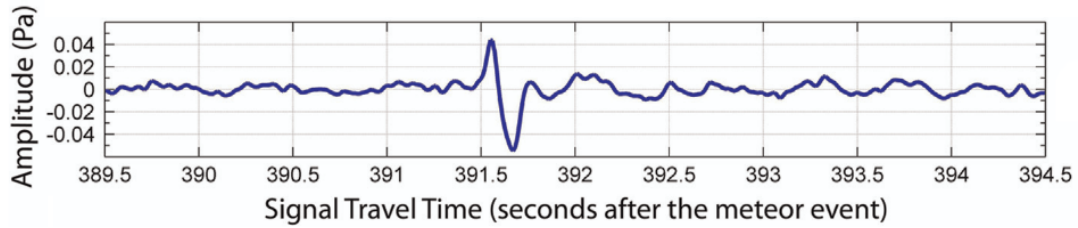


Figure 3.3.: Meteoroid-related acoustic N-wave with the signal onset roughly at 391.5 seconds travelttime, recorded at a microbarometer. Modified after Silber and Brown (2014)

for how the wave is observed at surface stations. Since the cone is almost closed at supersonic entry speeds, the half-angle is very small and the Mach wave impinges on the ground at a steep angle, almost vertically. As the meteoroid travels along its trajectory, the intersection of the cone with the ground surface traces a hyperbolic pattern. Stations aligned approximately perpendicular to the flight path receive the wave nearly simultaneously, whereas those along the flight azimuth record it with a characteristic time delay. This time delay reflects the faster-than-sound propagation of the source along the track. The entire ground coverage produced by the Mach cone is referred to as the Mach carpet and the corresponding arrival at a recording station at the Earth's surface is called a ballistic arrival. This arrival can also be identified by its linear moveout across a network (Edwards et al., 2008).

Atmospheric propagation further shapes the waveform and timing of the Mach wave before it reaches the surface. Vertical temperature gradients and horizontal wind speed gradients cause acoustic rays to refract. Under certain stratospheric or thermospheric conditions the wave can

be ducted over large horizontal distances, producing secondary or tertiary arrivals long after the direct ballistic phase (Le Pichon et al., 2010; Drob et al., 2003). The interaction of the wave with the local wind field modifies both the apparent propagation speed and the amplitude of the arrival. This sensitivity to wind structure is one of the reasons that accurate atmospheric models, including full horizontal wind fields, are indispensable when using infrasound and seismic data to reconstruct meteoroid trajectories. The Mach wave is the dominant acoustic source for meteoroids that do not significantly fragment and is generally the most persistent and spatially extensive of the three emission mechanisms. This makes it the primary source of trajectory information in acoustic-seismic records.

3.2.2. Explosive Fragmentation Events

Whenever the dynamic ram pressure p_{ram} , experienced by the meteoroid during atmospheric flight, exceeds the bulk tensile strength of its constituent material, the body undergoes rapid, catastrophic fragmentation. The dynamic ram pressure is given by:

$$p_{\text{ram}} = \rho_a v^2, \quad (3.2)$$

where ρ_a is the local atmospheric density. Since ρ_a increases approximately exponentially with decreasing altitude and v decreases only gradually until low altitudes, p_{ram} rises steeply as the meteoroid descends. Fragmentation normally occurs at a characteristic altitude determined by aerodynamic loading and material strength (Hills and Goda, 1993).

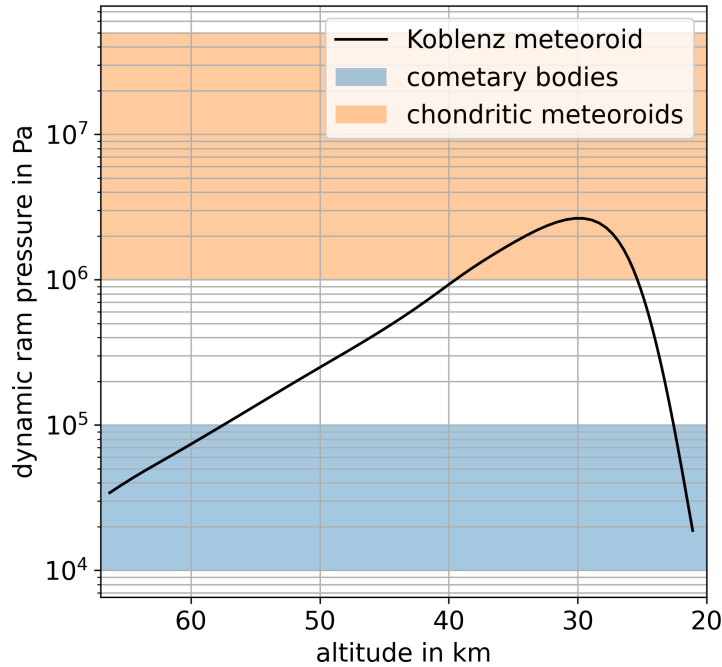


Figure 3.4.: Dynamic ram pressure p_{ram} versus altitude of the Koblenz meteoroid (see Chapter 7) shown as a black line. Also shown are the dynamic ram pressure ranges for chondritic stony meteoroids (orange area) and more loosely consolidated or cometary bodies (blue area). With this, the Koblenz meteoroid can be loosely categorised in the more consolidated, chondritic stony meteoroid regime.

Chondritic stony meteoroids, which constitute the majority of observed falls, have tensile strengths on the order of 1–50 MPa. In contrast, loosely consolidated or cometary bodies can fragment at pressures as low as 0.01–0.1 MPa (Ceplecha et al. 1998; see Figure 3.4). In contrast to the continuous, cylindrically symmetric emission of the Mach wave, fragmentation events are impulsive, nearly isotropic energy releases that radiate quasi-spherical pressure waves into the surrounding atmosphere. The resulting acoustic signal on the surface is an N-wave that is potentially shorter in duration and higher in peak amplitude than the ballistic arrival, depending on the energy released. The dominant period of the signal scales with the cube root of the released energy and inversely with distance from the source. These relationships follow from the scaling relations derived for point-source explosions (Kinney and Graham, 2013), and for energetic fragmentations they can be extended into the sub-Hz infrasound range.

The fragmentation process is rarely a single discrete event. Multiple fragmentation episodes at different altitudes are common and appears in infrasound and seismic records as a sequence of distinct, separated pulses (see e.g., Chapter 7, Figure 7.2). Progressive fragmentation models, in which the body breaks into a growing cloud of smaller fragments that continue to interact aerodynamically and ablate collectively, can significantly modify the energy deposition profile compared to a simple single-body model (Hills and Goda, 1993). The altitude at which the peak energy deposition occurs, termed the burst point, is a critical parameter for assessing the ground-level effects of the event. Fragmentations at low altitudes (below c. 30–40 km) can generate destructive overpressures at the surface, whereas those at higher altitudes dissipate their energy in the upper atmosphere, having little impact at the Earth’s surface (Boslough and Crawford, 2008).

The 2013 Chelyabinsk event is a well documented example of meteoroid fragmentation. The body underwent two major fragmentation episodes and several minor ones during its descent, culminating in the largest recorded fragmentation, estimated to have an explosive yield of around 500 kilotons of TNT at an altitude of approximately 30 km (Brown et al. 2013; Popova et al. 2013). The resulting blast wave caused widespread damage to windows across the southern Ural region and injured over 1,500 people, demonstrating that even sub-megaton fragmentations can have significant societal consequences (Kartashova et al., 2018). Fragmentation events are thus the main cause of meteoroid-related hazards at ground level, and due to their impulsive nature and high signal-to-noise ratio, they are often easily identifiable in acoustic-seismic recordings.

3.2.3. Impact

Meteoroids with sufficient pre-entry mass can survive the combined effects of aerodynamic drag and ablative mass loss during atmospheric passage, reaching the ground as one or more solid fragments referred to as meteorites. Before impact, the surviving fragments enter a phase of unpowered ballistic flight, termed dark flight. The luminosity of the meteor ceases at velocities of approximately $2.5\text{--}5\text{ km s}^{-1}$ and meteoroids are, albeit later in the flight, acoustically inactive at subsonic velocities (Ceplecha et al., 1998; Gritsevich et al., 2014; Towner et al., 2022). The terminal velocity at the end of the ablation phase depends on the fragment’s mass, cross-sectional area, drag coefficient, and the local atmospheric density. It is in the

range of c. $100\text{--}300\text{ m s}^{-1}$ for centimetre- to decimetre-sized objects (Ceplecha et al., 1998). During dark flight, the fragment trajectories are deflected by horizontal wind fields, which can displace the landing locations by several kilometres relative to the projected ballistic trajectory. This displacement is particularly pronounced for light fragments ejected at high altitudes. Accurately predicting the strewn field, the surface distribution of recovered meteorite material, therefore requires knowledge of the full three-dimensional atmospheric wind field integrated over the dark flight path (Gritsevich et al., 2014; Towner et al., 2022). This is one of the central motivations for incorporating realistic wind models into the inversion framework developed in this thesis (see Chapter 7).

When a meteorite fragment impacts the ground, it generates seismic waves through direct momentum transfer from the impacting body to the solid Earth. The seismic wavefield generated by impacts is dominated by compressional and shear body waves, as well as surface waves. The dominant frequencies are determined by the source duration and the local site response (Teanby and Wookey, 2011). In addition to the seismic energy generated directly, the impact excites an impulsive acoustic pulse in the overlying air column through ground-to-air coupling. This pulse can propagate as an infrasound signal to infrasound recording stations (Garcia et al., 2017). Similar seismic-to-acoustic coupling has also been observed for small earthquakes (Krenz et al., 2023).

For very large meteoroids, mainly those exceeding several tons of mass, the impactor does not decelerate to subsonic speeds and retains a significant fraction of its pre-Earth velocity at the moment of contact with the ground. In this case, the impact itself acts like a surface-level explosion, excavating a crater and radiating a seismic signal that far exceeds the amplitude of any in-flight acoustic-to-seismic sources (Melosh, 1989). Such events are geologically rare on human timescales, and the in-flight acoustic signals are secondary to the impact signals for these events. For the size range of meteoroids considered in this thesis, primarily objects weighing a few kilograms to a few tonnes, subsonic impacts represent the relevant end state. The impact seismic signal, while possibly detectable at short distances, is generally the weakest of the three emission mechanisms and is included here for completeness rather than being an often recorded observable.

3.2.4. Acoustic-to-seismic Coupling

Another mechanism, that is particularly relevant to the seismic detection of meteoroid events, is acoustic-to-seismic wave coupling. Acoustic waves generated by the flight of a meteoroid, including the Mach wave and fragmentation pulses, propagate to the surface and impinge on the ground. Some of their energy is then coupled into the subsurface and continues to propagate as seismic waves. This air-to-ground coupling produces a characteristic seismic phase whose wavelet matches a W-shape on vertical component seismic data (Edwards et al. 2008; Langston 2004; see Figure 3.5). The W-shape is generated by the acoustic wave deforming the Earth's surface and the resulting transformation of the signal from a pressure to a ground motion signal. This signal is the dominant meteoroid-related signal at the recording stations discussed in this work. Identifying and separating these acoustic-seismic phases from ambient

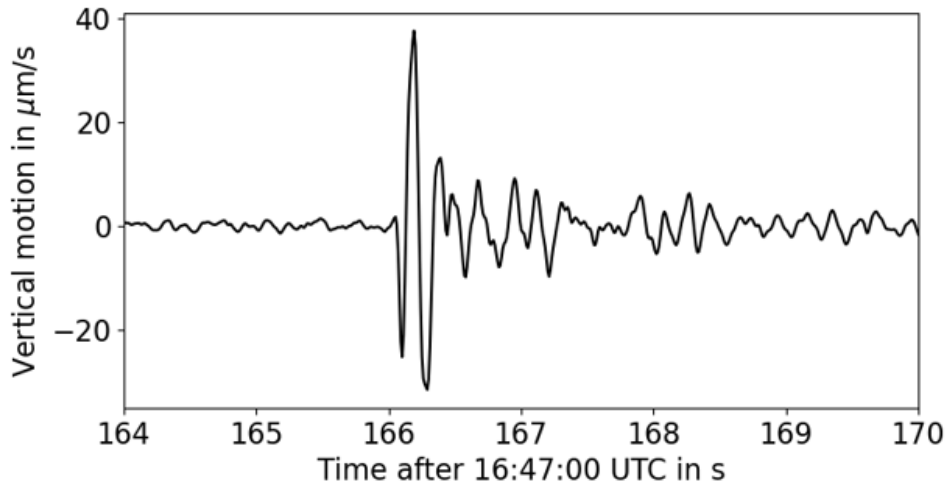


Figure 3.5.: W-shaped wavelet produced by a meteoroid over Germany recorded on the vertical component of the seismic recording station HS.GWBO. The onset time of the signal is roughly 166 seconds after the event time of 16:47 UTC.

seismic noise is one of the core challenges addressed by the deep-learning phase identification approach presented later in this thesis (see Chapter 5).

Coupling Efficiency

The impedance contrast between the two media is the fundamental control on how efficiently acoustic energy crosses the air-ground interface. The specific acoustic impedance of the atmosphere, $Z_{\text{air}} = \rho_{\text{air}} c_s$, is approximately 400 Pa s m^{-1} at sea level, while the seismic impedance of near-surface materials, $Z_{\text{ground}} = \rho_{\text{rock}} v_p$, ranges from approximately 10^5 Pa s m^{-1} for unconsolidated sediments to 10^7 Pa s m^{-1} or more for intact crystalline bedrock (Aki and Richards, 1980; Sheriff and Geldart, 1995). At normal incidence, the pressure and normal particle velocity must be continuous across the interface. Solving these two boundary conditions yields the plane-wave pressure transmission coefficient:

$$T_p = \frac{2Z_{\text{ground}}}{Z_{\text{air}} + Z_{\text{ground}}}, \quad (3.3)$$

which approaches the value of 2 in the limit $Z_{\text{ground}} \gg Z_{\text{air}}$. However, the fraction of incident energy transmitted into the ground is much smaller. Using the plane-wave intensity relation

$$I = \frac{p^2}{Z}, \quad (3.4)$$

and evaluating the transmitted and incident intensities with the impedance of their respective media, the energy transmission coefficient ζ becomes

$$\zeta = \frac{I_{\text{trans}}}{I_{\text{inc}}} = \frac{p_{\text{trans}}^2 Z_{\text{air}}}{p_{\text{inc}}^2 Z_{\text{ground}}} = T_p^2 \frac{Z_{\text{air}}}{Z_{\text{ground}}} = \frac{4 Z_{\text{air}} Z_{\text{ground}}}{(Z_{\text{air}} + Z_{\text{ground}})^2}. \quad (3.5)$$

This expression is symmetric in Z_{air} and Z_{ground} , as energy transmission across an interface must be. In the limit $Z_{\text{ground}} \gg Z_{\text{air}}$, the denominator reduces to Z_{ground}^2 and the coefficient simplifies to

$$\zeta \approx 4 \frac{Z_{\text{air}}}{Z_{\text{ground}}}. \quad (3.6)$$

This means that the energy transmission coefficient ζ scales with the ratio $Z_{\text{air}}/Z_{\text{ground}}$, implying that only a small fraction of the incident acoustic energy, commonly of the order of 10^{-3} to 10^{-4} enters the ground (Sabatier et al., 1986; Cates and Sturtevant, 2002). Despite this inherent inefficiency, the large acoustic amplitudes produced by energetic meteoroid events are sufficient to generate clearly measurable seismic signals at distances of tens to hundreds of kilometres.

The coupling efficiency varies with the angle of incidence of the acoustic wave at the surface, the frequency content of the signal, and the elastic properties of the near-surface medium. An important enhancement occurs at the critical angle θ_c , at which the horizontal phase velocity of the impinging acoustic wave matches the phase velocity of seismic surface waves in the near-surface material:

$$\sin \theta_c = \frac{c_s}{v_R}, \quad (3.7)$$

where v_R is the Rayleigh wave phase velocity. For usual near-surface Rayleigh wave speeds of $0.5\text{--}2 \text{ km s}^{-1}$, θ_c falls in the range of approximately $10\text{--}40^\circ$ from vertical (Sabatier et al., 1986).

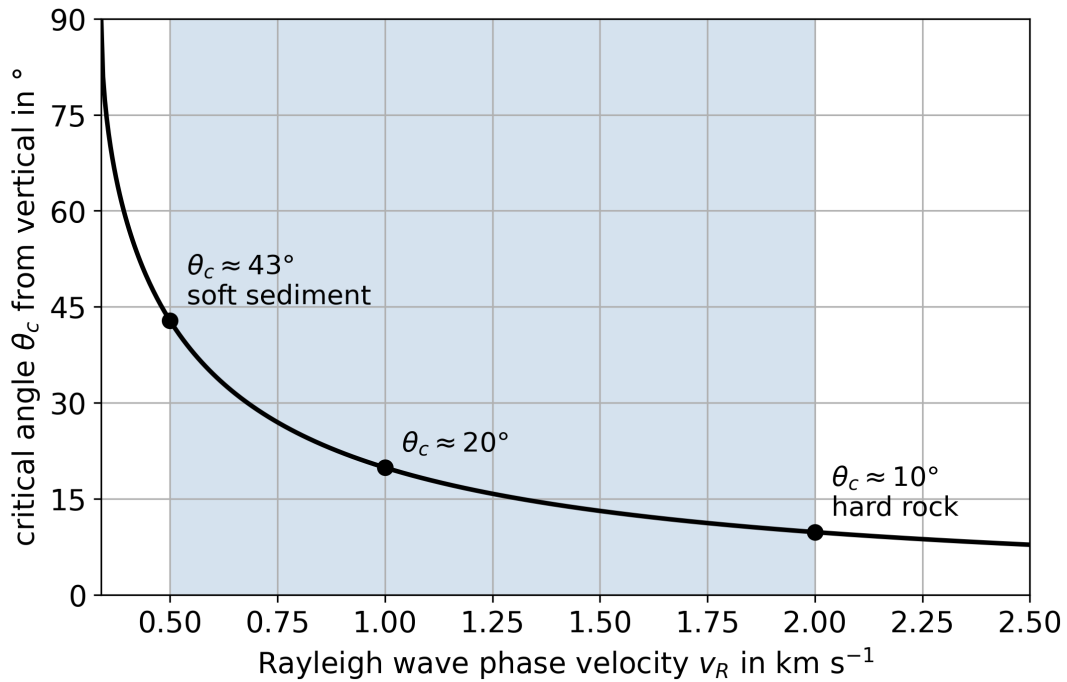


Figure 3.6.: Critical angle θ_c versus Rayleigh wave phase velocity v_R , indicating that the acoustic-to-seismic coupling unconsolidated surface material can yield to more efficient surface wave coupling for shallow incidence angles.

Acoustic arrivals close to this angle excite the subsurface resonantly and couple energy most

efficiently into the ground (see Figure 3.6). Soft, water-saturated sediments and unconsolidated soils exhibit more efficient coupling than hard rock, because their lower seismic velocities reduce the impedance contrast with air and shift θ_c to values more commonly achieved by atmospheric refracted arrivals (Cates and Sturtevant, 2002). Consequently, seismic stations installed on soft basin sediments are generally more sensitive to acoustically coupled meteoroid signals than those on hard bedrock sites. This effect must be accounted for when comparing signal amplitudes across a set of seismic stations (see Chapter 6).

Despite its importance as a scaling factor between observed seismic amplitudes and meteoroid source parameters, the efficiency of acoustic-to-seismic coupling is poorly constrained at virtually every seismic stations in practice. Unlike infrasound networks, for which calibration procedures have been developed to relate measured pressure amplitudes directly to source energies and, by extension, to meteoroid masses (ReVelle 1976; Edwards et al. 2008), seismic stations lack equivalent site-specific coupling calibrations. Targeted calibration measurements at individual stations, for example through controlled acoustic sources or well-characterised natural events are therefore needed to establish the transfer function from acoustic overpressure to ground velocity at each site. Only then can seismic amplitudes can be used reliably to infer source masses in a quantitative way (Fernando et al., 2024b). Such calibrations are important not only for meteoroid science, but also for planetary bodies where infrasound networks are absent. In these cases, acoustic-to-seismic coupling at broadband seismometers may be the only way to constrain the energy and mass of incoming meteoroids. This is demonstrated by the InSight seismometer deployed on Mars, where acoustic-to-seismic coupling has been identified as a key mechanism for detecting atmospheric events in the absence of a dedicated infrasound sensor (Kenda et al., 2020). Improving our understanding of coupling efficiency at terrestrial seismic stations would not only enhance mass estimates for future meteoroid events where infrasound data are unavailable or no a priori mass information exists, but also inform calibration strategies for future planetary seismic networks.

The W-Wavelet

The characteristic seismic waveform produced by the acoustic-to-seismic coupling of an meteoroid-generated acoustic wave is commonly described as a W-wavelet (see Fig. 3.5). It is a multi-lobed waveform with alternating polarity phases that directly reflects the time-domain structure of the impinging acoustic N-wave (Langston, 2004; Edwards et al., 2008). The acoustic N-wave consists of a compressional onset and a trailing return to ambient pressure. In this way bipolar pressure loading is applied to the ground surface. The seismic ground velocity response to this loading is approximately proportional to the time derivative of the acoustic pressure history. Differentiating the bipolar N-wavelet transforms it into a tripolar W-wavelet. On the vertical component of a seismometer, this results in a negative (downward) velocity pulse, which is associated with the sharp compressional onset and deformation of the subsurface from above. This is followed by a positive pulse as the overpressure reverses its polarity and a further negative pulse that is mostly a rebounding effect due to the subsurface returning to its initial state. The resulting three-lobed, W-shaped waveform is therefore an imprint of the

N-wave pressure structure convolved with the impulse response of the coupling medium and the seismometer transfer function (Langston, 2004).

The duration and dominant period of the wavelet are primarily determined by the acoustic N-wave period, which is proportional to the energy deposited per unit path length, and thus to the size of the meteoroid (ReVelle, 1974). For sub-tonne to multi-tonne meteoroids, which are the focus of this thesis, the dominant period of the coupled seismic signal usually falls within the range of 0.025–2 s (0.5–40 Hz). Compared to the acoustic N-wave, the frequency range is broader in the high-frequency direction. This is mainly due to coupling at the Earth’s surface and site conditions, which can modulate the frequency content of the seismic W-wavelet. Thus, the W-wavelet morphology can serve as a diagnostic waveform signature for acoustic-to-seismic coupling, which is one of the key features exploited by the deep-learning phase identification approach developed in this thesis to discriminate between meteoroid-related phases and other seismic arrivals and ambient seismic noise (see Chapter 5). The apparent propagation velocity of this phase across a seismic network reflects the acoustic rather than seismic wave speed, which provides an additional way to separate the different phases (Edwards et al., 2008).

Surface Wave Generation

Two distinct surface wave phenomena can be generated by acoustic-to-seismic coupling of meteoroid-generated shock waves. These are precursory Rayleigh waves and air-coupled Rayleigh waves (Edwards et al. 2008; see Figure 3.7).

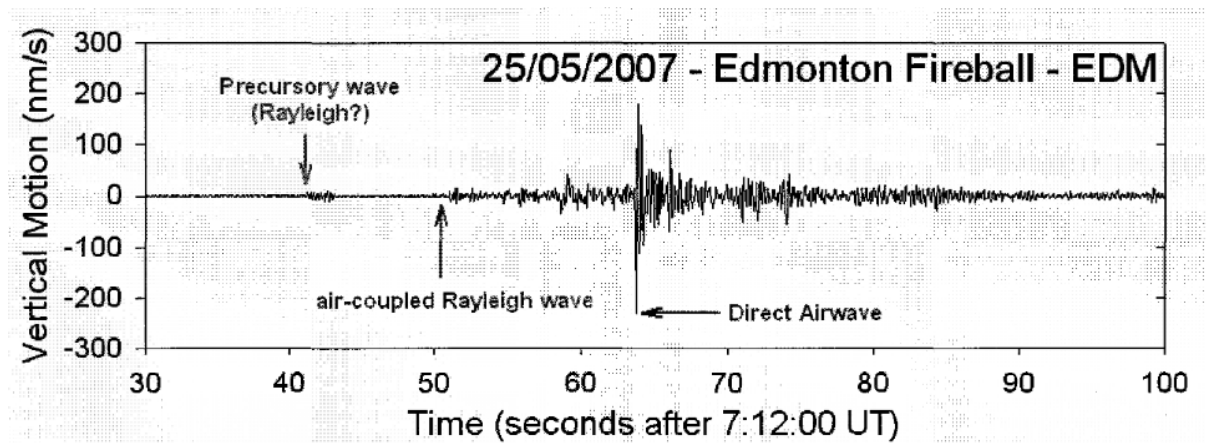


Figure 3.7.: A possible precursory surface wave arriving several seconds prior to the signals associated with the direct acoustic-to-seismic coupled wave. The air-coupled Rayleigh wave is shown here as well. Taken from Edwards (2008).

Precursory Rayleigh waves arise from a mechanism of resonant coupling that operates when the horizontal phase velocity of the incident infrasound wave, which is defined as the acoustic velocity divided by the sine of the incidence angle, precisely matches the horizontal phase velocity of a seismic surface wave mode in the near-surface medium. This most commonly occurs with the fundamental-mode Rayleigh wave (Ben-Menahem and Singh, 2012; Langston,

2004). Under these conditions, the acoustic wave excites an additional seismic wave, that propagates independently in the shallow ground surface. For a supersonic meteoroid producing a cylindrical Mach wave, this coupling leads to the formation of outward-propagating Rayleigh waves. These waves exhibit mirror symmetry along the projected ground path of the trajectory. For a quasi-spherical fragmentation source, outward-propagating waves with cylindrical symmetry are generated (Edwards et al., 2008). For precursory coupling to occur and Rayleigh waves to appear, the near-surface shear wave velocity v_s must exceed the ambient acoustic wave speed c_s . When the shear wave velocity is smaller than the acoustic wave speed, which is frequently observed in unconsolidated or water-saturated soils, where v_s can be as low as 100 m s^{-1} , this coupling mode is entirely suppressed (Edwards et al., 2008; Langston, 2004). This requirement to the near-surface geology might explain why precursory Rayleigh waves are not observed or documented often in meteoroid seismology (Anglin and Haddon, 1987; Tauzin et al., 2013; Olivieri et al., 2023). When Rayleigh waves are excited, they travel faster than the acoustic wave in air. This means that these seismic waves propagate ahead of the directly coupled acoustic wave, arriving at the recording station as precursor phases up to tens of seconds before the direct acoustic arrival.

The more commonly observed surface wave phenomenon is the air-coupled Rayleigh wave, which was first theoretically described by Press and Ewing (1951). In contrast to the precursory waves, the air-coupled Rayleigh waves are essentially attached to the acoustic wave as it propagates along the Earth's surface (Langston, 2004). Air-coupled Rayleigh waves are observed as a train of dispersive, periodic waveforms that appear immediately after the arrival of the directly coupled acoustic wave (Edwards et al., 2008). These waves are usually characterised by a retrograde elliptical particle motion in the vertical plane. However, prograde elliptical motion has also been observed and documented (Langston, 2004; Pujol et al., 2006). The position of the air-coupled surface wave train relative to the acoustic wave is determined by the relationship between the group velocity v_g and the apparent horizontal acoustic wave velocity. When v_g is less than the acoustic wave phase velocity, the surface waves trail behind. If v_g is higher the surface waves propagate faster than the acoustic wave (Edwards et al., 2008). Following the occurrence of the acoustic wave, the air-coupled Rayleigh wave amplitude decays. It can however, persist for a period ranging from tens of seconds to a few minutes (Edwards et al., 2008). Higher-order Rayleigh wave modes, alongside the fundamental mode, emerge in the presence of surface layering, which produces strong dispersion that results in a prominent Airy phase. This phase is a discrete, high-amplitude event that occurs at the minimum of the group velocity of the dispersion curve (Edwards et al., 2008). It can couple back from the ground into the atmosphere itself as an infrasonic signal. This phenomenon was observed during the re-entry of the Stardust sample return capsule in 2006, when the late arrival of the Airy phase was initially misidentified as a fragmentation event (Edwards et al., 2007). It is also notable that Love waves, which consist of SH waves, cannot be generated in an ideal homogeneous medium by meteoroid acoustic coupling, since atmospheric waves cannot produce transverse motion at the surface. In reality however, near-surface heterogeneities scatter the induced wavefield, producing a mixture of P/SV and SH motion, which makes these surface waves observable (Langston, 2004; Edwards et al., 2008).

In practice, it can be difficult to distinguish between precursory waves and air-coupled Rayleigh waves, since both can generate leading retrograde particle motion in front of the main Mach

cone signal. The distinction primarily depends on the physical relationship of each phase to the acoustic arrival. Precursory Rayleigh waves are seismic phases that have decoupled from the acoustic wave and travel at surface-wave velocity. They therefore arrive at a fixed seismic moveout and advance further ahead of the acoustic arrival as the source-receiver distance increases. In contrast, air-coupled Rayleigh waves remain tied to the acoustic wavefront and travel at acoustic trace velocity. This means that their timing relative to the direct acoustic arrival remains essentially constant regardless of distance.

3.2.5. Meteoroid Blast Radius

The concept of a blast radius was introduced by ReVelle (1974) in the context of the cylindrical blast wave generated by a supersonic meteoroid. It defines the radial distance from the meteoroid flight path at which the pressure perturbation becomes non-linear, meaning that its amplitude is approximately equal to or larger than the ambient pressure. Beyond this distance, the pressure perturbation decays as a linear acoustic wave. The blast radius is a key parameter in modelling infrasound sources and in predicting ground-level overpressure from fragmentation (ReVelle, 1976).

Following ReVelle's formulation, the blast radius R_0 is defined as the distance at which the peak overpressure ΔP equals the ambient atmospheric pressure p_0 . The blast radius of the cylindrical meteoroid source itself, which is approximated as a line source, is then:

$$R_0 = d_m M = d_m \frac{v}{c_s}, \quad (3.8)$$

where d_m is the meteoroid diameter and M is the Mach number at the relevant altitude (ReVelle 1974; Silber et al. 2015). Equivalently, R_0 can be expressed in terms of the energy deposited per unit path length E_l in J m^{-1} and the ambient atmospheric pressure in Pa:

$$R_0 = \sqrt{\frac{E_l}{p_0}} \quad (3.9)$$

This formulation is more useful observationally because E_l can be estimated from the measured light curve derived from camera measurements or from the period of a recorded infrasound wave (Edwards et al., 2009). For a 1 m diameter meteoroid travelling at 30 km s^{-1} at an altitude of 40 km (with $p_0 \approx 300 \text{ Pa}$), R_0 is of the order of 1-2 km. For the 2013 Chelyabinsk event R_0 , was in the order of 10 km at the peak fragmentation altitude (Silber et al. 2015; Brown et al. 2013). Beyond the blast radius, the overpressure amplitude decays as $r^{-1/2}$ for cylindrically spreading line source, transitioning to spherical spreading with r^{-1} at distances $\gg R_0$, where the curvature of the finite-length meteoroid trajectory becomes apparent (ReVelle, 1974). At the Earth's surface, the overpressure can cause structural damage. Window breakage begins at a pressure difference of approximately 200-500 Pa (Gi et al., 2018), while catastrophic structural damage requires $\Delta p \approx 1\text{-}10 \text{ kPa}$ (Glasstone, 1956). This means that the blast radius is directly connected to the footprint of the hazard of fragmentation events during meteoroid falls (Boslough and Crawford 2008; Glasstone 1956; Gi et al. 2018).

3.3. The Meteor Light Curve

This section as well as the following section 3.4 are part of my 3-month stay at the Meteor Physics Group of Prof. Peter Brown at Western University in Canada. His group focuses on the observation and modelling of meteoroids passing through our atmosphere and operates radar, optical, infrasound and all-sky camera systems. During this stay I compiled a 5-year dataset of infrasound array recordings for visually detected meteors around the Elginfield Observatory that belongs to Western University. The project idea is to constrain the luminous efficiency parameter (see section 3.4) through the combination of optical and infrasound records. For this I analysed more than 3000 meteoroid events. The luminous efficiency parameter is one of the most poorly constrained parameters in meteor science and mass estimations depend directly on it. Using infrasound data the luminous efficiency parameter can be constrained independently and be estimated over multiple events, potentially closing a long standing knowledge gap. Unfortunately, a full publication has not made it into my thesis, because I ran out of time. Nevertheless, I will stay in close connection with Prof. Brown during this and future projects.

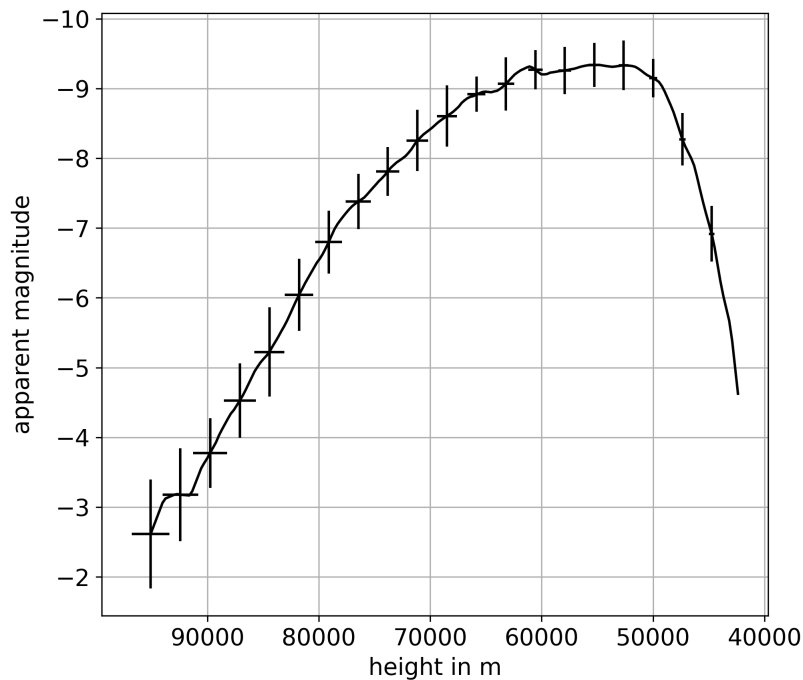


Figure 3.8.: Average light curve out of 4 cameras, with 1σ deviations, of a Geminid meteor, as measured by the Meteor Physics Group at Western University using their meteor camera network. The smooth light curve without short-duration brightness peaks indicates that this meteoroid did not undergo large scale fragmentation events.

The temporal profile of the optically radiated power, the light curve, encodes information about the ablation dynamics, fragmentation history, and physical structure of the meteoroid. For a single, unfragmented body undergoing steady ablation, the light curve is approximately bell-shaped, with a peak near the altitude of maximum energy deposition, and is approximately

symmetric (see Fig. 3.8). Fragmentation episodes produce distinct superimposed flares. These flares are short-duration brightness peaks and are one of the most important indicators of a meteoroid's mechanical strength and internal structure (Campbell-Brown, 2019). The total optically measured radiated energy E_{opt} , which is integrated over the light curve, is related to the kinetic energy input through the luminous efficiency τ , which is described in the following sections.

3.4. Luminous Efficiency

The luminous efficiency τ is the dimensionless ratio of the power radiated as visible light to the meteoroid's rate of kinetic energy loss. It is both the most important and one of the most poorly constrained parameters in meteor science (Subasinghe and Campbell-Brown, 2018). Optically derived quantities, such as meteoroid mass, mass flux and the ablation coefficient, depend directly on its assumed value (Subasinghe and Campbell-Brown, 2018).

3.4.1. Definition and the Luminous Equation

The physical parameters of meteoroids can be derived from their trajectories and from the meteor itself. The standard single-body meteoroid flight equations, which describe deceleration and mass loss, include the luminous equation (Ceplecha et al., 1998; Bronshten, 1983). This equation relates the instantaneous optical power (in watts) to the rate of kinetic energy loss:

$$I = -\tau \frac{d}{dt} \left(\frac{1}{2} m v^2 \right) = -\tau \left(\frac{1}{2} v^2 \frac{dm}{dt} + m v \frac{dv}{dt} \right) \quad (3.10)$$

where τ is the luminous efficiency, m is the meteoroid mass, v is its speed, $\frac{dv}{dt}$ its deceleration, and $\frac{dm}{dt}$ is the mass loss rate due to ablation (Ceplecha et al., 1998). The mass loss rate can be described as following:

$$\frac{dm}{dt} = -\Gamma A (\sigma + \sigma_e) \rho_a \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}} v^3 \quad (3.11)$$

with the meteoroid's drag coefficient Γ , atmospheric density ρ_a , the shape factor A , the meteoroid density ρ_m , the ablation coefficient σ (Ceplecha et al. 1998), and the erosion coefficient σ_e (Campbell-Brown et al. 2013). The deceleration can be written as follows:

$$\frac{dv}{dt} = \frac{\Gamma \rho_a v^2}{m} A \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}} \quad (3.12)$$

We can compare the deceleration and ablation term by taking their ratio, starting from the luminous equation:

$$\frac{m v \frac{dv}{dt}}{\frac{1}{2} v^2 \frac{dm}{dt}} = \frac{2 m \frac{dv}{dt}}{v \frac{dm}{dt}} \quad (3.13)$$

By substituting the equations 3.11 and 3.12, we can find that most of the equations cancel each other and both contain the factor $\Gamma A \rho_a (m/\rho_m)^{2/3}$. Writing the factor as Λ , equation 3.13 can be written as:

$$\frac{2\Lambda v^2}{v(\sigma + \sigma_e)\Lambda^3} = \frac{2}{(\sigma + \sigma_e)v^2} \quad (3.14)$$

This means that the deceleration term becomes negligible when $(\sigma + \sigma_e)v^2 \gg 2$, which holds for low ablation coefficients and moderate meteoroid velocities. When the deceleration term is negligible, the luminous equation can be simplified to:

$$I \approx -\frac{1}{2}\tau v^2 \frac{dm}{dt} \quad (3.15)$$

Integrating equation 3.15 and assuming τ and v vary slowly compared to the mass loss, the total radiated energy is $E_{opt} \approx \frac{1}{2}\tau v^2 \Delta m$. If the meteoroid is almost completely ablated then $\Delta m \approx m_0$ at nearly constant speed v_0 . The kinetic energy $E_k = \frac{1}{2}m_0 v_0^2$ can be related to the total optically radiated energy $E_{opt} = \tau \Delta E_k$, if all kinetic energy is deposited during atmospheric flight and the terminal mass and speed is negligible. This can be rearranged so that the initial mass m_0 of the meteoroid can, in principle, be estimated from the estimated light curve:

$$m_0 \approx 2 \frac{E_{opt}}{\tau v_0^2} \quad (3.16)$$

This demonstrates the direct and proportional conversion of uncertainties in luminous efficiency τ into uncertainties in the derived meteoroid mass and fluxes.

3.4.2. Dependence on Entry Velocity and Physical State

Luminous efficiency is not a material constant, but depends on several parameters. The most important of these is entry velocity. Physically, faster-moving meteoroids have more kinetic energy per unit mass, which heats the ablated material to higher temperatures. This, in turn, produces a more energetically excited, and therefore more luminous, plasma. Empirical and theoretical literature suggests an approximate power-law dependence:

$$\tau \propto v^a, \quad a \in [1, 2], \quad (3.17)$$

which is valid over the whole range of meteoroid velocities of 11-72 km s⁻¹ (Cepilecha and McCrosky 1976; Bellot Rubio et al. 2000; Subasinghe and Campbell-Brown 2018). Jenniskens (2007) reported a velocity dependence in the non-thermal component of meteor spectra, suggesting that the shock-heated plasma component obeys the relationship $\tau \propto v^2$. However, the velocity scaling of τ remains a subject of active debate, with published values ranging from 0.1-10% across the velocity range (Subasinghe and Campbell-Brown 2018; Gritsevich and Koschny 2011; Drolshagen, E. et al. 2021)

3.4.3. Practical Determination of Luminous Efficiency

Because luminous efficiency cannot be measured directly, it is usually estimated using one of four approaches. First, if meteorites can be recovered, their masses can provide a lower bound on the entry mass. Combined with photometric light curves, τ can be estimated. This approach yields $\tau \approx 0.5\text{--}7\%$ at meteoroid velocities of $15\text{--}20\text{ km s}^{-1}$ (Ceplecha and McCrosky 1976; Halliday et al. 1996). Another approach involves estimating τ by taking simultaneous optical and radar measurements. Radar provides an independent velocity measurement, as well as a mass estimate through the electron line density (Close et al. 2004), which is related to the mass of the meteoroid through the ionisation coefficient (Jones 1997; Campbell-Brown et al. 2012). Luminous efficiency can then be inferred from the ratio of the optical-to-radar estimated masses, or equivalently from the ratio of the radiated power to electron line density (Weryk and Brown 2013; Subasinghe and Campbell-Brown 2018). The third approach involves numerical modelling where fragmentation equations for single or multiple bodies are fitted to the light curve and the deceleration profile of the meteoroid. This yields self-consistent τ values (Vondrak et al. 2008). The fourth approach is optical-acoustic measurement, where the light curve and the corresponding blast radius estimation depend on the luminous efficiency. Through modelling, the measured infrasound frequency is then fitted to the luminous efficiency value (McFadden, 2025).

