

A Shallow Reflector Beneath Krafla Volcano, NE Iceland, Detected With Seismic Migration of Local Earthquake Phases

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Abstract This study explores passive reflection seismic imaging at Krafla volcano, NE Iceland. Krafla hosts a large geothermal area and is one of the few places on Earth where magma has been directly encountered through drilling, at depths of 2.1 km at the IDDP-1 borehole and 2.6 km at borehole KJ-39. These in-situ observations offer a unique opportunity to test seismic imaging in complex media, providing rare ground truth for evaluating imaging performance and resolution. We propose a new migration workflow for earthquake reflection imaging and apply it to local seismicity recorded by a dense array of short-period seismometers at Krafla. Additionally, by integrating synthetic data generated with SPECSEM2D, we benchmark our migration workflow and assess its limitations in the context of imaging with passive sources. Our results reveal a prominent reflector at 2.4 km depth in the area of the IDDP-1 borehole. Considering the uncertainties in earthquake locations, this reflector may correspond to the known magma body at IDDP-1 or outline a local brittle-ductile transition zone within the geothermal system. Key imaging challenges are related to the partially unknown source characteristics, the limited aperture of our station network, and the geological complexity of the subsurface. We show that if not properly accounted for, these factors can contribute to imaging artifacts such as spurious reflectors. Our findings provide new insights into the subsurface structure of the Krafla geothermal system and highlight the potential of migration techniques for passive seismic imaging in complex, highly scattering environments.

Non-technical summary Krafla volcano in NE Iceland is internationally known for the unexpected discovery of magma at shallow depths during geothermal drilling at boreholes KJ-39 and IDDP-1. Despite repeated attempts, these magma bodies have not been clearly resolved to date, highlighting the challenge of imaging small-scale geological structures in volcanic environments. The known location of magma and persistent seismic activity make Krafla an ideal natural laboratory for testing and improving imaging techniques. In this study, we adapt seismic migration, which is an imaging method more commonly used with controlled, human-generated seismic sources, for application to earthquake-generated seismic data. The workflow is benchmarked and findings are validated using numerical simulations. Applied to Krafla, this approach reveals a prominent shallow reflector near the depth where magma was encountered at the IDDP-1 borehole. This reflector may represent the top of the known magma-rock interface or a local brittle-ductile boundary within the Krafla geothermal system. Our analysis demonstrates the potential of adapting controlled-source techniques to earthquake data and provides new insights into the subsurface geology at Krafla.

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1 Introduction

Seismic imaging of magmatic systems remains a major challenge due to the complexity of subsurface structures and their dynamic processes. Rather than forming simple, coherent magma chambers, magma tends to exploit pre-existing crustal weaknesses such as faults and fractures, creating intricate networks of interconnected sills, dykes, and partially molten zones (e.g., Cas-truccio et al., 2017; Sparks et al., 2019; Paulatto et al., 2022). These features are often small in scale, irregular in shape, and have varying physical properties, posing significant challenges for conventional seismic imag-

ing techniques. For example, tomographic approaches often lack the resolution required to detect such fine-scale features and tend to smooth or smear sharp velocity contrasts (Lees, 2007; Paulatto et al., 2022). Seismic reflection methods, in principle, are well suited for imaging discontinuities in the crust due to their sensitivity to sharp impedance contrasts (i.e. contrasts in the product of seismic velocity and density). However, in highly heterogeneous volcanic environments, this sensitivity also becomes a limitation. Resolving small-scale structures requires high frequencies, but high frequency waves are more likely attenuated and scattered at fine-scale heterogeneities—especially in the shallow subsurface, which often comprises loose pyroclastic de-

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posits, fractured basalts, and lava flows. This multiply scattered energy arrives at the receivers simultaneously with deeper reflections, creating complex interference patterns that complicate the reliable identification of reflections. Passive-source imaging introduces additional challenges due to uncertainties in earthquake locations and variable (and often unknown) source characteristics, but offers advantages in terms of cost and logistics. Moreover, sources at depth, such as earthquakes, can be beneficial as the highly scattering near-surface layers are avoided on the source side, in contrast to active surveys where the sources are located on or close to the surface. In this study, we investigate passive reflection seismic imaging at Krafla volcano, NE Iceland, by applying seismic migration to local earthquake phases. Seismic migration aims to reconstruct subsurface reflectivity by back-propagating surface-recorded reflected wavefields to their origin at depth. When multiple sources and receivers are distributed across a 3D area, this approach enables the construction of three-dimensional images of subsurface structures. Krafla is globally recognized as one of the few locations where magma has been directly encountered by drilling (e.g., Elders et al., 2011). Persistent shallow seismicity and the confirmed presence of magma at shallow depths make Krafla an ideal natural laboratory for testing and advancing seismic imaging techniques in complex volcanic environments. Improved understanding of the subsurface at Krafla is directly relevant to geothermal energy operations, supporting safer drilling and more informed risk assessments.