Despite decades of research efforts to tightly constrain luminous efficiency, it remains uncertain to within a factor of two to three. Treating it as a constant across different meteoroid events is one of the main sources of error in estimates of meteoroid mass flux from optical observations, the most common technique for observing meteoroids (Ceplecha et al. 1998). This persistent uncertainty motivates the use of independent constraints derived from the acoustic and seismic signatures of meteoroids. Optical and radar methods, infer mass through the luminous efficiency and the ionisation coefficient, which are themselves functions of speed and composition. However, the optic-acoustic approach ties the energy deposition to the physically measurable blast radius and the dominant frequency of the resulting shock wave. As the luminous efficiency that determines the optical light curve also affects the conversion between deposited energy and acoustic source dimension, an event observed simultaneously by optical, infrasound and seismic instruments provides an over-determined system in which the luminous efficiency can be determined rather than assumed. Thus, acoustic and acoustic-seismic data are well positioned to help answer the long-standing open question of luminous efficiency, offering a route to mass and energy estimates that are independent of the optical luminous-efficiency assumption.

4. Data and Methods

4.1. Data

4.1.1. Atmospheric Data and Models

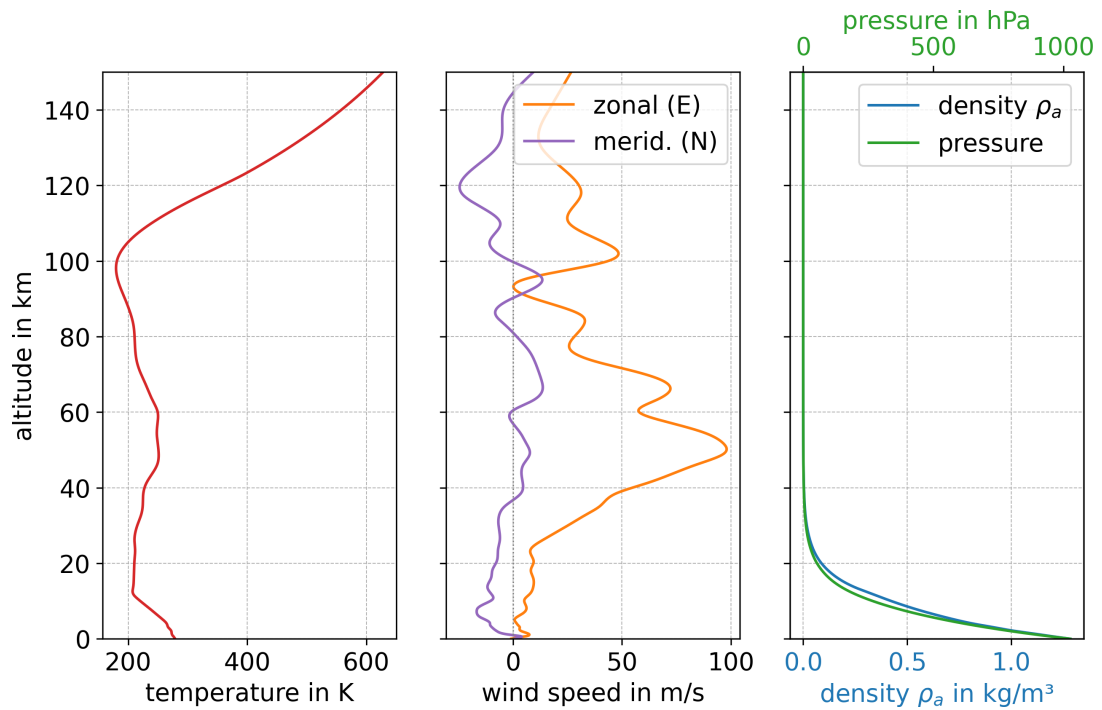


Figure 4.1.: Atmospheric specifications of a G2S model accessed with the `ncpag2s-clc` utility, for the date of Nov 14th 2017 at 17:00 UTC.

An accurate characterisation of the atmospheric state is essential for modelling the propagation of infrasound and understanding observed waveform features. In this work, two data sources are used to derive atmospheric specifications: empirical models and radiosonde observations. The empirical atmospheric models were obtained using the Ground-to-Space (G2S) model (Drob et al., 2003), which provides a continuous vertical profile of wind and temperature from the Earth’s surface to the upper thermosphere. This model merges lower-atmosphere numerical weather prediction models (e.g., ECMWF: Hersbach et al. 2020), with middle-atmosphere (approximately 30-80 km) weather prediction models (e.g., NAVGEM-HA: Eckermann et al. 2018) and empirical upper-atmosphere models (e.g., HWM: Drob et al. 2015; NRLMSISE-00: Picone et al. 2002). G2S profiles were generated using the `ncpag2s-clc` utility (Hetzer, 2024),

which queries the NCPA G2S server at arbitrary geographic locations and at hourly UTC intervals. The utility then returns profiles of temperature, wind (split into its meridional and zonal components), pressure, and density on a uniform 0.1 km vertical grid from the surface to an altitude of 150 km (see Figure 4.1).

In addition to modelled atmospheric profiles, radiosonde observations were used to constrain the structure of the lower-tropospheric wind and temperature, when available. The radiosonde data were obtained from the Integrated Global Radiosonde Archive (IGRA), which is operated by NOAA's National Centers for Environmental Information (NCEI) (Durre et al., 2006). These in situ measurements most often extend to altitudes of 30–35 km, providing high-resolution vertical profiles of temperature, humidity and horizontal wind components at launch intervals of at least twice daily. Where available in close temporal and spatial proximity to the events of interest, the lower-atmosphere portion of the G2S specifications was validated and replaced, where appropriate, using the radiosonde profiles.

4.1.2. Waveform Data

The waveform data used in the studies presented here include both seismic and infrasound recordings from a combination of global and local station networks. All waveform data were retrieved with computing routines using the ObsPy framework (Beyreuther et al., 2010), which provides a standardised interface to the Federation of Digital Seismograph Networks (FDSN) web services (Suárez et al., 2008). This approach ensures reproducibility and enables automated data retrieval across heterogeneous network operators.

4.1.2.1. Seismic Data

The waveform data used in the studies presented here include a global set of stations (see Chapter 5) and otherwise local station sets (see Chapters 6; 7). The seismic waveform data were retrieved from a combination of global and regional networks, which are accessible via FDSN web services, using the ObsPy FDSN client (Beyreuther et al., 2010). Data requests were issued to multiple FDSN data centres, primarily the EIDA-Federator (Strollo et al., 2021) and IRIS-DMC (Trabant et al., 2012), as well as national data centres where regional coverage was required. The federated nature of FDSN services enables a single, standardised query interface to span heterogeneous networks and operators, ensuring broad spatial coverage without the need for manual data handling. Three-component broadband and short-period waveform data were prioritised in the channel selection process. Following the SEED channel naming convention (IRIS, 2009), requests were submitted preferentially for high-broadband (HH[ZNE]), broadband (BH[ZNE]), and short-period (EH[ZNE]) channels, in which the first letter encodes the sampling rate band and the second letter denotes the instrument type. If the deployed seismometer provides the full three-component ground motion vector (vertical Z, north N, east E), both scalar amplitude analyses and polarisation-based signal characterisation are possible. Otherwise, only the vertical component is used.

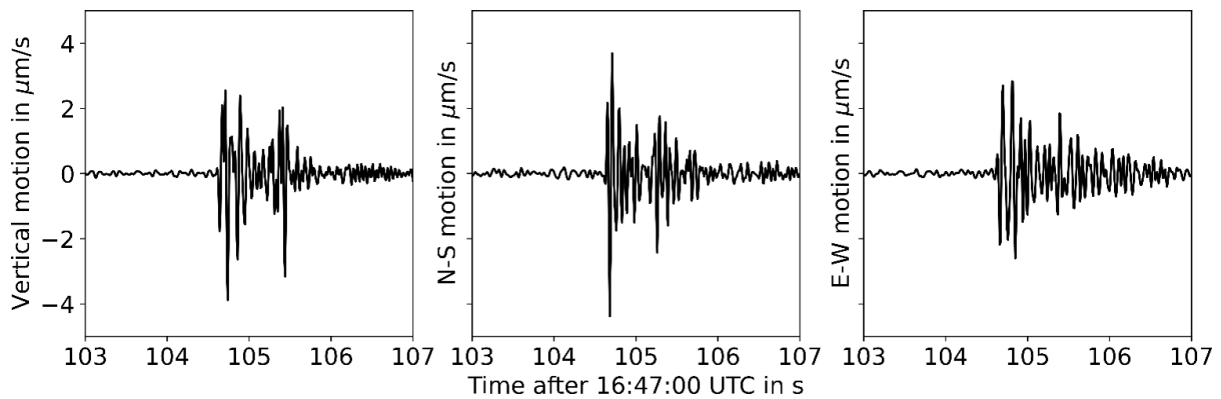


Figure 4.2.: High-broadband three component recording data of the seismic station KB.DEP07, bandpassed with corner frequencies of 1 and 20 Hz. The meteoroid-related signal starts at roughly 104.5 seconds after the event time of 16:47 UTC.

BH channels, sampled at 20–40 Hz, and HH channels, sampled at 100 Hz, correspond to broadband and high-broadband seismometer recordings, respectively. A 3-component broadband recording of a seismometer can be inspected in Figure . They provide waveform data across the frequency range most relevant to acoustic-to-seismic coupling, which is approximately in the range of 0.5–40 Hz. EH channels, which are sometimes sampled to up to 100 Hz but associated with short-period sensors, were included to extend coverage at regional networks where broadband instruments were not deployed.

4.1.2.2. Infrasound Data

Infrasound waveform data are utilised in two distinct configurations. These are single-station and array configurations, which each offer complementary observational capabilities.

Single-Station Infrasound. Single-station infrasound data were obtained from FDSN-accessible networks operating microbarometer sensors and were retrieved with computing routines via the ObsPy FDSN client (Beyreuther et al., 2010) following the same approach as for seismic data. Infrasound channels follow the SEED naming convention (IRIS, 2009), with the instrument code D denoting differential pressure sensors. The most commonly encountered channel designations are BDF and HDF, corresponding to broadband and high-broadband sampling rates, respectively. In recent years, the low-cost, citizen-science Raspberry Shake network has expanded significantly, and a combined seismic and infrasound instrument, which is called the Raspberry Shake & Boom, has become commercially available (Lamb et al., 2021). These instruments record a BDF pressure channel alongside one vertical component seismic channel, and are FDSN-accessible via the Raspberry Shake network AM (Raspberry Shake, S.A., 2016). Due to their rapid global increase in deployments, Raspberry Shake & Boom stations represent a substantial fraction of the single-station infrasound recording sites used in this work, particularly in regions lacking dedicated research infrastructure. Single-station infrasound recordings are especially valuable for characterising waveform amplitude and frequency

content at individual sites, and are incorporated here for events where array deployments were unavailable within the region of interest.

Infrasound Array Data. Infrasound array data were collected from stations belonging to the International Monitoring System (IMS) infrasound network (Christie and Campus, 2009), as well as regional infrasound arrays available in the respective research area. The IMS, which is operated by the Comprehensive Nuclear-Test-Ban Treaty Organisation (CTBTO), as of 2022, contains 53 certified array stations distributed globally (Hupe et al., 2022). A subset of IMS array waveform data is FDSN-accessible in near real-time and additional regional infrasound arrays were accessed through FDSN services where available. Infrasound arrays are designed to suppress incoherent wind noise while preserving coherent infrasonic signals. They generally consist of four to eight elements distributed over 1-3 km apertures (Christie and Campus, 2009), with individual elements recording either BDF or HDF differential pressure channels. The spatial distribution of array elements enables array processing techniques such as beamforming (see Chapter 4.2.2.3), yielding estimates of the signal’s apparent velocity and back-azimuth. These array-derived parameters provide direct observational constraints on the origin and propagation path of the measured signal, and serve as key observables in the propagation and source analyses described in subsequent chapters.

4.2. Methods

4.2.1. Acoustic-Seismic Waveform Processing

Prior to any analysis, all seismic and infrasound waveforms undergo a series of standard pre-processing steps to ensure data quality and comparability across different sensors. The first step involves demeaning and detrending the raw time series. Demeaning removes the mean offset of the signal, which can arise from instrumental DC bias or digitiser offsets, which would otherwise introduce artefacts in subsequent spectral analysis (Havskov and Alguacil, 2004). Detrending additionally removes linear trends from the data, which can result from slow instrumental drift, temperature-dependent sensor response, or tilt effects on broadband seismometers (Bormann, 2002). Both of these operations are essential prerequisites for any subsequent filtering or spectral processing step, as uncorrected low-frequency polynomial trends can leak across the entire frequency spectrum and severely degrade signal quality if left uncorrected (Baisch and Bokelmann, 1999). Following demeaning and detrending, a bandpass filter is applied to retain signal energy within the desired frequency band, which in the case of meteoroid-related signals is the range of 0.5-40 Hz. The lower cutoff of 0.5 Hz suppresses long-period noise sources, including microseismic (Webb, 1992) and microbarometric (Waxler and Gilbert, 2006) noise, atmospheric pressure fluctuations, and instrumental tilt noise, which dominate below roughly 1 Hz at most seismic and infrasound stations (Peterson, 1993). The upper cutoff of 40 Hz removes high-frequency instrumental noise, electronic noise, and other artefacts that are unrelated to the signals under investigation. This frequency band is well suited to capture the characteristic signal content of meteoroid-generated infrasound and its

coupled seismic expression. Meteoroid fragmentation events predominantly radiate acoustic energy in the infrasonic range from below 1 Hz up to several tens of Hz. The spectral peak depends on the meteoroid's size and velocity (ReVelle, 1997; Brown et al., 2002; Edwards et al., 2008). The seismic wave, after coupling, retains energy up to approximately 40 Hz, so the high-frequency cutoff preserves both the direct infrasound arrival and the acoustic-to-seismic coupled counterpart. This also enables joint waveform analysis across co-located sensors. In this study, a fourth-order minimum phase Butterworth filter is used to avoid phase distortion of the waveform, which could impair arrival time identification (Scherbaum, 2006). The final preprocessing step is the removal of the instrument response, also referred to as restitution. Each sensor records a convolution of the true ground motion or pressure signal with the instrument's frequency-dependent transfer function. If this response is not corrected for, the recorded waveform amplitudes are expressed in raw counts and are specific to the particular sensor model, gain settings, and digitiser used. Deconvolving the pole-zero instrument response from the raw data, converts seismic waveforms to ground velocity in units of m s^{-1} and infrasound waveforms to acoustic overpressure in Pa (Havskov and Alguacil, 2004; Edwards et al., 2009). This step is essential for two reasons. First, it renders waveform amplitudes physically meaningful and directly interpretable in terms of ground motion or pressure amplitude. This enables direct estimation of acoustic-to-seismic coupling coefficients (Edwards et al., 2007; Novoselov et al., 2020). Second, it enables quantitative comparison of amplitudes recorded across the heterogeneous sensor networks, which is the case for most meteoroid events. All preprocessing steps are implemented using the ObsPy framework (Beyreuther et al., 2010), which provides standardised routines for waveform I/O, filtering, and response removal compliant with the SEED/StationXML metadata standards.

4.2.2. Trajectory Reconstruction

4.2.2.1. Geometric reconstruction Method (based on Pujol et al. (2005))

The standard acoustic-seismic trajectory reconstruction method for supersonic objects flying through the atmosphere was developed by Pujol et al. (2005). In this method, the meteoroid trajectory is defined by 6 parameters: its azimuth φ , the angle from the vertical ϑ , the horizontal coordinates (x_0, y_0) of the point P where the extrapolated trajectory intersects the Earth's surface (see Fig. 4.3), the corresponding time t_0 , and the meteoroid velocity v . For a station at position $\mathbf{r}_s$, the arrival time of the shock wave is:

$$t = t_0 - \frac{d_t}{v} + \frac{d_p \cos \theta_M}{c_s}, \quad (4.1)$$

where $d_t = \overline{PQ}$ and $d_p = \overline{SQ}$ are the distances along and perpendicular to the trajectory from P to the closest point on the Mach cone (see Fig. 4.3), c_s is the speed of sound, and $\theta_M = \sin^{-1}(c_s/v)$ is the Mach angle. Because Eq. 4.1 is non-linear in the trajectory parameters,

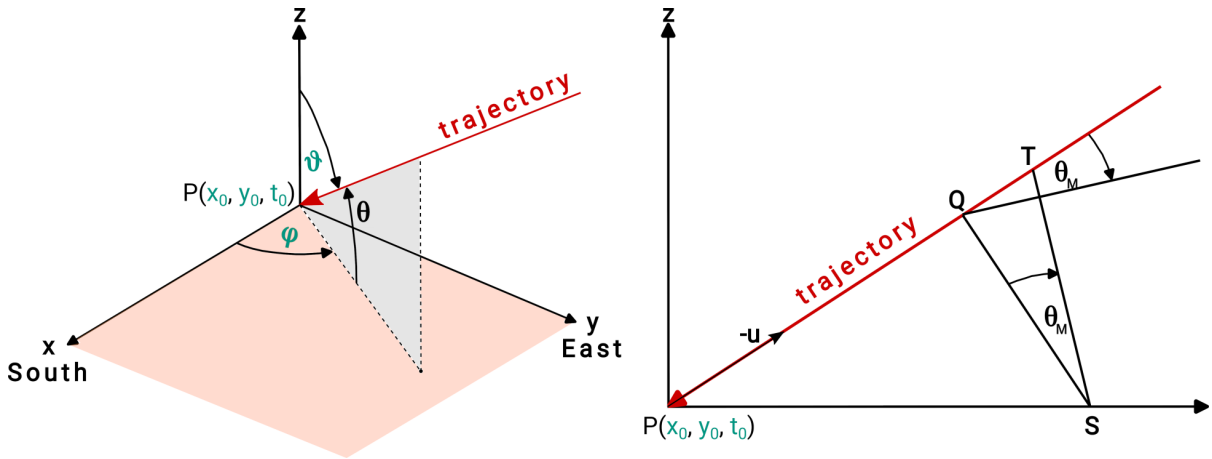


Figure 4.3.: The coordinate system that is used for the trajectory definition of a moving source. P indicates the intersection of the Earth's surface with the trajectory. ϑ is the inclination angle θ the elevation angle and φ the azimuth angle, positive measured from South. Modified after Pujol et al. (2005).

the inversion is linearised via a Taylor series expansion. This gives:

$$\mathbf{r} = \mathbf{t}^{\text{obs}} - \mathbf{t} = \frac{\partial t}{\partial v} dv + \frac{\partial t}{\partial x_0} dx_0 + \frac{\partial t}{\partial y_0} dy_0 + dt_0 + \frac{\partial t}{\partial \varphi} d\varphi + \frac{\partial t}{\partial \vartheta} d\vartheta, \quad (4.2)$$

which is solved iteratively using either damped least squares or a generalized inverse (singular-value decomposition). The misfit is quantified by the root-mean-square residual over all stations. While this method can provide reasonable trajectory solutions, it is limited by the simplifications that are made to linearise the system of equations and make the inversion feasible. There are three simplifying assumptions that become increasingly problematic as the spatial scale of the network grows:

1. **Constant meteoroid velocity.** The meteoroid decelerates significantly as it passes through the atmosphere. In the Morávka case (e.g., Brown et al. 2003) for example, the main fragment slowed from ~ 22 km/s to ~ 4 km/s over its visible path. Pujol's method uses a single effective velocity, which causes v and t_0 to trade off with each other and renders both essentially undetermined. Nevertheless, the trajectory angles and the ground intersection coordinates are somewhat robust.
2. **Straight ray paths.** Both the meteoroid trajectory and the acoustic ray paths from the Mach cone to the seismic stations are assumed to be straight lines. In reality, sound speed gradients and temperature stratification refract acoustic rays, an effect that grows with propagation distance. The straight-ray approximation is adequate for small inter-station distances but must be considered for larger networks.
3. **No atmospheric winds and constant sound speed.** The speed of sound c_s varies with altitude (by ~ 15 -20% over 0-70 km), and horizontal wind fields introduce asymmetric travel-time perturbations that are not captured by the model. Using a height-averaged value of c introduces systematic errors in the inferred coordinates x_0, y_0 and in this way also directly into the angular variables φ and ϑ . Additionally, via the trade-off described above, in v and t_0 more errors are introduced, when not accounting for wind

and atmospheric stratification. Partial corrections, such as altitude-dependent averages or displacing stations by the wind-drift vector, can reduce but not eliminate these errors.

Additionally, a sufficient initial parameter guess must be provided for the inversion to converge. This can be achieved through a grid search over all six dimensions of the trajectory parametrisation, or via a particle swarm optimisation approach, which was initially applied to this trajectory inversion scheme by Vida (2023).

4.2.2.2. Ray based reconstruction Method

Instead of assuming linear ray paths, ray tracing can be used as a high-frequency approximation of the acoustic waves produced by meteoroids, accounting for temperature and wind stratification, as well as sound speed gradients in the atmosphere, thus overcoming the limitations of the geometric method. A supersonic object travelling at velocity v in an atmosphere with a local sound speed c_s radiates energy perpendicular to its Mach cone. The ray direction

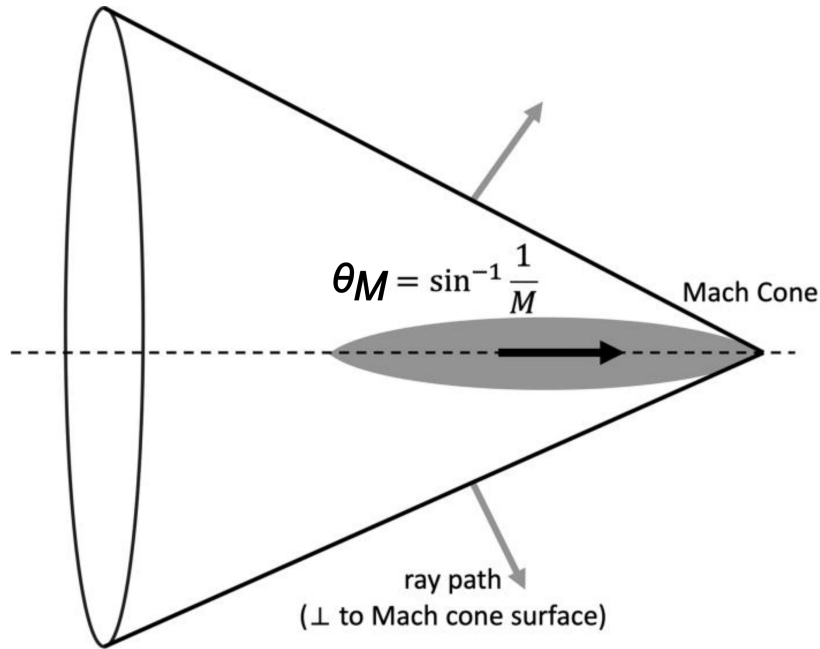


Figure 4.4.: The geometry of ray paths produced by a Mach cone. The black arrow indicates the direction of motion $\hat{e}_y$ and the grey arrows indicate the ray directions $\hat{n}_0(M, \lambda)$. From Blom et al. (2024).

$\hat{n}_0$ is perpendicular to the cone surface and, for a source moving along $\hat{e}_y$, can be expressed as follows:

$$\hat{n}_0(M, \lambda) = \sqrt{1 - \frac{1}{M^2}} \sin \lambda \hat{e}_x + \frac{1}{M} \hat{e}_y + \sqrt{1 - \frac{1}{M^2}} \cos \lambda \hat{e}_z, \quad (4.3)$$

where $\lambda \in [0^\circ, 360^\circ]$ sweeps the full cone. Arbitrary elevation and azimuth angles for a given meteoroid trajectory are handled by two successive rotation matrices $\mathbf{R}_x$ and $\mathbf{R}_z$:

$$\mathbf{R}_x(\theta) = \begin{pmatrix} 1 & 0 & 0 \\ 0 & \cos \theta & -\sin \theta \\ 0 & \sin \theta & \cos \theta \end{pmatrix}, \mathbf{R}_z(\varphi) = \begin{pmatrix} \cos \varphi & -\sin \varphi & 0 \\ \sin \varphi & \cos \varphi & 0 \\ 0 & 0 & 1 \end{pmatrix} \quad (4.4)$$

with θ being the elevation angle, measured from the horizontal and φ being the azimuth angle. This yields $\hat{n}(\theta, \varphi, M, \lambda) = \mathbf{R}_z(\varphi)\mathbf{R}_x(\theta)\hat{n}_0(M, \lambda)$. The Mach number M itself is re-evaluated at each point along the trajectory using the local atmospheric sound speed, so a meteoroid decelerating from tens of km/s to a few km/s is handled consistently throughout its path.

Pujol's model simplifies the acoustic rays to straight lines. Instead, Blom et al. (2024) launch the rays defined by Eq.4.3 into the infraGA/geoac spherical-geometry ray-tracing solver, which integrates the full Eikonal equations through a altitude-stratified atmosphere. Temperature and wind gradients naturally refract ray paths, reproducing phenomena entirely absent from the straight-ray model, such as tropospheric and stratospheric ducting, thermospheric returns, and the focusing of energy into distinct atmospheric waveguides.

Rather than using a single average value for the speed of sound across the entire atmosphere, a full ground-to-space (G2S) atmospheric specification is used, providing temperature, density, and both zonal (U_w) and meridional (V_w) wind profiles from the surface to well above 100 km. Wind enters via the effective sound speed $c_{\text{eff}}(z, \varphi) = c_s(z) + U \sin \varphi + V \cos \varphi$, which determines whether a ray launched at a given elevation and azimuth is ducted back to the ground or escapes into the upper atmosphere.

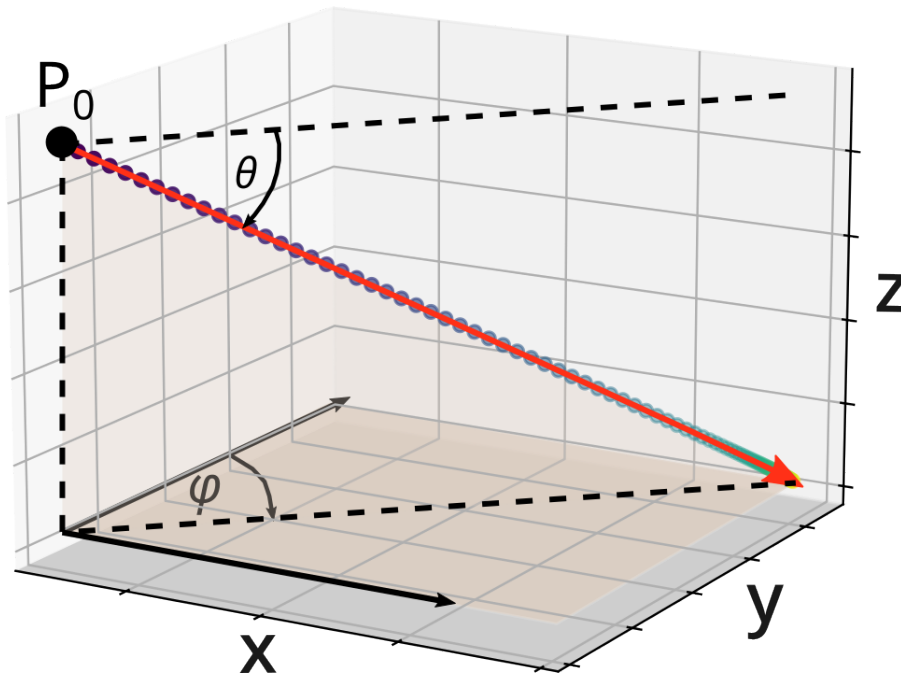


Figure 4.5.: infraGA trajectory parametrisation. $P_0(x, y, z)$ marks the onset point of the trajectory from which the trajectory is calculated with the direction defined by the azimuth angle φ , the inclination angle θ , and the meteoroid velocity v .

For the ray based trajectory reconstruction the trajectory parametrisation differs slightly from the Pujol et al. (2005) definition. Instead of using the piercing point where the trajectory intersects Earth's surface, the trajectory is defined by its highest altitude point P_0 (see Fig. 4.5). From there the trajectory is calculated by using the equations of motion following

Campbell-Brown et al. (2013), accounting for deceleration instead of assuming an overall constant meteoroid velocity:

$$\frac{dv}{dt} = \frac{\Gamma \rho_a v^2}{m} A \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}} \quad (4.5)$$

With the meteoroid velocity v , drag coefficient Γ , atmospheric density ρ_a the meteoroid mass m , the shape factor A and the meteoroid density ρ_m . The calculated trajectory can then be evaluated, and the residual arrival time between the observed and measured arrivals can be determined. By evaluating many trajectories in this way, a parameter spread that best describes the measured data can be found. However, compared to the inversion method described in Pujol et al. (2005), this approach is computationally very expensive, with a single trajectory taking minutes to calculate on a standard home computer. Achieving a satisfactory fit to the measured data can take several days to weeks of high performance computing (HPC) time, in the case of the direct search for the Starlink satellite trajectory using 96 cores (see Chapter 6). Nevertheless, the fact that this method results in a more physical trajectory solution involving realistic ray-traced acoustic waves justifies its use. Additionally, the slow performance of direct ray tracing can be eliminated by applying a look-up table approach, as discussed in Chapters 5 and 7.

4.2.2.3. Using Infrasound Array Data and Beamforming

Infrasound arrays consist of spatially distributed microbarometers whose recordings can be combined through beamforming. Beamforming is a time-delay-and-sum technique that steers a virtual beam across candidate slowness vectors to identify the most coherent incoming wavefront (Rost and Thomas, 2002). This yields, at each time step, an estimated back-azimuth and apparent velocity (related to the angle of incidence) of the infrasonic signal (see Fig 4.6).

A widely used beamforming implementation is the Progressive Multi-Channel Correlation (PMCC) algorithm (Cansi, 1995), which is part of the standard Comprehensive Nuclear-Test-Ban Treaty (CTBT) infrasound array meteoroid analysis, for example. The algorithm detects and characterises coherent wave packets across the array in both time and frequency. One of the main benefits of signal-processing of array data rather than a single sensor is the increase in signal-to-noise ratio (SNR) gain. For an N -element array, coherent beamforming improves the SNR by a factor of $\sqrt{N}$ (Rost and Thomas, 2002). This enables the detection of meteoroid events at regional to global distances of several thousand kilometres (Brown et al. 2002; Le Pichon et al. 2010). Because the back-azimuth evolves continuously as the meteoroid passes through the atmosphere, a single well-sited array can track the apparent source direction over time, providing a time-parametrised evolution of signal back-azimuths (Ens et al. 2012; Silber and Brown 2014). When signals from two or more arrays are available, the intersection of their respective back-azimuth great-circle planes can be used to constrain the full 3D trajectory in a way that is similar to triangulation. This has been demonstrated for the 2013 Chelyabinsk event (Le Pichon et al. 2013; Brown et al. 2013). However, this technique has important limitations. First, a single array is inherently azimuth-only and without range information, it cannot uniquely locate the source along the beam direction.

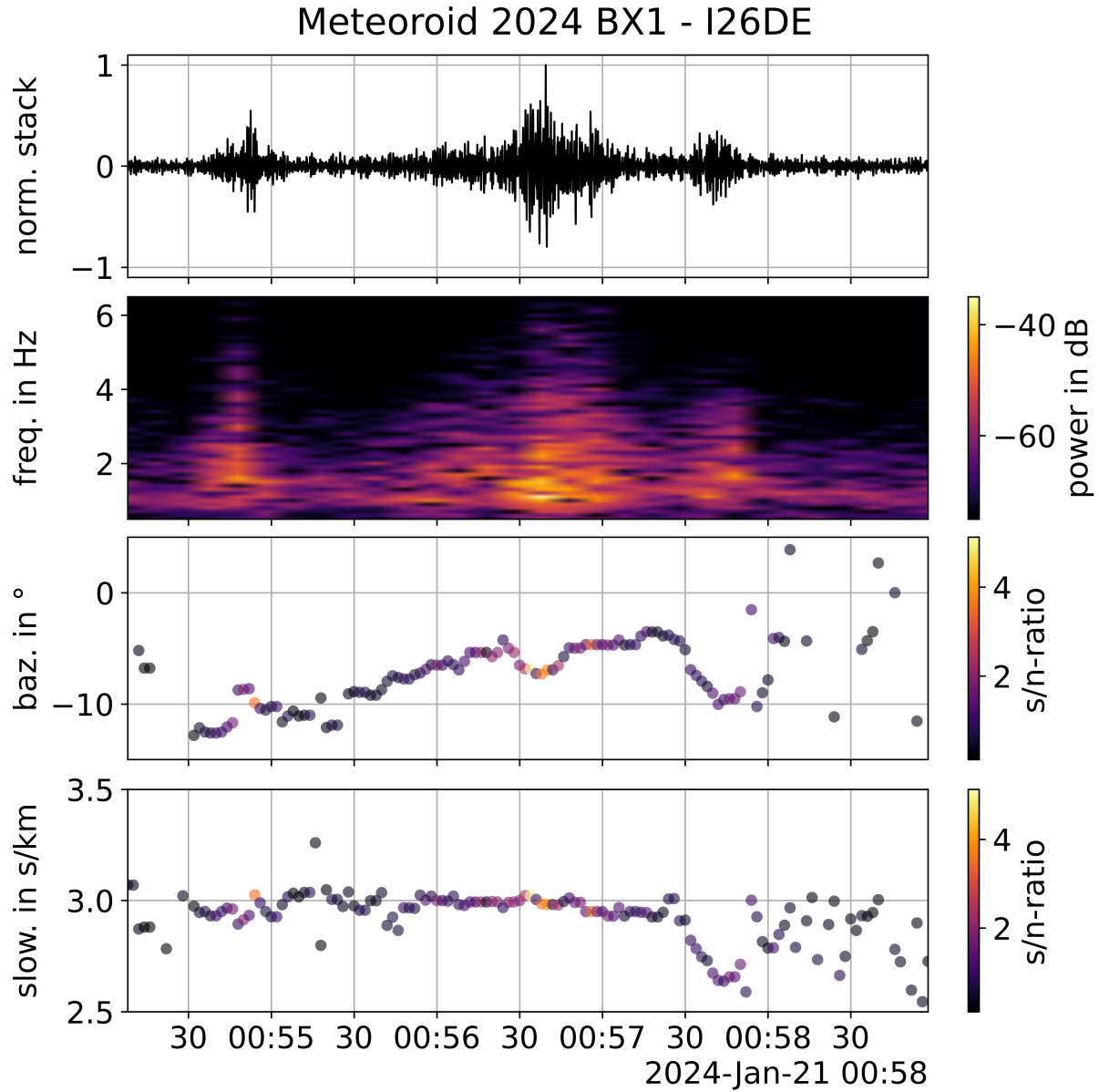


Figure 4.6.: Beamforming result of a meteoroid event close to Berlin, as recorded by the IMS array I26DE. The top panel displays the normalised stacked trace of all array sensor recordings. The second panel visualises the frequency content of the stack. The third panel indicates the back azimuth evolution, while the fourth panel shows the evolution of the signal’s slowness over time. The overall distance of over 400 km between the event and the infrasound array leads to a complex and multipathed set of meteoroid-related arrivals that are distributed over approximately 3 minutes.

Second and more critical are atmospheric heterogeneities (i.e., horizontal wind gradients and temperature structures), which leads to ray bending that can distort the observed back-azimuth by several degrees relative to the true source bearing (e.g., Drob et al. 2003; Silber 2025). Correcting for this requires accurate atmospheric specifications such as ECMWF operational analyses or empirical models like HWM (Drob et al. 2008) and NRLMSISE-00 (Picone et al. 2002). However, these models are themselves subject to uncertainty. The array aperture

establishes a minimum resolvable wavelength, and small or poorly designed arrays can suffer from spatial aliasing (Rost and Thomas 2002; Christie and Campus 2009). Finally, multi-path propagation, where stratospheric and thermospheric returns arriving at different times and apparent azimuths (Sutherland and Bass 2004), can complicate or corrupt the beamformed solution if not carefully identified and separated (see Fig. 4.6).

4.2.3. Markov chain Monte Carlo and Bayesian Inference

Bayesian inference provides a probabilistic framework for estimating parameters by combining prior knowledge with observed data. Given a forward model parametrised by a vector $\theta \in \mathbb{R}^N$, and a set of observations $\mathbf{d}$, Bayes' theorem states

$$p(\theta | \mathbf{d}) = \frac{p(\mathbf{d} | \theta) p(\theta)}{p(\mathbf{d})}, \quad (4.6)$$

where $p(\theta | \mathbf{d})$ is the posterior probability of the parameters given the data, $p(\mathbf{d} | \theta)$ is the likelihood, $p(\theta)$ is the prior, and $p(\mathbf{d})$ is the Bayesian evidence, acting as a normalisation constant (Sivia and Skilling, 2006; Gelman et al., 1995).

The prior encodes all knowledge about the parameters before observing the data. This can take the form of uniform priors over physically motivated bounds, Gaussian priors derived from previous measurements, or log-uniform priors for scale parameters. The posterior then represents the updated belief about the model θ after incorporating the data. This then fully characterises parameter uncertainties and degeneracies without reducing the overall solution to a single best-fit value (Trotta, 2008).

For problems with $N \gtrsim 4$ dimensions, direct numerical integration of equation (4.6) becomes impractical. Markov chain Monte Carlo (MCMC) methods overcome this issue by constructing a Markov chain whose stationary distribution is the posterior itself. This method draws samples $\{\theta_i\}_{i=1}^S$ that asymptotically represent $p(\theta | \mathbf{d})$ (Metropolis et al., 1953; Hastings, 1970). For problems in the $N = 8-10$ dimensions considered here, standard single-walker Metropolis-Hastings sampling can exhibit slow mixing and strong sensitivity to the proposal covariance. Ensemble sampling, introduced by Goodman and Weare (2010) and implemented in the widely used EMCEE package (Foreman-Mackey et al., 2013), addresses this issue by evolving an ensemble of $K \gg 2N$ parallel walkers through affine-invariant stretch and walk moves (Goodman and Weare, 2010). Because these moves are invariant under linear transformations of the parameter space, the sampler remains efficient even for highly correlated or anisotropic posteriors, eliminating the need for manual tuning of a proposal matrix. In practice, convergence is assessed via the integrated autocorrelation time τ_{int} , ensuring that the ensemble has adequately explored the posterior.

A key advantage of Bayesian MCMC over classical single-best-fit approaches (e.g. least-squares minimisation) is that it yields the full N -dimensional posterior. This can be used to obtain marginalised uncertainties, credible intervals, and parameter covariances directly (Hogg et al., 2010). Single-best-fit methods only provide a point estimate and a local curvature-based error. This error is unreliable for multimodal or non-Gaussian posteriors and it cannot capture degeneracies between parameters.

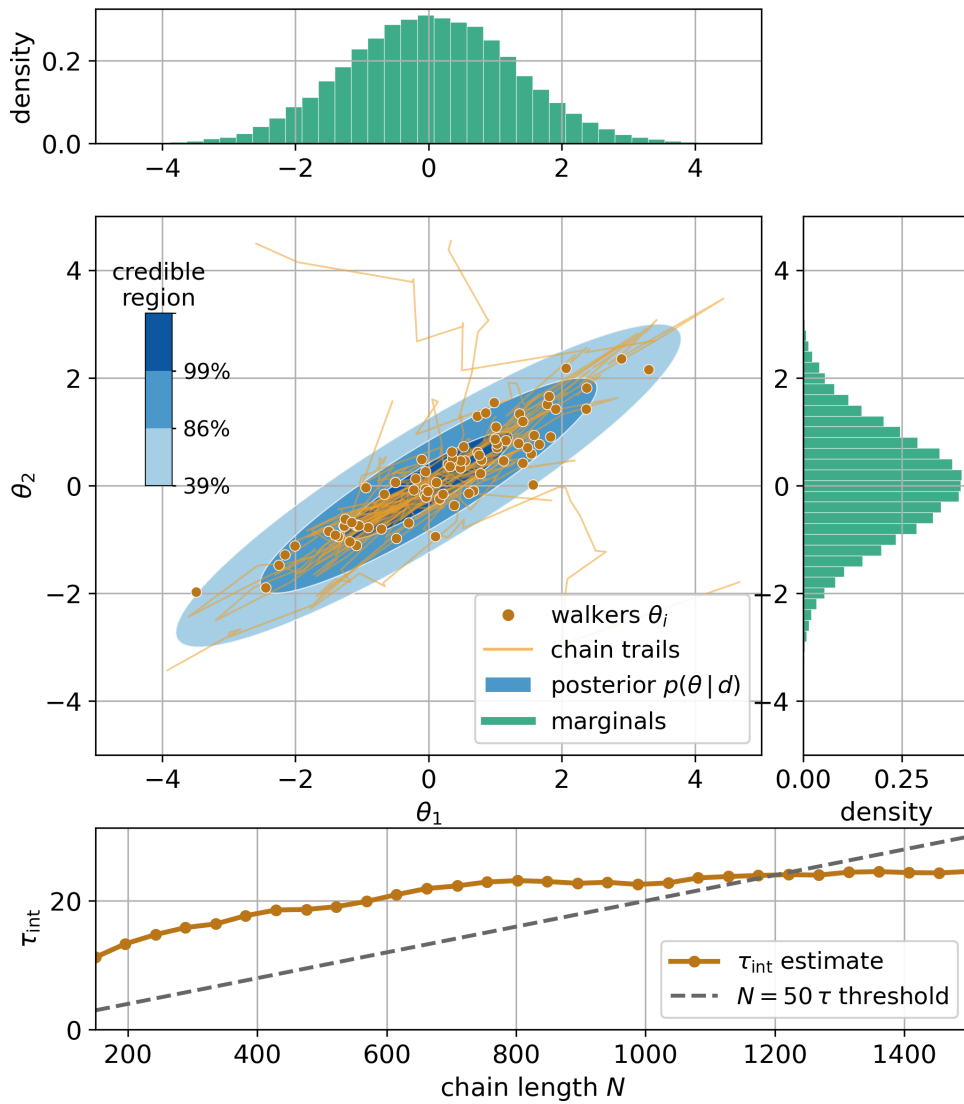


Figure 4.7.: Ensemble MCMC sampling of a correlated, anisotropic posterior distribution. The middle panel displays the target posterior $p(\theta | d)$ as filled credible regions (with the blue shading enclosing the 1-3 σ probability levels for a 2D Gaussian). An ensemble of affine-invariant walkers (dark orange dots) and their chain trails (orange lines) can be seen converging towards the high-probability region in the middle. The chain trails are shown for a subset of the walkers. The top and right panels show the 1D marginalised posteriors of θ_1 and θ_2 recovered from the post-burn-in samples in the form of histograms. The bottom panel indicates the integrated autocorrelation time τ_{int} versus chain length N . The estimate is considered reliable once it falls below the $N = 50 \tau$ threshold (black dashed line) and flattens.

In the context of meteoroid trajectory reconstruction, Bayesian MCMC ensemble sampling can be used to obtain statistically robust estimates of the modelling parameters. In our application, it is based on the ray-based trajectory reconstruction method using infraGA. However, as this is computationally expensive and single trajectory evaluations can take seconds to minutes on HPC infrastructure, ensemble sampling with $\sim 10^6$ proposal trajectories becomes unfeasible. For MCMC sampling to be a realistic alternative, the time taken for trajectory evaluation must be reduced. We achieve this by pre-computing the rays with G2S atmospheres at different

altitudes and by reducing ray tracing to efficient interpolation, which results in only a minimal loss of resolution (see Chapter 7 for more details). This decreases the required time to evaluate a single trajectory to approximately one second on a standard home computer. This process can then be parallelised on HPC infrastructure, which allows for the evaluation of $\sim 10^6$ trajectories in around 5 hours using 60 parallel threads.

4.2.4. Extending the inferred Meteoroid Trajectory

4.2.4.1. Strewn-Field Calculations

Accurate prediction of the strewn field, i.e. the spatial probability distribution of meteorite landing locations, is essential for recovering meteorites following a meteoroid event. Without this prediction, systematic ground searches are unfeasible and valuable scientific material may be lost (Moilanen et al. 2021; Sansom et al. 2015). Forward integration to produce the strewn field starts at the point where the calculated trajectory ends. This point is the last reliable observation point, which is located at or below the ablation termination height. Here, the meteoroid has decelerated to near-terminal velocity (Ceplecha et al., 1998; Borovička et al., 2013). From this initial state, the equations of motion are integrated forward in time through the lower atmosphere until each fragment reaches the ground. The dynamical model governing the dark-flight phase includes aerodynamic drag, gravitational acceleration, and horizontal wind advection. The deceleration of a fragment with an instantaneous mass m and a bulk density ρ_m is given by:

$$\begin{aligned}\frac{dv_x}{dt} &= -\rho_a \sqrt{(v_x - U_w)^2 + (v_y - V_w)^2 + v_z^2} (v_x - U_w) \Gamma_A m^{-1/3} \rho_m^{-2/3}, \\ \frac{dv_y}{dt} &= -\rho_a \sqrt{(v_x - U_w)^2 + (v_y - V_w)^2 + v_z^2} (v_y - V_w) \Gamma_A m^{-1/3} \rho_m^{-2/3}, \\ \frac{dv_z}{dt} &= -g - \rho_a \sqrt{(v_x - U_w)^2 + (v_y - V_w)^2 + v_z^2} v_z \Gamma_A m^{-1/3} \rho_m^{-2/3},\end{aligned}\tag{4.7}$$

where U_w is the zonal wind component and V_w is the meridional wind component, which are both interpolated from a radiosonde or reanalysis atmosphere profile. ρ_a is the ambient atmospheric density, g is the gravitational acceleration, and $\Gamma_A = \frac{1}{2} C_D A^*$ is the combined drag-shape parameter with dimensionless drag coefficient C_D and mass- and volume-scaled cross-sectional area $A^* = A V^{-2/3}$, with $V = m/\rho_m$. The drag coefficient is treated as Mach number dependent, following empirical fits for spherical and cubic bodies (Carter et al., 2011), while in practice a continuous shape-factor parameter interpolates between these two limiting geometries. During the early dark-flight phase, ablation is neglected, as it becomes negligible at the lower velocities (Towner et al., 2022). The coupled system of equations is integrated numerically using an explicit Runge-Kutta method with adaptive step control (Dormand and Prince, 1980).

As the initial conditions derived from the reconstructed trajectory are subject to finite uncertainty and as the atmospheric wind field is not perfectly known, a Monte Carlo ensemble approach is used to propagate these uncertainties into the predicted landing region (Towner

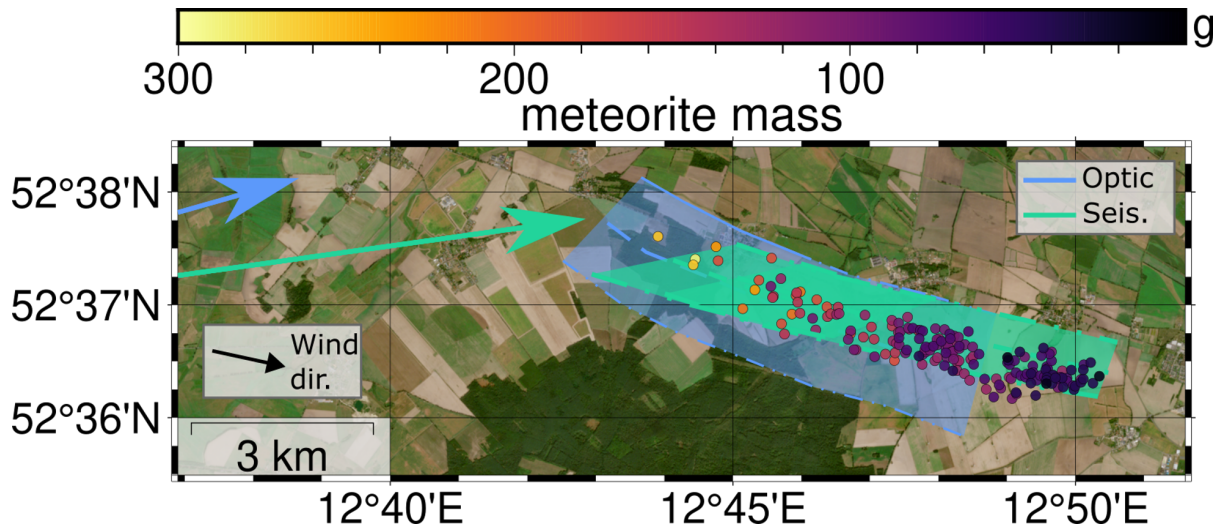


Figure 4.8.: Strewn field from the optical (light blue) and seismic (light green) trajectory solutions. The arrows indicate the endpoints of the trajectories, and the coloured areas represent the predicted strewn field. The coloured circles represent the recovered meteorite masses, ranging from heavy fragments (yellow) to light fragments (purple-black), which are taken from Rendtel et al. (2024). Note that the heavy fragments land closer to the line projection of the trajectories, while the lighter fragments are carried further downrange due to the SE heading wind.

et al., 2022). A large number of independent trajectory simulations are performed by perturbing all entry-state parameters, including velocity, elevation angle, azimuth, geographic coordinates, entry height, ablation coefficient and shape factor, within their respective uncertainty bounds. At the same time, independent wind perturbations are sampled from the atmospheric uncertainty envelope. The set of ground-impact points obtained across the ensemble defines the strewn field. For each mass class, the ensemble distribution is characterised by its mean landing position and a covariance ellipse whose semi-axes quantify the positional uncertainty in the along-wind and cross-wind directions (Drolshagen et al., 2017; Gritsevich, 2009; Towner et al., 2022). Heavier fragments, due to their higher mass are less sensitive to atmospheric drag and wind deflection and land closer to the linear projection of the entry trajectory, which can lead to smaller uncertainty ellipses. In contrast, lighter fragments are transported further downrange, producing the characteristic elongated strewn field observed for fragmented meteorite falls (Borovička et al., 2013; Sansom et al., 2015; Moilanen et al., 2021; Spurný et al., 2024). For an example in low wind conditions (see Fig. 7.3). However, in the case of strong cross wind conditions the distribution fragments must not follow this inertia dominated ordering (see Fig. 4.8).

4.2.4.2. Pre-Earth Orbit Determination

Recovering the pre-Earth heliocentric orbit of a meteoroid from ground-based observations requires converting a locally measured velocity vector into an orbit expressed in the solar reference frame. This is achieved through a sequence of physical corrections and coordinate transformations, which can be grouped into three stages:

- a) correcting the measured velocity for Earth's gravitational pull,
- b) transforming the corrected velocity through a chain of reference frames into a geocentric inertial frame, and
- c) combining it with Earth's motion to obtain the heliocentric orbit. Each stage is described in turn below.

a): The starting point is the observed geocentric speed v_{obs} and apparent radiant, measured at the meteoroid's highest constrained altitude. As the meteoroid approaches Earth, it is gravitationally accelerated, so both the observed speed and the apparent zenith distance of the radiant are inflated relative to their pre-atmospheric values. Removing this effect is known as the zenith attraction correction (Ceplecha, 1987; Ceplecha et al., 1998). The pre-atmospheric speed v_{∞} is recovered by subtracting the kinetic energy gained from Earth's gravitational potential during infall. Equivalently, the observed speed is the quadrature sum of the pre-atmospheric speed and the local escape velocity, so that

$$v_{\infty}^2 = v_{\text{obs}}^2 - v_{\text{esc}}^2(r_{\text{obs}}), \quad (4.8)$$

where $v_{\text{esc}}(r_{\text{obs}}) = \sqrt{2GM_{\oplus}/r_{\text{obs}}}$ is the escape velocity evaluated at the observation radius $r_{\text{obs}} = R_{\oplus} + h$ (Ceplecha, 1987), with $M_{\oplus}$ being the mass of the Earth. The same deceleration also shifts the radiant. Conserving angular momentum along the hyperbolic approach trajectory gives, to first order,

$$\sin z_{\infty} = \frac{v_{\text{obs}}}{v_{\infty}} \sin z_{\text{obs}}, \quad (4.9)$$

where z_{obs} and z_{∞} are the apparent and corrected zenith distances of the radiant. Since $v_{\text{obs}} > v_{\infty}$, the correction moves the true radiant away from the zenith (Ceplecha et al., 1998; Borovicka, 1990).

b): The corrected speed and radiant define a three-dimensional velocity vector, which now will be expressed in a geocentric inertial frame. To do this, the vector is first constructed in the local East-North-Up (ENU) frame at the observer's geodetic position. It is then rotated into the International Terrestrial Reference System (ITRS) using the standard ENU-to-ITRS rotation matrix, parametrised by geodetic latitude and longitude (Engel, 1994). Finally, the ITRS state is transformed into the Geocentric Celestial Reference System (GCRS), an Earth-centred inertial frame whose axes are aligned with the International Celestial Reference System (ICRS). This last step accounts for Earth's rotation, precession, nutation, and polar motion at the time of observation using standard IAU reduction procedures (Capitaine et al., 2003; Fukushima, 2003). The result is the geocentric position and velocity ($\mathbf{r}_{\text{geo}}, \mathbf{v}_{\text{geo}}$) in GCRS.

c): Now the heliocentric orbit can be calculated. For this, the heliocentric state of the meteoroid has to be found. It can be calculated by adding Earth's own heliocentric state to the geocentric vectors of the meteoroid,

$$\mathbf{r}_{\odot} = \mathbf{r}_{\oplus} + \mathbf{r}_{\text{geo}}, \quad \mathbf{v}_{\odot} = \mathbf{v}_{\oplus} + \mathbf{v}_{\text{geo}}, \quad (4.10)$$

where $(\mathbf{r}_{\oplus}, \mathbf{v}_{\oplus})$ is the heliocentric position and velocity of Earth, taken from a planetary ephemeris (Folkner et al., 2014). Treating the resulting state $(\mathbf{r}_{\odot}, \mathbf{v}_{\odot})$ as a classical two-body

problem with the solar gravitational parameter $\mu_{\odot} = GM_{\odot}$ gives the full set of Keplerian elements: semi-major axis a , eccentricity e , inclination i , longitude of the ascending node Ω , argument of perihelion ω , and true anomaly ν at the time of impact (Curtis, 2019; Bate et al., 2020). These elements characterise the pre-Earth heliocentric orbit of the parent body and allow its dynamical history to be traced through comparison with known asteroid families and meteor stream databases (Jopek et al., 2008; Granvik and Brown, 2018). This pre-Earth orbit determination framework is applied to the Koblenz meteoroid and can be The application of this pre-Earth orbit determination framework to the Koblenz meteoroid is given in Chapter 7.5.

4.2.5. Deep-Learning Phase Identification with PhaseNet and SeisBench

The automated identification and precise timing of seismic phase arrivals is a fundamental problem in observational seismology. In the past it has been addressed by short-term-average over long-term-average (STA/LTA) detectors or template matching methods (Allen, 1978; Gibbons and Ringdal, 2006). While these approaches perform well in scenarios with a high signal-to-noise ratio, they are less effective for noisy data, when phases overlap, or when the measured signals diverge too far from the reference template. Deep-learning methods based on encoder-decoder architectures can be used to overcome these issues. These methods can significantly improve picking accuracy and generalise across diverse seismic environments, even in conditions with a low signal-to-noise ratio (Ross et al., 2018; Zhu and Beroza, 2019; Mousavi et al., 2020). PhaseNet (Zhu and Beroza, 2019) is a U-Net-type encoder-decoder network (Ronneberger et al., 2015) that takes a fixed-length, multi-component seismic data as input and outputs the probability of detecting each target phase for every time sample of the input data. The encoder part of the network compresses the waveform step by step into a hierarchical representation. This starts from short impulsive transients for the shallow layers to longer coda envelopes for the successive deeper layers. The decoder then upsamples this representation back to the original resolution to produce a dense, per-sample estimate. Skip connections pass high-frequency onset detail directly from the encoder to the decoder, which preserves information that would otherwise be discarded by downsampling of the encoder (Ronneberger et al., 2015). Finally, a softmax step converts the network’s unnormalised per-class scores into per-sample class probabilities that are non-negative and sum to one (Bridle, 1990). Onset times are then identified at the peaks of these probability series, when they exceed a fixed threshold τ , which converts the continuous probabilities into binary detections (Fig. 4.9).

Training requires labelled waveform-onset pairs. Rather than marking each onset as a single hard point, the labels are Gaussian-smoothed pulses centred on the manual selection, which acts as a soft target reflecting manual phase identification uncertainties and making optimisation easier (Zhu and Beroza, 2019). The network learns by minimising a cross-entropy loss, which measures the difference between the probabilities predicted by the model and the true probabilities, and increases when the model assigns a low probability to the correct class. As noise samples can outnumber the desired phase arrivals in practice, the loss is class-weighted to give rarer phase arrival picks more influence, ensuring that the network

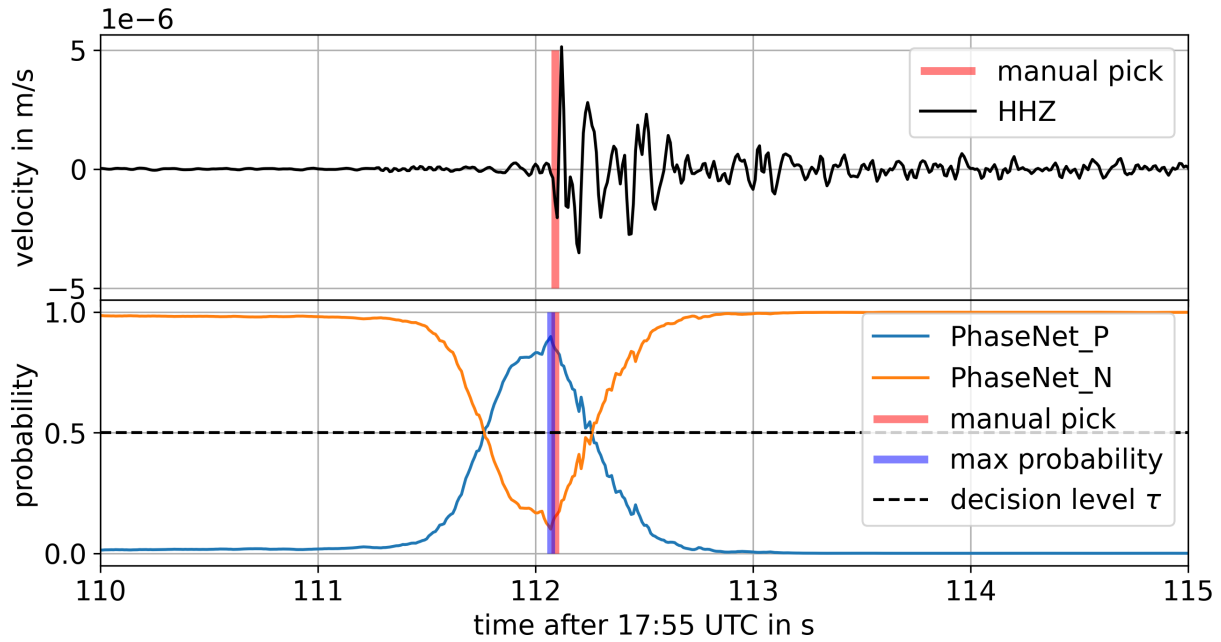


Figure 4.9.: Application of the trained PhaseNet picker (see Chapter 5) to 3-component seismic waveform data. In the upper panel the vertical component of the seismometer KB.DEP52 (black line), with a meteoroid-related signal at roughly 112 seconds after 17:55 UTC, can be inspected. The manually identified phase pick is marked with the red line. The lower panel indicates the PhaseNet P and N (P for the compressional wave and N for the noise) probabilities as the blue and orange lines, respectively. The maximum probability binary detection is indicated by the blue vertical line, which is 0.02 seconds before the manual pick. An exemplary decision level $\tau = 0.5$ is indicated by the black dashed line.

does not simply default to predicting noise everywhere (Buda et al., 2018). The optimisation of the network is usually performed using an algorithm that combines adaptive per-parameter learning rates with decoupled weight decay regularisation. This improves the convergence stability over stochastic gradient descent for high-dimensional parameter spaces (Loshchilov and Hutter, 2017). To stabilise the training even more, gradient-norm clipping is applied, that rescales the gradient vector, whenever its L_2 -norm exceeds a set threshold. This prevents unstable parameter updates that can be caused by very large gradients (Pascanu et al., 2013). Learning rate scheduling via cosine annealing (Loshchilov and Hutter, 2016) allows for better exploration during the initial stages of training, as well as for progressively finer adjustments as the optimisation process nears convergence. To improve the generalisation of the model the data can be augmented. This can include adding Gaussian noise, randomly scaling the amplitudes, and applying random horizontal-component rotations to expand the data to unseen source-receiver geometries. These augmentations are applied on-the-fly during training to artificially expand the effective dataset size and to discourage memorisation of station-specific waveform artefacts (Zhu and Beroza, 2019; Mousavi et al., 2020; Eickhoff et al., 2026b).

SeisBench (Woollam et al., 2022) provides a unified software infrastructure that covers all stages of the deep learning workflow. This includes dataset handling and converting datasets, as well as pre-processing pipelines, defining models, training loops, and evaluation utilities. It includes PhaseNet and a growing catalogue of additional architectures (e.g., EQTransformer Mousavi et al. 2020; GPD Ross et al. 2018). SeisBench allows for systematic benchmarking

and transfer learning across different waveform datasets by centralising these components within a common interface. Waveform datasets are packaged with their metadata, pick times and quality flags. They are then also partitioned into training, development and test subsets, mostly at the station or event level, to prevent leakage from near-duplicate segments (Sasse et al., 2025).

The utility of PhaseNet can be extended beyond earthquake seismology to any signal with a repeatable, characteristic onset and a sufficient signal-to-noise ratio. Meteoroid acoustic-seismic arrivals are therefore a suitable candidate, because the coupling between the acoustic and seismic waves produced by the supersonic meteoroid results in a distinct W-shaped wavelet on the vertical seismic component. This wavelet exhibits dominant energy between approximately 0.5 Hz and 40 Hz, with an impulsive negative-polarity onset (Edwards et al., 2008; Langston, 2004; Cepolecha et al., 1998). Since the morphology remains mostly consistent across events and stations, the network can learn a generalisable template rather than memorising individual signatures. Training PhaseNet within SeisBench using a manually labelled database of meteoroid arrivals therefore applies the full methodology of modern deep-learning seismology to the automated detection of meteoroids. This in turn, yields onset times with uncertainties in timing that can be used for subsequent acoustic-seismic trajectory inversion (see Chapter 5).

5. Seismo-Acoustic Meteoroid Observation Recording Database (SMORD): A Global Dataset and Deep-Learning Phase Picker for Meteoroid-Generated Air-to-Ground Coupled Seismic Waves

The following section describes the Seismo-acoustic Meteoroid Observation Recording Database (SMORD). It consists of 307 meteoroid events and 9 re-entry events (316 total) and 3,295 labeled acoustic-to-seismic coupled arrivals across a global set of seismic stations. The labeled arrivals are used for training a PhaseNet picker in SeisBench. The performance of the trained picker is evaluated and autonomous meteoroid seismic arrival picking is performed on a recent meteoroid event in Italy. The following section is currently in press as (Eickhoff et al., in press).

My contributions to this study, following CRediT, are:

1. Conceptualisation,
2. Data Curation,
3. Formal Analysis,
4. Investigation,
5. Methodology,
6. Software,
7. Validation,
8. Visualisation,
9. Writing (original draft, review & editing)

5.1. Abstract

Meteoroids impacting Earth’s atmosphere generate acoustic waves that can couple into the ground and can be recorded by dense, globally distributed seismic networks. Thus, these records complement optical and radar observations, especially since seismic stations also operate in cloudy weather conditions and during daytime. However, open datasets that link meteoroid events to labeled seismic waveforms are scarce, limiting the development of automated detectors for meteoroid-induced seismo-acoustic signals. We introduce the **Seismo-acoustic Meteoroid Observation Recording Database (SMORD)**, compiled by cross-referencing public meteoroid catalogs (International Meteor Organization fireball reports; NASA CNEOS fireball catalog) with seismic archives. Continuous waveforms are manually labeled for the first clear meteoroid-related onset of air-to-ground coupled seismic waves using a three-level pick-quality scheme. SMORD v1.0 contains 307 meteoroid events and 3,295 labeled arrivals across a global station set. Using SMORD labels, we train a PhaseNet picker in SeisBench with station-level splits and augmentation. On test data, the model achieves 91% precision and 94% recall at a 0.5 decision threshold (area-under-curve value 0.89), with median absolute timing error of 0.02 s (90% within c. ± 0.3 s). We demonstrate automated onset detection and trajectory reconstruction for an April 2025 Adriatic fireball, highlighting the values of SMORD for rapid post-event analysis.

5.2. Non-technical summary

When a bright meteor, or fireball, enters the Earth’s atmosphere, it generates a powerful acoustic wave in the air. This wave can shake the ground slightly, and sensitive earthquake-monitoring stations can detect it. Since these stations operate day and night and are unaffected by clouds, they can detect fireballs even when camera systems or radars miss them.

However, the lack of a well-organized, shared collection of these fireball signals in seismic data is a major challenge. To address this, we created a **Seismo-acoustic Meteoroid Observation Recording Database (SMORD)**. SMORD is an open database that links reported fireball events from public sources with the corresponding seismic recordings from nearby earthquake stations. We gathered relevant seismic recordings and marked the first onset time of each fireball signal, which can be used to calculate their trajectories.

SMORD currently includes 307 fireball events and 3,295 marked signal arrivals from seismic stations worldwide. Using these examples, we train a machine-learning tool that can identify fireball-related signals in new seismic data with high reliability and precise timing. We demonstrate how this tool can quickly detect a fireball and estimate its travel path through the atmosphere, supporting rapid follow-up studies after an event.

5.3. Introduction

Earth is constantly subjected to an influx of interplanetary matter and most of this mass arrives as interplanetary dust (Drolshagen et al., 2017). As the size and mass of this matter increases, penetration depth into Earth's atmosphere increases as well (Gritsevich, 2009, e.g.). Surpassing $30\text{ }\mu\text{m}$ these objects are classified as meteoroids (Koschny and Borovicka, 2017). Meteoroids impacting Earth are studied by different techniques: optical observations by ground based camera networks (e.g., FRIPON see Colas et al. 2020), space based optical detections (e.g., the GOES Geostationary Lightning Mapper (GLM) see Jenniskens et al. 2018), reports from human observers (e.g., IMO Fireball Program see Hankey and Perlerin 2014; International Meteor Organization (IMO) 2025b), radio forward scatter and radar systems (e.g., CMOR see Webster et al. 2004; Jones et al. 2005; BRAMS see Lamy et al. 2022), infrasound measurements (e.g., Ens et al. 2012; Ott et al. 2019; Pilger et al. 2020a; Silber 2024), and seismo-acoustic methods that track ground coupled airwaves at seismic stations (e.g., ReVelle 1976; Edwards et al. 2008; Neidhart et al. 2021).

Optical and radio techniques rely on ionization produced along the meteoroid trail, whereas seismo-acoustic methods are based on the shock wave induced by a meteoroid while it travels at supersonic speed. Atmospheric entry speeds of meteoroids range from about 11 km s^{-1} to 72 km s^{-1} (Ceplecha et al. 1998; Moorhead et al. 2019) considerably exceeding the velocity of sound in the atmosphere (c. $290 - 350\text{ m s}^{-1}$). After coupling into the solid Earth, seismic waves are excited with much higher propagation velocities compared to the velocity of sound, while still being significantly slower than the meteoroids themselves. In Fig. 5.1, three events with recording examples of meteoroid-induced signals are presented.

A key advantage of seismo-acoustic observations is the abundance and global distribution of seismic sensors, which provide broad and dense spatial coverage compared with the much smaller number of dedicated meteor cameras. These cameras, in turn, greatly outnumber meteor radar systems (e.g., Andrioli et al. 2013; Vida et al. 2021; Ringle et al. 2022). Additionally, seismo-acoustic stations are independent of cloud coverage and light conditions, which especially inhibit optical detections. Previous studies have demonstrated that meteoroids produce infrasound that can be readily detected (ReVelle 1976; Silber et al. 2015) and that a portion of this acoustic energy couples into the ground and can be recorded with seismic instruments (Edwards et al. 2007; Edwards et al. 2008; Novoselov et al. 2020). Observational studies have shown that infrasound sensors can reliably detect acoustic waves from meteoroids with masses ranging from grams to tens of kilograms (Silber and Brown, 2014). Seismic stations, however, record ground motions that have undergone acoustic-to-seismic or air-to-ground coupling in the case of meteoroid signals. In this study, we focus on the so-called direct coupled airwave (e.g., Edwards et al. 2008) that produces characteristic W-shaped wavelets on seismic recordings, because they are easy to identify and exhibit the best signal-to-noise ratios. They are called direct coupled airwaves because the coupling into the ground happens close to the seismometer. Estimates of acoustic to seismic coupling efficiency commonly lie at the $\approx 1\text{--}3\%$ level (Novoselov et al., 2020); this variability is closely linked to near surface site conditions such as the stiffness of the local geological material and other surface conditions, e.g., topography (Edwards et al. 2008; Novoselov et al. 2020). Another factor is the incidence angle of the meteoroid: steeply incident meteoroids radiate acoustic waves nearly parallel

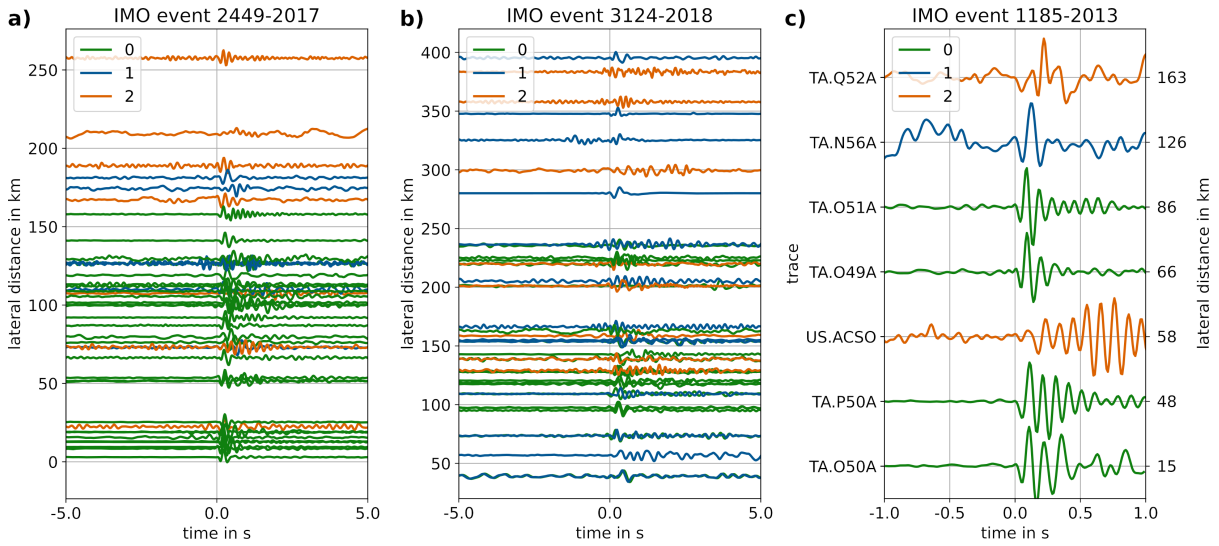


Figure 5.1.: Vertical component waveform examples of meteoroid signals. Lateral distance is relative to the center of the trajectory, and time zero corresponds to the onset time of each signal arrival which is manually picked. The waveforms are colored with respect to the pick quality (0/green: high pick quality, 1/blue: moderate pick quality, 2/orange: low pick quality, see Table 1). Here seismo-acoustic onsets can be identified at lateral distances of a) up to 260 km for an event above Washington State and b) up to 390 km for an event above Italy. c) Detailed view of waveform examples for the different pick qualities for the IMO event 1185-2013 over Ohio State. The Y-axis annotations represent the network-station-code combination (FDSN) for each seismogram. Time (along the X-axis) is relative to picked onset time. The waveforms are bandpass filtered from 1–10 Hz and the traces are normalized.

to the Earth’s surface and, therefore, are harder to detect (Neidhart et al., 2021). Neidhart et al. (2021) identified meteoroid related signals on seismic sensors for meteoroid masses ranging from 1 g to 10^5 g. Their study determines a lower limit of meteoroid masses that can be recorded with seismometers. It shows that the lower mass end of recordable meteoroids is well below 10^3 g, which is the order of magnitude at which larger meteoroids begin to pose potential hazards at Earth’s surface, highlighting the importance of rapid and reliable detection. Thus, seismological studies can contribute to quantify the meteoroid flux to assess risks associated with these extraterrestrial bodies. In addition, a rapid calculation of the trajectory, which is the path the meteoroid takes through the atmosphere, can help to recover the extraterrestrial material for further studies. Meteoroid trajectories have been reconstructed by seismo-acoustic signals before (e.g., Arrowsmith et al. 2007; Vera Rodriguez et al. 2023) and there exist geometrical reconstruction methods (e.g., Pujol et al. 2005), as well as ray-based approaches that can include the effects of atmospheric winds (e.g., Blom et al. 2024).

For these reasons, we focus on meteoroid events from publicly accessible catalogs which contain significant events: the International Meteor Organization (IMO) fireball event lists derived from crowdsourced visual human observer reports (International Meteor Organization (IMO), 2025b) and the NASA Center for Near-Earth Object Studies (CNEOS) fireball catalog compiled from U.S. Government sensors (NASA/JPL Center for Near-Earth Object Studies (CNEOS), 2025).

Open datasets linking meteoroid events to seismic waveforms and arrival onsets remain scarce, and automated detectors trained specifically on meteoroid-generated signals do not exist

yet or are unknown to the authors. Therefore, we introduce the Seismo-acoustic Meteoroid Observation Recording Database (SMORD), compiled from cross referencing public meteor event catalogs with regional seismic observations and by manually labeling meteoroid related phase arrivals. We describe the database, data sources, and labeling protocol. Furthermore, we develop and evaluate a phase-arrival picker based on PhaseNet via SeisBench (Zhu and Beroza 2019; Woollam et al. 2022) trained on SMORD labels. We report accuracy, precision, and recall on test data, which was not part of the training. Finally, we demonstrate automatic phase detection for meteoroid acoustic to seismic signals and trajectory reconstruction using SMORD, as well as its current limitations.

5.4. Data

5.4.1. Meteoroid event catalogs

IMO Fireball Program: The event catalog of the IMO fireball program is a community driven, publicly browsable database of fireball events reported by human observers (Hankey and Perlerin, 2014). For our selection we restrict IMO events to those with ≥ 100 independent reports to prioritize well constrained trajectories. We analyze all IMO events from 2010–2023 and selected events from 2024. Because the database relies on human witnesses, the events cluster near population centers and during times of day when many people are awake. For each IMO event, we take the reported trajectory beginning and end points and calculate the midpoint along the trajectory. Then we search for seismic stations within a great-circle distance of 3° from this midpoint. The catalog also includes nine reentry events of space debris, which produce seismo-acoustic signals analogous to meteoroids (e.g., Fernando and Charalambous 2026; Fernando et al. 2024b); these are included in SMORD but flagged for optional filtering. **CNEOS fireball catalog:** We use all CNEOS events from 1988–2024 with reported estimated yields ≥ 0.3 kt TNT equivalent to prioritize energetic airbursts expected to radiate strong acoustic waves. The CNEOS events are based on U.S. Government space-based sensors and provide a near-global coverage. As with IMO, we take the reported event location, but we increase the search radius for seismic stations to 5° to reflect (1) weaker ties of CNEOS detections to population centers and dense seismic networks and (2) generally higher signal-to-noise (SNR) ratios for higher-yield events.

5.4.2. Waveform retrieval and preprocessing

Continuous waveforms are downloaded using public FDSN services (e.g., the EIDA Federator see Strollo et al. 2021; the IRIS DMC see Trabant et al. 2012; the NCEDC see Northern California Earthquake Data Center 2014; the SCEDC see Southern California Earthquake Center 2013; Raspberry Shake see Raspberry Shake, S.A. 2016), with ObsPy (Beyreuther et al., 2010). Network, station, location, and channel codes are stored in SMORD. Preprocessing for picking includes instrument response removal (to ground velocity), detrending, and bandpass

filtering (varied in the frequency band 0.5–40 Hz) to emphasize the onsets of the acoustic-to-seismic signals. Time windows are aligned relative to catalog event times with padding to accommodate atmospheric propagation delays and catalog uncertainties in location and origin time.

5.5. Methods

5.5.1. Manual phase picking and labeling protocol

We visually inspect time windows around expected wave arrivals at each recording station and manually pick the first clear onset attributed to the meteoroid related wave. The time windows are selected around the arrival time of an acoustic wave traveling with 330 m/s with the time window width increasing with increasing origin-station distance. The time windows are additionally expanded in both directions by 120 s to account for the uncertainty of the vertical location of the meteoroid source. Selection criteria include an impulsive, negative polarity onset on the vertical component, with a dominant frequency between 1 - 20 Hz, consistent with ground coupled airwaves (i.e., W-shaped wavelets; see Edwards et al. 2008). For further verification, we check for a plausible moveout of an acoustic wave, cross component coherence and exclusion of disturbing local anthropogenic or tectonic phase arrivals. Each pick receives a quality flag of 0 to 2 (best to worst) that is chosen depending on onset clarity and signal-to-noise ratio (SNR) (see Figure 5.1c; Table 5.1). In the case of onsets with characteristics of two different qualities, e.g., a clear onset with only moderate amplitude, the resulting pick quality will be the lower one of the two quality identifiers; in this exemplary case a pick quality of 1 would be attributed. Metadata are stored per pick including event and station identifiers as well as locations. Picks have an associated pick time in UTC, network/station/location code, channel, and quality flag (see Table 5.2).

Pick quality	High (0)		Moderate (1)		Low (2)
Onset clarity (I)	clear wavelet	W-shaped	slightly distorted or wavelet with multiple peaks/troughs		wavelet with multiple peaks/troughs or no clear onset; emergent waveforms
SNR (II)	W amplitude is elevated significantly above the noise		W amplitude is elevated moderately above the noise		W amplitude is close to the noise

Table 5.1.: Guidelines for decision-making for pick quality.

Metadata type	Value
Event id	IMO 1185-2013
Event origin time	2013-05-31T03:03:00.000
Event location	40.014 °N; 83.598 °W
Station network code	TA
Station code	O94A
Trace channels	BH*
Trace sampling rate	40 Hz
Station location	40.188 °N; 84.335 °W; 292 m
Phase pick quality	0
Trace start time	2013-05-31T03:06:46.000
P phase arrival sample	4006
Split	Train

Table 5.2.: Metadata of a phase pick stored in SMORD with exemplary values. The P phase arrival sample marks the waveform sample at which the meteoroid related onset is picked. The split, either train, dev or test is the separation of the dataset into the three groups used for the PhaseNet training (see Section 3.2).

5.5.2. Model training with SeisBench PhaseNet

To automatically find seismic signals of meteoroids we trained a probabilistic seismic phase picker using PhaseNet (Zhu and Beroza, 2019), as implemented in the SeisBench framework (Woollam et al., 2022). SeisBench provides standardized access to waveform datasets, pre-processing routines, and benchmark models for seismological machine learning (Woollam et al., 2022). PhaseNet is a deep convolutional neural network that maps multi-component waveform windows to phase-probability time series (Zhu and Beroza, 2019). This routine makes it well-suited for robust P- and S-phase picking in noisy environments. In our setup, we train the network on positive windows centered on analyst labeled arrivals, off-center positive windows as well as negative windows. Training proceeds over multiple epochs, during which the model updates its parameters iteratively to minimize a weighted cross-entropy loss that encourages sharp phase probability estimates. This strategy leverages the strengths of neural networks to learn complex nonlinear relationships in seismic data while controlling overfitting through data augmentation and repeated exposure to diverse positive examples.

SMORD waveforms and labels are packaged as a SeisBench WaveformDataset (Woollam et al. 2022; Münchmeyer and Woollam 2025). For training, the raw, unprocessed time series are resampled to a common 100 Hz sampling rate, which is consistent with the standard configuration of PhaseNet (Zhu and Beroza, 2019). The dataset is partitioned into disjoint training, development, and test subsets with an 80/10/10 split at the station level. This ensures that all waveform windows derived from a given station across all meteoroid events appear in only

one subset. This grouping prevents train-to-test contamination, i.e., data leakage, in which nearly identical waveform segments would be present in both the training and evaluation data. Contamination would lead to overly optimistic validation performance and poor generalization (Sasse et al., 2025), which is mitigated by our strategy.

Since meteoroid events are usually observed at multiple stations, the station-level split allows each subset to sample a variety of events while keeping station-specific effects separate. This better reflects the intended application to unseen stations and follows common machine learning practice for grouped or correlated data, where samples sharing a common identifier are kept within the same split to obtain an unbiased estimate of model performance (Sasse et al., 2025). To train the network, we use three different complementary generators that let the network learn the pattern of wave onsets, while being robust to different source-receiver orientations and noise:

1. positive windows centered on labeled meteoroid onsets,
2. augmented positive windows which are identical to 1 but with on-the-fly data augmentation; random rotations in the horizontal plane and additive Gaussian noise, and
3. noise-only windows drawn uniformly from training records and explicitly labeled as “N” (noise) to control false positives.

The first generator allows the network to learn the typical waveform patterns associated with arrivals of meteoroid signals, e.g., onset shape, polarity and relative timing between components. Using 1) as a distinct generator ensures that enough genuine onsets are contained in every batch, so that the model can learn a sharp phase likelihood for true arrivals. The second generator uses the same positive samples but with random horizontal component rotations, that simulate different source-receiver geometries. The additive Gaussian noise used in this generator teaches the network to keep detecting onsets even as the SNR diminishes. Because these augmentations are re-applied during each epoch, the model effectively sees a much larger variety of examples, which reduces overfitting and improves generalization. The third generator teaches the network to identify noise, which reduces false positives. The three generators are concatenated to form the training stream used by PyTorch (Paszke et al., 2019) and increase the amount of training data for each epoch. In this way the data is increased from 2,637 signals for only generator 1 to over 7,911 signals with all three generators. Time windows are 6,000 samples long (60 s at 100 Hz), with the labeled onset randomly positioned within the window followed by a zero padding to maintain fixed length across all windows. We train a PhaseNet architecture from SeisBench configured for three per-sample classes P/S/N (with S unused here) from scratch, because meteoroid air-to-ground coupled phases differ fundamentally from earthquake P and S phases and no suitable pre trained models exist for this signal class. Inputs are three component seismograms (Z, N, E) if available, otherwise vertical (Z) recordings only. To encourage orientation invariance, we apply random rotations in the horizontal plane, and to simulate a variable noise floor we add Gaussian noise which scales up to $\approx 10\%$ of the normalized amplitude for each epoch for generator 2. All inputs are demeaned and peak normalized within the window.

Labels are produced as probabilistic time series with a Gaussian smoothing across 20 samples (0.2 s at 100 Hz). For positive examples we use the single manually picked meteoroid signal onset and map it to the P-wave channel (the S-wave channel remains unused, as there are no

labelled S-phase picks associated with acoustic meteoroid-induced seismic waves; no S-phases are contained in SMORD). Noise only windows are labeled $N = 1$ everywhere. The network thus learns a per sample posterior over P/S/N tailored to meteoroid air-to-ground coupled seismic wave onsets.