1.1 Geological setting

As one of five active volcanic systems in Iceland's Northern Volcanic Zone, Krafla is located along the onshore continuation of the Mid-Atlantic ridge (Figure 1). The volcano spans approximately 20 km in diameter and features a 8–10 km wide central caldera, formed around 110,000 years ago (Sæmundsson, 1991). A 90 km-long fissure swarm, trending NNE–SSW, intersects the volcano and is characterized by numerous surface fissures and predominantly normal faults aligned with the rift axis (e.g., Einarsson, 1978; Mortensen et al., 2010, 2014). Eruptions at Krafla occur every 300–1000 years (e.g., Sæmundsson, 1991; Ármannsson et al., 2014). The most recent eruptive episode, known as the Krafla Fires, occurred between 1975 and 1984 and was characterized by a series of basaltic fissure eruptions accompanied by substantial ground deformation (e.g., Einarsson, 1978; Brandsdóttir et al., 1997). The system hosts a NW–SE elongated geothermal field covering roughly 10 km², which has been actively exploited for energy production since 1974 (Mortensen et al., 2010). More than 40 wells have been drilled—some reaching depths of up to 2900 m—to tap the hydrothermal reservoir (e.g., Mortensen et al., 2014). Krafla is internationally recognized as one of the few locations where magma has been directly intersected by drilling: in 2008, borehole KJ-39 encountered magma at 2.6 km depth (e.g., Mortensen et al., 2010), followed by a second intersection at 2.1 km in the IDDP-1 well in 2009 (e.g., Elders et al., 2011). Core

samples and borehole logs indicate that the uppermost kilometer of the subsurface is composed of alternating basaltic lava flows and hyaloclastites. At greater depths, the geology is dominated by basaltic intrusions (e.g., Mortensen et al., 2014; Scott et al., 2019; Árnason, 2020). The characteristics and broader structure of the Krafla system have been investigated using a wide range of seismic methods, including seismic tomography (Arnott and Foulger, 1994; Schuler et al., 2015; Glück et al., 2025), shear wave splitting (Tang et al., 2008), moment tensor inversion (Mildon et al., 2016; Schuler et al., 2016; Guðnason et al., 2023), and seismic reflection imaging with both active sources (Brandsdóttir et al., 1997; Reiser et al., 2020; Millett et al., 2020; Kästner et al., 2020) and local earthquakes (Brandsdóttir and Menke, 1992; Kim et al., 2017, 2020; Maass et al., 2025). A magma chamber beneath the caldera, first proposed by Einarsson (1978) with its upper boundary at approximately 3 km depth and later estimated to be 0.75 to 1.8 km thick, is believed to supply heat to the overlying hydrothermal system (Brandsdóttir et al., 1997; Mortensen et al., 2014). More recent studies have imaged seismic reflectors at depths near the magma encountered at IDDP-1 (Kim et al., 2020; Reiser et al., 2020; Maass et al., 2025). However, no study to date has conclusively resolved the full extent or geometry of the shallow magma body, highlighting the challenges of imaging magmatic intrusions in volcanic settings.

1.2 Study objectives

Building on the efforts of previous studies, we apply reverse vertical seismic profiling (RVSP) to a new dataset acquired in 2022 with a densely spaced seismic network. We adapt and parallelize the migration algorithm in a new Python-based framework optimized for passive-source imaging. The goals of this study are to (1) image the structure beneath the IDDP-1 borehole, and (2) assess the limitations and challenges of passive-source reflection seismic imaging at Krafla. We also incorporate synthetic waveforms to benchmark our methodology and validate results. Several examples of earthquake wavefields are shown throughout this work to illustrate the variability and complexity of seismic data recorded at Krafla. Unless otherwise noted, all depth values in this study refer to depth below the surface, defined as the elevation of the IDDP-1 wellhead at 554.16 m above sea level.

2 Station setup and earthquake locations

Between June and July 2022, we deployed 114 short-period (5 Hz) seismic nodes at Krafla volcano as part of the IMPROVE project (Figure 1). The network consists of two dense linear profiles with an inter-station spacing of 30 m: Line 1 (L1) spans approximately 1 km with 33 stations, while Line 2 (L2) extends about 2 km with 71 stations. The two lines intersect at the IDDP-1 borehole. The 21 northernmost stations on line L2 and the 14 northernmost stations on line L1 record three components (3C), the remaining stations provide only one-