We use the AdamW optimizer from PyTorch (Paszke et al., 2019) with a learning rate of 10^{-3} and a weight decay of 10^{-2} . These are unitless hyperparameters that control update magnitude and regularization strength, respectively. The AdamW optimizer combines adaptive step sizes, or per-parameter learning rates based on running estimates of gradient mean and variance, with decoupled weight decay. This improves the behavior of weight decay for adaptive optimizers compared to classic L2-penalty formulations (Loshchilov and Hutter, 2017). We use cosine annealing to schedule the learning rate across 500 epochs, yielding large updates early on and progressively smaller updates later on for fine-tuning (Loshchilov and Hutter, 2016). To stabilize optimization, we apply global gradient-norm clipping with an L_2 -norm threshold of 1.0. This means that if the overall L_2 -norm of the gradients across all parameters exceeds 1.0, the gradients are rescaled to prevent excessively large parameter updates that could destabilize the training (Pascanu et al., 2013). For each time sample, the model outputs softmax probabilities for the P/S/N classes (e.g., Bridle 1990). We train with class-weighted cross-entropy (weights of 1.0, 0.0, and 0.4 for P, S, and N, respectively) and then average the loss over all time samples and across the batch. Class weighting is a standard approach for imbalanced classes (i.e., the P and N classes in this case) and per-sample supervision (Buda et al., 2018). We monitor the loss for each epoch on the development split of the dataset, and we save the model with the best loss.

5.6. Results

SMORD currently contains 307 distinct meteoroid and 9 re-entry events as well as 3,295 labeled seismo-acoustic picks across a global station set of 1,772 seismological recording stations (SMORD v1.0). Of the 316 events, 262 are taken from the IMO Fireball event list ($\approx 83\%$), accounting for 3,083 associated picks ($\approx 94\%$). CNEOS events represent $\approx 17\%$ of the database and they contribute to $\approx 6\%$ of labeled picks. The contribution of picks from the CNEOS database is lower, because events can be identified over the open ocean far away from seismometers. Here, we only include the events where we identified at least one meteoroid-related phase. Events cluster around densely populated regions of central Europe and the eastern United States (Fig. 5.2), a consequence of the eyewitness driven IMO contribution. We partly mitigated this bias by including the CNEOS list, which is less location and time biased. Thereby events are added from the southern hemisphere as well.

Of the 307 meteoroid events, 216 occurred in North America, 60 in Europe, and 31 elsewhere. Only 18 events were reported from the southern hemisphere. Figure 3 summarizes key features of SMORD metadata. Before 2010 only CNEOS events are included in SMORD, because no IMO events reached the threshold of ≥ 100 observers. The CNEOS list provides on average one event per year to SMORD. After 2010, as the IMO program grew, the number of SMORD events per year rose markedly. Figures 3b-c show the event time distribution over UTC and

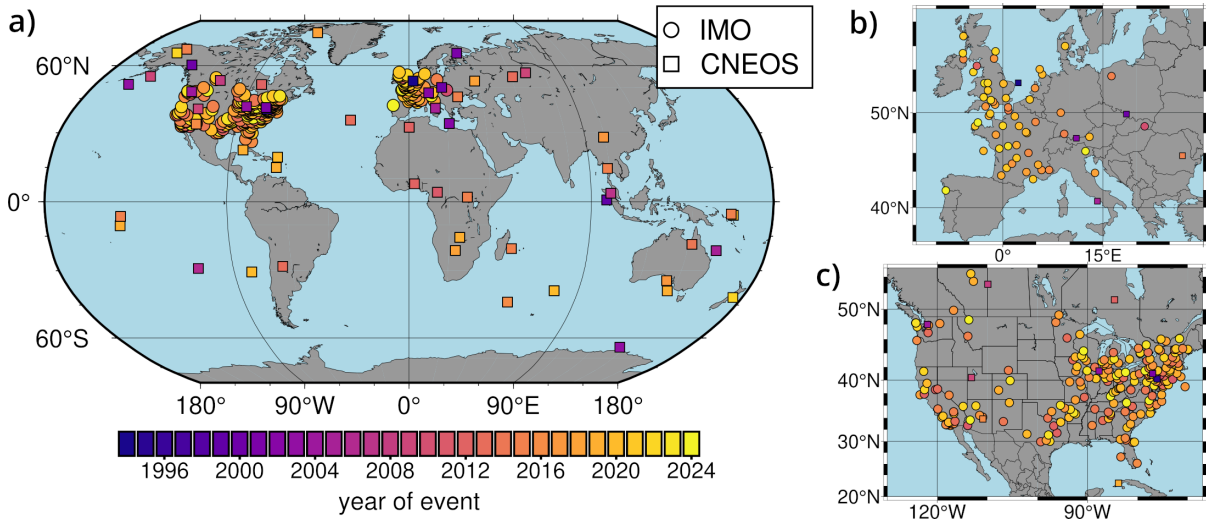


Figure 5.2.: a) World map of SMORD events. Colored circles mark IMO (2025) events and colored squares mark CNEOS (2025) events. Events are colored with respect to the year of their respective occurrence. The same coloring and marking are used in b) and c). b) Zoomed-in map with Central European events. c) Zoomed-in map with North American events.

local time. The high number of events at 0 - 4 h UTC time coincides with evening to night times in North America and Western Europe, respectively, which indicates the diurnal bias of the dataset. This bias becomes much clearer when the event times are given in local times for the event locations (see Fig 3c). There, the paucity of 0 - 6 h local time events (Fig. 3c) reflects the eyewitness nature of IMO events, as at this time most people are asleep. Observational studies indicate that meteoroid flux at night is approximately uniform with enhancements near dusk and dawn (e.g., Brown et al. 2007; Silber et al. 2015). This means that SMORD is currently heavily biased towards human wake-sleep cycles as near-dusk events are mostly well represented, but near-dawn events are underrepresented.

Figure 5.3d shows the temporal distribution of SMORD meteoroid events by month and year. About two-thirds of all detections occur between October and February. This pattern aligns with the longer winter nights in the Northern Hemisphere, which increase the likelihood of observing meteoroid events. In contrast, August shows an unusually low number of detections even though the Perseid meteor shower peaks in that month. We interpret this as an indication that most SMORD events do not arise from dedicated meteor-observing campaigns scheduled around major showers. Instead, events are observed opportunistically when observers are awake and conditions are favorable. In such a scenario, the shorter nights and increased daylight in August naturally lead to fewer sightings, resulting in the comparatively weak representation of Perseid activity in our sample. However, due to the uncertainties in the IMO and CNEOS trajectories, the meteoroids in the SMORD database are not associated with meteor showers. The distribution of the number of individual phase picks against the lateral distance from the center of the meteoroid trajectory is given in Fig. 5.3e. Because event heights are not uniformly reported in the IMO and CNEOS catalogs, we opt to consider only the lateral distance, meaning we ignore height information in the distance calculation. Instead, we calculate distance using the horizontal offset to the trajectory midpoint. The number of

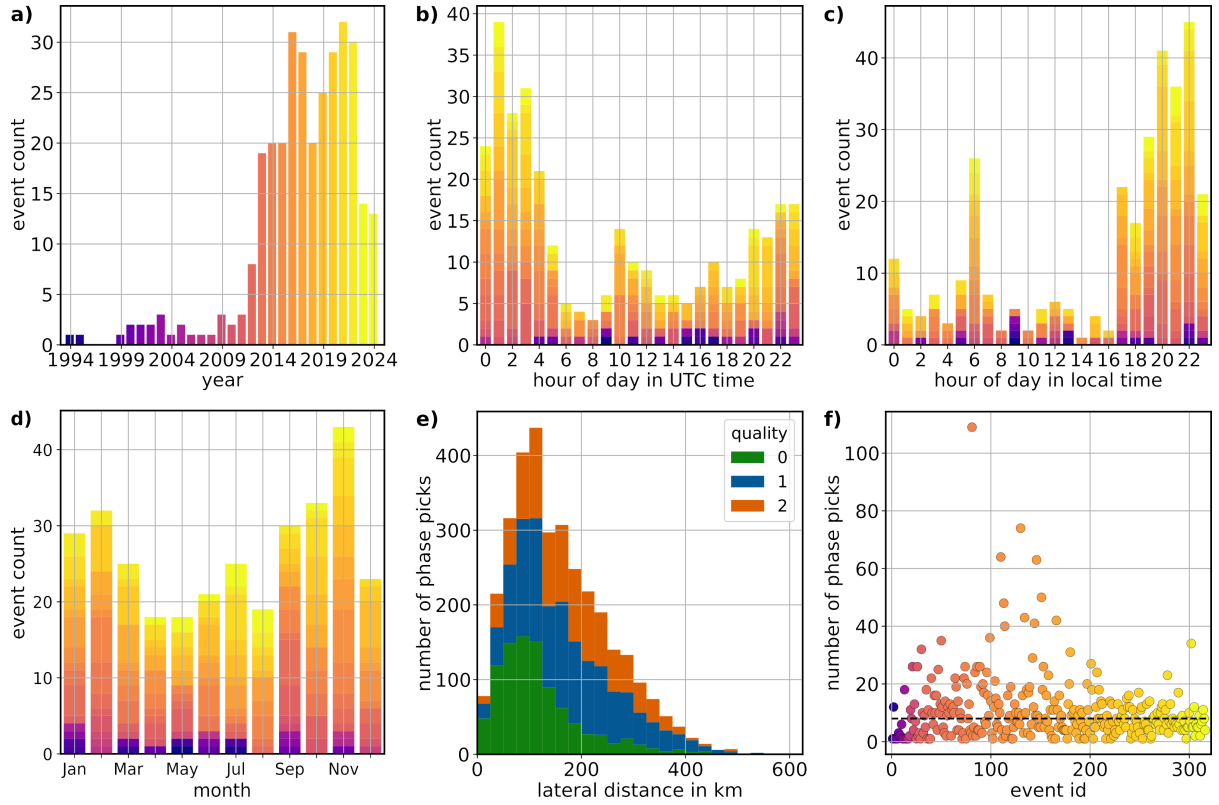


Figure 5.3.: Statistics of SMORD metadata. a) Count of meteoroid events (at least one picked seismic arrival exists) versus year of the event. Purple to yellow coloring is the same as in Figure 5.2 and Figs. 5.3b, 5.3c, 5.3f. b) Count of meteoroid events versus UTC event time. c) Count of meteoroid events versus local event time. d) Count of meteoroids versus month of the event. e) Count of individual phase picks versus their lateral distance from the trajectory midpoint. The coloring indicates the quality of the individual phase pick from 0 (best/good) to 2 (worst/poor). f) Number of overall phase picks versus individual event id. The dashed black line represents the median phase pick number of 7.5 picks per event.

phase picks decreases with increasing station range with a peak at a lateral station-event distance distinct for each quality: this peak is around 80 km distance for the good (0) quality picks. The medium quality (1) peaks at a lateral distance of c. 140 km, while the phase picks of poor (2) quality peak at a lateral distance of c. 160 km. The offset of the peak in number of phase picks relative to 0 km lateral distance at the center of the trajectory may result from limitations in the distance definition and altitude-dependent source strength (see Figure 5.3e). This projection may shift the apparent location picks relative to the midpoint. Additionally, acoustic emissions produced during the latter stages of flight at lower altitudes are expected to yield higher signal-to-noise ratios at the seismic stations, which increases the likelihood of confident detections (e.g., Silber 2025). However, signals from meteoroid fragmentation cannot easily be discriminated from signals due to Mach cone waves, based only on individual waveforms. Instead a source determination is required to discriminate between the underlying point or line source. The observed clustering of picks at $\approx 80 - 160$ km distance is consistent with a combination of geometric bias and enhanced detectability of signals from the lower portion of the trajectory. The decrease of phase picks with distance aligns with the expected decrease in SNR over greater distances and thus also increasing travel times (Fig. 5.3e). Across

all events, the median number of picked arrival phases per event is 7.5 (Fig. 5.3f). The values in Fig. 3f are stable over time, with outliers for large bolides over dense seismic networks which, of course, produce more signals with high amplitudes, e.g., event id 81 (IMO 2015 657; 109 picks, event above Switzerland) and id 130 (IMO 2017 1298; 74 picks, event above California). We evaluate the performance of the PhaseNet-based picker (hereafter PN picker) using standard binary-classification metrics derived from the entries of the confusion matrix, following e.g., Sokolova and Lapalme (2009), with true positives (TP), false positives (FP), true negatives (TN) and false negatives (FN), which are computed by comparing predicted onset detections to the manually picked reference labels. Because the PN picker outputs a probability for the presence of a seismo-acoustic onset, a fixed probability decision threshold τ is used to convert probabilities into binary predictions (i.e., onset vs no onset). We measure overall performance of the PN picker at this fixed τ with the metrics given in Table 5.3.

Metric	Formula	Measure
precision	$\frac{TP}{TP+FP}$	fraction of correctly predicted onsets
recall	$\frac{TP}{TP+FN}$	fraction of correctly detected true onsets
F1 score	$\frac{Precision \cdot Recall}{Precision + Recall}$	balance between precision and recall
AUC (area under the curve)	$\frac{1}{2} \left(Recall + \frac{TN}{TN+FP} \right)$	threshold independent discrimination ability between positives and negatives

Table 5.3.: PN picker performance metrics (modified after Sokolova and Lapalme 2009). TP: true positives, FP: false positives, TN: true negatives, FN false negatives.

With the test subset, the PN picker achieved 91 % precision and 94 % recall for identifying seismo-acoustic meteoroid signal onsets for a fixed decision threshold of 0.5 (see Figure 5.4a), that was selected on the development subset. An AUC value of 0.89 (see Figure 5.4b) indicates a discriminative power, that is well above the 0.5 chance level of a random picker. The AUC value is calculated by first calculating the receiver operating characteristic (ROC) over a range of decision thresholds. The ROC is calculated by comparing the true positive rate with the false positive rate. The true positive rate indicates the proportion of positive cases that are correctly identified as such, while the false positive rate indicates the proportion of negative cases that are incorrectly classified as positive. The high AUC value implies that, operationally, we can raise or lower the decision threshold to prioritize fewer false alarms or fewer misses without collapsing the performance. Pick timing uncertainty as displayed in Figure 5.4c is likewise robust over the test subset and different meteoroid events. Picking uncertainties cluster tightly around zero with a median absolute uncertainty of 0.02 s and with 90 % of the uncertainties between -0.31 s and +0.29 s. These uncertainty values are quite low for inverting source parameters, because wave propagation is applied in a low-velocity acoustic medium (290 – 350 m/s), quite different from solid Earth situations with higher seismic velocities.

Figure 5.4d shows a comparison between the manually picked onsets (blue) and PN predicted probability distributions (red) as a record section over a random assortment of traces. These visual comparisons as well as the performance metrics and timing uncertainty of the PN picks confirm that the trained model not only separates signal from noise reliably but also localizes the time onset to within a few tenths of a second with a success rate of roughly 90 % over all phase picks.

The performance of the PN picker is robust for a large lateral distance range (Fig. 5.5), and we do not find a performance degradation directly related to increased distance ranges, which could be expected as the SNR generally decreases with distance (Fig. 5.5a). PN picks with an increased pick uncertainty ($> |0.5|$ s) do not seem to cluster either at large lateral distances or for low SNR signals, while pick probability correlates with pick uncertainty. In Figure 5.5b picks with a higher uncertainty do not cluster below a specific SNR and seem to be erratically distributed with distance to the source.

5.7. Application

The picked arrival times can be used to calculate the meteoroid trajectory which is important to estimate the possible strewn field on Earth's surface or the meteor origin in space. Here, we present an application to the fireball event listed as IMO Event 2025-2077 (International Meteor Organization (IMO), 2025a) and FRIPON Event 24644 (Fireball Recovery and InterPlanetary Observation Network (FRIPON), 2025). This fireball crossed the Adriatic Sea coming from Southeastern Europe towards Northern Italy on 12th April 2025 at 18:58 UTC. IMO received 401 reports from visual observations and the FRIPON camera network had 12 recordings. These data sources are used for a comparison with our trajectory solutions (Table 4).

The following 4-step workflow is applied:

1. Download of seismic waveform data for recording stations within 2.5° distance around the average midpoint between the FRIPON and IMO of 45°N , 13.5°E .
2. Prediction of onset times automatically with the PN picker using time padding around the modeled acoustic travel time from the reported trajectory to each station.
3. Calculation of 100 bootstrapped fast trajectory fits using particle swarm optimization (PSO) (e.g., Kennedy and Eberhart 1995; Miranda 2018) based solely on a random subset of 80 percent of PN picks at a fixed decision threshold of 0.3, 0.4, and 0.5.
4. Repetition of step 3 after a manual screen of the PN picked subset of waveforms and adjusting two outlier picks (seismic stations PE3 and SSFR).

As there are two outliers in the data, we call case 1 the one without correction and case 2 the other one for which we checked the picks and corrected the two pick times manually (Fig. 5.6b). The two manually corrected picks significantly improve the model fit based on the RMSE measured as difference of measured and modelled arrival times. Based on that, we find the best fitting solution for $\tau = 0.4$ with an RMSE of 2.2 s. Increasing the decision threshold to $\tau = 0.5$ slightly degrades the RMSE and results in larger azimuth and elevation angle uncertainties

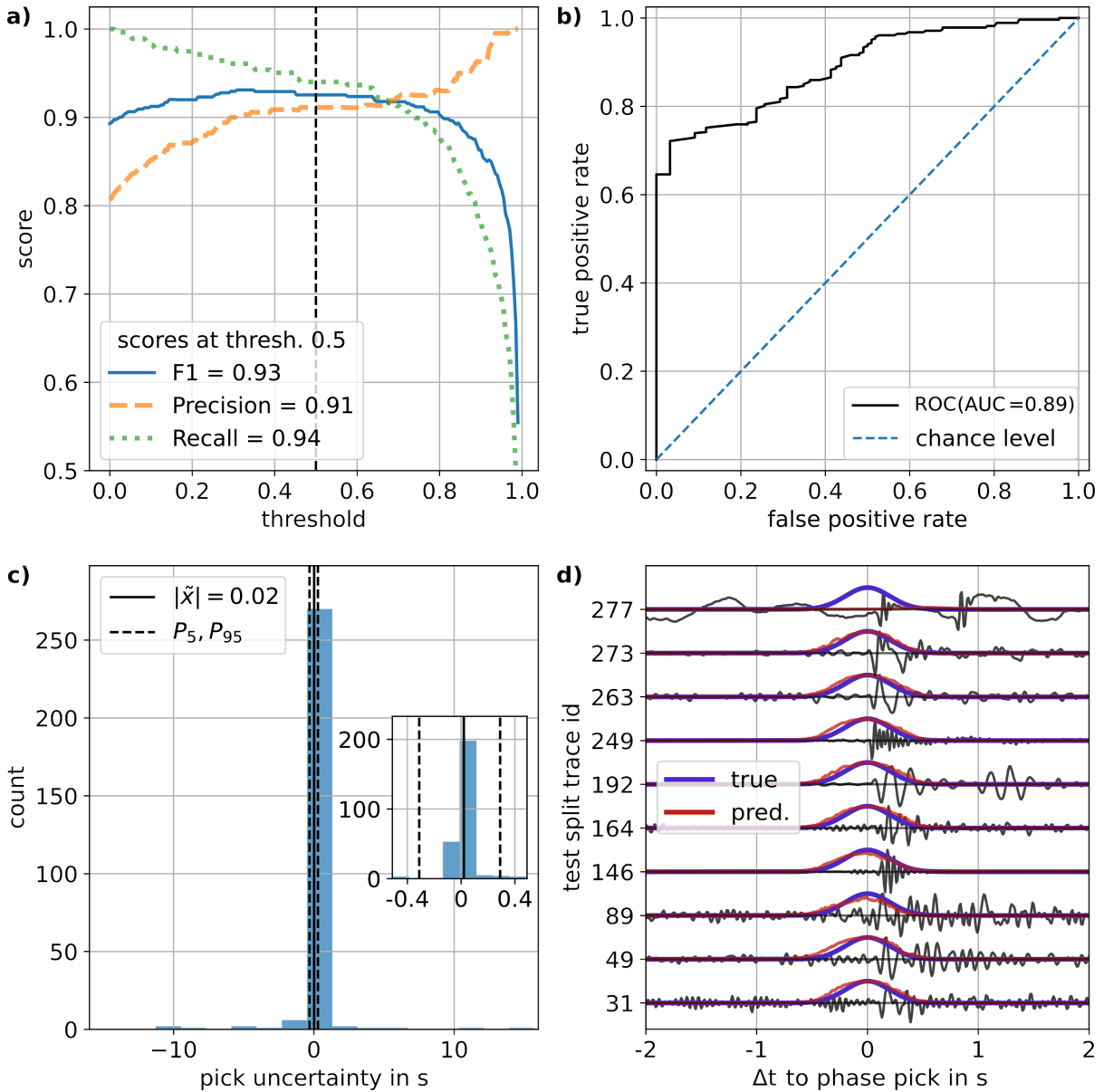


Figure 5.4.: Training results. a) Precision (orange), Recall (green) and F1 (blue) scores versus the decision threshold at which predicted picks are counted as successful hits. The threshold of 0.5 is chosen on the development split. b) Receiver-operating characteristic plot. True positive versus false positive rate of the PN picker (ROC: black line) to measure the overall success of the phase pick probability predictions. The AUC value of 0.89 indicates the overall goodness of the fit independent of the chosen decision threshold. A true random picker would predict at equally true and false positive rates, breaking down the success of phase pick predictions to a chance level (50 per cent). This is shown here as the blue dashed diagonal. c) Histogram of the pick uncertainty in seconds versus the time accuracy of the trained PN picker. The median absolute pick uncertainty value is 0.02 s and 90 per cent of pick uncertainty values are found between $P_5 = -0.31$ s and $P_{95} = 0.29$ s. The inset histogram shows a zoomed in portion around P_5 and P_{95} with thinner bars. d) Section plot of normalized vertical component seismic traces of the test split arranged by their test split trace id shifted so that $t = 0$ is at the time of the phase pick. In blue and red, the true picks after applying the Gaussian window and the predicted phase pick probabilities are given, respectively. Note the overall good match between truths and predictions, but also how the PN picker failed to identify the phase for trace id 277.

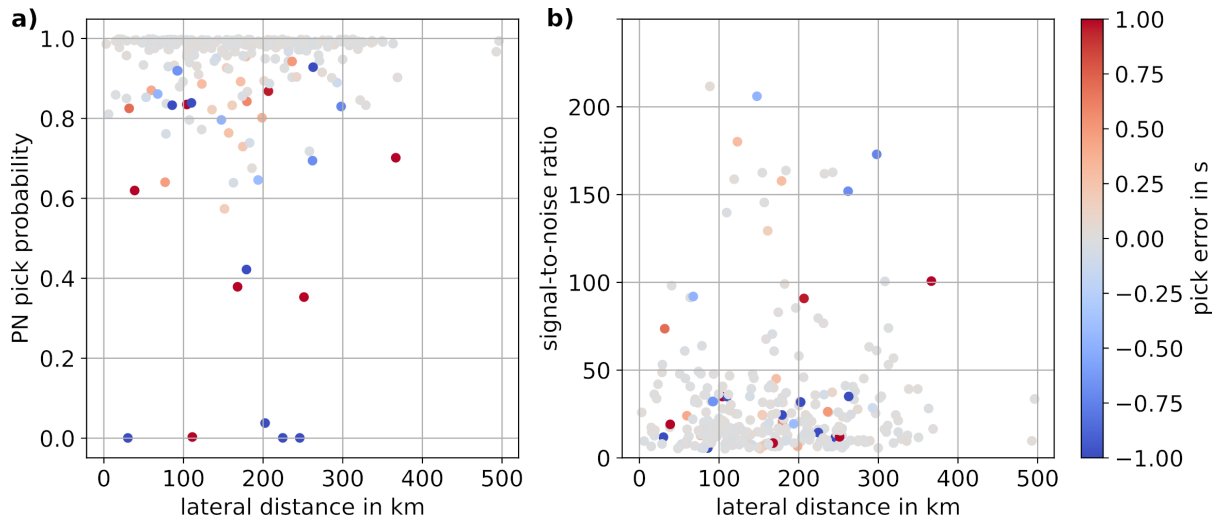


Figure 5.5.: a) Distribution of PN pick probabilities versus lateral distance to the respective meteoroid trajectory midpoint. Color indicates the picking uncertainty of any given PN pick. b) Distribution of SNR of picked phases versus lateral distance and associated PN pick uncertainty.

with slightly narrower meteoroid velocity uncertainty, compared to $\tau = 0.4$. The solutions at decision threshold $\tau = 0.3$ for cases 1 and 2 result in a substantially wider spread of solutions over the 100 bootstrapped runs at 80 per cent of the amount of PN picks compared to the solutions at higher thresholds.

The applied PSO method for reconstructing the trajectory uses 3-D ray tracing with *infraGA* (Blom, 2014). By applying *infraGA*, we can incorporate complex ground-to-space atmosphere models (Hetzer, 2024) that include wind to model the acoustic wave propagation in a heterogeneous moving atmosphere. The modelling results in Fig. 5.6 illustrate that high quality PN picks provide robust constraints. In addition, slight manual curation of onset picking can improve trajectory fits at present SNR and model training levels. For truly independent and automatic picking without further manual inspection, we suspect that SMORD must be expanded with more training data. Nevertheless, SMORD and the trained PN picker can already be used during manual phase picking as a tool to easily identify seismo-acoustic onsets in waveforms and, thereby, help to quickly find a trajectory and possible strewn field.

5.8. Discussion

SMORD provides a valuable opportunity for future studies on meteoroids and other seismo-acoustic phenomena. Still, there are only few systematic studies on meteoroid falls observed with seismic networks (e.g., Edwards et al. 2007; Neidhart et al. 2021). None of these previous studies provide easy access to the original waveforms which are necessary to repeat earlier findings or to test new methods and compare their outcome with previous results. Our data compilation of meteoroid falls, whose seismo-acoustic signals were observed by seismic stations, is unique worldwide. Thereby SMORD will make it easier to study meteoroid properties in details such as trajectory determinations (e.g., Vida 2023), energy and mass

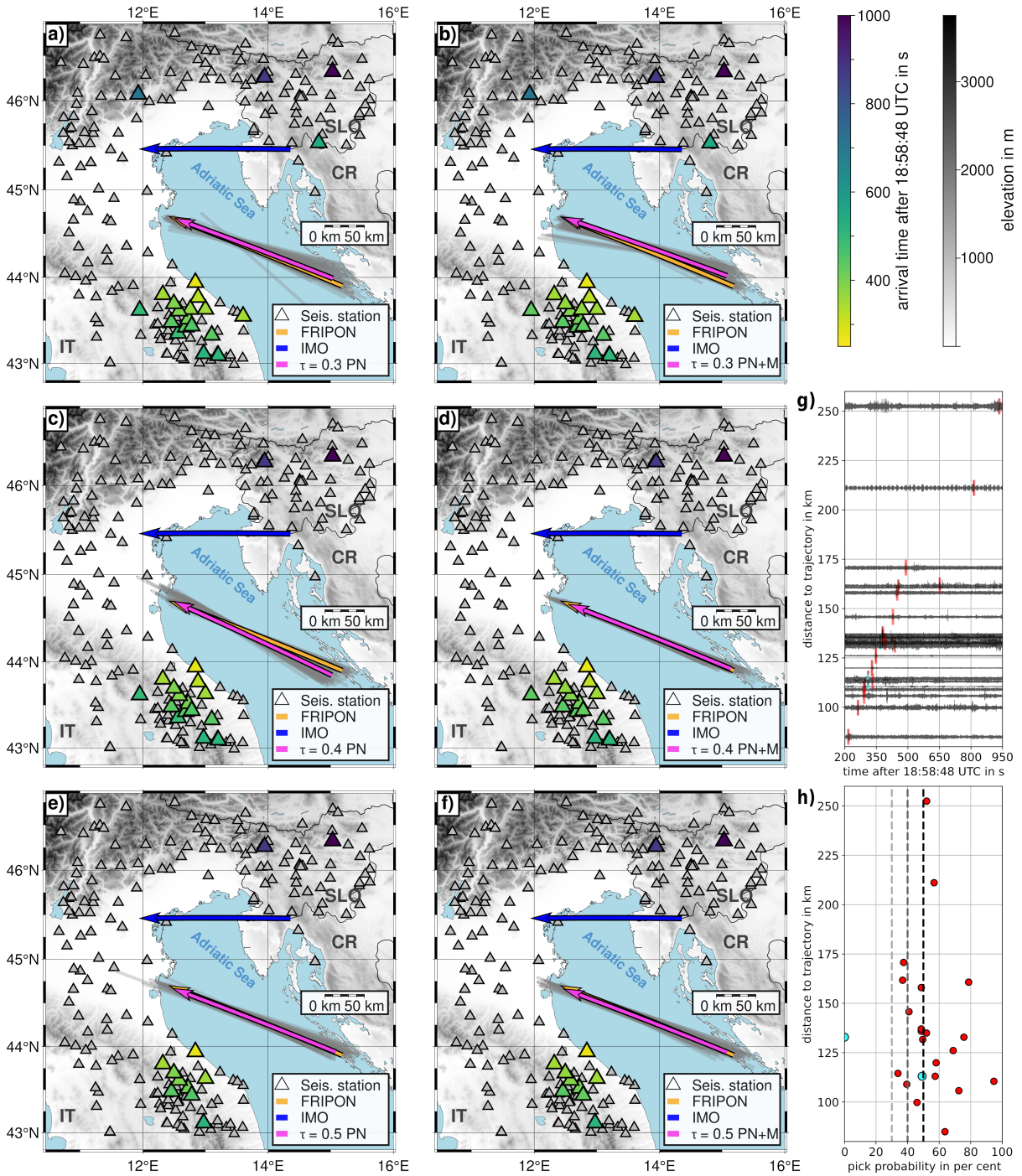


Figure 5.6.: Fireball event on 12th April 2025 across the Adriatic Sea. a-f) FRIPON camera trajectory (yellow arrow) and IMO observation report trajectory (blue arrow). Gray lines show 100 bootstrapped runs, and the median seismic PSO solution trajectory is given as a purple arrow. Seismic station distribution as triangles, colored with respect to the identified arrival time. Gray triangles mark stations that were discarded during PSO modeling, because they do not contain a seismic onset surpassing the given pick decision threshold. a, c, e) show the distribution of used stations, bootstrapped PSO runs and median PSO solution for case 1 (no manual correction, see text) with pick decision thresholds τ of 0.3, 0.4, and 0.5, respectively. b, d, f) same as a, c, e), but for case 2 with PN picks manually verified and two corrected PN picks. g) Section plot of the seismic waveforms: Red bars show the onsets of the PN picks and blue bars show the two manually corrected picks. h) Pick probability versus distance to the trajectory midpoint. Red circles mark the PN pick probabilities, and blue circles mark the pick probabilities at the manually set pick. The probability at one of the manually set picks is 0, meaning the PN picker failed to identify the phase. The pick decision thresholds 0.3, 0.4, 0.5 are drawn as dashed gray to black lines.

Trajectory source	Median azimuth in °	az-angle	Median elevation angle in °	Median velocity in km/s	Median RMSE in s
IMO Event 2077-2025	270.9	-	-	-	-
FRIPON Event 24644	292.9	12.1	26.3	-	-
PSO solution $\tau = 0.3$ PN (21 picks, case 1)	290.6 ± 4.5	13.6 ± 2.7	24.4 ± 2.5	26.9	
PSO solution $\tau = 0.3$ PN (21 picks, case 2)	288.6 ± 5.3	12.8 ± 3.5	26.1 ± 2.6	18.5	
PSO solution $\tau = 0.4$ PN (17 picks, case 1)	294.9 ± 5.1	12.8 ± 1.2	26.6 ± 2.5	18.0	
PSO solution $\tau = 0.4$ PN (17 picks, case 2)	291.6 ± 1.8	12.8 ± 0.9	26.4 ± 3.9	2.2	
PSO solution $\tau = 0.5$ PN (11 picks, case 1)	291.2 ± 2.4	12.4 ± 1.7	26.4 ± 2.9	10.8	
PSO solution $\tau = 0.5$ PN (11 picks, case 2)	291.1 ± 2.0	12.7 ± 1.7	26.4 ± 2.3	2.7	

Table 5.4.: Comparison of FRIPON trajectory parameters with automatic PSO solutions (case 1) and with corrected data (case 2 with two manual pick corrections). Trajectory uncertainties are calculated from the 90 percent confidence interval of the 100 PSO solutions. The RMSE is the misfit between modelled and observed arrival times and the median RMSE is the misfit over all 100 PSO trajectories.

estimations, or ablation parameters (e.g., [Campbell-Brown et al. 2013](#)). For larger objects the trajectory can be used for refined strewn field mapping before going into the field to search for precious extraterrestrial material. The search for meteoroid-related seismo-acoustic signals can be quite difficult, because the recorded amplitude depends on numerous factors: source strength, distance or geometrical spreading, atmospheric winds and temperature and frequency-dependent attenuation, background noise, site effects, etc. Due to low SNRs, such signals are often not identified. Another point is that the moveout of seismo-acoustic waves is quite different from earthquake waves, and therefore, seismic data analysts often fail to recognize such waves crossing a seismic network. These issues can be treated with SMORD and our first trained PhaseNet based phase picker to automatically identify meteoroid-induced seismo-acoustic signals. PhaseNet was successfully trained on a global dataset containing a wide variety of seismic sensors at different sites (achieving >90% precision and recall at a decision threshold level of 0.5). This achievement streamlines the process of phase identification when the trained PN picker is used as a tool to aid manual picking routines. The common approach of training PhaseNet for a specific network is not possible in our case, because there are not enough observations per network. However, we find that the PN picker performs well for meteoroid-related signals with different frequency contents, regardless of specific sites (Fig. 5.4d). In our application the PN predictions can be followed over a distance range of up

to approximately 250 km to the FRIPON meteoroid trajectory (see Figs. 5.6g, 5.6h). This aligns well with the distance ranges we encountered during the compilation of SMORD (see Figs. 5.1a, 5.1b, 5.5a, 5.5b). We find that the capability to identify seismo-acoustic arrivals largely depends on the source characteristics, recording conditions, as well as atmospheric conditions. Uncertainties related to the PN picks are well below the accuracy required for precise trajectory reconstruction. At a decision level of 0.5, 90% of picks are within ± 0.3 s. Together with an acoustic wave velocity of c. 330 m/s this corresponds to a distance uncertainty of just ± 100 m. Here we note that these pick timing uncertainties are negligible compared to modelling uncertainties due to wind effects on sound propagation or modelling limits such as spatial discretization for wavefield modelling. Meteoroid event trajectories and velocities from existing meteoroid catalogs contain uncertainties; CNEOS records depend on sensor characteristics and processing whereas IMO eyewitness reconstructions are biased towards populated regions and wake-sleep cycles. Thus, their combination is helpful for narrowing down uncertainties. Our manual labeling during the initial phase identification of SMORD may omit weak or ambiguous arrivals and inter analyst variability remains but can be mitigated by guidelines. Atmospheric state variability (winds and temperature stratification) was not modeled explicitly during the initial phase identification of the seismo-acoustic onsets and likely explains some false detections of the PN picker. A major future application could be autonomous, continuous monitoring, detection, and localization of meteoroid falls using seismic data. The 4-step workflow we presented can be automated almost completely and the only user input would be origin time and location, which would allow for a rapid meteoroid trajectory calculation. The observational basis, continuous online waveform data, already exists in many continental regions due to open data centers such as EIDA (Strollo et al., 2021) or IRIS (Trabant et al., 2012). An extension of these observational capabilities can be expected for the future based on previous experience. Of course, detectors need to be improved to account for wavefields with acoustic moveout, but such issues can be solved and coincidence detection with acoustic moveouts is possible in principle. One feature which could be used is the addition of the typical waveform pattern, associated with meteoroid related onsets, which is W-shaped on the vertical component in the beginning of the wavetrain with a downward motion due to the downward propagating pressure front (Kanamori et al. 1991; Edwards et al. 2008). This enhanced detection approach would be especially valuable for detecting meteoroids during the daytime, when optical methods struggle due to the daytime light conditions which lead to oversaturation of the cameras. Another crucial application would be signal detection during the dark flight phase at the end of the trajectory which is important for estimating strewn fields (Devillepoix et al. 2018; Moilanen et al. 2021). Another possible future application of SMORD is the evaluation of site effects and local coupling effects at seismological recording stations. Shallow geological features as well as above surface features such as buildings or steep topography affect the efficiency with which the acoustic wave can couple into the subsurface and convert into seismic waves. Depending on the local geology, seismo-acoustic wave amplitudes might be dampened or amplified and the waveform clarity can be affected by the site effects as well (Edwards et al. 2008; Novoselov et al. 2020). In a future application the measurement of seismic amplitudes and their comparison with amplitudes from acoustic infrasound measurements can be used to better estimate seismo-acoustic coupling coefficients (Novoselov et al. 2020; Bishop et al. 2022), because amplitude information helps to determine the energy or mass of a supersonic object (e.g., Edwards et al. 2007). As there are many more seismic stations

than infrasound sensors available for scientific and forensic studies, seismological analysis will enable better investigations of events with strong acoustic emissions, e.g., accidental explosions (Hinzen, 2007) or combat actions (Dando et al., 2023).

5.9. Conclusions

We introduce SMORD, a curated dataset of 3,295 seismo-acoustic picks from 316 meteoroid events and demonstrate that a PhaseNet model, trained on SMORD labels, can detect meteoroid air-to-ground coupled seismic waves with >90% precision and recall on a test subset of events. SMORD and the associated phase-identification model provide a reproducible foundation for automated meteor signal detection in seismic data and for multimodal bolide analyses. The trained picker enables rapid scanning of regional seismic archives following reported bolides, supporting time critical meteorite searches and initial trajectory inference. Integration with GLM detections (Jenniskens et al., 2018), with radar (CMOR/BRAMS), and with infrasound arrays will enable joint constraints on source time-height functions and energy deposition. As GLM and global seismic recording coverage continue to grow, SMORD-like labels can power near real time pipelines. Future work will focus on integrating atmospheric propagation modeling with the automatic identification of seismo-acoustic arrivals. Coupling the current workflow with atmospheric propagation models (e.g., ray tracing with *infraGA* (Blom, 2014) constrained by reanalysis wind fields (Hetzer, 2024)) is expected to improve the physical consistency of predicted travel times and phase associations. Additionally, we plan to expand the compiled database by lowering the thresholds of 100 IMO sightings and the 0.3 kt CNEOS yields, as well as incorporating additional meteoroid observations from external networks, such as FRIPON (Colas et al., 2020) and the Global Meteor Network (GMN) (Vida et al., 2021). This will increase the number and diversity of events available for training and validation. Regarding machine learning, we will investigate training strategies dedicated to infrasound data. This will include separate models for array-based recordings and individual stations to better capture differences in waveform characteristics and noise conditions. Finally, we will explore joint seismic-acoustic phase identification to leverage complementary information from ground motion and pressure waveforms, aiming to improve robustness and reduce picking ambiguities.

6. Re-entry and Burn Up of Starlink-2382 Satellite: Estimating Trajectory and Ablation Coefficient from Acoustic and Coupled Seismic Waves

The following section describes the acoustic-seismic signature of the re-entry of Starlink-2382 satellite on August 27th, 2024, at c. 19:30 UTC. The work presented in this study shows how reliable satellite re-entries can be observed with acoustic and seismic stations and how its trajectory can be reconstructed across a >700 km long ground track. It is also presented how the main fragment of the Satellite lost mass over the course of its atmospheric path and how its acoustic waves couple into the subsurface at collocated acoustic-seismic receiver locations. The following section is completely published as [Eickhoff et al. \(2026a\)](#).

My contributions to this study, following CRediT, are:

1. Conceptualisation,
2. Data Curation,
3. Formal Analysis,
4. Investigation,
5. Methodology,
6. Software,
7. Validation,
8. Visualisation,
9. Writing (original draft, review & editing)

6.1. Abstract

On August 27th, 2024, at approximately 19:30 UTC, the Starlink-2382 satellite entered the Earth's atmosphere following an uncontrolled re-entry manoeuvre over Central Europe. This event resulted in a relatively low-angle re-entry of the satellite into the atmosphere, which might have provided sufficient time to burn up the satellite before reaching the Earth's surface. This study employs acoustic-seismic (A-S) data from 226 recording stations to analyse the trajectory of Starlink-2382's re-entry, utilizing 3-D atmosphere models including wind data and acoustic ray tracing methods. To identify signals emitted by the falling satellite, we process A-S recordings of Austrian, French, German, Italian, Slovenian, and Swiss regional seismic networks. We compute the satellite trajectory with a novel ray-based direct-search optimization method and find an azimuth angle of $120.5^\circ \pm 0.4^\circ$ from North and an initial elevation angle of $1.5^\circ \pm 0.7^\circ$, together with an entry velocity of approximately $8.9 \pm 0.7 \text{ km s}^{-1}$. Our findings indicate that this acoustic-seismic approach, including travel time effects due to wind, achieves a better fit to our large dataset compared to the trajectory solutions from optical methods in this specific context. Furthermore, we calculate an effective ablation coefficient of $0.11 \pm 0.02 \text{ s}^2 \text{ km}^{-2}$. Within the limits of this estimate, this is consistent with a scenario in which the satellite, with a mass of c. 100 kg could have experienced near-complete ablation during atmospheric descent. Finite-difference modelling illustrates the complex acoustic wavefield resulting from the satellite's deceleration and shows the expected widening of the Mach Cone. This effect highlights the importance of accounting for trajectory curvature and time-varying Mach angles when modelling acoustic wave propagation from low-angle re-entering objects. For recording sites with both, acoustic (infrasound) and seismic sensors, the acoustic-to-seismic ground coupling coefficients are determined. These vary up to three orders of magnitude, from $4.31 \times 10^{-10} \text{ m s}^{-1} \text{ Pa}^{-1}$ to $5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$ across our station sites, which is primarily explained by differences in stiffness of surface rocks.

Key words

Infrasound; Numerical modelling; Wave propagation; Seismic Instruments; Acoustic-gravity waves; Acoustic emission

6.2. Introduction

The presence of uncontrolled satellites in Low-Earth Orbits (LEOs) poses considerable economic as well as social risks. This threat manifests in the form of potential hazards to existing satellites, manned space missions, infrastructure, and even lives on Earth's surface (Patera 2008; Ailor 2019; Omalý et al. 2025). Related risks increase with the rapidly accumulating number of objects occupying LEOs (Lawrence et al., 2022). The recent development of megaconstellation communication-satellite networks, which provide commercial satellite-based internet access, resulted in a significant increase in the number of LEO objects (McDowell, 2020). A leading driver to provide satellite-based internet access is the U.S. company SpaceX with about 7,000 satellites for the Starlink service (SpaceX, 2025). On August 27th at approximately 19:30 UTC, the Starlink-2382 satellite re-entered Earth's atmosphere above France (see Figure 6.1b). This event occurred after SpaceX lost control over the satellite that led to an uncontrolled

re-entry of the spacecraft. This event occurred after SpaceX lost control over the satellite that led to an uncontrolled re-entry of the spacecraft over Central Europe in contrast to the target region of open oceans, away from populated islands and heavily trafficked airline and maritime routes (SpaceX, 2025). SpaceX asserts that a Starlink satellite should completely burn up during an atmospheric re-entry (SpaceX, 2025). Here we want to use this uncontrolled satellite re-entry to investigate the likelihood of a full burn-up during re-entry and compare the satellite's rate of ablation with other re-entry events. Our study benefits from a high number of recording stations in Central Europe and a low elevation angle which causes a strong wave emission towards the Earth's surface. The partly known parameters of the satellite e.g., mass, offer an advantage compared to meteoroid observations. Such a priori information helps to test methods which are also applied to study meteoroid properties such as mass or kinetic energy. When an object enters Earth's atmosphere with a velocity v exceeding the local speed of sound (v_s), acoustic Mach waves are generated. For $v \gg v_s$ the wavefronts of these Mach waves travel almost parallel to the fast-moving object producing the waves (e.g., Edwards et al. 2008). The following interaction and deformation of the Earth's crust, induced by these acoustic waves and recorded with seismometers, can be used to invert for trajectory parameters of the flying object (e.g., Kanamori et al. 1992; Arrowsmith et al. 2007; Edwards et al. 2008; Fernando and Charalambous 2026). In the region of Central Europe, beneath the re-entry trajectory of Starlink-2382 satellite, there are some of the world's most extensive seismic station networks (Fig. 6.1a), a circumstance that facilitates a comprehensive study of the re-entry. The geology underneath the recording stations can be broadly categorized into poorly consolidated sediments in the northern and southern foreland basins of the European Alps and hard rock in the mountainous Alpine region (Pfiffner, 2021). This petrophysical difference may cause varying coupling coefficients for the conversion of acoustic to seismic waves (e.g., Edwards et al. 2008; Novoselov et al. 2020; Fernando et al. 2024b). Previous investigations (Kanamori et al. 1992; Qamar 1995; Cates and Sturtevant 2002; Langston 2004; Pujol et al. 2005; de Groot-Hedlin et al. 2008; Edwards et al. 2008) have demonstrated that the use of seismic recordings can serve to constrain meteoroid as well as spacecraft trajectories. In a similar manner, acoustic recording data from infrasound stations can be utilized (e.g., Edwards 2009; de Groot-Hedlin and Hedlin 2014; Pilger et al. 2020b). One objective of this study is to compare the accuracy of different acoustic-seismic (A-S) trajectory solutions, such as geometrical and ray-based methods together with finite-difference wavefield modelling, with the results provided by human observations, camera recordings, and the two-line element set (TLE) data of the satellite. The acoustic waveform data can be also used to extract the (kinetic) energy of the moving object. In this study, another objective is to investigate the energy of the moving source and the rate at which acoustic waves induce seismic waves into the subsurface. This is achieved by analysing co-located A-S recording stations. First, the change of the source energy along the satellite's fall trajectory will be analysed and then the A-S coupling efficiency. Additionally, the change of energy of the source will be related to a change of satellite mass due to the ablation processes. Understanding the local A-S coupling coefficients is important to constrain future space-debris or meteoroid masses based on ground motion measurements and relationships between acoustic wave amplitudes and source energies (ReVelle, 1997).

Acoustic and Seismic Data

The A-S data for this study is obtained through the International Federation of Digital Seismo-

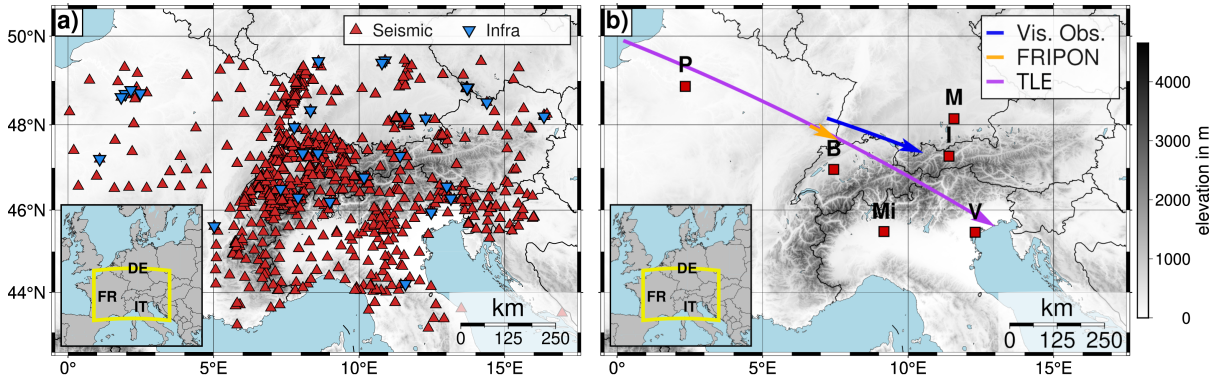


Figure 6.1.: a) Map of the target region in Central Europe. Red triangles indicate seismic stations, while blue inverted triangles denote infrasound stations from which data were retrieved. The inset map illustrates the geographical location of the target region within Europe. b) Three trajectories are given, derived from human observations in blue (International Meteor Organization (IMO) 2024), the meteor observation camera network FRIPON (2024) in yellow, and satellite TLE data in purple (Satcat 2024). The inset map indicates the position within Europe. Larger cities for orientation are Paris (P), Bern (B), Milano (Mi), Innsbruck (I), Munich (M), and Venice (V).

graph Networks (FDSN, see Network References), and the study area encompasses a total of 1376 seismic recording stations and 34 infrasound recording stations from 36 networks. Ten infrasound stations are co-located with seismic sensors. We use the TLE data of the satellite to further constrain the search area. We restrict the analysis to stations where the expected acoustic wave arrival is within 1000 s after 19:28:00 UTC. This reduces the dataset of 1410 individual stations to 611 seismic and 12 acoustic stations. The vertical recording component of the seismic data was analysed to determine the first onset of the A-S coupled wave. A deconvolution was applied to transform the waveform amplitudes to true ground motion velocity. Afterwards a non-zero phase high-pass filter with a cut-off frequency of 1 Hz was applied to the seismic waveforms. The infrasound recording data was processed in a similar manner and the waveform amplitudes were transformed to pressure. Due to the significantly lower frequency nature of the acoustic waves, the data was high-pass filtered with non-zero phase and a corner frequency of 0.1 Hz (see Figure 6.2). The lower frequencies of the infrasonic acoustic waves also result in detection ranges over thousands of kilometres (ReVelle 1976; de Groot-Hedlin and Hedlin 2014).

Out of the 611 seismic recording stations, the A-S coupled wave arrivals could be identified at 218 of these stations. This results in a recovery rate of approximately 36%. For shielding reasons, seismological recording stations are often placed in locations where short-period pressure differences, such as acoustic waves, should not disturb the instruments. This is a critical aspect of earthquake seismology, but it hinders the detection of acoustic non-earthquake events, such as this satellite fall. Notably, the density of seismic recording stations in this area is exceptionally high, particularly given that it is an unplanned re-entry, as opposed to the planned recording efforts for missions such as OSIRIS-REx (e.g., Fernando et al. 2024b, Fernando et al. 2024a, Silber et al. 2024), Hayabusa (e.g., Yamamoto et al. 2011; Ishihara et al. 2012) and Hayabusa2 (e.g., Nishikawa et al. 2022; Sansom et al. 2022). In terms of infrasound recording stations, these demonstrate clearly enhanced efficiency as no A-S coupling is involved: of the

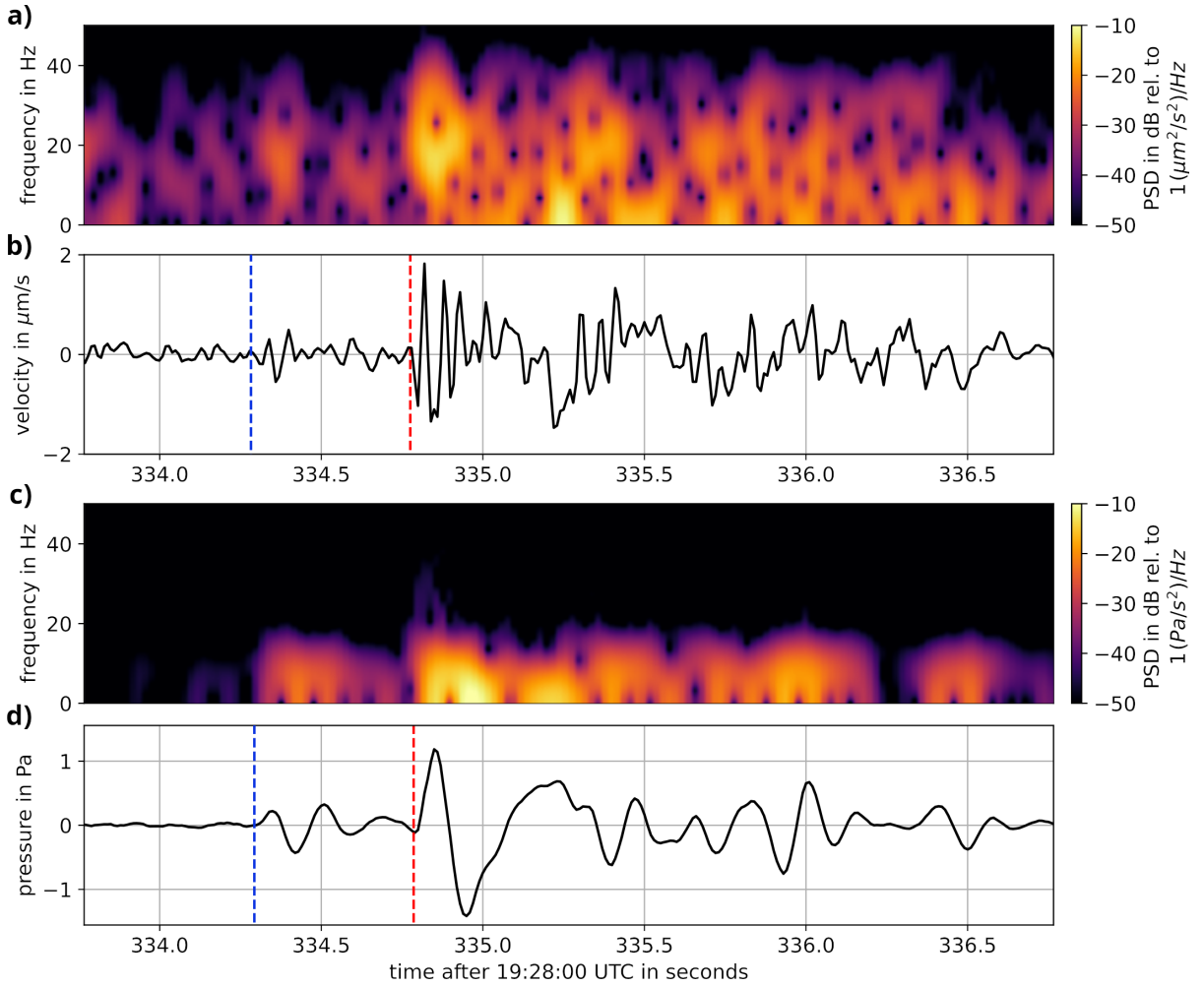


Figure 6.2.: Observed recordings at Raspberry Shake & Boom station AM.R03EA, located within 10 km distance of the centre line (see Fig. 6.3). a) Spectrogram of the vertical component of the seismic recording. Colours are scaled to the maximum measured spectral amplitudes. b) Waveform of the seismic recording data in $\mu\text{m/s}$. The red dashed line marks the onset of measured W-shaped wavelet associated with the A-S coupled wave. The blue dashed line marks the onset of a wave due to a possible smaller fragment of the satellite which is ca. 0.5 s ahead, an air-coupled Rayleigh wave (e.g., Edwards et al. 2008) or a faster multipathing event. c) Spectrogram of the infrasound recording. Colours are scaled to the maximum measured spectral amplitudes. d) Waveform of the infrasound recording data in Pa. The red dashed line marks the onset of measured N-shaped wavelet associated with the acoustic wave.

12 infrasound stations, eight were able to detect clear acoustic wave signals, resulting in a recovery rate of approximately 66%.

In summary, we identify signals related to the satellite re-entry at 218 seismic and 8 infrasound stations, yielding 226 total arrival times as data distributed across Austria, France, Germany, Italy, Slovenia, and Switzerland. From the isolines of the signal-arrival times (see Figure 6.3a), we derive an initial approach trajectory based on A-S data alone. This trajectory fits quite well with the projected trajectory of the TLE data of the satellite (Satcat 2024). A significant non-random systematic offset is observed when comparing the average apparent acoustic wave velocity for stations located north and south of the initial trajectory (Fig. 6.3b and 6.3c).

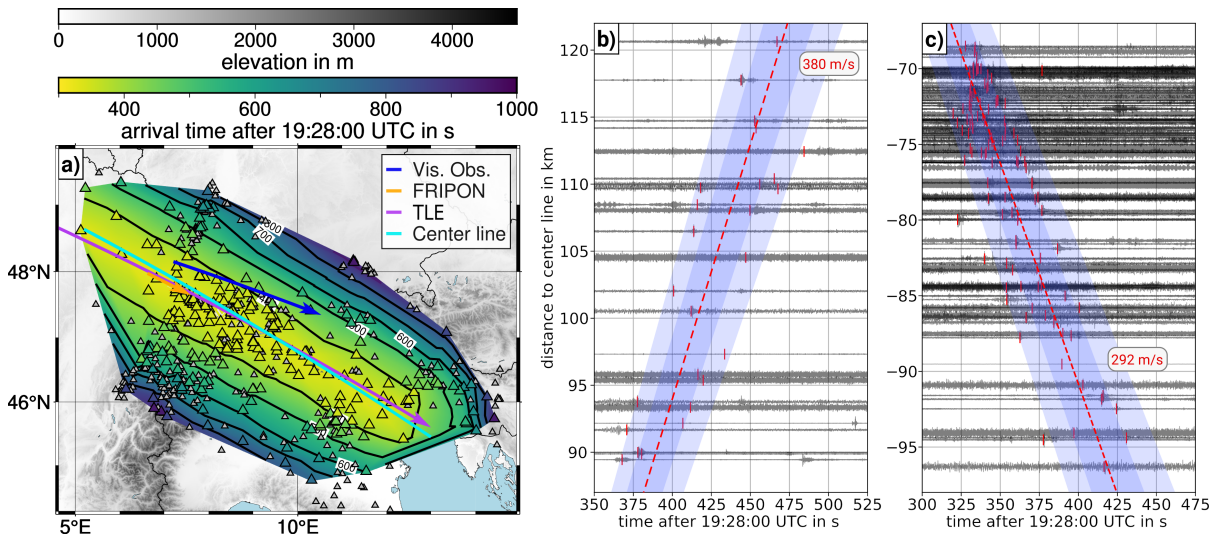


Figure 6.3.: a) Map of the target region displaying the locations of acoustic and seismic recording stations (triangles). The triangles are colour-coded according to the arrival time of the acoustic or seismic wave, respectively. Grey triangles are stations where no signal could be identified. Contour lines represent interpolated arrival times, and the colour gradient gives a smooth picture of the spreading wave. Trajectories of the International Meteor Organization (IMO) 2024 (blue), FRIPON (Fireball Recovery and InterPlanetary Observation Network) 2024 (yellow), and Starlink-2382 satellite TLE data (purple; (Satcat, 2024)) are also displayed. Additionally, a centre line (cyan) is derived geometrically from the minimum of the arrival time contours. b) Record section with acoustic and seismic waveforms located to the north of the centre line. The first onset times of the acoustic wave signals are marked with a red vertical bar, and the average apparent propagation velocity is marked with a red dashed line. Dark blue areas indicate deviations of the average velocity of up to 5%, while light blue areas indicate deviations of up to 10%. c) The same as b), but for waveforms located to the south of the centre line.

This offset suggests that the signals arrive faster at stations north of the trajectory, hence with a higher apparent average velocity along their respective wave travel paths compared to stations south of the trajectory (see Figure 6.3b, 6.3c). This offset could be related to a SW-NE wind below ca. 50 km height (Fig. 6.4) with an upwind effect towards SW slowing down acoustic waves and a downwind effect to NE with acoustic waves becoming faster. In order to better understand the origin of this offset, it is necessary to incorporate more complex atmospheric models for acoustic velocity that include a stratified atmosphere and wind for the trajectory-reconstruction approach. To this end, a 3-D grid compiled from 1-D ground-to-space atmosphere models is utilized (Drob et al. 2003; Drob et al. 2010). The highest wind speeds occur in the thermosphere and mesosphere (above 50 km altitude), with speeds of up to 50 m/s and a uniform north-westerly direction at all grid points (Fig. 6.4). In the stratosphere (20 - 50 km altitude), the wind direction shifts to south-southwest, with average speeds of 10 m/s. Below altitudes of 20 km, the wind direction appears to be more random, which is expected for the troposphere as it is dominated by smaller-scale weather phenomena (e.g., (Huber et al., 2001); (Mahrt, 2011)).

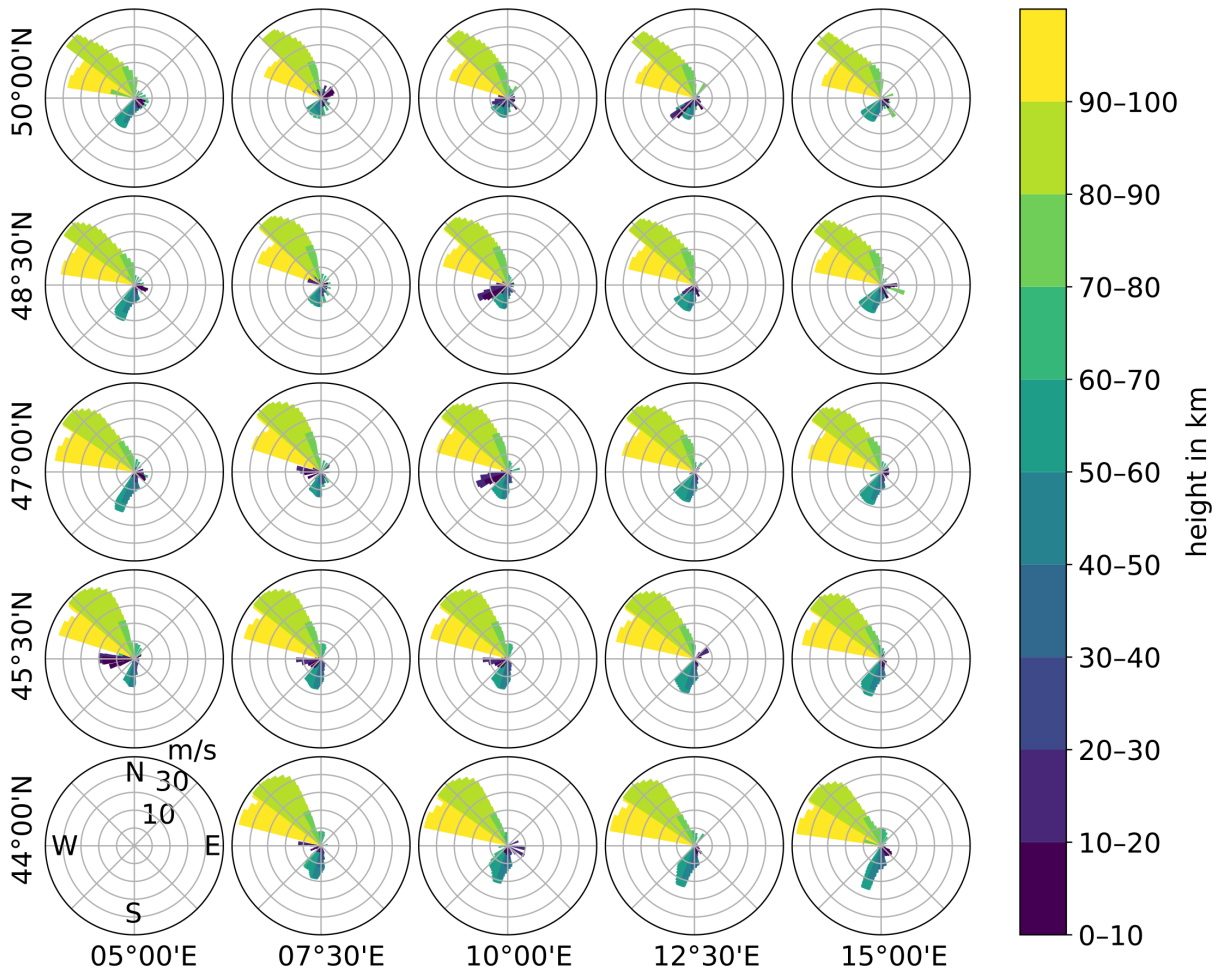


Figure 6.4.: Grid of polar plots with 1.5° distance in N-S direction and 2.5° distance in W-E directions displaying wind direction, speed and height (colour). The direction of the colour-coded bins corresponds to the wind direction. The length of the bars denotes wind speed, ranging from 0 m/s to 50 m/s (one circle corresponds to a 10 m/s increment). Each position of a polar plot within this grid corresponds to the location of its respective one-dimensional atmospheric model, which was employed to calculate these wind directions and speeds. Data is provided by Hetzer 2024.

6.3. Methods

To determine the optimal trajectory solution, the particle swarm optimization (PSO) method proposed by Vida (2023) is applied, which is an improved application of the inversion derived by Pujol et al. (2005). We call this method, that derives trajectories from geometrical rays, the geometrical model in the following. This method is implemented to identify a linear trajectory that minimizes the arrival time residual over all acoustic and seismic travel time readings. The trajectory is constrained by 6 parameters: The puncture point P of the trajectory with the Earth's surface as $P(x_0, y_0, t_0)$, as well as the elevation angle θ , the azimuth angle φ , and the velocity v . The analysis minimizes the cost function L to find the optimal trajectory solution:

$$L = \sum_{i=1}^N 2 \left(\sqrt{1 + (t_{a,i} - t_{o,i})^2} - 1 \right) \quad (6.1)$$

Here the analytical arrival time $t_{a,i}$ is defined as the time when the acoustic or seismic signal arrives at the observation point i , as described by Pujol et al. (2005) and Vida (2023) for a Mach Cone in the atmosphere. The observed arrival time $t_{o,i}$ is defined as the onset time of the identified acoustic or seismic wave arrival at the observation station i . To determine the optimal trajectory that explains our arrival time measurements, we applied the PSO as implemented in the PySwarms (Miranda, 2018) Python library. However, it should be noted that this implementation of the geometrical method is subject to certain limitations and assumptions: 1) a constant sound velocity in the atmosphere which was fixed at 320 m s^{-1} , 2) linear travel paths from the trajectory to the A-S stations, and 3) a straight trajectory with a constant velocity (e.g., Pujol et al. 2005; Eickhoff et al. 2023; Föst et al. 2023; Eickhoff et al. 2024; Föst, Eickhoff & Ritter (in preparation)). The geometrical model is applied as a two-step process: in the first step, the cost function is minimized with all 226 arrival time observations. In the next step, arrival times with a residual greater than 20 s are eliminated, and then the geometrical model is reiterated. The initial parameters for both geometrical model runs of a given iteration are both defined by the same geometrical trajectory (see Fig. 6.3a). For our application we use 216 arrival times and invert for 6 parameters, which is a well-posed inversion problem.

To find the optimal trajectory, 500 two-step geometrical model runs are performed, generating 500 trajectories. The goodness of their fit is measured by their arrival time residual. These are displayed as blue bars in Figures 6.5a, 6.5c and 6.5d within the target region. The best-fit geometrical model trajectory has azimuth and elevation angles of approximately 120° and 12° , respectively, and an average satellite velocity of approximately 8 km s^{-1} . The RMSE of our geometrical model is approximately 10.9 s based on 161 arrival times. In most of the geometrical model runs, stations close to the geometrical centre line are not excluded after the first step. This is expected, as only for more distant stations the assumption of linear rays leads to large arrival time residuals that the method cannot balance anymore. This is further illustrated by stations located at the periphery of the modelling region, which are identified as potential candidates for removal following the initial PSO step, because their residual often exceeds 20 s (see Figure 6.5b). This outcome is attributed to the violation of assumptions 1) constant velocity and 2) straight ray paths. For stations distant from the trajectory, the average velocity of a pressure wave arriving at the station is likely to deviate from the anticipated 320 m s^{-1} , as atmospheric conditions such as wind significantly influence the propagation speed. As demonstrated in Figure 6.4, the wind velocity in the target region is in the order of magnitude of tens of m s^{-1} . For long travel paths, the impact of wind disturbances on the average velocity becomes more pronounced. Assumption 2) is violated due to the impact of winds on the propagation of the pressure wave, resulting in a curvilinear travel path. Furthermore, Figure 6.5e reveals systematic deviations in the arrival time residuals. Negative residuals are observed to the north of the trajectory, indicating that the acoustic wave velocity exceeds the average speed of sound of 320 m s^{-1} . Conversely, the stations positioned south of the trajectory predominantly exhibit positive arrival time residuals, indicating that the acoustic wave propagates at a velocity that is less than 320 m s^{-1} . The gridded wind model (Fig. 6.3) suggests that this spatial distribution of arrival time residuals can be attributed to the preferential winds with a predominant north-northwest orientation within the lower 50 km in the atmospheric layers. Consequently, an improved approach to reconstruct the

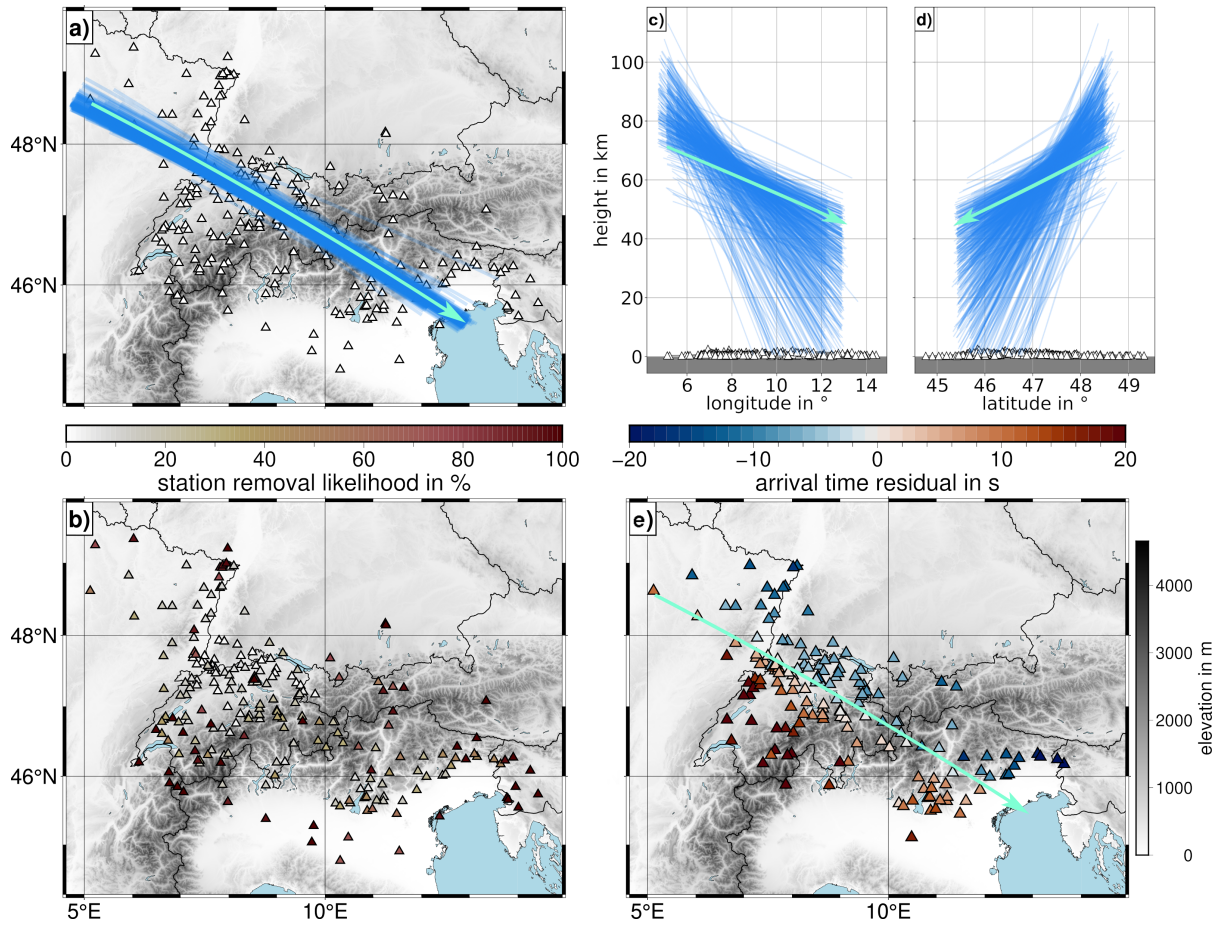


Figure 6.5.: a) Map showing different trial trajectories of the geometrical model reconstruction method in transparent blue. The optimal geometrical model trajectory is highlighted in cyan, while white triangles indicate seismic and acoustic recording stations. The colour-coding of trajectories and symbols is consistent with Figures 6.5c-e. b) Map of A-S stations and their likelihood of removal during the calculation of the geometrical model solution. Figures 6.5c and 6.5d illustrate the E-W versus height and N-S versus height distribution of the trial trajectories, respectively. Subfigures c) and d) are shown with a c. 6 times exaggerated height scale. Figure 6.5e shows a map of the arrival time residuals, with blues representing negative and reds representing positive travel time residuals for the optimal geometrical model trajectory.

satellite trajectory was developed that includes the wind parameters. This approach is based on infraGA/GeoAc (Blom, 2014) to enhance the solution of the geometrical model. The program infraGA encompasses a ray tracer that allows the use of complex atmospheric weather models as input. We use a native functionality of infraGA that enables the simulation of supersonic trajectories and their associated arrivals from an approximated Mach wave at the Earth's surface (Blom et al., 2024). We apply an iterative damped direct search to identify new trajectory candidates that are then evaluated with the infraGA to achieve a so-called ray model. We use the best-fitting geometrical model trajectory solution as the initial guess to calculate rays that minimize the wind effects on our trajectory solution using the gridded 3-D wind model shown in Figure 6.4. This approach allows us to use: 1) a realistic atmospheric model, 2) curved ray-paths that adhere to the physics of atmosphere and can be bent by wind and velocity changes, and 3) trajectory geometries that allow for deceleration and bending of the trajectory

due to wind and gravity. By taking advantage of these features, we are able to overcome the restrictive assumptions of the geometrical model and model a more realistic re-entry trajectory of Starlink-2382 satellite and its associated A-S signals. One of the drawbacks, however, is that including the topography and thus the elevation of the A-S stations leads to an increase in computing time of roughly ten times. Because of that, we manually correct the wave arrival times for elevation, by adding the travel time of an inclined acoustic wave between the station elevation and sea level, similar to station corrections used in seismological applications (Kissling et al., 1995). To do this we use the PSO solution and calculate the Mach Cone for that trajectory. With the inclination angle of the nearest ray arrival to each station, we correct their elevation with the assumption that trajectory geometry will not change significantly between different trajectory evaluations. To better understand the wavefield generated by the downward bending trajectory of the Starlink satellite and the progressive opening of the Mach Cone as the satellite decelerates, we extended our analysis with a finite-difference (FD) wavefield modelling (short: FDM). This approach follows the methodology described by Föst et al. (in preparation) for meteoroid-induced acoustic and seismic signals. The wavefield of the last segment of the trajectory (c. 360 km) was modelled using the WAVE-Toolbox (Qin and Bohlen, 2022), applying a three-dimensional FD scheme to solve the elastic wave equation on a high-performance computing grid. The moving supersonic source was implemented as a sequence of time-shifted point sources along the precomputed trajectory, capturing the generation and evolution of the Mach Cone and its interaction with the Earth's surface. The time shift is calculated accordingly from the velocity and flight path of the object along the trajectory. For the point source function, we used a Fuchs-Müller wavelet (Fuchs and Müller, 1971), with a centre frequency of 0.05 Hz. The low frequency makes it possible to set the grid distance to 250 m avoiding grid dispersion effects. However, in this way we just model the lowest frequencies from the shock wave, which is sufficient enough to investigate the wave propagation and the first arrival behaviours. For the interpretation of the modelled signals this frequency limit should be accounted for. As velocity model, we used the atmospheric model described in section 6.2, excluding wind effects. To model the coupling into seismic waves, a solid Earth layer was included in the velocity model with a P-wave velocity of 4700 m s^{-1} and an S-wave velocity of 3000 m s^{-1} . Modelling parameters were selected based on grid stability conditions to ensure minimal numerical dispersion, following Föst et al. (in preparation). The used parameters and modelling details are provided in the supplement.

6.4. Trajectory and Ablation Modelling

To calculate the trajectory of the satellite we begin with calculating its speed. For that we follow the methodology of Campbell-Brown et al. (2013) and Cepplecha et al. (1998), where the speed is calculated based on the conservation of momentum:

$$\frac{dv}{dt} = \frac{\Gamma \rho_a v^2}{m} A \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}}, \quad (6.2)$$

with the shape factor A , the drag coefficient Γ , and the time dependent mass m of the satellite:

$$A = S \left(\frac{\rho_m}{m} \right)^{\frac{2}{3}}, \quad (6.3)$$

where ρ_a is the atmospheric density, v is the speed of the satellite, ρ_m is the density of the satellite, and S is the cross-sectional area of the satellite. For simplicity we assume that the satellite spins randomly while re-entering and that we can assume a spherical body, which is mostly made of aluminium (Boley and Byers, 2021) with a density of c. 2700 kg m^{-3} . As shown by Carter et al. (2011) the drag coefficient is Mach number dependent, and we use their derived functions to calculate the drag coefficient Γ dynamically:

$$\Gamma(M) = 0.5 \begin{cases} 0.45M^2 + 0.424 & \text{if } 0 \leq M \leq 0.722 \\ 2.1e^{-1.2(M+0.35)} - 8.9e^{-2.2(M+0.35)} + 0.92 & \text{if } M > 0.722 \end{cases} \quad (6.4)$$

where M is the Mach number. To account for ablation in our modelling we use the classical approach, by which mass loss can be described by ablation and fragmentation. For this we follow Campbell-Brown et al. (2013) and define mass loss as:

$$\frac{dm}{dt} = -\Gamma A (\sigma + \sigma_e) \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}} \rho_a v^3, \quad (6.5)$$

where the ablation coefficient is σ . The starting parameter for the ablation of $\sigma = 0.1 \text{ s}^2 \text{ km}^{-2}$ is chosen to reflect a complete mass loss, where the end particle mass is $< 1\%$ of the original mass. The satellite separated into multiple individual fragments during re-entry (see A.1, A.1), but because there is little detailed information available on this and especially the detached masses are unknown, we opt to model the Mach wave of the main satellite fragment. Because of this, we set $\sigma_e = 0$ to disallow further disintegration of the main fragment. The original wet mass of the satellite is 260 kg. Because the satellite was roughly halfway into its lifecycle and the dry mass is 248 kg, we assume that the satellite has a mass of at least 254 kg. We also assume that about 95% of the original mass should be accounted for along the portion of our trajectory. As the atmospheric density increases exponentially (and thus ablation increases) with decreasing altitude, we extrapolate the geometrical model solution parameters to a higher altitude of 85 km.

There the mass loss from ablation begins to significantly affect the satellite (Campbell-Brown et al., 2013). We subtract an additional estimated 10%, or about 25 kg of the satellite mass to account for those parts of the satellite which will break off in the beginning during re-entry, namely the solar panels. This estimation leads to an approximate initial mass of 220 kg of the main body before the start of our modelled trajectory. Additionally, we analyse available photos and videos of the re-entry to estimate fragmentation. In Supplementary Information S1 the observational points and in A.1 the identified fragments are listed. We find that the satellite separated into multiple fragments during its atmospheric flight path and identified one main fragment which seems to encompass about half of the original size. Thus, we assume a mass of c. 100 kg for the main fragment, to estimate the mass loss due to ablation. Furthermore, the satellite's distinct fragmented masses may have caused the precursor signal identified in Figure 6.2. Using equations (6.2-6.5) we determine gravity-, wind- and ablation-affected trajectories

of this main fragment and use it as input for the ray model. In contrast to the gravity and wind effects the ablation of the main fragment is additionally inverted for in the ray model. To fully define the modelled trajectory of a given iteration we define a starting point of the trajectory as well as an initial azimuth, elevation angle, and initial velocity (see Table 1). The solution of the geometrical model results in an averaged velocity for the main fragment of the satellite along the trajectory. For the ray modelling we use the starting velocity of 8 km s^{-1} , which is also in agreement with the velocity calculated from the TLE (see Table 1). In the ray modelling we fix the start time to 85 s after the reference time of 19:28 UTC using the event time gathered from the camera network FRIPON (2024) to reduce the number of free parameters by one. The inversion for the trajectory calculation contains not only the ablation σ , but also an additional parameter β which corrects the vertical velocity change due to a lift-like force, i.e., the Magnus effect, which is related to the possible uncontrolled spinning of the satellite (e.g., Elliott et al. 2024; Mancuso et al. 2024). This modifies eq. 6.2 and we obtain the following expression for our improved ray modelling:

$$\frac{dv_z}{dt} = -g - \frac{\Gamma \rho_a v v_z}{m} A \left(\frac{m}{\rho_m} \right)^{\frac{2}{3}} - \beta v_z, \text{ with } v = \sqrt{v_x^2 + v_y^2 + v_z^2} \quad (6.6)$$

6.5. Results

6.5.1. Trajectory

Using the ray model, including the effects of wind and a stratified atmosphere, we find the optimal trajectory solution with eq. 6.6 and the eight parameters defined as in Table 1 including ablation. This solution has a minimum arrival time residual value after evaluating approximately 4,500 trial trajectories. The arrival time residuals of the best-fit ray model trajectory (Fig. 6.7b) do not have the systematic north-south deviations with respect to the geometrical model trajectory (Fig. 6.5d). Instead, they seem to have a NW-SE (positive to negative to positive residuals) trend along the trajectory. The solution can be evaluated with a correlation analysis (see Figure 6.6a) which indicates that most parameter pairs are at most weakly related to each other, suggesting that most of the parameters are approximately uncorrelated. One exception is a positive association ($r = 0.71$) between start elevation and elevation angle θ of the modelled trajectory. Additionally, some moderate correlations are observed between elevation and velocity ($r = -0.31$) as well as elevation and the vertical lift parameter ($r = -0.33$). These correlations indicate that parameters with respect to elevation are partly coupled and hence cannot be resolved completely independently; however, their correlation is low except for θ (Figure 6.6a). The ablation parameter σ has near-zero correlations with the other parameters (maximum $|r|$ 0.17), which indicated that it contributes mostly independent information. The lift-parameter β has only weak-to-moderate correlations except for elevation, indicating its role as a nearly independent parameter in the modelling. The covariance structure (see Figure 6.6b) is consistent with the correlation patterns. Due to the different physical units of the parameters, there are scaling effects which complicate the

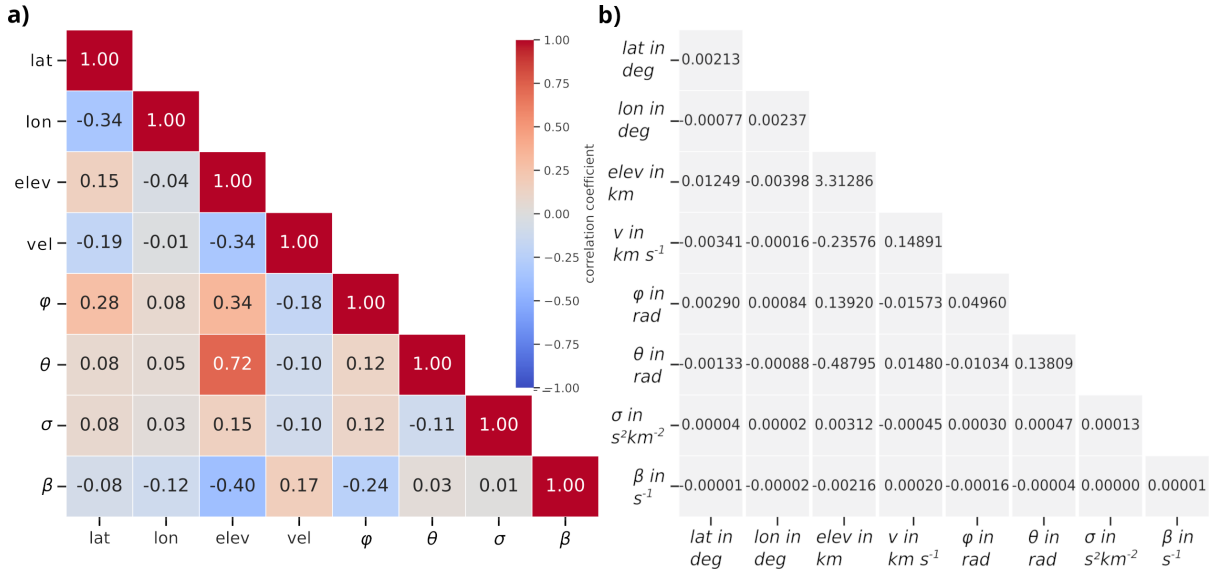


Figure 6.6.: a) Correlation matrix for the eight inverted parameters of the satellite trajectory (lat: latitude; lon: longitude; elev: elevation; vel: velocity; φ azimuth angle; θ elevation angle; σ : ablation parameter; β : lift parameter). b) Covariance matrix for the eight inverted parameters of the satellite trajectory. Units at the axes are the ones used for calculating the covariances. Note that due to the different units and scales of each parameter, the resulting off-diagonal covariances are difficult to interpret.

interpretation of absolute values. Again, the largest relative covariances involve elevation with a positive co-variation with elevation angle ($cov \approx 0.488$) and azimuth ($cov \approx 0.139$) and negative co-variation with velocity ($cov \approx -0.236$). In contrast, the ablation and lift parameters have near-zero covariances with respect to most other parameters, which indicates a largely independent resolution behaviour. The final average arrival time residual (RMSE), using the modified ray model approach, is 6.4 s, compared to 10.9 s using the geometrical method, achieving a c. 40% improvement of the fit. In addition, the ray model can explain 41 additional arrival times (altogether 202 data), using the same culling technique as for the geometrical model with discarding arrival time residuals exceeding 20 s, while achieving the lower RMSE. If we allow an RMSE of 10.9 s as for the geometrical model solution, the ray model can explain even 216 out of 226 arrival times. This difference corresponds to a 34% enhancement of usable input data in the ray modelling relative to 161 usable data for the geometrical model and results in a RMSE of 10.7 s. This comparison demonstrates the superior robustness of the ray model relative to the geometrical model due to the inclusion of wind, a stratified atmosphere and more realistic ray paths taking into account β and σ (for selected ray paths see Supplementary Information A.2). The spatial data distribution (see Fig. 6.7) and the ratio of 27 between data samples and parameters also indicates a stable and well-posed inversion problem.