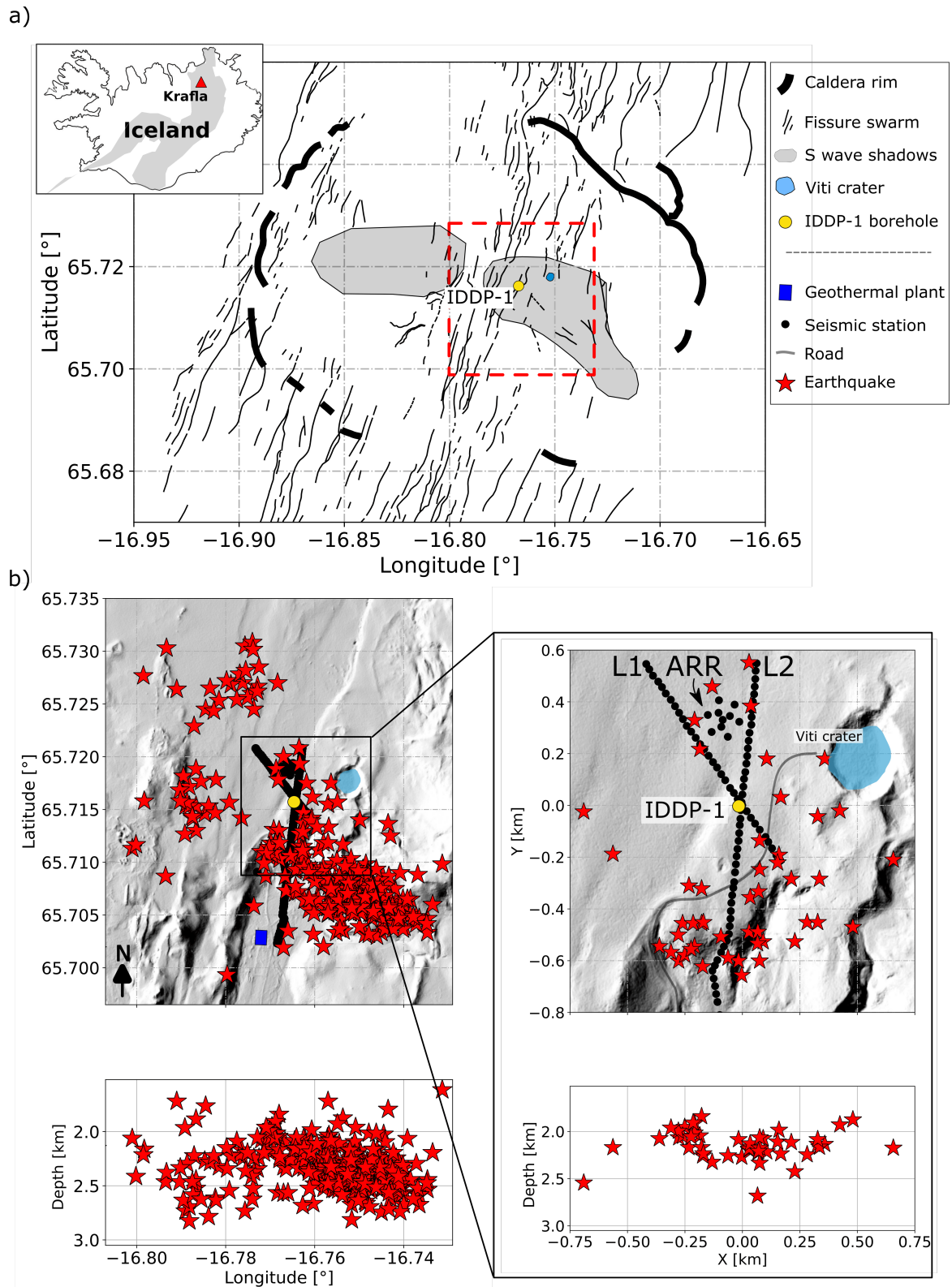


Figure 1 Study area. a) Location of Krafla volcano in Iceland. The black line outlines the caldera rim. The red dashed line marks the area investigated in this study shown in b). NNE–SSW trending fissures and faults traverse the region (e.g., [Sæmundsson, 1991](#)). Areas shaded with gray color represent shadow zones from [Einarsson \(1978\)](#), which have been interpreted as potential magmatic bodies. b) Distribution of seismic stations deployed during a 2022 field campaign (black dots) and the locations of over 300 local earthquakes from two earthquake catalogues ([Guðnason et al., 2023](#); [Glück et al., 2025](#)), recorded during the deployment period (red stars). The black box marks the region analyzed in this study and is shown in detail in the inset. The station profiles (L1 and L2) intersect at the IDDP-1 borehole (yellow dot), and an additional array (ARR) was deployed north of the borehole. 48 local earthquakes located within a 0.7 km radius of the IDDP-1 with a median depth of 2.1 km are included in the imaging analysis.

component records (1C). In addition, we deployed ten 3C-short period (1 Hz) seismometers as an array (ARR) with an aperture of 150 m north of IDDP-1. The sampling rate of the seismic nodes was 250 Hz, while the array stations recorded data at 200 Hz. Across the network, the maximum difference in elevation was 90 m. For the imaging workflow presented in this study, we used all stations of station profile L1 as well as the northernmost 58 stations of station profile L2. Stations further south on L2 were excluded from the analysis due to poor data quality resulting from the interference of noise from the nearby power plant (Figure 1). During the five-week recording period, more than 300 local earthquakes were recorded. Local 1D-velocity models as well as an earthquake catalogue with 306 earthquakes, located with 12 broadband stations across the Krafla region, were kindly provided by ÍSOR (Guðnason et al., 2023, and references therein). These events were manually selected and refined from automatic SeisComp locations. In addition, we use a refined catalogue by Glück et al. (2025), comprising 223 events located with a dense local network of 100 stations using the NonLinLoc software (Lomax et al., 2000, 2009). Both earthquake catalogues provide absolute locations. For events appearing in both catalogues, we use the locations from Glück et al. (2025) because of their reduced uncertainties due to the denser network coverage. To image the subsurface structure around the IDDP-1 borehole, we selected 48 local earthquakes that occurred within a 0.7 km radius of the borehole (Figure 1c). These events have focal depths ranging from 1.8 km to 3.6 km with a median depth of 2.1 km. Uncertainties of event longitudes, latitudes, and depths vary between 200 m and 500 m. Local magnitudes range between -0.47 and 0.33.

3 Data

3.1 Real data

The recorded seismic wavefields at Krafla are complex and strongly affected by local site effects. This is illustrated in Figure 2, which shows vertical-component (Z) recordings from all stations of profile L1 for an earthquake which occurred at 2.0 km depth. Data were bandpass-filtered between 5 and 25 Hz using a second-order Butterworth filter. Theoretical traveltime curves for direct P- and S-waves were calculated using a ray-tracing algorithm (Margrave, 2000) with 1D velocity models (Figure 2e) and are overlaid on the time series for comparison. Despite the close station spacing of 30 m, waveform amplitudes vary significantly across the profile—especially in the S-wave coda (Figure 2a). This variability reflects differences in coupling conditions and local site effects, highlighting the lateral geological variability of the shallow subsurface. To quantify wavefield coherency, we apply a beamforming approach (e.g., Rost and Thomas, 2002). A grid search over slowness values ranging from -0.1 s/km to 0.1 s/km is performed to account for a range of plausible (apparent) velocities. At each time step, coherency is evaluated within a 0.2 s sliding window and defined as the

maximum semblance (Neidell and Taner, 1971) across all tested slowness values. Coherency increases with the arrival of the direct P-wave, gradually decreases in its coda, and rises briefly with the direct S-wave arrival before dropping to low values throughout the S-wave coda (Figure 2b). The excitation of incoherent energy in the S-wave coda is indicative of scattering from heterogeneities in the lithosphere (e.g., Aki and Chouet, 1975; Sato et al., 2012). Additionally, with increasing time, cross-talk between different phases becomes more likely, as multiple arrivals following increasingly complex and diverse propagation paths interfere at the receivers. This can further degrade wavefield coherency. Maass et al. (2025) showed that seismic waveforms at Krafla are largely dominated by heterogeneity in the shallow near-surface, with local seismic resonances contributing to site effects at specific stations. In Figure 2d, each trace is normalized by its maximum absolute amplitude to improve the visibility of weaker signals, with insets highlighting the P- and S-wave coda (Figure 2f and g). While the theoretical traveltimes match the P-wave arrival reasonably well, they do not perfectly align with the observed onsets. This is likely due to inaccuracies in earthquake locations and/or discrepancies between the 1D velocity model and the actual seismic velocity structure. In contrast to the S-wave coda, the P-wave coda contains coherent phases that propagate along the station profile, carrying valuable information about subsurface structures at Krafla. In this study, we exploit the coherent energy in the P-wave coda to image the subsurface structure near the IDDP-1 borehole.

3.2 Synthetic data

To benchmark our methodology and validate results, we use the spectral-element code SPECFEM2D (Tromp et al., 2008) and simulate the wavefield from each earthquake recorded at the receivers. We choose a Dirac delta function as source time function, consistent with the broad frequency content observed in real data (Maass et al., 2025) and the small magnitudes of the events, which suggest short rupture durations. To produce reflected phases, we embed a flat reflector at a depth of 2.4 km within the 1D velocity model provided by ÍSOR (blue dashed line in Figure 2e). The rationale for this specific reflector depth will become clear in later sections. Although real-world structures such as magmatic sills may be very thin, the reflector is modeled as a low-velocity boundary over a halfspace for simplicity. The synthetic data are used to benchmark the method and assess its limitations, rather than to directly invert for the dimensions of subsurface structures. After simulation, synthetic waveforms are bandpass-filtered between 5 and 25 Hz to match the frequency content of the observations.

4 Methods

4.1 The migration approach

Seismic migration aims to construct a three-dimensional image of subsurface reflectivity by