Uncertainty bounds for the ray model solution are estimated based on the 500 best-fitting models with the lowest arrival time residuals. The distribution of the input parameters within this subset is analysed to derive the uncertainty bounds for each parameter. This approach assumes that models with lower misfits are more likely to reflect realistic parameter combinations. Because the input parameters are normally distributed, we use the standard

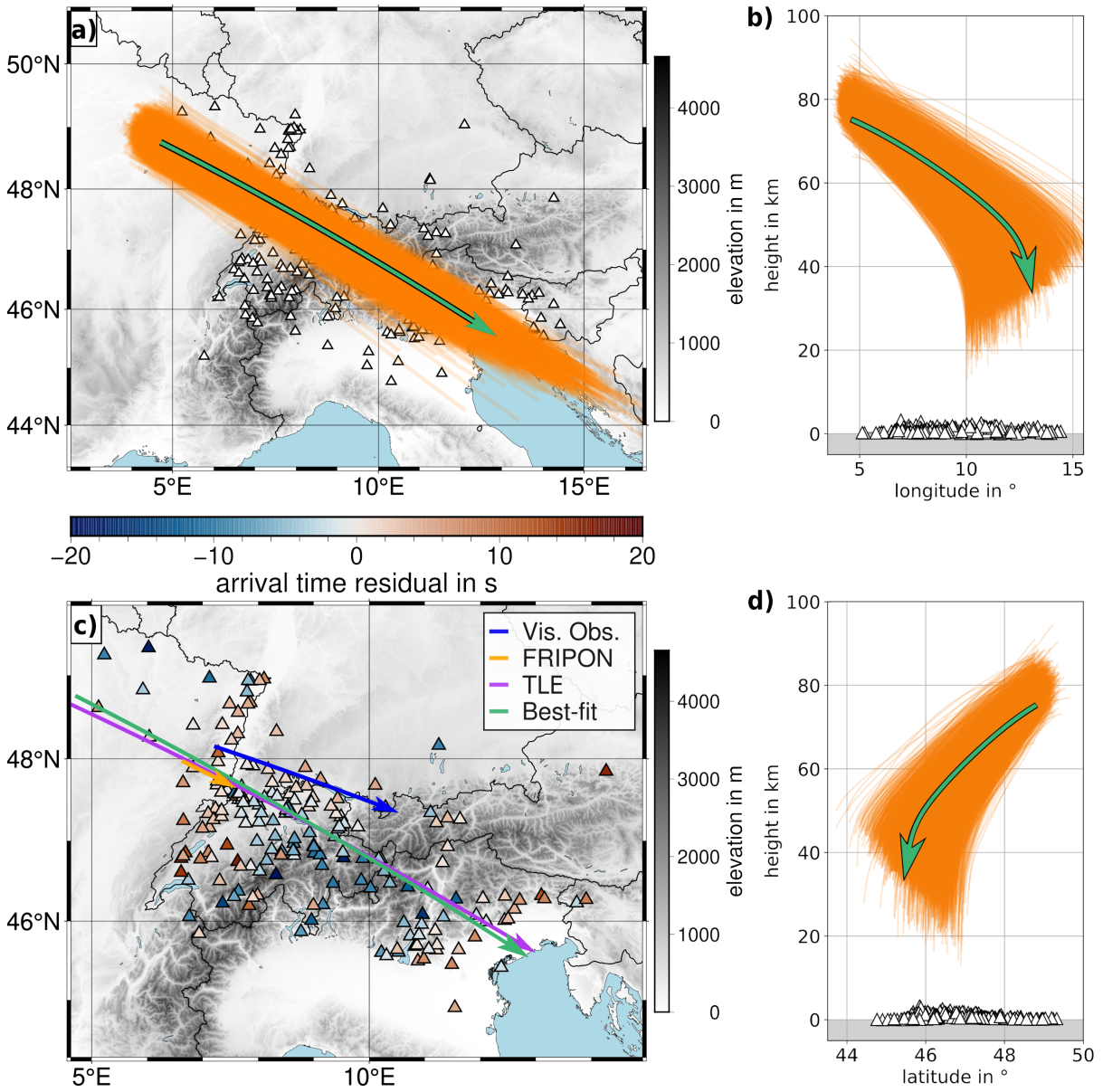


Figure 6.7.: a) Map showing different trial trajectories of the ray model reconstruction method in transparent orange. In green the best-fit ray model trajectory is shown. White triangles mark seismic and acoustic stations. This colour-coding of trajectories and symbology is the same in b) and d) as well. b) and d) show the E-W and N-S versus height distribution of trajectories, respectively (with a c. 6 times exaggerated height scale). c) map of the arrival time residuals. Blues represent negative and reds represent positive arrival time residuals for the best ray model trajectory.

deviation of the parameters to estimate the uncertainty bounds (see Table 1). Both A-S models, geometrical and ray, capture most of the length of the satellite re-entry trajectory, while both optical trajectory reconstructions, International Meteor Organization (IMO) 2024 and FRIPON (FRIPON (Fireball Recovery and InterPlanetary Observation Network) 2024; Colas et al. 2020), yield only short subsets of the satellite trajectory. The IMO trajectory most likely fails to cover the whole length of the re-entry trajectory, due to a limited number of observations at the edges

of the trajectory, while the FRIPON trajectory is clearly limited by the low number of active cameras (2; see FRIPON (Fireball Recovery and InterPlanetary Observation Network) 2024) and the very shallow elevation angle of the re-entry. Using the IMO and FRIPON trajectories, we evaluate their fit to our acoustic-seismic dataset. These calculations result in RMSE values of 2803 s and 2415 s for all 226 arrival times, respectively. Compared to our ray and geometrical models with RMSE around 10 s, we propose that the latter ones should be preferred, because they should better recover the whole length of the re-entry trajectory.

	IMO	FRIPON	TLE	Geometrical Model	Ray Model
Lon _{start} in °E	4.69 [*]	4.69 [*]	4.69 [*]	4.8	4.69 ± 0.08
Lat _{start} in °N	48.76 [*]	48.63 [*]	48.4 [*]	48.6	48.76 ± 0.06
alt _{start} in km	80	74.8 ^{**}	n/a	75.6	74.0 ± 3.2
v_{start} in km s ⁻¹	n/a	c. 10	c. 7.8 ^{***}	8	8.9 ± 0.6
φ in °	108.5	116	113	120	120.5 ± 0.4
θ in ° ^{****}	19.5	n/a	n/a	12	1.1 ± 0.7
σ in s ² km ⁻²	n/a	n/a	n/a	n/a	0.11 ± 0.02
β in s ⁻¹	n/a	n/a	n/a	n/a	0.02 ± 0.005

^{*} Lat/Lon values were matched to the start locations of best-fit trajectory.

^{**} At the start of trajectory (Figs. 6.1b, 6.5b).

^{***} From TLE data.

^{****} For IMO and the geometrical model the elevation angle is constant along the trajectory.

Table 6.1.: Comparison of trajectory parameters start longitude, start latitude, start altitude, start velocity, azimuth angle φ , elevation angle θ , and ablation coefficient σ for the different methods: International Meteor Organization (IMO) (2024), FRIPON (Fireball Recovery and InterPlanetary Observation Network) (2024), TLE (Satcat 2024), our Geometrical and Ray models.

6.5.2. Finite difference modelling of wave propagation

We use 3-D acoustic and seismic wavefield modelling to analyse the effects of deceleration and the resulting bending of the trajectory due to gravity with our best-fit ray model trajectory (Fig. 6.7c, 6.7d). Because the angle with which the acoustic Mach waves are radiated from the satellite is inversely related to the satellite velocity, a decrease in satellite velocity leads to an opening of the Mach Cone (Sasoh, 2020). We can observe this change in opening angle in the ray-, as well as the wave-based approach (see Figs. 6.6b, 6.6c, 6.8). Modelling details are given in Supplementary Information A. As expected, the FDM results reveal that as the satellite slows down, the Mach Cone opening angle increases, consistent with the theoretical expectation that the angle depends inversely on the Mach number. The synthetic wavefield snapshots (Figures 6.8a - d) illustrate the broadening of the cone and the resulting wider ground

footprint of the direct acoustic wave. At 173.8 s (Fig. 6.8a) there is still a near cylindrical Mach Cone wave front with a narrow opening angle. After about 213.8 s the satellite reaches the end of the trajectory, see red dots in Fig. 6.8a-d, where most of the material is ablated and the velocity reaches subsonic speeds. The modelled wavefield confirms that the trajectory bends downwards more sharply during the final re-entry phase due to increased drag and deceleration, as the shock front weakens and the supersonic Mach carpet (see A.3) decreases in size. The combination of the downward bend and the opening leads to an asymmetric Mach Cone shape. After 312.8 s this bending results in a shift of the first ground coupling area away from the end of the trajectory towards the bend direction (Fig. 6.8c). Simplified trajectory estimations (e.g., Pujol et al. 2005) ignore this effect which gets quite significant depending on the incidence angle and velocity of the object. Figure 6.8d shows the produced seismic signals which travel much faster than the acoustic waves. This modelling supports the trajectory reconstruction and explains arrival time discrepancies that arise, if Mach Cone opening is neglected as well as the complexity of the wavefield.

Figure 6.9 summarizes how the FDM compares with the ray model for the best-fit trajectory. In Fig. 6.9a the coloured triangles represent the ray-traced arrival times while the colored circles/area show the FDM arrivals sampled on a regular grid. Both methods reproduce the characteristic Mach carpet pattern of arrival times, confirming that the overall source–receiver geometry is correctly captured. A slight, nearly uniform negative time shift of the FDM arrivals with respect to the ray model solution is evident, which we attribute primarily to the resolution limit of the FDM. Figure 6.9b maps the residuals (twave – tray) between the wave and the ray models. Despite the fundamentally different nature of the two approaches, the resulting residuals remain small (typically within 1 - 2 s, with larger residuals in the area of the over-the-top Mach Cone arrivals) but become increasingly negative toward the eastern, terminal part of the path. This difference may be due to numerical resolution. In the ray model, which is a linear, high-frequency approximation, the Mach-cone opening angle is determined by the Mach number along the trajectory. In contrast, the finite-difference simulation, which is also linear in acoustics, forms the Mach cone wavefront as a superposition/interference pattern from distributed source points along the path, while propagating the full 3-D wavefield explicitly on the numerical grid. With a near-surface sound speed of 344 m s^{-1} , a 1 s travel-time residual corresponds to about 344 m of path length, i.e., approximately 1.38 times the receiver grid spacing (250 m). This places the typical misfit close to the effective resolution limit of the wave simulation. Overall, the small residuals indicate that the ray model can accurately capture the far-field travel times of the sonic boom in the linear acoustic regime. However, close point-by-point comparisons are sensitive to modelling choices, such as source discretization, receiver spacing, and the imposed versus emergent representation of the Mach cone wavefront, particularly in regions dominated by complex wave arrivals.

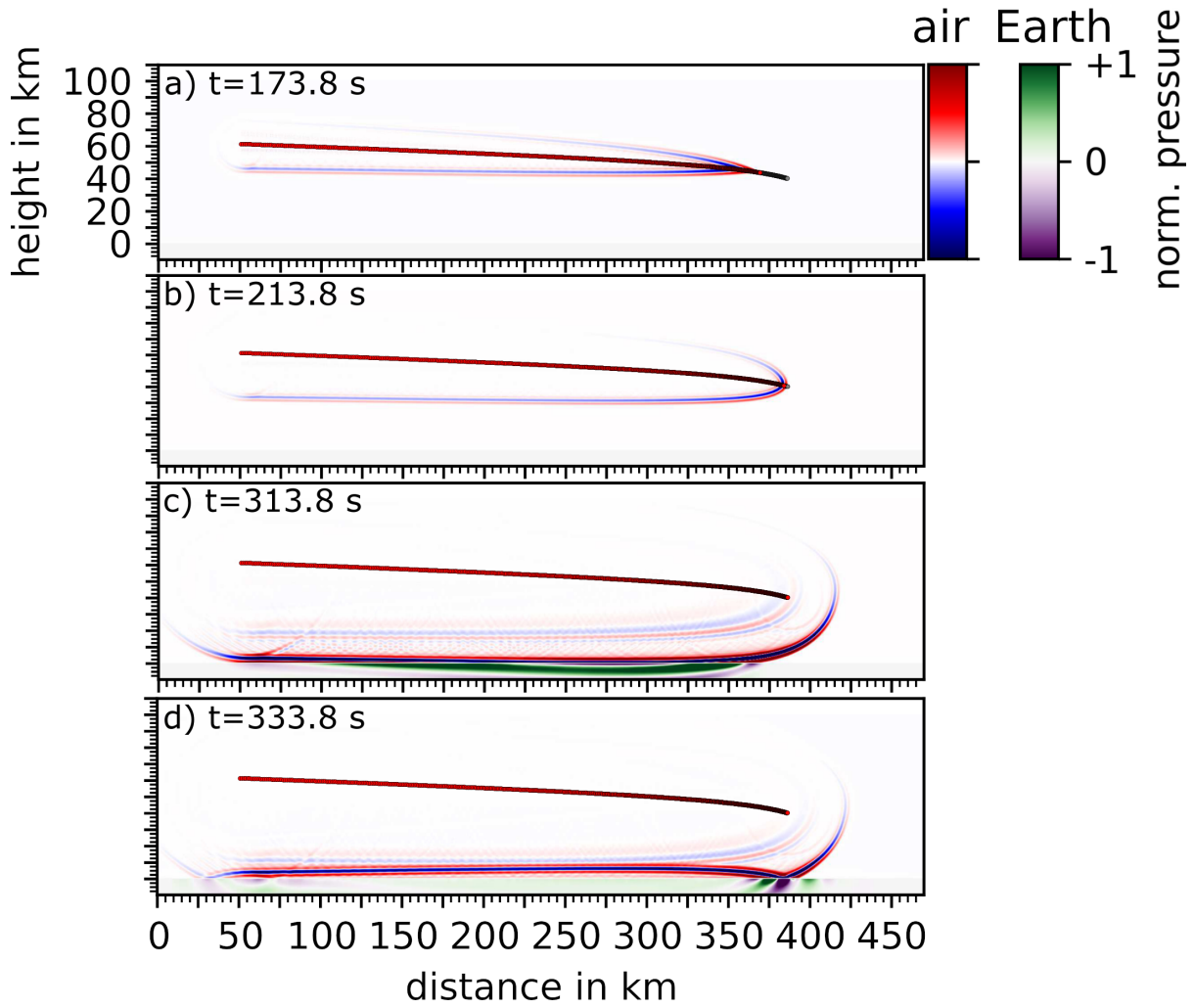


Figure 6.8.: Time evolution of the wavefield generated by the Starlink Satellite while traveling through the atmosphere, here in a vertical plane along the trajectory (distance vs. height). Normalized pressure is visualized at four time snapshots. The time t is relative to 19:28 UTC: a) $t = 173.8$ s: the object travels at high altitude with a narrow Mach Cone, indicating supersonic speed. b) $t = 213.8$ s: the trajectory curves downward and deceleration begins, resulting in a broader Mach Cone. c) $t = 313.8$ s: the broadened Mach Cone leads to the first acoustic coupling with the Earth's surface, occurring away from the trajectory end point. The acoustic wave arrives after 312.8 s at the surface. d) $t = 333.8$ s: seismic waves travel through the Earth. Red dots mark the already triggered source points along the trajectory (flightpath and position). White dots mark the following source points (future trajectory).

6.5.3. A-S Ground Coupling

To estimate the minimum ground compliance K_v , we follow the approach by Fernando et al. (2024b) and Kenda et al. 2020 for six collocated A-S station pairs (see Table 2, Fig. 6.10). K_v is given by:

$$K_v = 2c \frac{1 - v^2}{E}, \quad (6.7)$$

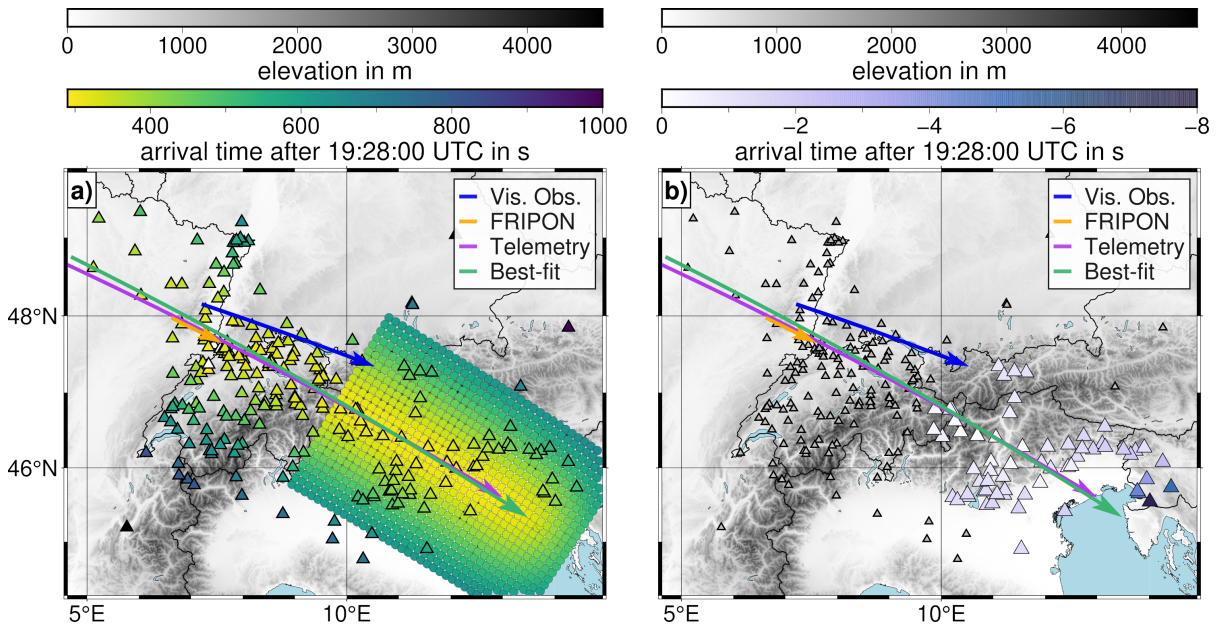


Figure 6.9.: a) Comparison between the ray model arrival times (coloured triangles) and the FDM arrival times (coloured area). For this model comparison a vertically layered atmosphere model without wind is used. The arrival pattern from the Mach cone is well reproduced by the modelling, although slightly shifted, compared to previous figures because of the deliberate omission of wind effects. b) Arrival time difference between ray and FDM (bluish shading). There is a small systematic offset, due to the fundamental difference between high-frequency ray approximation and low-frequency FDM. The arrival time differences towards the end of the trajectory show that the ray modelling might not be able to account for the full complexity of the wavefield due to the Mach Cone opening (Fig. 6.8). Trajectories as in Fig. 6.7d.

where E is the Young's modulus, ν is the Poisson ratio, and c is the advection speed of the pressure loading. To choose the elastic properties for the station pairs we separate them into two groups: case 1) unconsolidated soft rock, which is found in the Molasse region north of the Alps and case 2) consolidated hard rock, which is found inside the mountain range of the Alps.

Following Fernando et al. (2024b), we use a speed of the local sound at the Earth's surface of 343 m s^{-1} . Using eq. 6.6, the values of Table 6.2, and the measured peak ground velocity amplitudes (Fig. 6.10) of the vertical components of the respective seismograms we can derive the minimum A-S ground coupling coefficient (c_{AS}) for each station site. We find that the c_{AS} values range between $0.98 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$ and $5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$ for soft rock sites and between $4.31 \times 10^{-10} \text{ m s}^{-1} \text{ Pa}^{-1}$ and $3.96 \times 10^{-8} \text{ m s}^{-1} \text{ Pa}^{-1}$ for hard rock sites. Thus, at unconsolidated, soft rock sites, there seems to be a 1 - 3 order of magnitude increased amplitude excitation. This agrees well with the results of Fernando et al. (2024b) and Edwards et al. (2007) who calculated minimum c_{AS} values of $1.4 \times 10^{-6} \text{ m s}^{-1} \text{ Pa}^{-1}$ and $7.3 \pm 0.2 \times 10^{-6} \text{ m s}^{-1} \text{ Pa}^{-1}$, respectively, for even softer material at the Earth's surface. Novoselov et al. (2020) found c_{AS} values of $1.99 - 2.74 \times 10^{-6} \text{ m s}^{-1} \text{ Pa}^{-1}$ for water-saturated limestone-rich material.

Knowledge about the acoustic-to-seismic coupling efficiency at seismic stations can help to constrain the masses of future events (ReVelle 1976; Edwards et al. 2008) where a priori

Surface Material	Station pairs	E in GPa	ν	K_v in $\text{m s}^{-1} \text{Pa}^{-1}$ at $c = 343 \text{ m s}^{-1}$	c_{AS} in $\text{m s}^{-1} \text{Pa}^{-1}$
Unconsolidated soft rock	AM.R2F1D & CH.DAGMA; AM.R03EA & CH.ZUR	4	0.25	1.6×10^{-7}	$0.98 \times 10^{-7} - 5.86 \times 10^{-7}$
Consolidated hard rock	AM.R489A & CH.SARD; AM.R54F9 & OX.BOO; AM.R06CE & CH.SGRA; AM.RCDB6 & CH.SUSI; AM.RA038 & OE.WTTA	55	0.2	1.2×10^{-8}	$4.31 \times 10^{-10} - 3.96 \times 10^{-8}$

Table 6.2.: Parameters to estimate ground compliance.

information of the object mass such as for meteoroid events is unknown and infrasound is not available.

6.5.4. Ablation estimation

The unique Starlink-2382 dataset can be used to study the mass loss (see equation 6.5) of the satellite during the re-entry to investigate the premise that it should fully burn up. SpaceX targets a kinetic impact energy of $<3 \text{ J}$ (SpaceX, 2025). As we use acoustic wave generation, we chose a slightly different metric in determining a full burn up of the 100 kg main fragment. We consider the satellite to be fully burned up, as soon as it falls below 1% of the original mass which is 2 kg in mass as per our modelling. For infrasonic data it can be assumed that objects with masses $>1 \text{ kg}$ are reliably visible on acoustic recordings (ReVelle 1976), to account for the seismic stations we adjust our lower limit mass threshold to 2 kg. However, a 2 kg fragment at terminal velocity of $c. 90 \text{ m s}^{-1}$ would have a kinetic energy of $8,100 \text{ J} \gg 3 \text{ J}$. This means that even if the satellite loses 99% of its original mass, it is still possible that some material survives to the end of the modelled satellite fragment trajectory, at around 40 km altitude. For the modelled best-fit trajectory the main satellite fragment reaches this threshold after 180 s after the reference time of 19:28:00 UTC. After $c. 200 \text{ s}$ the mass of the satellite diminishes

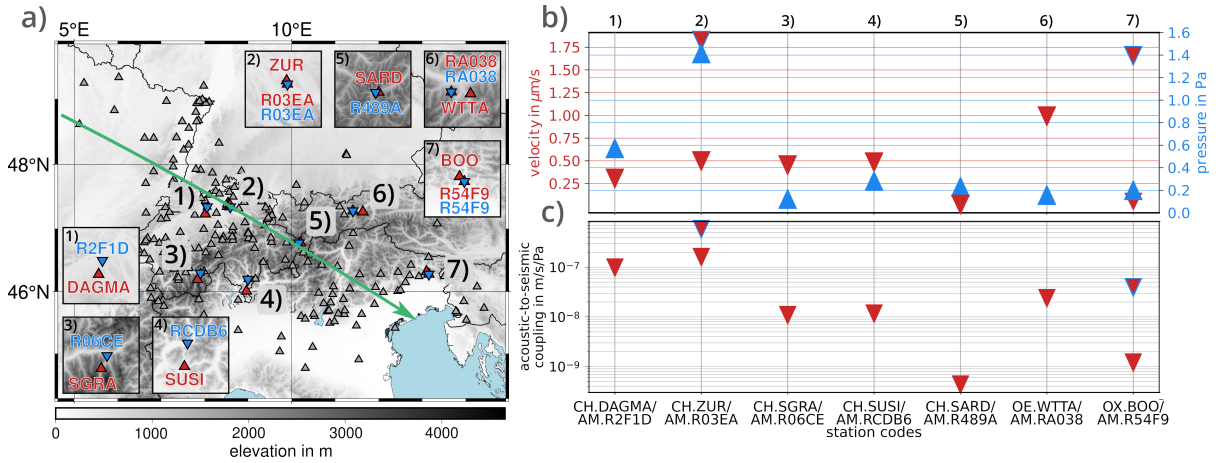


Figure 6.10.: a) Map of collocated A-S recording sites numbered from 1 to 7 with detail-view in related inset plots. Seismic stations are indicated with red triangles and infrasonic stations with inverted blue triangles. b) Diagram of measured seismic ground velocity (red triangles) and acoustic overpressure (blue inverted triangles). Raspberry Shake & Boom stations with both, seismic velocity and acoustic overpressure measurements, are marked with blue-framed red triangles. The station pairs are ordered according to the respective origin time of the associated acoustic wave. c) Diagram of acoustic-to-seismic coupling coefficients for each recording pair. The colour-coding is the same as in b).

to a mass of ≈ 1.5 kg (see Fig. 6.11a). Considering the uncertainty for estimating the ablation coefficient σ , we find that the lower end of our retrieved range, i.e. $\sigma < 0.09 \text{ s}^2 \text{ km}^{-2}$ produces a terminal mass > 2 kg. Therefore, it is yet inconclusive if the satellite fully burned up from our modelling alone. Nevertheless, we consider the occurrence of satellite fragments with masses > 2 kg to be unlikely, because:

1. the use of a purely spherical body is too simplistic, a more irregular rectangular shape would increase the drag coefficient, resulting in a greater mass loss for the same ablation coefficients,
2. as the re-entry was uncontrolled, it might make sense to try to model an erratic tumbling motion of the main satellite fragment which also increases its drag coefficient and mass loss. This is indirectly covered with the introduction of the lift parameter β , that is based on the spinning motion of the satellite fragment. While higher β values decrease the mass loss over time, its value has negligible effect on the final mass of the satellite fragment (see Fig. 6.11c),
3. video material of the re-entry shows an increase of individual fragments with time during the re-entry (Table A.1 in SI), which means that neglecting fragmentation of the satellite fragment might be too simplistic. Even though the mass loss attributed to the break-off of the solar panels is factored in already and we deliberately model only the main mass, we cannot be certain that not more material is subject to breaking off from the main mass during re-entry. This would decrease the mass of the main body that is most likely generating the acoustic waves that we analysed in this study. A mass decrease would in turn then result in a more likely burn-up with the same ablation coefficient.

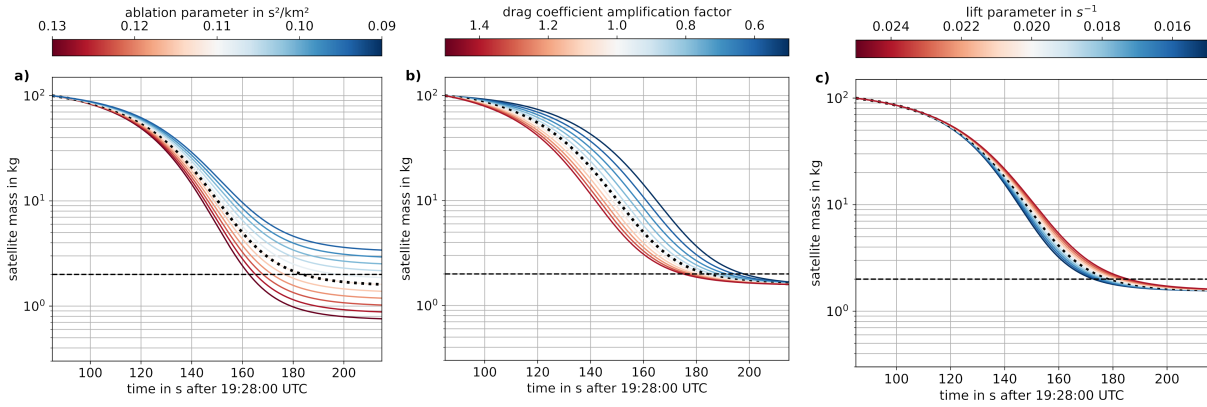


Figure 6.11.: a) Influence of the ablation parameter σ on the mass of the satellite fragment versus time. The different colours represent the ablation parameter within the error bounds (shown in Table 1). The dotted line represents the mass over time behaviour of the best-fit run with $m = 100$ kg, $\sigma = 0.11$ s² km⁻², and a lift parameter $\beta = 0.02$ s⁻¹. The horizontal dashed black line represents the cutoff mass of 2 kg. Lighter line colours represent closer values to the best-fit ray model trajectory. This representation is the same in b) and c). b) shows the influence of the drag coefficient on the satellite fragment mass-time distribution. Different coloured lines show different drag coefficient amplification values. c) shows the influence of the lift parameter β on the satellite fragment mass-time distribution. Different coloured lines show different lift parameter values.

Due to 1) the best-fit ablation parameter should be viewed as an effective ablation parameter for the Starlink satellite. The true ablation parameter can only be derived from modelling, like the one presented in this study, if the shape and composition of the satellite is known and accounted for more thoroughly. Figure 6.11 shows the influence of the ablation parameter σ , the drag coefficient Γ and the lift parameter β on the satellite fragment mass over time based on the best-fit ray model trajectory. To simulate different drag coefficients, we vary these by $\pm 50\%$ and find that this results in ca. ± 15 s time for burn-up. To estimate the effect of the lift parameter on the mass loss of the satellite fragment, we simulate lift parameter values within the uncertainty bounds between 0.015 s⁻¹ and 0.025 s⁻¹. We find that an increased lift parameter results in a decreased mass loss rate along the trajectory. This is due to the increased travel time through the less dense higher atmosphere, when the satellite fragment experiences the additional lift. The estimated best-fit effective ablation coefficient value of 0.11 s² km⁻² is within the upper range of values typically found for natural meteoroids with typical ablation coefficients between 0.014 s² km⁻² and 0.19 s² km⁻² (Ceplecha et al., 1998). Table 3 provides a comparative analysis. Starlink-2382 displays effective ablation coefficient values that are an order of magnitude higher, peaking at approximately 0.11 s² km⁻² compared to the Chelyabinsk and Neuschwanstein meteoroids at c. 0.034 s² km⁻² and 0.018 s² km⁻², respectively. It aligns better with meteoroids with higher ablation coefficient values and low bulk densities like e.g. meteoroid 39048 (see Table 3). This is consistent with the design intent of Starlink satellites to fully disintegrate upon re-entry (SpaceX, 2025), suggesting a higher surface area-to-mass ratio and greater susceptibility to drag-induced ablation. Additionally, the simplifying assumptions in our model, e.g., spherical symmetry, likely underpredict the true mass loss. This can be further illustrated by the comparison of the Starlink satellite to the Hayabusa sample return capsule, which was engineered for controlled re-entry and landing, with its low ablation coefficient of only 0.0014 s² km⁻² (Borovicka et al., 2011).

Event	Initial mass in kg	Initial velocity in km s ⁻¹	Ablation coefficient in s ² km ⁻²
Starlink-2382 main fragment (this study)	c. 100	c. 10.5	0.11 ± 0.02
39048 ¹	c. 12	22.143	0.19
Chelyabinsk meteoroid	1.3e7 ± 1e2 ²	19.16 ± 0.3 ²	0.034 ³
Neuschwanstein meteoroid ⁴	600	21	0.018 ± 0.002
Morávka meteoroid ⁵	1500	22.5	0.003
Hayabusa sample return capsule ⁶	20	12	0.0014 ± 0.001

Table 6.3.: Comparison of ablation coefficients σ between re-entry events and meteoroids. Sources: ¹ Ceplecha et al. 1998, ² Popova et al. 2013, ³ Trigo-Rodríguez et al. 2021, ⁴ Spurný et al. 2003, ⁵ Borovička and Kalenda 2003, ⁶ Borovička et al. 2011

6.6. Discussion

The reconstruction of the Starlink-2382 re-entry event trajectory is significantly improved by the ray model incorporating atmospheric stratification, wind effects, the ablation of the satellite and a vertical lift parameter β . In comparison with the geometrical model (Pujol et al. 2005; Vida 2023), the application of the ray model achieves a reduction in RMSE residuals of approximately 40% (from 10.9 s to 6.4 s), whilst simultaneously expanding the number of usable arrival times by over 30% (see Table 1). This enhancement is of particular significance when considering the complex acoustic propagation environment encountered during re-entry, where assumptions of a uniform atmosphere or linear ray paths frequently result in systematic misfits, especially for flat-angle incident objects. The incorporation of the β parameter, a vertical perturbation term, yields a trajectory more consistent with the wave arrivals observed at the surface. The parameter helps to mitigate residual trends along the trajectory. Although its physical origin remains partly uncertain, β may reflect aerodynamic lift or torque-related forces such as the Magnus effect arising from satellite spin (Elliott et al. 2024; Mancuso et al. 2024). Because rotational state and torque are unknown to us and thus not explicitly modelled here, β should be interpreted as an effective parameter. Future investigations could focus on isolating this effect through controlled experiments or additional modelling of the rotational effects. The trajectory uncertainty analysis, based on 500 models, provides a stable best-fit and its narrow range further supports the robustness of the ray model result within the assumptions and trajectory parameterization. As demonstrated in Table 6.1, the initial parameters, including longitude, latitude, and velocity, exhibit a high degree of convergence with low uncertainty. This finding suggests that infrasound and seismic data, when integrated with realistic trajectory physics, has the capacity to provide precise constraints on initial re-entry conditions. Our finite-difference wavefield simulations confirm the ray-based modelling and demonstrate that the downward bend in the final trajectory phase, combined with the Mach Cone opening, results in a complex acoustic wave arrival pattern (Figs. 6.8, 6.9). This provides a plausible

explanation for the spatial mismatch between expected and observed arrivals when static Mach angles are assumed. It also provides a physical explanation for several anomalous arrival times, especially at the end of the trajectory. However, these interpretations should be regarded as indicative rather than definitive because the current finite-difference simulations cannot include all relevant physical processes, such as wind affecting acoustic wave propagation and full 3-D thermodynamic variability. Incorporating these processes into future models is important for quantifying their relative contributions and avoiding systematic biases in source location estimates for artificial and natural atmospheric entry events, particularly at low entry angles. Our analysis of acoustic-to-seismic coupling across seismic stations reveals strong site-dependent variability in the acoustic-to-seismic ground coupling coefficient c_{AS} . Using elastic parameters and peak ground motion amplitudes (Table 2), we estimate c_{AS} values ranging from $0.98 \times 10^{-7} - 5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$ on soft, unconsolidated rock material, compared with significantly lower values ($4.31 \times 10^{-10} - 3.96 \times 10^{-8} \text{ m s}^{-1} \text{ Pa}^{-1}$) on hard rock. These findings are consistent with previous studies ((Fernando et al., 2024b); (Edwards et al., 2007); (Novoselov et al., 2020)) and emphasize the role of surface geology in modulating signal amplitude. This variability must be accounted for when using seismic amplitudes to infer source parameters such as energy or mass (ReVelle 1976). Ignoring local geological effects at a seismometer site may introduce large systematic errors, particularly for events lacking infrasound or optical data. While not explicitly observable in our data, the c_{AS} values are possibly frequency- and azimuth-dependent. To better understand these coupling effects, additional studies are needed, however. Our modelling indicates an effective ablation coefficient for Starlink-2382 that is one to two orders of magnitude higher than for typical meteoroids. This relatively high value is consistent with the satellite's deliberate design to disintegrate efficiently during re-entry. However, our modelling indicates that some material could have survived into the late stages of the modelled trajectory, and our rough kinetic energy estimate may be orders of magnitude above the upper limit reported by SpaceX (SpaceX 2025). However, this comparison is highly sensitive to our modelling assumptions and should be treated with caution. Although our modelling cannot confirm complete disintegration, the high effective ablation coefficient supports the likelihood of a 99% mass loss of the satellite during re-entry (Fig. 6.11). Furthermore, our use of a simplified spherical geometry likely underestimates atmospheric drag and mass loss by ablation. Better estimates of satellite-specific geometry, tumbling behaviour, and fragmentation, would likely increase the effective ablation rate. These results emphasize the necessity of incorporating object-specific physical characteristics in ablation models. As an artificial satellite can reach the upper values of the expected ablation efficiency of natural meteoroids, this implies that meteoroid modelling assumptions cannot be universally applied to artificial re-entry events. This distinction is crucial for future atmospheric entry reconstructions involving space debris.

6.7. Conclusions

This study presents a refined reconstruction of the atmospheric trajectory and acoustic wave generation related to the re-entry of the Starlink-2382 satellite in August 2024. The high seismic station density in Central Europe and the low elevation angle with a downward

wave emission both lead to a very high number of 226 usable observations which allows a well-constrained trajectory reconstruction approach. Our preferred reconstruction is based on a novel ray-based iterative damped direct search approach using the *infraGA* software (Blom, 2014). Incorporating a 3-D atmospheric wind model, ablation to model the time-dependent mass, and a lift-like force enables us to achieve a significantly better fit to acoustic and acoustic-to-seismic coupled arrival times compared to geometrical trajectory modelling. The ray model reduces average arrival time residuals by 40% compared to the more commonly used geometrical model approach (Pujol et al. 2005; Vida 2023) with linear ray geometries. The more realistic 3-D ray path modelling also fits more data (34% in our case with a cut-off at 20 s residual), enabling a more robust and comprehensive trajectory estimation. In this specific case the acoustic-seismic trajectory reconstruction yields a better fit to the measured data than optical trajectory estimates. Finite-difference simulations of the emitted acoustic wavefield provide an independent benchmark for the dynamic ray-based reconstruction of the trajectory. Consistency between both approaches indicates that the ray model can accurately capture the far-field travel times of the sonic boom in the linear, acoustic regime. In this context, the shadow zones, which can arise in *infraGA* due to ray approximations and low-velocity layers, do not appear to significantly impact the results. The wavefield simulations also reveal the Mach Cone opening and the influence of the trajectory bending. These effects can significantly affect acoustic arrival patterns, and their travel time anomalies must be incorporated into future models to avoid a mislocation of extra-terrestrial sources or a misinterpretation of the related phase arrival times. However, an acoustic wavefield modelling including wind is currently not feasible, due to the lack of modelling capabilities at present. The bending during the terminal flight, often without light emissions and thus called dark-flight phase, is especially important for objects with a low entry angle and could be observed with acoustic and seismic measurements. Such objects, i.e., shallow-entry meteoroids do not completely burn up and can thus be picked up at the Earth's surface. Nevertheless, correct modelling using the acoustic rays can narrow down the strewn field, optimizing search areas to recover precious meteorite material. This is particularly relevant for cosmic studies, such as astrobiology, where the swift recovery of meteorites is crucial for studying extraterrestrial amino acids (e.g., Glavin et al. 1999) or rare isotopes before terrestrial weathering begins. Our determined effective ablation coefficient of $0.11 \pm 0.02 \text{ s}^2 \text{ km}^{-2}$ corresponds to the upper end of values expected for typical meteoroids. Such a high value is consistent with an intentional design that allows for substantial burn-up upon re-entry. Since we chose to model a spherical body, so that we could apply Mach number dependent drag coefficients, the ablation coefficient remains an effective ablation coefficient for the modelled satellite fragment. To find the true ablation coefficient of the satellite more detailed studies must be made. More such studies should be done in order to better calibrate the determination of ablation for meteoroids with acoustic-seismic data for which the mass, geometry, and material properties are not known. The found variation in the coupling coefficient of $4.31 \times 10^{-10} - 5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$ depends on the local surface geology of the recording site. Of course, there remain uncertainties, particularly in how coupling efficiency depends on wavefront incidence angles and azimuth. To improve the use of amplitude information from measured extraterrestrial material, e.g. to estimate their energy or mass, more controlled-source experiments could be carried out to study the azimuth dependency of the ground coupling coefficients. More acoustic sensors at seismic sites should be used for the determination of empirical coupling coefficients at selected sites. A significant

advantage of our studied event is that the initial mass of the Starlink-2382 satellite is precisely known (approximately 260 kg in a fully fuelled state) in contrast to natural meteoroid entries. This unique opportunity allows us to calibrate models of acoustic energy generation and mass loss against a controlled reference, however the fragmentation (see Supplementary Information A.1, A.1) poses an uncertainty in the mass parameter. The increasing number of satellites in low Earth orbit is likely to result in a higher frequency of uncontrolled re-entry events of similar nature. Overall, a future challenge is the calibration of seismic stations for consistent retrieval of source energy or mass of extra-terrestrial objects with acoustic waves. Signal amplitudes are affected by surface geology, atmospheric path effects, wavefront geometry, and coupling dynamics. This study establishes a benchmark for future calibration efforts of this kind and demonstrates the value of combining multidisciplinary observational data with advanced modelling techniques.