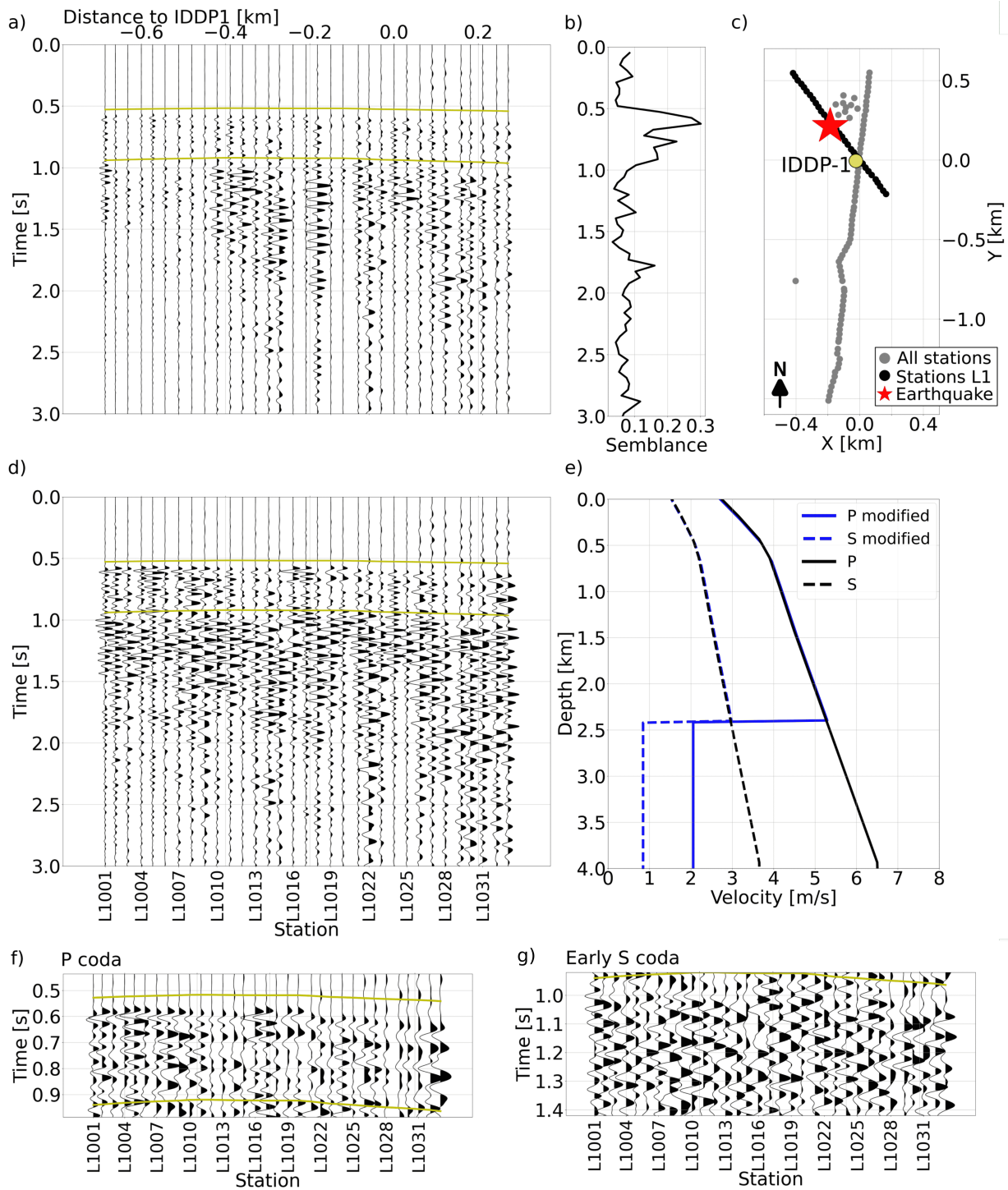


Figure 2 Example earthquake (Z-component records) recorded by all stations of station line L1. The event occurred on 2022-07-02 at 07:40:04 am at a depth of 2 km with a local magnitude of -0.35. Data are bandpass-filtered between 5 and 25 Hz. a) Raw vertical-component recordings, highlighting significant amplitude variations across the profile despite a small interstation distance of 30 m. Yellow lines indicate theoretical arrival times for direct P- and S-waves, calculated using a raytracing algorithm with the local 1D velocity model (see main text). b) Semblance values indicating the coherency of phases across the profile as a function of time. c) The black dots show seismic stations of profile L1, and the red dots marks the epicenter of the event. d) Same wavefield as in a) but with each trace normalized by its maximum absolute amplitude, enhancing the visibility of weak phases in P- and S-codas. e) 1D velocity models for P- and S-waves provided by ÍSOR (black lines), and modified velocity models used for synthetic modeling of seismic waves using SPECFEM2D (blue lines). It should be noted that the low-velocity body introduced in the model is not expected to be as thick as 1.5 km. f) and g) Insets highlighting the P-wave coda and early S-wave coda. While the P-wave coda shows coherent energy across stations, the S-wave coda is more complex and largely incoherent across the traces.

backpropagating recorded seismic energy to its origin at depth. A comprehensive overview of different migration methods can be found in Gray et al. (2001). While migration is traditionally applied in controlled-source seismology (e.g., Gazdag and Sguazzero, 1984; Gray et al., 2001; Hloušek et al., 2015; Martuganova et al., 2022), it has also been successfully used with passive sources on local (e.g., Stroujkova and Malin, 2000; Reshetnikov et al., 2010; Nakata and Shelly, 2018; Quiros et al., 2017) and regional to global scales (Hutko et al., 2006; Luo et al., 2022; Wolf et al., 2025). In this study, we apply a reverse vertical seismic profiling (RVSP) approach, which is based on summing amplitudes from multiple seismic traces along predicted reflection trajectories, under the assumption that reflected energy originates from a common subsurface region. This method is conceptually similar to the normal moveout (NMO) correction and the common-depth-point (CDP) binning and stacking widely used in controlled-source seismology (e.g., Yilmaz, 2001). While RVSP can in principle be applied to any wave type — including converted phases and multiples — we restrict our analysis to P–P reflections. In the following, we outline the methodology which has been adapted for use with earthquakes. Parameter choices specific to our setting will be discussed in Section 4.2.5.

4.1.1 Raytracing with velocity models

In the case of a horizontally stratified Earth and sources on the surface, the reflection (or “bounce”) points at varying depths lie along a vertical line midway between the source and receiver (e.g., Schneider, 1984; Yilmaz, 2001). However, this is not the case when the sources are located at depth (e.g., Inamori et al., 1992; Quiros et al., 2017). Figure 3a and b show that in the case of earthquakes, the reflection points for a given source–receiver pair lie along a curved, hyperbolic trajectory. The exact shape of this trajectory is governed by the wave type, epicentral distance, source depth, and the seismic velocity structure. Notably, the horizontal spacing between bounce points decreases with increasing reflector depth (Figure 3b), leading to more tightly clustered reflection points at greater depths. To enable the grouping of seismic traces based on common reflection points and to accurately sum amplitudes along predicted reflection paths, both the bounce points and traveltimes of reflections must be computed. The first step in our migration workflow is therefore the calculation of reflection bounce points and traveltimes for each earthquake–station pair. Assuming flat reflectors at discrete depths down to 4.5 km, we perform raytracing using the CREWES-Software (Margrave, 2000) with a local 1D velocity model (Section 2). Each reflection is uniquely characterized by its spatial bounce point (x_R, y_R, z_R) and its traveltime t_R . We emphasize that the Raytracer is utilized only for computing traveltimes and not for amplitude modeling.

4.1.2 Binning and stacking

Following raytracing, the study area is discretized into a regular 3D grid. The orientation and resolution of the

grid are chosen based on the dominant seismic wavelength, the distribution of bounce points, and the desired spatial resolution (Section 4.2.5). For each grid cell, or bin, the traces (source and receiver combinations) are gathered which contain reflections whose estimated bounce points fall within it. For example, as illustrated in the schematic sketch in Figure 3c, two earthquakes produce reflections that intersect a particular grid cell and are recorded by two surface stations. The corresponding seismic traces are grouped into a ‘common-image-point’ (CIP) gather, analogous to a CDP gather in controlled-source surveys (Figure 3d). Within each CIP gather, amplitudes at the predicted reflection arrival times are extracted from the real data, and their mean is computed. This average amplitude is used as a proxy for the local reflectivity at the center of the grid cell. The procedure is repeated for all bins in the 3D matrix, resulting in a migrated image volume in which each voxel represents the stacked amplitude—i.e., the estimated reflectivity—at that location in the subsurface. In principle, various waveform attributes could be stacked, including squared amplitudes and envelopes. However, in this study, we use raw amplitudes—including both positive and negative values—for reasons that will be discussed in Section 6.3.

4.2 Data pre-processing and parameter choices

4.2.1 Filtering and data normalization

As we focus on P–P reflections, we restrict our analysis to raw vertical (Z-component) recordings. After testing multiple frequency bands, we applied a first-order Butterworth bandpass filter between 5 and 25 Hz. Frequencies below 5 Hz were excluded due to contamination from low-frequency noise generated by the nearby geothermal power plant. Frequencies above 25 Hz were discarded due to incoherent scattering and noise, which degraded the imaging quality. Prior to filtering, all time series were tapered and their mean was subtracted. To account for the significant amplitude variability observed in the wavefields due to coupling conditions, site effects (Figure 2) as well as varying earthquake magnitudes across events, we normalized each trace by its root-mean-square (RMS) amplitude.

4.2.2 Alignment with velocity models and trace selection

As discussed in Section 3.1, observed direct P-wave arrivals do not align perfectly with predictions from the 1D velocity model due to discrepancies between the 1D model and the actual velocity structure, uncertainties in earthquake locations, or a combination of both. However, the migration workflow relies on accurate reflection traveltime calculations and amplitude stacking under the assumption of a 1D velocity structure. Therefore, direct P-waves in the recorded data must be aligned with the predictions based on the 1D velocity model. To achieve this, we apply a data-driven correction. First, an STA/LTA algorithm (Earle and Shearer, 1994) is used to detect the onset of direct P-waves in