7. The Koblenz Meteoroid from Orbit to Strewn Field with Acoustic-Seismic Data

The following section describes the end-to-end analysis of the recent Koblenz meteoroid fall. It demonstrates how acoustic-seismic data from 95 stations can be used to constrain the trajectory in a statistically robust way with Bayesian Inference. It allows to carry uncertainties downstream to the derived meteorite ground locations and the pre-Earth orbit. This is enabled by the use of a Lookup-table approach to speed up the trajectory inversion. The study shows how rapidly after a given meteoroid event, acoustic-seismic data can be used to derive the meteoroid's trajectory, strewn-field and also orbit. The following section is submitted as (Eickhoff and Ritter, submitted).

My contributions to this study, following CRediT, are:

1. Conceptualisation,
2. Data Curation,
3. Formal Analysis,
4. Investigation,
5. Methodology,
6. Software,
7. Validation,
8. Visualisation,
9. Writing: original draft

7.1. Abstract

Every year, thousands of meteoroids enter Earth's atmosphere, but recovering the resulting meteorites before they start weathering depends on the rapid reconstruction of where the fragments fell. Mostly, optical fireball networks are used, yet these are limited by clouds, daylight, and coverage gaps. Continuously operating seismic and infrasound networks offer a time- and weather-independent alternative for recording the shock waves generated during

supersonic flight and fragmentation. However, existing acoustic-seismic methods typically yield single best-fit trajectories and rarely fully account for atmospheric winds or propagate observational uncertainties. Here we demonstrate that probabilistic Markov chain Monte Carlo inversion can reconstruct the trajectory and five distinct fragmentations which characterise the disintegration of the 8 March 2026 Koblenz meteoroid, recorded at 95 stations. Additionally, we recover the meteoroid's heliocentric orbit and strewn field, incorporating the full wind field for modelling the acoustic wave propagation and propagating uncertainties into ensemble-based impact ellipses that are validated directly against meteorite fragments that struck a building in Koblenz-Güls, Germany. This demonstrates that the analysis of regional near-real time available seismic recordings can deliver statistically rigorous meteoroid fall reconstructions. Our approach paves the way toward automated, rapid meteorite recovery and routine characterisation of meteoroids similar to automated earthquake detection.

7.2. Introduction

Recovering fresh extraterrestrial material is a race against time. Weathering begins as soon as a meteorite impacts the Atmosphere (Bland and Artemieva, 2006), which means that finding fragments before they degrade depends on quick and accurate reconstructions of the meteoroid's atmospheric trajectory. While optical fireball networks are the established tool for this purpose, they are rendered ineffective under cloud cover, during the day and in areas with sparse camera coverage (Moilanen et al., 2021; Vida et al., 2021; Devillepoix et al., 2019; Colas et al., 2020; Borovička et al., 2015). Continuously operating acoustic-seismic monitoring overcomes these limitations by providing time and weather-independent recordings available in near-real time (Edwards et al., 2008; Eickhoff et al., 2026b). This approach is based on the acoustic waves produced by a meteoroid during supersonic flight, which form a cylindrical Mach cone along its trajectory and quasi-spherical fragmentation events where shear stresses exceed the meteoroid's structural strength (ReVelle, 1976; Edwards et al., 2008; Silber and Brown, 2014; Silber et al., 2018). These signals reach the ground as infrasound in the 0.01–20 Hz range, which is recorded by microbarometers and as air-to-ground coupled seismic waves measured by seismometers (ReVelle, 1976; Edwards et al., 2008; Langston, 2004; Pujol et al., 2005). Seismic networks are particularly valuable because the stations tend to be densely and evenly distributed. This provides reliable arrival time measurements for ground-coupled phases and helps to determine fragmentation timing even when acoustic monitoring coverage is sparse or infrasound backazimuths are distorted by atmospheric winds (Edwards et al., 2008, 2009; Silber, 2025). Meteoroid trajectories have been modelled using seismic records within a few degrees of uncertainty for several meteorite falls, including Morávka (2000), Neuschwanstein (2002), Villalbeto de la Peña (2004), Jesenice (2009), and Košice (2010) (Brown et al., 2003; Borovička and Kalenda, 2003; Revelle et al., 2004; Llorca et al., 2005; Spurný et al., 2010; Borovička et al., 2013; Kalenda et al., 2014). Combining seismic arrival times with infrasound backazimuth estimates further refines fragmentation heights and source energetics (Brown et al., 2003; Vera Rodriguez et al., 2023; Stich et al., 2022; Hicks et al., 2023). Yet existing methods (Pujol et al., 2005, 2006; Edwards and Hildebrand, 2004; Ens et al., 2012) primarily provide single best-fit solutions, mostly without fully accounting for wind during

modelling (Vera Rodriguez et al., 2023). Here we present and apply a rigorous framework for reconstructing trajectories, propagating uncertainties and predicting probabilistic strewn fields using acoustic-seismic data that takes into account the full wind field, as well as the stratified temperature of Earth’s atmosphere at the time of the fall. The exceptional fall of the Koblenz meteoroid over Central Europe on 8th of March 2026 at approximately 17:55 UT was recorded at 95 seismic stations (see Fig. 7.1), capturing both Mach cone arrivals and discrete fragmentation sequences with unusual clarity (see Fig. 7.1, Fig. 7.2). Low wind conditions throughout the lower 20 km of the atmosphere minimised acoustic distortion (see Figs. 7.1b and 7.3; B.2). Meteorite fragments struck a residential building, providing rare, precisely geolocated ground truth for constraining the terminal trajectory endpoint. The recovered fragments were identified as possibly being eucrite meteorites (Meteorite Studies, 2026). We use a Markov Chain Monte Carlo (MCMC) trajectory inversion framework (Foreman-Mackey et al., 2013) that, for the first time, enables comprehensive statistical analysis of meteoroid trajectories from acoustic-seismic observations, incorporating reanalysed atmospheric profiles and winds (see 7.7). This methodology propagates observational uncertainties into ensemble-based strewn field estimates, which we validate directly against the documented impact site. Additionally, five fragmentation explosions can be clearly correlated in the seismic records for the first time by an eigenray search of secondary signal arrivals (see Fig. 7.2, 7.7).

7.3. Trajectory reconstruction and fragmentation localisation

The trajectory model consists of nine free parameters, and their maximum a posteriori (MAP) values are listed in Extended Data Table B.1. The full marginal, and their joint posterior distributions are given in Extended Data B.1. The MCMC inversion converged on a modelled trajectory onset point located above the Luxembourg–France border region at an altitude of $c. 66^{+3}_{-1}$ km (see Fig. 7.1, Extended Data Figure B.1, 7.7). The meteoroid travelled roughly in northeast direction and descended at a shallow elevation angle of $c. 16 \pm 0.2^\circ$ relative to the horizontal. This trajectory is consistent with the wide geographic extent of visual and audible reports by human observers (International Meteor Organization (IMO), 2026). The modelled onset velocity of $16.7^{+2}_{-1.6}$ km s⁻¹ places the event within the intermediate velocity range for meteoritic fireballs which is consistent with an asteroidal source on a low inclination orbit (Terai et al., 2013). The inferred effective ablation coefficient of 0.04 ± 0.01 s² km⁻² is roughly consistent with values reported for stony meteoroids of carbonaceous chondrite composition (Ceplecha et al., 1998) but is higher than typical values measured for achondrites such as eucrites at $c. 0.005\text{--}0.02$ s² km⁻² (Ceplecha et al., 1998). However, the derived ablation coefficient should be interpreted with caution as ablation by thermal erosion and fragmentation are not treated separately during modelling, because intensity of fragmentations and the associated discrete mass loss is unknown. Because our modelling provides an exceptional fit to the measured data, we assume that fragmentations did not affect the trajectory significantly and that they did not produce major sub-bodies. The initial mass at the onset point is estimated to be 305^{+180}_{-90} kg, before significant mass loss from ablation. With

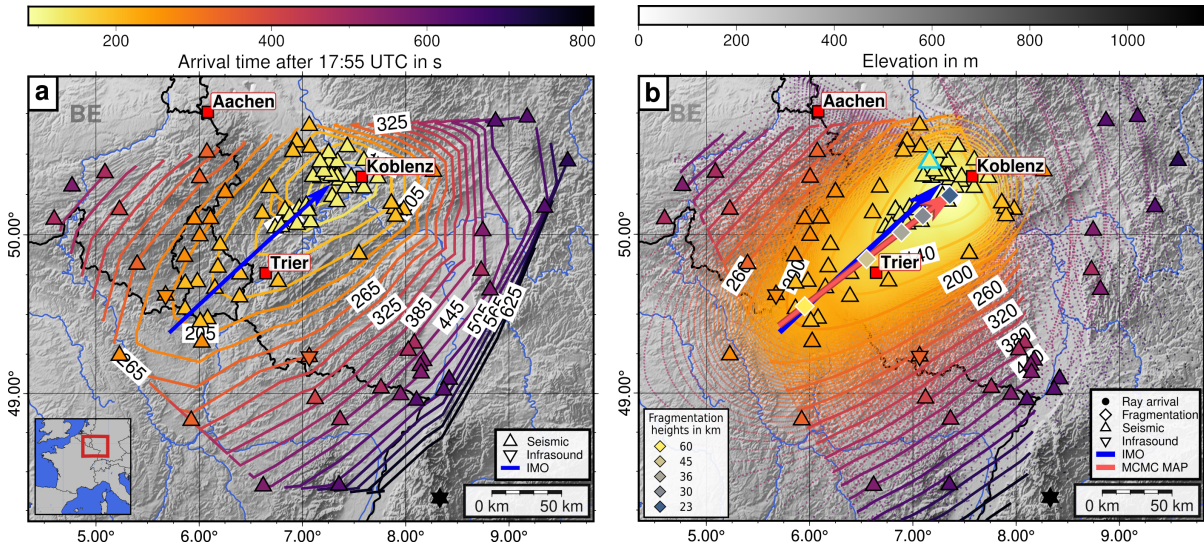


Figure 7.1.: Sensor distribution, trajectory and acoustic wave of the Koblenz Meteoroid. a) Seismic (triangles) and infrasound (inverted triangles) sensors with first arrival time contours of meteoroid-related signals. The trajectory determined by the International Meteor Organization (International Meteor Organization (IMO), 2026) is shown in blue. b) Arrival time contours of the modelled Mach wave, dots represent ray-based first arrivals at Earth's surface, coloured with respect to their arrival time. The trajectory from the maximum a posteriori (MAP) estimate of the MCMC sampling with an RMSE of 1.3 s is shown in red. Diamonds mark identified fragmentation events along the trajectory at altitudes between 23 and 60 km (see scale on the bottom left). The seismic recording station KB.DEP21, west of Koblenz, is highlighted in cyan (see Fig. 7.2).

an assumed bulk density of $3,100 \text{ kg m}^{-3}$, within the range of eucritic meteorites (Macke et al., 2011; Matlovic, P. et al., 2026), this suggests a meteoroid radius of approximately 0.3 m at the onset point.

Video recordings (Molau, 2026) indicate fragmentation events which should emit explosion- or point-source acoustic signals. Fragmentation sources along the MCMC trajectory were identified through an eigenray search using infraGA (Blom, 2014). For each potential fragmentation height along the maximum a posteriori (MAP) estimate trajectory, synthetic infrasound propagation paths were computed considering the stratified atmospheric wind and temperature profiles derived from the ground-to-space (G2S) weather model (see Extended Data Figure B.2). The fragmentation episodes were identified by matching the predicted eigenray arrival times to the observed signal onsets at each station. Five distinct fragmentation sources were identified at altitudes of 23–60 km, indicating a progressive disruption sequence as the meteoroid descended through regions of increasing dynamic pressure (see Fig. 7.1b; Fig. 7.2).

The identification of five distinct fragmentation sources between altitudes of 23 km and 60 km reveals a progressive disruption sequence that is consistent with the dynamic pressure increase, while the meteoroid descends. Following Spurný et al. (2024), the dynamic pressure at the lowest fragmentation altitude of c. 23 km can be estimated as $P_{dyn} = \rho_a v^2 \sim 0.05\text{--}1.8 \text{ MPa}$, depending on the local atmospheric density ρ_a and the decelerated velocity v of the meteoroid at that height. This range aligns well with terminal dynamic pressures of similar achondrite falls, such as asteroid 2024 BX1 at c. 1.8 MPa at 24 km altitude (Spurný et al., 2024) and the Bunburra Rockhole meteorite fall at c. 0.9 MPa at 31 km altitude (Spurný et al., 2012).

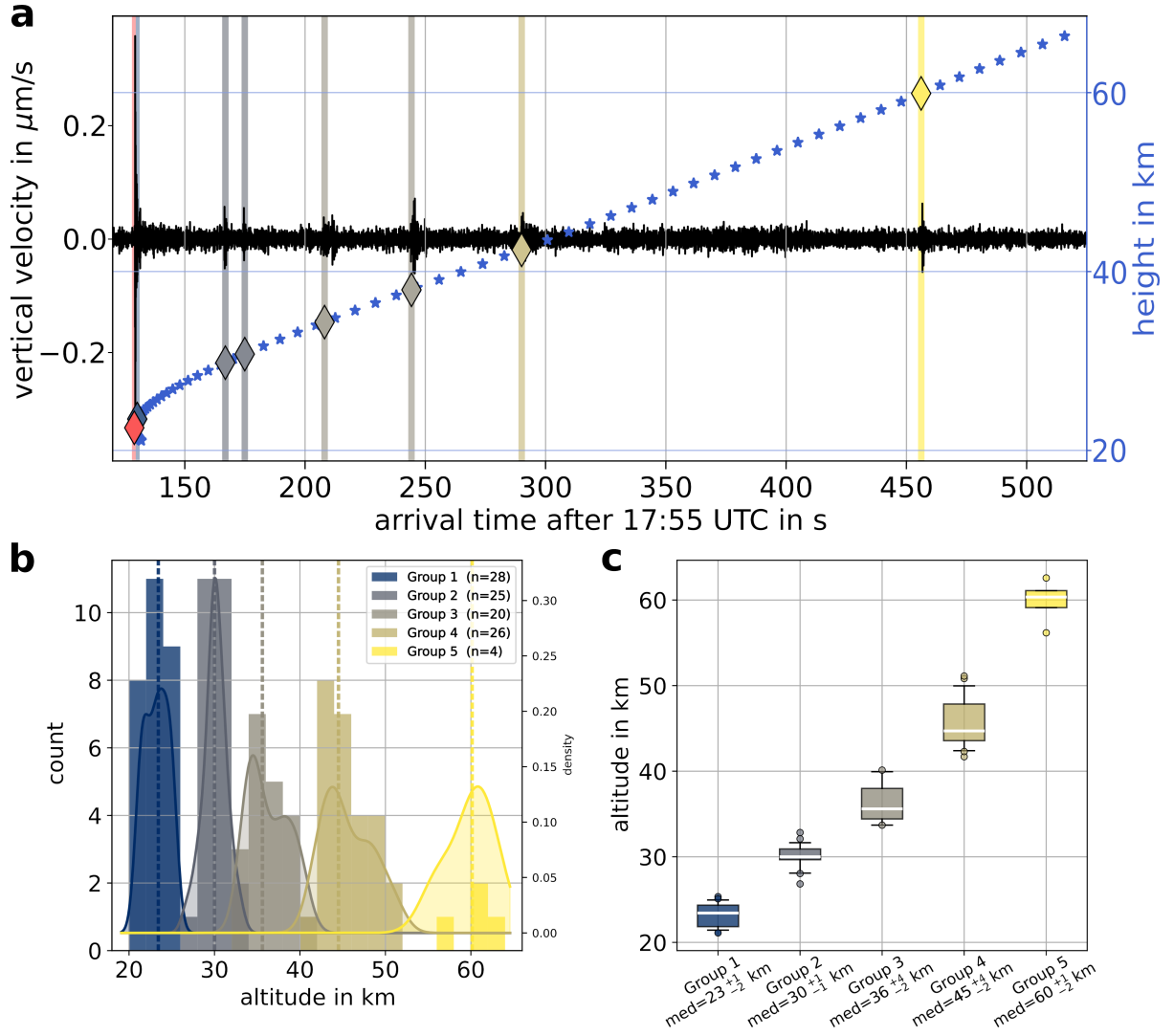


Figure 7.2.: Fragmentation events of the Koblenz Meteoroid. a) Vertical component of the ground motion velocity waveform at the seismic recording station KB.DEP21 (see Fig. 7.1b) bandpass- filtered between 1–10 Hz. Blue stars indicate eigenray arrival times versus emission altitude connecting the MAP estimate meteoroid trajectory with the seismic station. The red diamond and bar mark the Mach wave, while fragmentation related signal arrivals use the same symbols and are coloured with respect to their altitude grouping. b) Histogram of altitude measurements using 2 km altitude bins, overlaid with kernel density estimates for the individual *k*-means cluster groups. The dashed vertical lines mark the cluster medians. c) Boxplot of clustered altitude distributions. Medians are shown as white lines and the whiskers represent the 1σ uncertainties. Outliers are marked as coloured circles.

The progressive rather than catastrophic nature of the fragmentation, spread over c. 37 km of altitude, further suggests a heterogeneous internal structure, which is consistent with a weakly consolidated or brecciated body.

7.4. Strewn Field

The predicted strewn field is calculated using the full MCMC ensemble uncertainties for fragment masses ranging from 1 g to 1 kg (see Fig. 7.3, 7.7). The simulation begins at the point along the MAP trajectory at which the remaining meteoroid mass is 2 kg. This value reflects the upper limit of the cumulative mass of all the simulated fragments combined, and it is chosen to be well above the sub-kilogram lower limit of seismic meteoroid detectability (Neidhart et al., 2021). This ensures that the simulation begins in a physically meaningful regime. Separate ballistic trajectories are computed for the fixed fragment masses of 1 g, 10 g, 100 g, 500 g and 1,000 g by continuing the numerical integration of motion under gravitational acceleration and aerodynamic drag, using the G2S weather model (see Extended Data Figure B.2). No further ablation is applied, as it becomes negligible at the low velocities before surface impact. A predicted landing coordinate is retrieved for each member of each mass class. To propagate the trajectory uncertainties into the strewn field, the simulation is repeated 100 times for each mass class. This results in a probabilistic footprint for each mass class, formed by the spread of landing coordinates across the ensemble, from which 1σ and 2σ landing uncertainty ellipses are derived. An additional horizontal positional uncertainty of ± 100 m per mass class arises from an estimated atmospheric wind uncertainty of $\pm 5 \text{ m s}^{-1}$ (Assink et al., 2014), which is added as an independent random perturbation at each altitude level of the G2S model. The modelled strewn field is oriented approximately along the flight azimuth due to the low wind conditions. The resulting landing ellipses have a 1σ uncertainty semi-major axis ranging from 2.4 km to 3.3 km for the 1 g and 1 kg mass classes, respectively. The strewn field altogether ranges from approximately 50.19°N , 7.26°E to 50.36°N , 7.6°E within a c. 2 km wide corridor just north of the Moselle River (Fig. 7.3), and the largest surviving fragments are predicted to be located at the north-eastern, downrange end of the strewn field. The dominant contribution to the along-track extent of the uncertainty ellipses is due to the velocity and elevation angle posteriors. The strewn field is validated using the fragments that were recovered from the impact in a residential building in Koblenz-Güls (Fig. 7.3). The building falls within the 2σ landing ellipse for the 1 kg mass class, located approximately 4.7 km from the centre of that ellipse. This provides rare, precisely geolocated ground truth for the terminal trajectory. The coincidence between the predicted landing zone and the documented impact confirms that the uncertainties in the MCMC-derived trajectory are neither over- nor underestimated, and that the G2S wind profile adequately captures the lower-tropospheric conditions at the time of the fall.

7.5. Pre-atmospheric heliocentric orbit

The pre-atmospheric heliocentric orbit is computed from the MCMC entry state vector, following Cepplecha (1987). The geocentric radiant is first derived from the observed azimuth, elevation and entry velocity, applying corrections for diurnal aberration and zenith attraction to account for the gravitational deflection of the incoming body near Earth Cepplecha (1987). Then corrections are made for gravitational focusing, yielding geocentric and heliocentric speeds of 12.7 km s^{-1} and 36.1 km s^{-1} at 1 AU, respectively. The geocentric state vector is

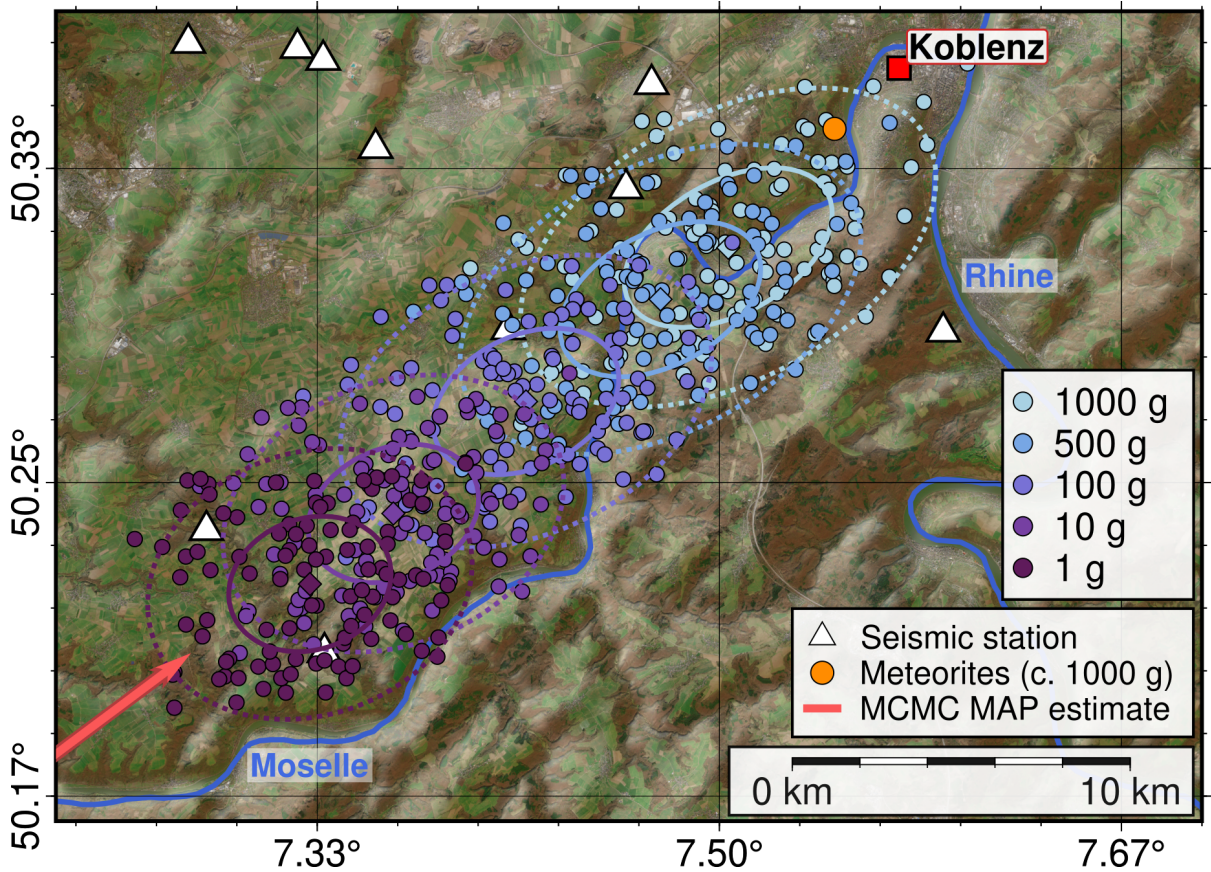


Figure 7.3.: Meteorite strewn field. Circles, which are coloured by mass, represent ground locations of modelled meteoroid fragments. Seismic stations are represented by white triangles. Uncertainty ellipses of each mass class are drawn for 1σ (solid lines) and 2σ (dashed lines). The meteoroid trajectory MAP estimate is displayed as a red arrow and the building, where the approximately 1 kg cumulative meteorite mass caused an impact, is shown as an orange circle. The background image is from Copernicus Sentinel data 2026, overlaid onto the local digital elevation model.

then rotated into the heliocentric ecliptic frame using the position and velocity of the Earth's barycentre from the DE430 planetary ephemeris (Folkner et al., 2014), evaluated at the observation epoch of 2026-03-08 17:55 UTC. At this epoch, osculating Keplerian elements are computed. The orbital elements and their 1σ quantiles are summarised in Extended Data Figure B.3, and the corresponding orbits are shown in Figure 7.4. The orbit has a semi-major axis a of 1.85 AU, an eccentricity e of 0.537, and an ecliptic inclination i of 13.66° , with a perihelion distance q of 0.86 AU and an aphelion distance Q of 2.84 AU, placing the parent body in the outer main asteroid belt. With a resulting orbital period of approximately 2.5 years, as well as the combination of $a > 1$ AU and $q < 1.017$ AU, the meteoroid is categorised as an Apollo-class near-Earth object (Shoemaker et al., 1979). The Tisserand parameter with respect to Jupiter, $T_J = 3.79$, places the orbit in the asteroidal regime ($T_J > 3$), which is consistent with an origin in the main belt rather than a cometary source (Levison, 1996). The low inclination and moderate eccentricity are consistent with an origin from an S-type main-belt source region via resonant transport (Wetherill and ReVelle, 1981). The aphelion distance of 2.84 AU lies in the vicinity of the 5:2 mean-motion resonance with Jupiter at 2.82 AU, which

is one of the principal dynamical pathways by which outer main-belt material is injected onto Earth-crossing orbits (Morbiddelli and Gladman, 1998; Bottke et al., 2002). This is more consistent with an S- or C-complex outer-belt origin than with a Vesta family origin (DeMeo and Carry, 2014). However, numerical integration studies have shown that fragments from (4) Vesta and the Vestoid collisional family can have their eccentricities secularly pumped to large values when injected into the ν_6 secular resonance or the 3:1 mean-motion resonance from the inner belt, producing present-day aphelia beyond 2.5 AU (Migliorini et al., 1997; Bottke et al., 2002). Therefore, an HED (howardite, eucrite, diogenite) provenance cannot be excluded on orbital grounds alone, and further dynamical backward integration of the full MCMC orbit ensemble would be needed to quantitatively assess the relative likelihood of inner- versus outer-belt delivery.

Our study demonstrates that combining the probabilistic MCMC inversion framework with existing regional seismic and infrasound networks can enable the reconstruction of the trajectory, fragmentation history, heliocentric orbit, and spatially calibrated strewn field of a single well-recorded meteoroid event. This can be achieved within a short amount of time, a few hours, after the meteoroid fall due to the online, near-real-time availability of seismic recordings (Strollo et al., 2021) and automatic phase identification procedures (Eickhoff et al., 2026b). This means that fully automatic routines for meteoroid characterisation will be available soon, comparable to those already used for earthquake monitoring.

7.6. Data

Atmospheric profiles. Temperature and wind profiles were obtained from the Ground-to-Space (G2S) model via the NCPAG2S command-line client (Hetzler 2024), downloaded for 2026-03-08 18:00 UTC at 50.3 ° N, 7.3 ° E, covering altitudes up to 150 km at 100 m vertical spacing. Observationally constrained atmospheric specifications are preferred over climatological averages, as they more accurately reproduce effective sound speeds and infrasound travel times (Drob et al., 2003; Le Pichon et al., 2005; Assink et al., 2014).

Seismic and infrasound data. Continuous waveforms from seismic stations within 3.5 ° distance from Koblenz were retrieved via FDSN/EIDA (Strollo et al., 2021), Raspberry Shake (Raspberry Shake, S.A., 2016), and KIT-GPI services using ObsPy (Beyreuther et al., 2010), supplemented by data from the LU, HS, BQ, and LE networks. Waveforms were pre-processed by removing instrument responses, converting the waveforms into physical units (infrasound to Pa; seismometers to $\mu\text{m s}^{-1}$), detrending, and bandpass filtering (0.5 – 40 Hz). Meteoroid onsets were manually identified from characteristic impulsive overpressure arrivals: N-shaped wavelets in infrasound records and W-shaped wavelets in seismograms resulting from acoustic-to-seismic coupling (Langston, 2004). Arrivals were classified into Mach cone and fragmentation groups using ray tracing with infraGA/GeoAc (Blom, 2014) to model expected Mach wave first arrival times. Stations for which no predicted Mach ray arrival could be computed were assigned to the fragmentation group.

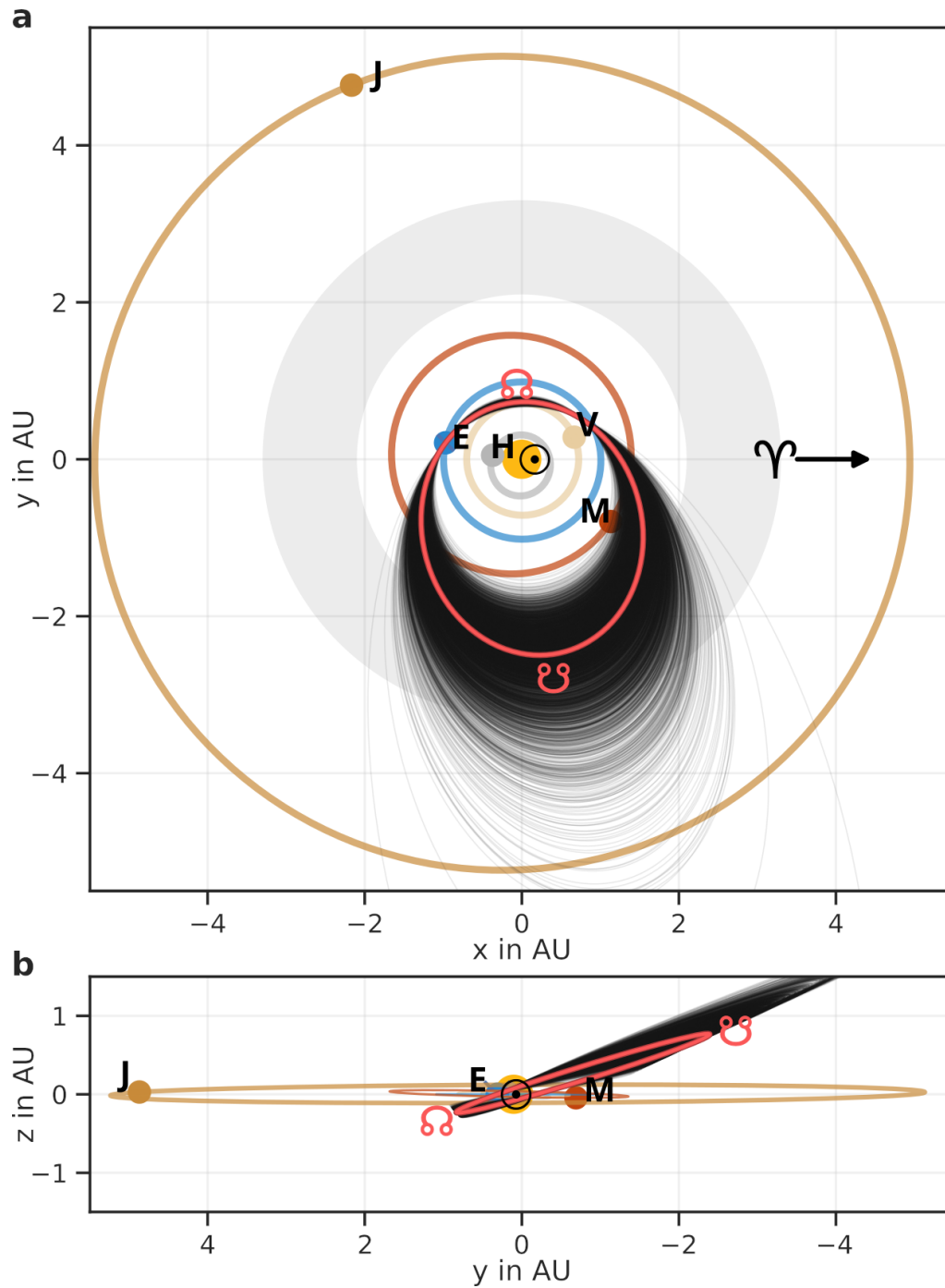


Figure 7.4.: Orbit views of the Koblenz meteoroid. a), Map view of the ecliptic plane with the Sun in the centre, major planets up to Jupiter, and the asteroid belt in grey at the time of the impact on Earth. The orbits of meteoroid candidates within the uncertainties of the MCMC ensemble are shown as black lines with the MAP estimate orbit shown as red line. This colouring is repeated in Panel b. b), Side view of the orbits.

7.7. Methods

Acoustic ray tracing and lookup table (LUT). The Mach cone shock wave emitted by a supersonic meteoroid can be approximated using the 3-D ray-tracer *infraGA/GeoAc* (Blom, 2014), which supports complex stratified atmospheric models and accounts for wind effects, which is a controlling factor in acoustic-seismic trajectory reconstruction (Eickhoff et al., 2026a; Fernando and Charalambous, 2026). Direct numerical ray tracing for every MCMC proposal using *infraGA* is, however, computationally prohibitive: a full inversion requires 10^5 – 10^6 forward evaluations, and each *infraGA* call is too slow to make ensemble sampling practical on standard high-performance computing infrastructure. Therefore, we precompute a lookup table (LUT) of acoustic ray arrivals from point sources, reducing each forward evaluation to a fast interpolation step. The LUT is constructed for discrete source altitudes at integer kilometre levels (15 – 100 km) and a regular sampling of ray take-off azimuth and inclination angles at 0.5° resolution. Each table entry stores the horizontal ground-intercept offsets relative to the source point in kilometres and the ray propagation time in seconds. For each proposed trajectory point during an inversion, we compute the local Mach number by dividing the trajectory speed by the sound speed from the G2S atmospheric model (Extended Data Figure B.2) and determine the acoustic ray emission directions by mapping the Mach cone outward normals into ray azimuth and inclination following Blom et al. (2024). The cone opening angle is parameterised around the instantaneous velocity vector at 0.5° sampling. To evaluate LUT values at arbitrary azimuth-inclination combinations, we apply k -nearest-neighbour inverse-distance weighting interpolation (Shepard, 1968) on the angular grid, with a maximum nearest-neighbour distance threshold of 1.5° to prevent extrapolation into unsupported angular regions. For source altitudes between stored integer levels, the two adjacent LUT slices are evaluated and the results are linearly interpolated. This architecture reduces each Mach cone forward evaluation to less than one second while retaining the full wind and temperature structure of the G2S profile.

Trajectory inversion An initial trajectory estimate is obtained via particle swarm optimization using *PySwarms* (Miranda, 2018) over the LUT forward model, providing a computationally efficient starting point for the full probabilistic inversion. Proposals are parametrised by the proposal vector containing the nine parameters. These are start longitude, latitude and elevation, which build the starting point P where the modelling begins. Then azimuth angle (measured clockwise from North) and elevation angle (measured relative to the horizontal) and entry velocity build the initial direction of the trajectory. Finally, a time shift t is used to account for uncertainties in the event time and additionally two physical meteoroid parameters are included: the ablation coefficient σ_{eff} and the initial mass. From there meteoroid trajectories are computed by integrating the equations of motion from Campbell-Brown et al. (2013), accounting for deceleration:

$$\frac{dv}{dt} = \frac{\Gamma \rho_a v^2}{m} A \left(\frac{m}{\rho_m} \right)^{2/3}, \quad (7.1)$$

with the meteoroid velocity v , drag coefficient Γ , atmospheric density ρ_a the meteoroid mass m , the shape factor A and the meteoroid density ρ_m . The shape factor is a geometric constant

and relates the meteoroid's cross-sectional area to its mass and volume. We consider the drag coefficient and shape factor together and set $\Gamma A = 0.7$, following Spurný et al. (2024), in our modelling. Mass loss is calculated by:

$$\frac{dm}{dt} = -\Gamma A \sigma_{eff} \left(\frac{m}{\rho_m} \right)^{2/3} \rho_a v^3, \text{ with } \sigma_{eff} = \sigma + \sigma_e, \quad (7.2)$$

with the ablation coefficient σ (Ceplecha et al., 1998) and the erosion coefficient σ_e (Campbell-Brown et al., 2013). Each trajectory, calculated from the proposal parameter vector is then discretized into 100 source points and evaluated against observed arrival times. Posterior sampling is performed with an affine-invariant ensemble MCMC (using the Python-package `emcee`: Foreman-Mackey et al. (2013)) using 60 walkers and a mixture of stretch, differential evolution, and snooker moves (Nelson et al., 2014; Ter Braak and Vrugt, 2008) to handle correlated parameters. Convergence is assessed via the integrated autocorrelation time, where sampling terminates once the chain length exceeds 50 times the maximum autocorrelation time. Underperforming walkers are re-seeded by cloning high-acceptance walkers with added Gaussian jitter.

Fragmentation localisation Arrivals not consistent with Mach cone ray paths are attributed to discrete fragmentation bursts, which emit quasi-spherical acoustic waves. For these, we search the MAP estimate trajectory, discretized into 100 source points, for the source location that minimizes travel-time residuals using the eigenray method in `infraGA` (Blom and Waxler, 2017). In this way, the arrivals can be assigned a most likely ray emission height (see Figure 7.2). We repeat the process for all fragmentation arrivals at all recording stations and separate the fragmentation height candidates into clusters by binning them (see Figure 7.2b, 7.2c).

Strewn field modelling As part of the modelling, we simulate the meteoroid mass along the trajectory. For the strewn field, i.e., the corridor where meteoritic material is found on Earth's surface, we use the meteoroid trajectory MAP estimate and use the point where the modelled meteoroid mass is 2 kg. From this point on, meteoroid trajectories are computed by using equations 7.1-7.2 for masses fixed at 1 g, 10 g, 100 g, 500 g, and 1000 g until the fragments impact the Earth's surface. This procedure is repeated within the MCMC modelling uncertainties to derive the elliptical footprints for each mass class (see Fig. 7.3).

8. Summary and Outlook

This thesis presents three studies which progressively build a single, coherent workflow for using densely and globally distributed seismic and infrasound networks to detect supersonic atmospheric meteoroid entries, reconstruct their trajectories, and enable the rapid recovery of the resulting meteorite material. This workflow transforms continuously operating acoustic-seismic networks, originally built for earthquake monitoring, into a time- and weather-independent alternative to optical fireball cameras, which are ineffective under cloud cover, during daylight and in regions with sparse coverage. The first study (Eickhoff et al., in press) establishes the observational and detection foundation. The second study (Eickhoff et al., 2026a) advances the physical modelling of the trajectory inversion. While the third study (Eickhoff and Ritter, submitted) unifies these advances into a probabilistic, end-to-end framework. Taken together, the key contribution of this thesis is the transformation of acoustic-seismic meteoroid analysis from isolated, single-event case studies into a reproducible, statistically rigorous and almost fully automatable process. As the underlying waveform data is already available in near real-time across many continental regions, this workflow paves the way for the rapid, autonomous calculation of meteorite strewn fields, the recovery of meteorites and the systematic characterisation of meteoroids, comparable to the automated routines already established in earthquake monitoring (see Fig. 8.1).

8.1. Summary

At the beginning of this thesis, I outlined the key motivations for applying acoustic-seismic methods to meteoroid analysis. Here, I revisit these motivations and discuss how the studies presented in this thesis, along with their results and conclusions, can address existing research gaps, and how they can be applied now and in the future. The first motivation was that acoustic-seismic data can close spatial gaps in operational meteor detection networks. The Starlink satellite re-entry (see Chapter 6) demonstrates this in practice. There, human observer reports were numerous, but only two dedicated camera stations observed the event optically. The trajectories derived from the human observer reports and the meteor camera network could not capture the full extent of the satellite re-entry when compared to the acoustic-seismic solution (e.g., Fig. 6.7c). This is not because one method holds a methodological advantage over the other, but because in this case the acoustic-seismic method had far more data available to constrain the trajectory. The associated research gap was the absence of infrastructure for reliably analysing meteoroid-related signals in seismic waveform data. This has been addressed by the PhaseNet-based meteoroid arrival picker, trained on a manually compiled seismic recording database (see Chapter 5).