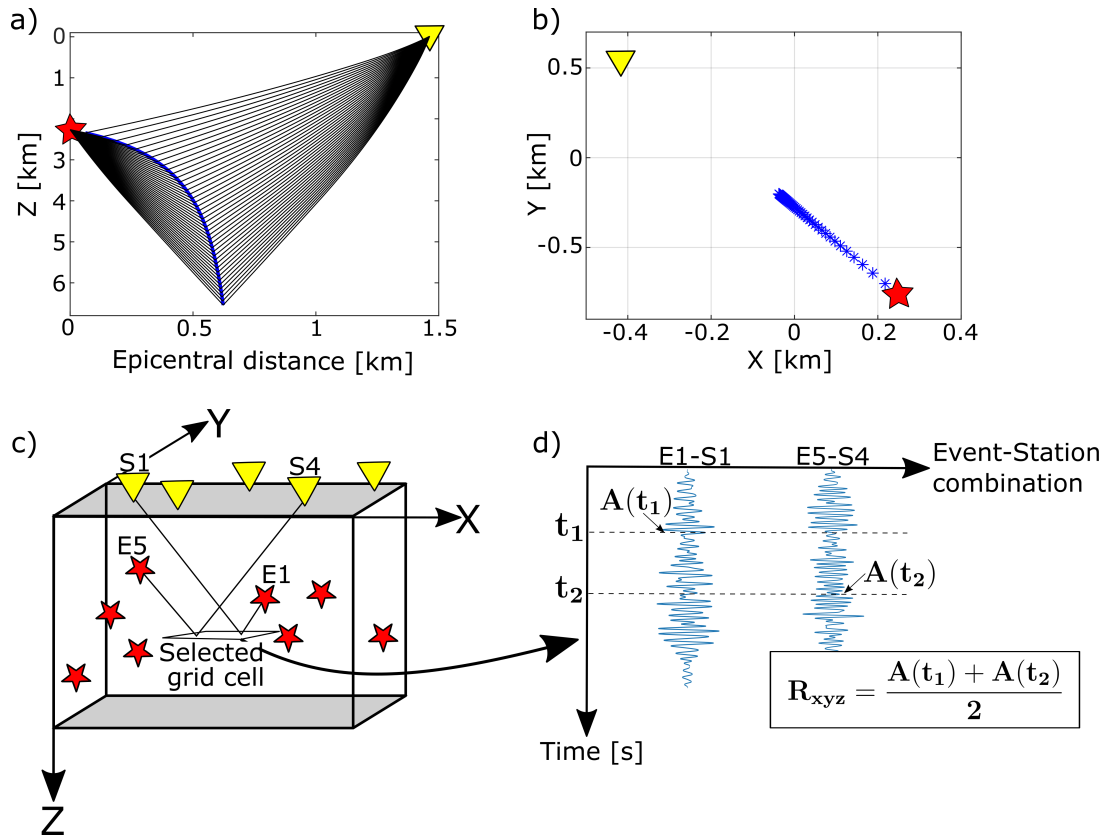


Figure 3 Illustration of the migration method using local earthquakes. Red stars represent earthquakes, yellow triangles indicate seismic stations. a) Trajectories of reflections for an arbitrary earthquake and station from our dataset, as a function of epicentral distance and depth (Z). The reflection bounce points lie on a hyperbola (blue line). Note that we only show every 5th trajectory for visualization purposes. b) Same earthquake, station and reflection bounce points as in a) but in the X - Y plane. The horizontal spacing between bounce points decreases as the epicentral distance increases. c) Conceptual sketch of the binning procedure. A grid cell, defined through its coordinates (x, y, z) in our model is selected. Two event-station combinations (E1-S1, E5-S4) yield reflections whose bounce points fall in the selected grid cell. d) The corresponding seismic traces are selected and the amplitudes at the reflection traveltimes are stacked in order to obtain the reflection strength R_{xyz} within the selected grid cell.

the real data. A time window of approximately one P-wave period is then cross-correlated with a synthetic Ricker wavelet generated at the expected arrival time t_P from the 1D velocity model. Each trace is subsequently shifted based on the time lag corresponding to the cross-correlation peak. After the alignment of the direct P-wave arrivals with the 1D velocity model, we apply an additional time shift $t_{\text{diff}} = t_P - 0.5$ to each trace, aligning direct P-waves with a fixed reference time of 0.5 s, which is an arbitrary choice. This ‘flattening’ of the direct P-wave simplifies polarity analysis and allows for the fast and efficient incorporation of different velocity models, as all reflected phases can now be expressed relative to a fixed reference time. Following the alignment, all reflection traveltimes are updated accordingly via $t'_R = t_R - t_{\text{diff}}$. We also visually inspect each trace and discard those where the direct P-wave arrival cannot be reliably identified due to high noise levels. By correcting for differential traveltime errors between observed and predicted P-wave arrivals, we reduce the impact of model inaccuracies and inaccurate earthquake locations. In Figure 4a and b, the original data are compared with a preprocessed version after waveform alignment and removal of severely noise-contaminated traces for

station profile L2 and an example earthquake located south of IDDP-1.