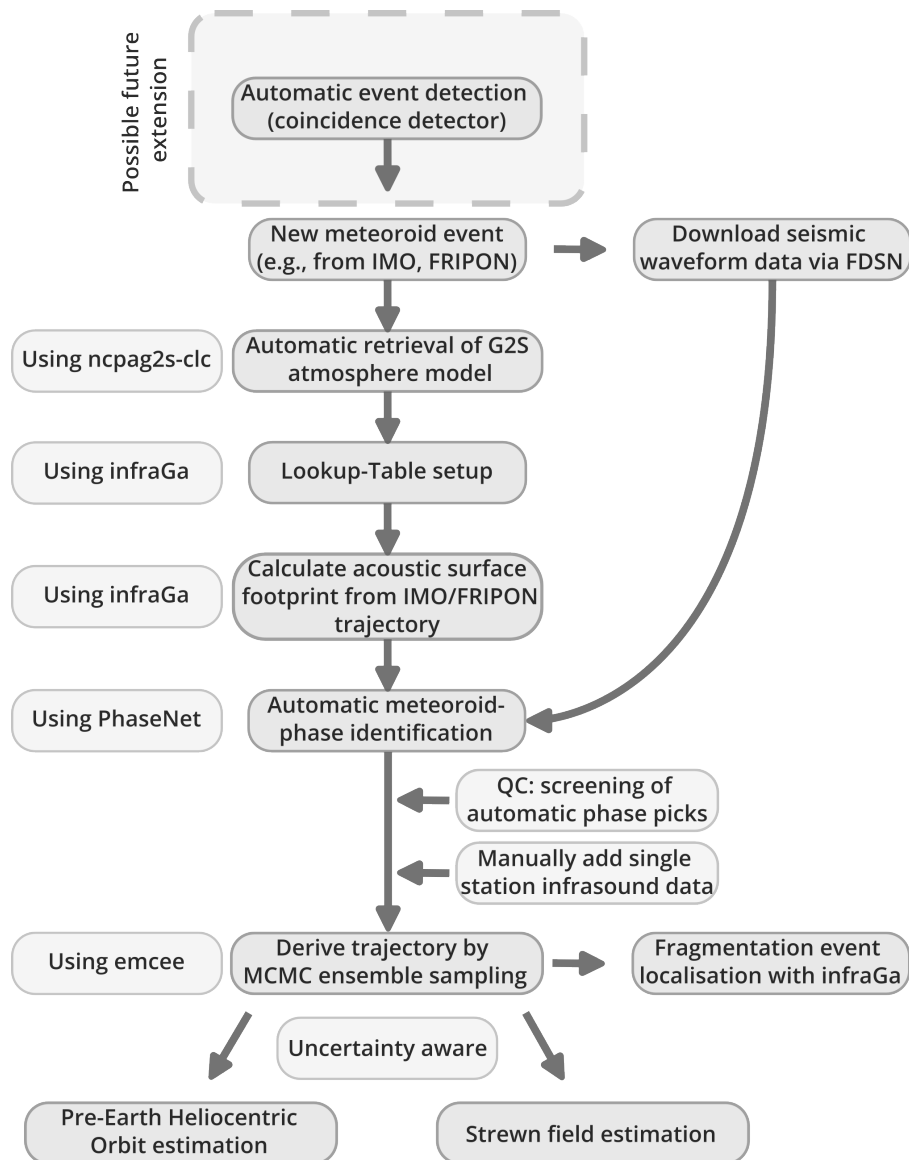


Figure 8.1.: End-to-end seismic meteoroid trajectory characterisation framework. For now meteoroid phases in infrasound waveform data have to be identified manually and the seismic PhaseNet picks have to be quality-controlled.

The second motivation was that seismic and infrasound stations operate continuously and under all atmospheric conditions. Although I analysed data from daytime meteoroids while compiling the SMORD (Chapter 5, Fig. 5.3c), this major advantage over optical trajectory detection could not be demonstrated directly, because the initial trajectory guess still relies on optically identified meteors. However, now that the acoustic-seismic meteoroid characterisation framework is almost complete, this reliance on optical initial solutions will diminish once automatic meteoroid event detection becomes practically feasible. This will, in turn, enable a more complete catalogue of daytime meteoroid activity that camera networks cannot cover.

The third motivation was that acoustic-seismic observations are physically distinct from, and independent of, optical and radar measurements, and that they can be used jointly with these to constrain meteoroid trajectories. For the Adriatic fireball, which served as a PhaseNet model benchmark (see Fig. 5.6), optical constraints from the FRIPON trajectory solution were placed on the acoustic-seismic trajectory. While this does not constitute a full joint inversion, it nevertheless yielded a more tightly constrained trajectory. In the future, when the acoustic-seismic meteoroid framework is improved to include fragmentation directly in the trajectory calculation, the light curve from optical measurements will provide valuable information about the timing of fragmentation events.

The fourth motivation was that acoustic-seismic methods are well suited to detecting meteoroids in the mass range from 10^3 g to 10^6 g, and that such observations can help to better quantify both the meteoroid mass flux and the physical parameters of individual meteoroids. Routine mass estimation from seismic data is not yet feasible, owing to the acoustic-seismic coupling and the associated uncertainties of up to three orders of magnitude, which depend on local geology. It is, however, possible to constrain the masses of individual, well-recorded events, such as the Koblenz meteoroid (see Chapter 7 and Fig. B.1). There, for the first time, it was possible to place reasonable constraints on the initial mass of a meteoroid and on its mass loss during atmospheric passage. Fully systematic mass estimation from seismic data alone remains out of reach for now, but databases such as the SMORD lay the groundwork for this type of physical meteoroid characterisation.

The fifth motivation, and one of the most important for the scientific community, was the calculation of strewn fields from acoustic-seismic trajectory products. This was fully realised with the framework presented in the study of the Koblenz meteoroid (see Chapter 7, Fig. 8.1). For this event, the application led to directed meteorite search expeditions, though these did not recover any meteorite material. For other events, such as the fall of meteoroid BX1 near Berlin, the acoustic-seismic strewn field matched the location and extent of the one derived from camera observations (see Fig. 4.8), demonstrating that the predicted ground locations are of comparable quality. The Koblenz event provided a further validation: a roughly 1 kg fragment struck a residential building, tearing a hole in the roof. This known impact location fell within the predicted strewn field for the 1 kg mass class, confirming that the method correctly predicts meteorite ground locations.

The sixth motivation was that collocated acoustic-seismic sensors can improve our understanding of acoustic-to-seismic coupling in the near surface. The limited knowledge of this coupling efficiency is the main research gap preventing the estimation of meteoroid masses and energies from seismic data alone. As shown in the Starlink study (see Chapter 6), the coupling efficiency spans up to three orders of magnitude, which translates directly into a comparable uncertainty in derived source parameters. Closing this gap is therefore a necessary step towards a seismic meteoroid magnitude scale. While this gap could not be closed within my thesis, it does lay the groundwork for closing it in the future. Through the seismo-acoustic meteoroid recording database (see Chapter 5) and the results of the Starlink study, it became clear that fully understanding the acoustic-to-seismic coupling will require controlled-source experiments.

In addition to the reevaluation of the initial motivations and research aims, I will continue with summarising the main conclusions of each study.

The first study introduced SMORD, the Seismo-acoustic Meteoroid Observation Recording Database, the first openly accessible dataset that links reported meteoroid events to manually labelled seismic waveforms (Eickhoff et al., in press). It is compiled by cross-referencing the crowd-sourced IMO fireball reports and the sensor-based NASA CNEOS catalogue with global seismic archives. SMORD contains over 300 meteoroid events observed at seismic stations and 3,295 quality-flagged air-to-ground coupled seismic phase arrivals. The study also demonstrates that a PhaseNet model trained on these labels via SeisBench can identify meteoroid-induced seismo-acoustic onsets with a precision and recall greater than 90 %, and a median absolute timing error of 0.02 s. 90 % of the picks fall within ± 0.3 s uncertainty. Additionally, this performance remains consistent across different sensor types and recording station sites. This level of timing accuracy converts to a distance uncertainty of only about ± 100 m, which is well below the modelling uncertainties introduced by atmospheric winds. The main takeaway is that seismic networks offer far denser and more uniform spatial coverage than dedicated meteor cameras or radar in many regions. The seismic stations operate independently of cloud cover and daylight and stream continuous, near real-time, accessible, open data across many continental regions. By providing a reproducible labelled dataset and a working phase detector, this study reduces the barriers to exploiting the underused seismic meteoroid observation technique. Furthermore, applying this to the April 2025 Adriatic fireball (see Fig. 5.6) confirms the practicality and ease of use of the PhaseNet model. There, a nearly fully automated four-step workflow produced a trajectory consistent with independent FRIPON solutions, requiring only the event's origin time and location as user input.

The second study analysed the uncontrolled re-entry of the Starlink-2382 satellite over Central Europe on 27 August 2024, using an exceptionally large set of 226 acoustic-seismic arrivals across six countries (Eickhoff et al., 2026a). Its main methodological contribution is a new ray-based, iterative damped direct-search inversion of the trajectory built on infraGA. This method includes a 3-D atmospheric wind model (see Fig. 6.4), accounts for time-dependent mass loss from ablation, and includes a lift-like force (see Fig. 6.11). This approach reduced average arrival-time errors by about 40% compared to the traditional geometrical method by Pujol et al. (2005). It also fits significantly more observations, resulting in a well-defined trajectory that outperforms the optical solutions in this low-elevation-angle setting. The known initial mass of the satellite (≈ 260 kg) provides a rare calibration reference that is not available for natural meteoroids. This allows the determination of an effective ablation coefficient of $0.11 \pm 0.02 \text{ s}^2 \text{ km}^{-2}$ (see Table 6.3), consistent with near-complete burn-up. Finite-difference wavefield modelling independently supports the ray-based reconstruction. It reveals the widening of the Mach cone and the impact of trajectory curvature, showing that time-varying Mach angles must be considered when modelling low-angle entries. Finally, at collocated acoustic and seismic sites the study constrain acoustic-to-seismic coupling coefficients spanning three orders of magnitude (4.31×10^{-10} to $5.86 \times 10^{-7} \text{ m s}^{-1} \text{ Pa}^{-1}$; see Table 6.2). The coupling efficiency is mainly controlled by the stiffness of the surface rocks. Re-entries like this can be used as 'controlled-source' experiments to improve the understanding about acoustic-to-seismic coupling and towards a standard for retrieving the source energy and mass for meteoroids from acoustic and seismic measurements.

The third study combines the previous advances into a single, end-to-end, probabilistic framework (see Fig. 8.1 and applies it to the Koblenz meteoroid recorded on 8 March 2026 at 95 seismic stations (Eickhoff and Ritter, submitted). The study’s key contribution is a Markov chain Monte Carlo trajectory inversion which, for the first time, provides a statistically rigorous reconstruction from acoustic-seismic observations while taking into account the entire wind field and the atmosphere’s stratified temperature structure (see Fig. 7.1). A precomputed acoustic-ray look-up table reduces each forward evaluation to less than one second on a standard home computer. This enables ensemble sampling while retaining realistic propagation physics. The framework simultaneously recovers the trajectory and resolves five discrete fragmentation events between 23 km and 60 km altitude via an eigenray search. Additionally, this approach enables the parameters to be regularised, providing a mass estimate of the meteoroid from acoustic-seismic data alone for the first time. It also propagates observational uncertainties into ensemble-based landing ellipses and pre-Earth heliocentric orbits. The predicted strewn field was validated directly against meteorite fragments that struck a residential building in Koblenz-Güls, which fell within the 2σ uncertainty landing ellipse for the 1 kg mass class (see Fig. 7.3). The struck building provides rare, precisely geolocated ground truth, confirming that the MCMC uncertainties are neither over- nor underestimated. The derived orbit places the parent body on an Apollo-class asteroidal-regime orbit in the outer main belt. The more general implication is that regional seismic data can deliver a complete fall reconstruction with quantified uncertainties, independently of any other observational data. The reconstruction can be achieved within hours of the fall, provided that near real-time data is available, demonstrating that near real-time automatic meteoroid analysis will soon be possible.

8.2. Outlook

Building on the detection foundation of the first study (Eickhoff et al. 2026b; Chapter 5), the most obvious extension is to grow and diversify the SMORD database by relaxing the current selection thresholds. These are the 100-report IMO cut-off and the 0.3 kt CNEOS yield cut-off. Additionally, the database can be extended by including events from additional networks such as FRIPON, the Global Meteor Network, as well as the AllSky7 network. This would result in a larger, more varied training set, which is the prerequisite for nearly autonomous picking with minimal manual curation. Instead of focusing on seismic data only, dedicated training strategies for infrasound data should be carried out. Separate models for array-based and single-station recordings could be trained. Together with the seismic counterpart, joint seismic-acoustic phase identification could be studied. This would exploit the complementary information in ground-motion and pressure waveforms to better understand local acoustic-to-seismic coupling.

To improve the physical consistency of predicted travel times and phase associations and to streamline the extension of SMORD, atmospheric acoustic wave propagation modelling could be used. This can be done with ray tracing using infraGA constrained by reanalysis wind fields. This would support the longer-term goal of autonomous, continuous monitoring and localisation of meteoroid falls. In this way coincidence detectors, similar to those in

earthquake seismology, adapted to recognise the characteristic acoustic moveout and the diagnostic W-shaped waveform, could be constructed. This would be especially valuable for daytime events and for the dark-flight phase that mainly controls the strewn-field estimation, because for both only sparse or no optical data is available.

The physical modelling developed in the second study naturally suggests several improvements. One clear next step is to extend acoustic wavefield modelling to include wind, which is not yet feasible with present capabilities but is required to fully reconcile far-field travel times for strongly curved, low-angle trajectories and to avoid mislocation of the source. The effective ablation coefficient derived from the satellite re-entry is based on a simplified spherical- and single-body assumption. To overcome this, more detailed studies that account for true geometry and material properties are needed to convert such effective values into physically meaningful ablation estimates and to calibrate the method for natural meteoroids of unknown mass and composition. The strong site dependence of the acoustic-to-seismic coupling coefficients highlights the need for dedicated controlled-source experiments to characterise their dependence on wavefront incidence angle, azimuth and local topography. These experiments should be accompanied by the deployment of more acoustic sensors co-located at seismic sites, with the aim of building an empirical catalogue of coupling behaviour. As the number of satellites in low Earth orbit increases and uncontrolled re-entries become more frequent, these events will provide recurring, partially calibrated references. Their systematic analysis will continuously improve the consistency with which seismic stations can be used to retrieve mass and energy of meteoroids.

The most direct extension of the probabilistic framework of the third study is full automation, which involves combining near real-time waveform retrieval and the automated phase picker from the first study with the MCMC inversion to create an operational pipeline (see Fig. 8.1). This pipeline can be triggered automatically in response to reported events and provides rapid strewn-field predictions to guide meteorite searches before the material is heavily degraded by weathering. Currently, ablation and fragmentation are treated through a single effective coefficient. In future applications, however, thermal ablation must be separated from discrete fragmentation mass loss. This would improve both the mass estimates and the physical interpretation of the disruption sequence. Additionally, the fragmentation analysis could be extended beyond locating discrete bursts to inferring source energies and the meteoroid's structural strength. Both of these improvements could in principle be realised by an iterative approach that balances trajectory inference with fragmentation location. Such an algorithm would perform best when coupled with optical data, but does not necessarily depend on it.

In conclusion, applying the analytical framework established throughout this thesis systematically to many events, as enabled by the expanding SMORD database and growing network coverage, promises to transform the analysis of acoustic-seismic meteoroids from individual case studies to population-level statistics on meteoroid flux, fragmentation behaviour and source-region distributions. This would realise the automated characterisation of meteoroid falls envisioned throughout this thesis. It would be a significant and necessary step towards developing a seismic meteoroid magnitude scale, which could offer a valuable means of directly

estimating meteoroid entry energies from seismic data. It would simplify the characterisation of new events across different seismic networks, allowing for systematic and statistical comparisons between otherwise unique meteoroid events.

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A. Supplement to chapter 6

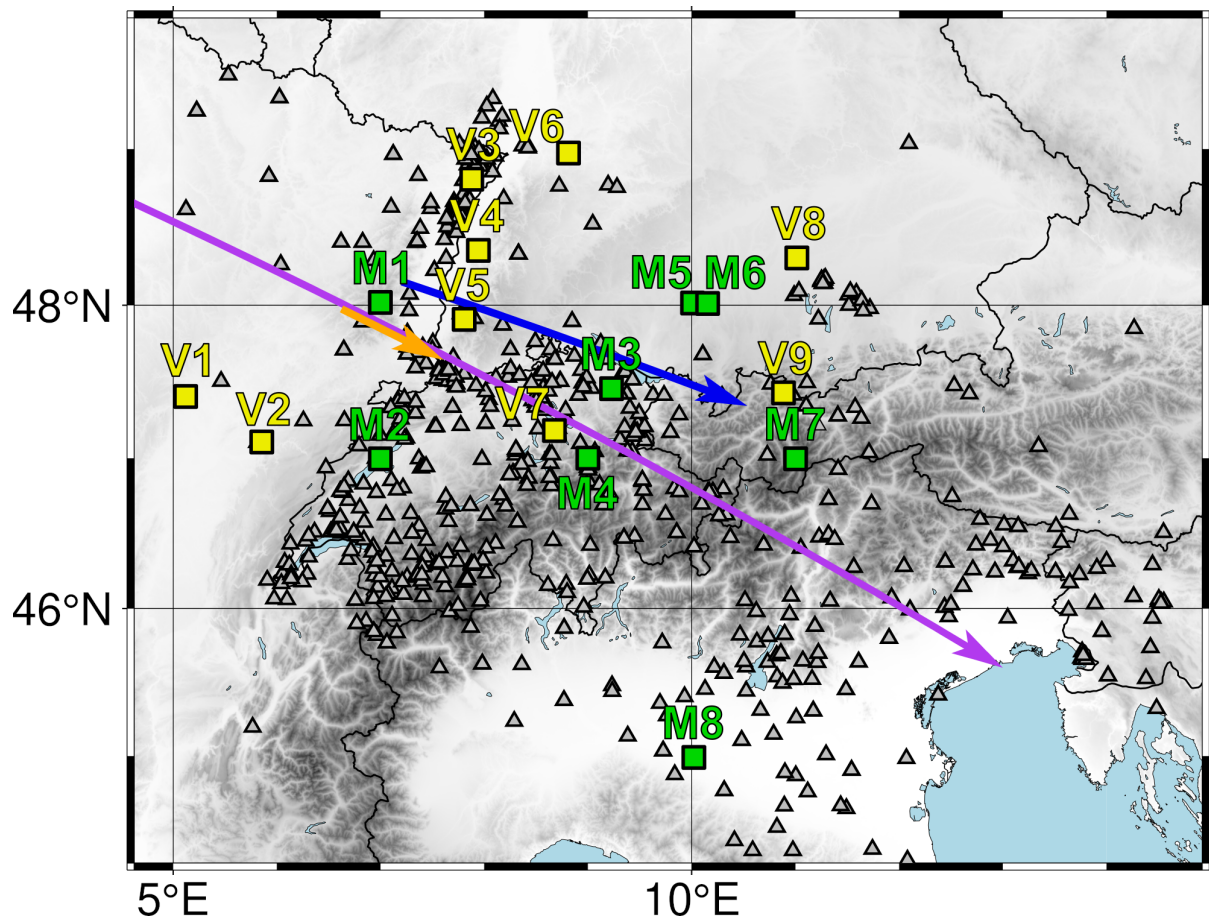


Figure A.1.: Map of photograph (M) and video (V) locations that captured the re-entry of Starlink satellite-2382. Note the increase of location number with distance along the trajectory.

The visual observations of the satellite fall indicate a possible increase of fragments along the trajectory (Table A.1). The analysis of photographs shows 1 – 11 fragments. Note that higher location numbers (M1 – M8) are related to a greater distance along the trajectory (observation locations are orthogonal to the trajectory). For video observations (V1 – V9) there is a similar trend (Table A.1). However, it must be noted that the different angles of view may hide objects and that the different observations vary in resolution.

Location	No. of frag- ments	Observations	Duration of video
M1	~5	small leading fragments with main fragment tailing behind	n/a
M2	~6	same as M1	n/a
M3	~7	same as M1	n/a
M4	~9	same as M1 but with increased separation between smaller fragments	n/a
M5	~5	same as M4	n/a
M6	~9	same as M4	n/a
M7	~11	same as M4 but with even more separation between fragments	n/a
M8	~7	same as M7	n/a
V1	1	only main fragment visible	>10
V2	~4	main fragment with one leading and two side fragments	>30
V3	1	only main fragment visible	>14
V4	~3	main fragment with one leading and one side fragments	>5
V5	1	only main fragment visible	>7
V6	~4	main fragment with one leading and two side fragments	>18
V7	~10	main fragment with 5 leading and 4 side fragments	>12
V8	~3	main fragment with one leading and one side fragments	>12
V9	~5	many fragments with no clear main fragment	>15

Table A.1.: Table of satellite fragments identified in photographs (M) and videos (V) provided by IMO (2024).M1: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17465M3: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17486M4: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17488M5: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17468M6: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17482M7: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17460M8: https://fireballs.imo.net/members/imo_photo/view_photo?photo_id=17477V1: https://fireballs.imo.net/members/imo_video/view_video?video_id=16081V2: https://fireballs.imo.net/members/imo_video/view_video?video_id=16058V3: https://fireballs.imo.net/members/imo_video/view_video?video_id=16028V4: https://fireballs.imo.net/members/imo_video/view_video?video_id=16038V5: https://fireballs.imo.net/members/imo_video/view_video?video_id=16046V6: https://fireballs.imo.net/members/imo_video/view_video?video_id=16056V7: https://fireballs.imo.net/members/imo_video/view_video?video_id=16068V8: https://fireballs.imo.net/members/imo_video/view_video?video_id=16077V9: https://fireballs.imo.net/members/imo_video/view_video?video_id=16034

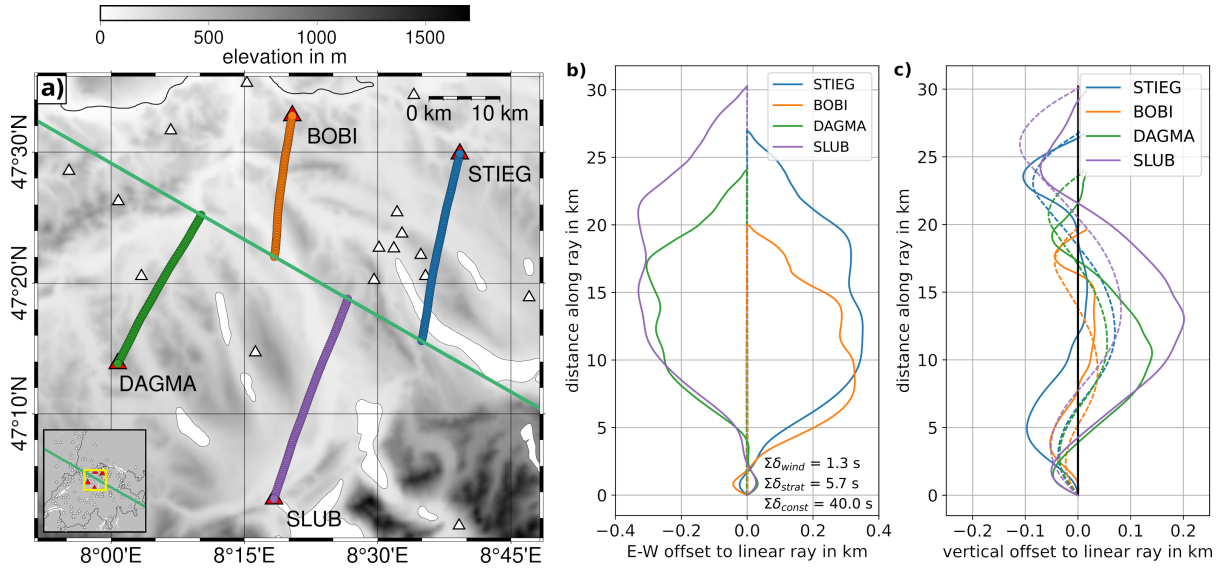


Figure A.2.: a) Map with four selected ray paths from the trajectory to the recording sites DAGMA, BOBI, SLUB, and STIEG. b) Line graph showing horizontal deviation of rays modelled with wind (solid lines) and constant atmosphere, and modelled rays with a stratified atmosphere only (dashed lines in the centre) compared to a constant atmosphere. The ray paths were rotated by their azimuth relative to north. c) The same as b) but for vertical offset of the rays compared to constant atmosphere rays.

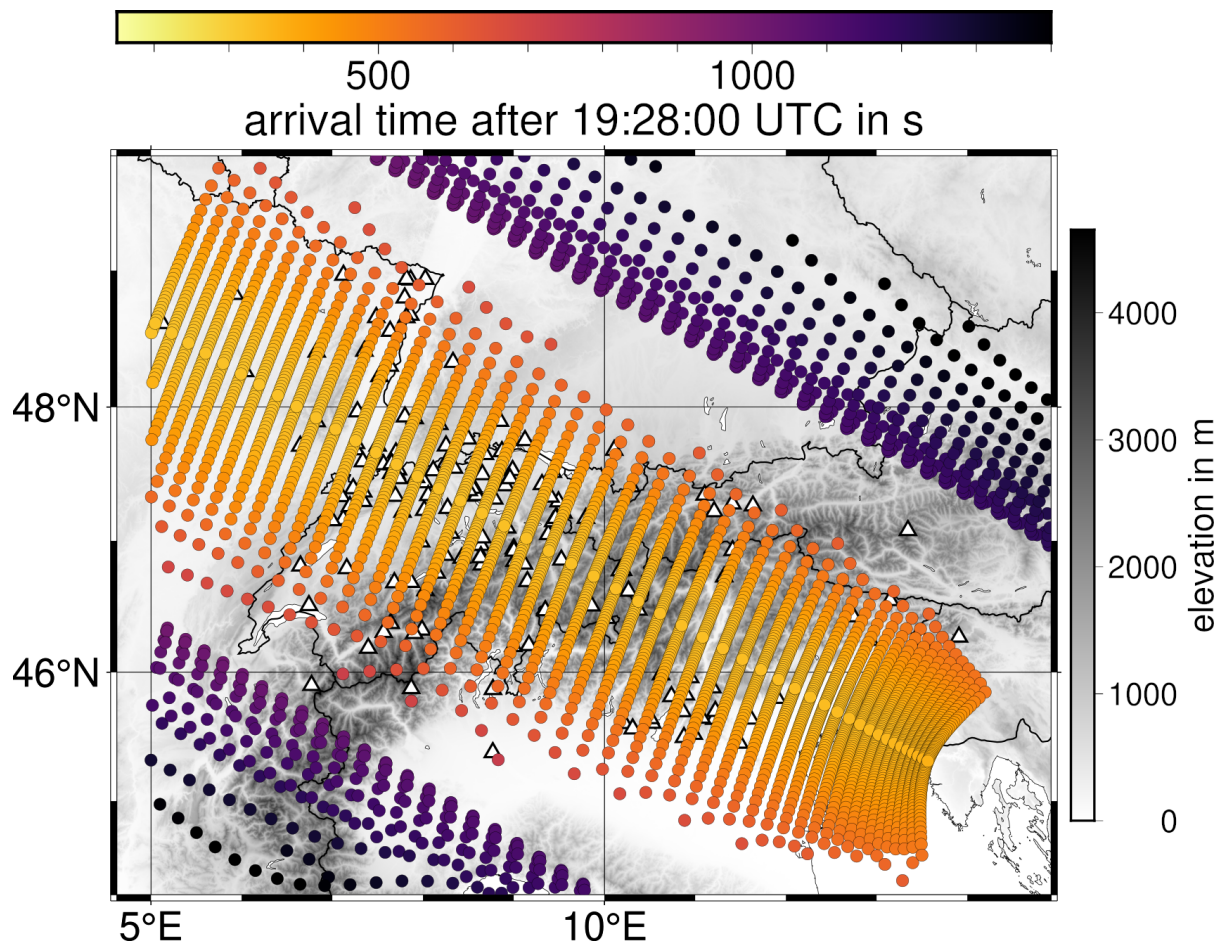


Figure A.3.: Modelled Mach carpet from infraGA ray arrivals at the Earth's surface. Coloured dots represent ray arrivals with an associated arrival time (colour value). Note the area where no rays arrive at the Earth's surface. Even in these areas, we can measure A-S coupled waves at recording stations (white triangles).

A4: Parameters, grid details, and HPC details for wavefield modelling with finite differences
The code WAVE-Simulation (Qin and Bohlen 2022) was used with the following modelling parameters: The 3-D wavefield modelling involved a grid of 1880 x 440 x 660 points, with a grid spacing of 250 m. We calculated 700 s of wave propagation with a temporal resolution of 0.005 s. The width of CPML layer was 15 grid points and the order of the spatial FD operator was 4. The dedicated high-performance computer (HPC) at KIT Geophysical Institute was used for the computations, which specifically included two HPC servers each with the following properties:

1. CPU sockets: 2
2. Cores per socket: 96
3. Threads per core: 2 (hyperthreading enabled)
4. Total number of physical cores: 192
5. CPU type: AMD EPYC 9654 @ 2.4 GHz (CPU max. up to 3.7 GHz)
6. The available memory sums up to 768 GB, providing 4 GB per physical core
7. These nodes are connected with each other by 100 Gbit Ethernet

B. Supplement to chapter 7

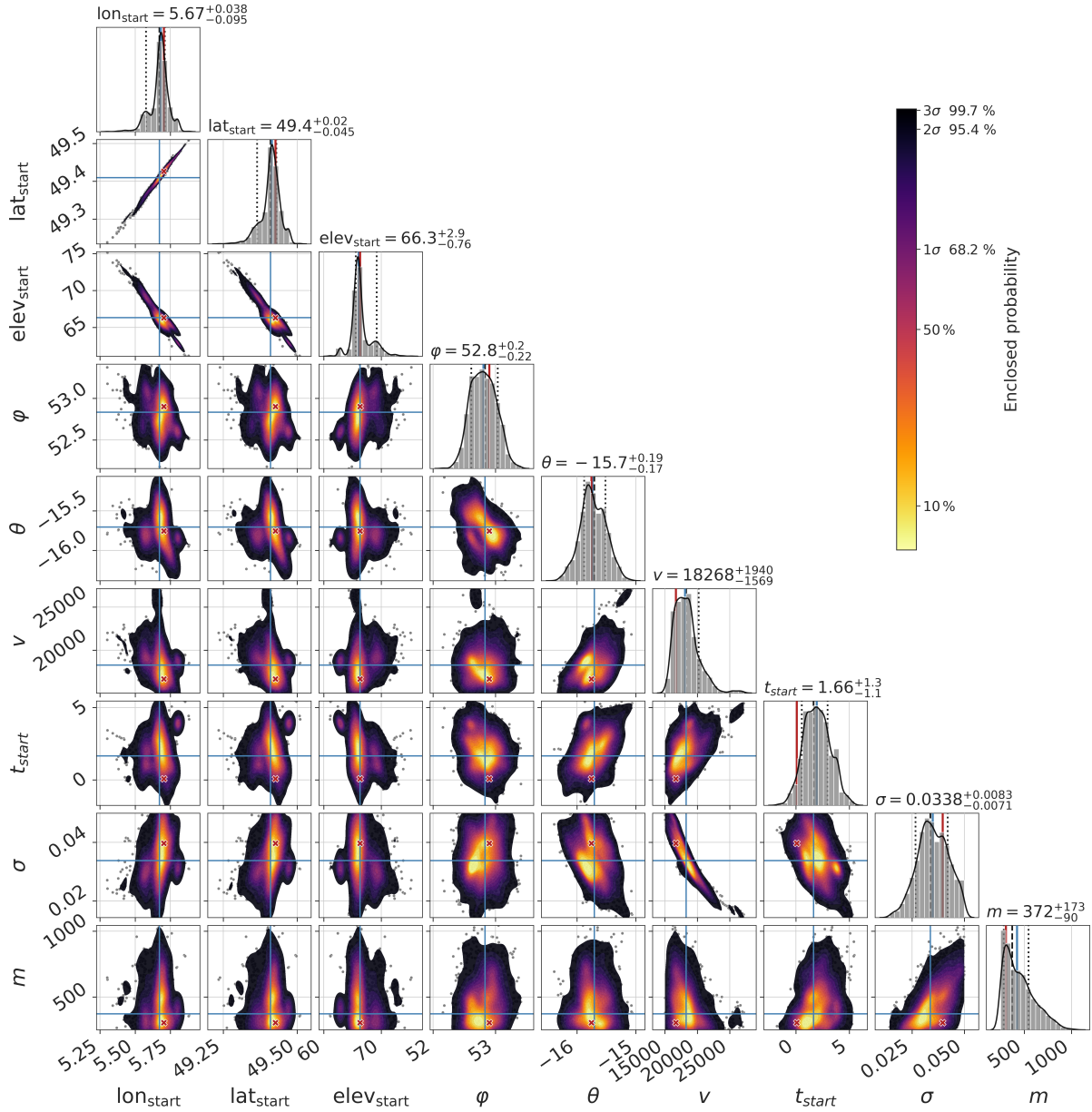


Figure B.1.: MCMC Corner Plot. 1-D (diagonal) and joint posterior distributions of the nine trajectory parameters. In the 1-D distributions the median values (blue line), the 1σ confidence intervals (dashed black lines) and the maximum a posteriori estimates (red line) are shown. The 2-D distributions indicate the median values (blue lines) and the maximum a posteriori estimates (red cross). The colour mapping refers to the enclosed probability by calculating the highest posterior density.

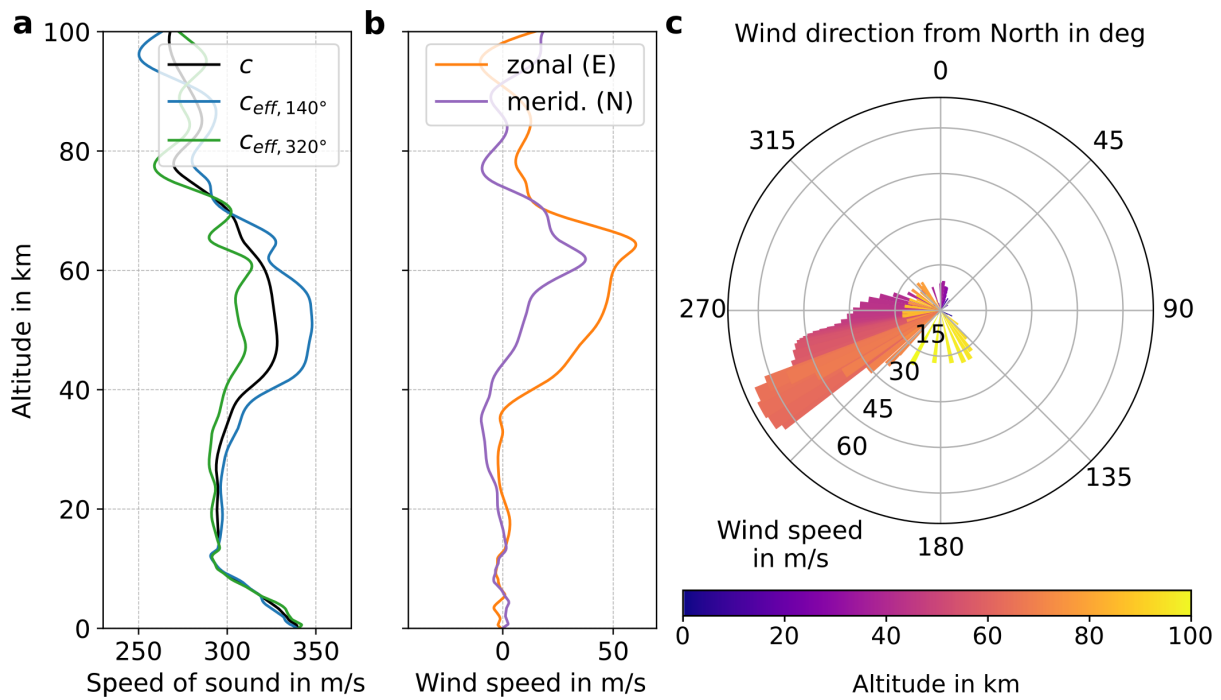


Figure B.2.: Atmospheric state. a) Speed of sound versus altitude (black line) and effective speed of sound perpendicular to the meteoroid trajectory azimuth. Blue: SE direction (140°). Green: NW direction (320°). b) Zonal (orange) and meridional (violet) wind speed components. c) Wind direction from North coloured with respect to altitude, with line lengths representing wind speed.

Parameter	Value
Lon_{start} in deg E	5.7 +0.04/-0.09
Lat_{start} in deg N	49.42 +0.02/-0.05
Alt_{start} in km	66 +3/-1
Azimuth angle φ in deg	52.9 $\pm$ 0.2
Elevation angle φ in deg	15.9 $\pm$ 0.2
Velocity v in km ⁻¹	16.7 +2/-1.6
t_{shift} in s	0.04 +1.3/-1.1
Ablation coefficient σ in s ² km ⁻²	66 +3/-1
$Mass_{start}$ m in kg	305 +180/-90
Bulk density in kg m ⁻³ (Macke et al., 2011)	3,100

* The bulk density is constant during the modelling.

Table B.1.: Trajectory maximum a posteriori (MAP) estimate with 1σ uncertainties.

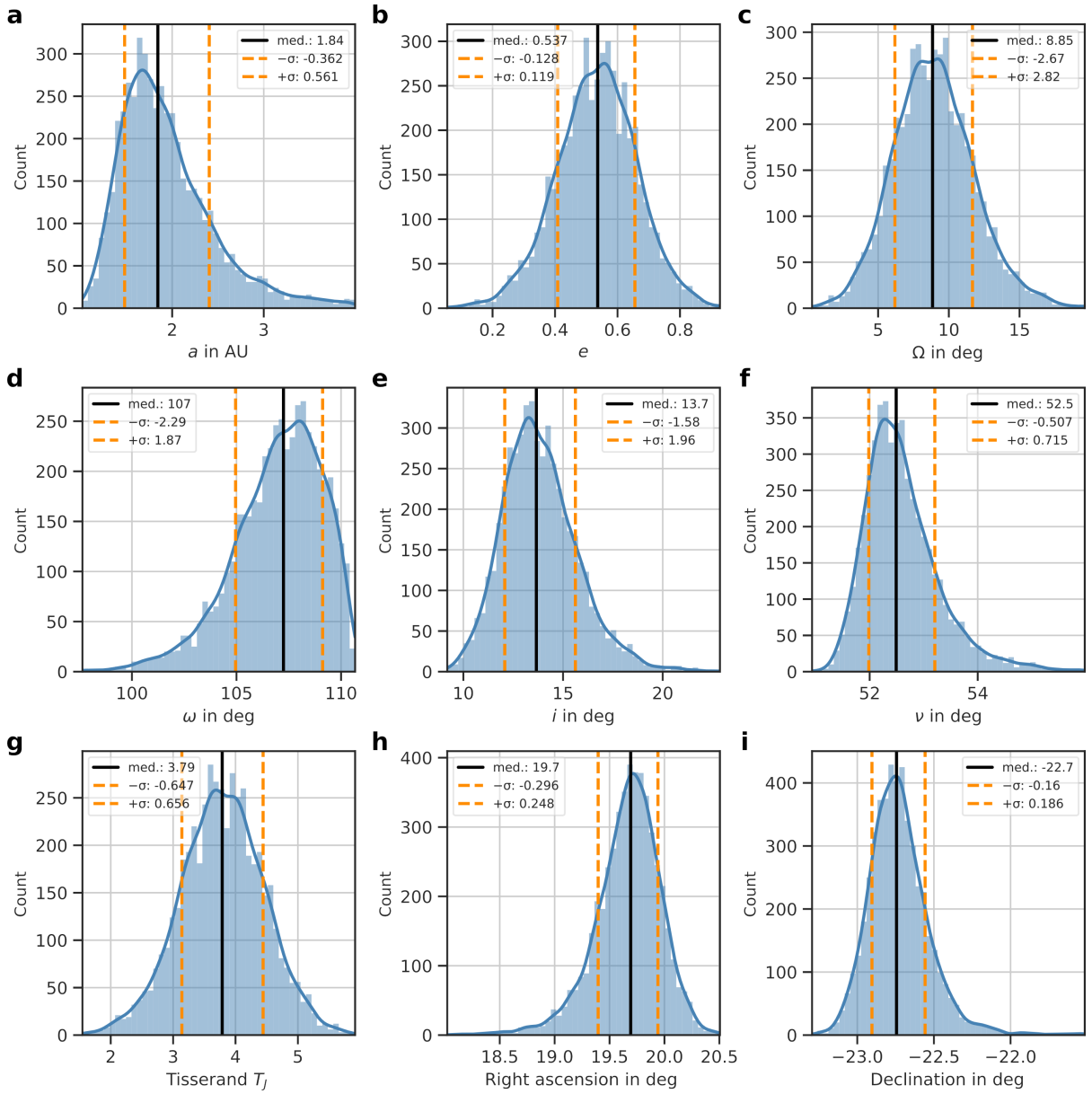


Figure B.3.: Orbital elements and auxiliary orbit-defining parameters. All parameters are presented as histograms (blue bars), with kernel density estimates shown as solid blue lines. The median of each distribution is represented by a black line, and the 1σ uncertainties are shown as orange dashed lines. a) Semi-major axis. b) eccentricity. c) Longitude of ascending node. d) Argument of periapsis. e) Inclination. f) True anomaly. g) Tisserand parameter. h) Right ascension. i) Declination.