4.2.3 Accounting for varying focal mechanisms

Variations in focal mechanisms across earthquakes lead to differences in radiation patterns, and potentially varying polarities of reflected phases. If left uncorrected, these variations can result in destructive interference during stacking, reducing the amplitude of reflections and the continuity of reflectors in the migrated image. To address this, we apply a polarity correction and event-clustering strategy. First, all wavefields are visually inspected. Since direct P-waves are aligned to 0.5 s, a simple linear stack across traces provides a robust indicator of the first-motion polarity (Figure 4c). Traces which show negative first motions are multiplied with -1 to ensure uniform positive polarity across the dataset. Aligning the polarity of direct P-waves does, however, not guarantee consistent polarity for reflected phases. For example, due to the different outgoing directions, the direct and reflected P-waves may fall to the different sections of the radiation pattern, leading to the opposite polarities of these waves. This is particularly expected for reflections from deeper structures

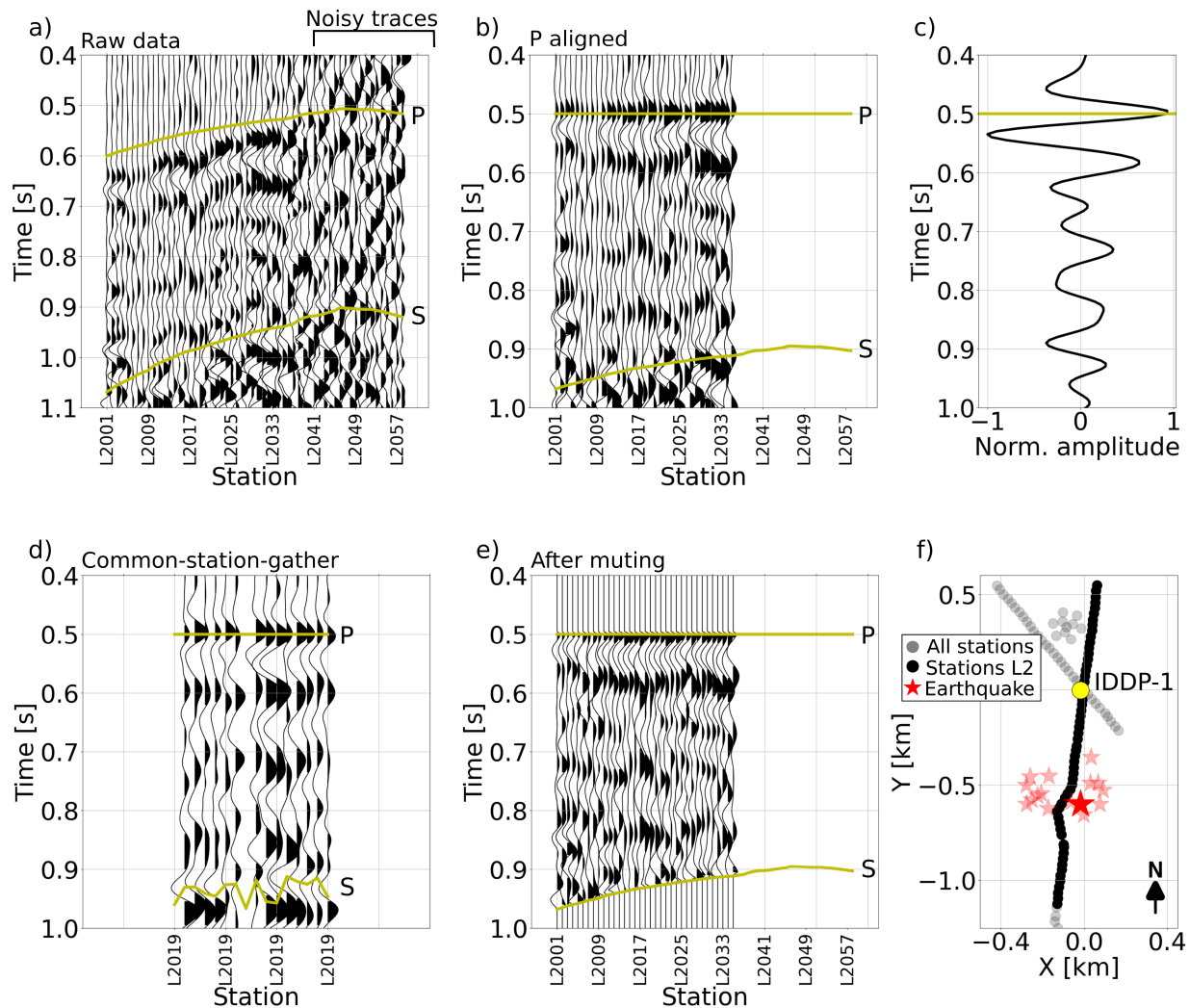


Figure 4 Pre-processing of data for an example earthquake, recorded by stations of the station profile L2. Yellow lines mark theoretical arrival times for direct P- and S-waves. a) Raw data, filtered between 5 and 25 Hz. Each trace was normalized by its RMS value. b) Data after alignment to the direct P-wave arrival and removal of traces with poor data quality. c) Linear stack of the gather shown in b), indicating the positive polarity of the direct P-wave. d) Common-station gather for an arbitrary station on profile L2. e) Data after muting energy prior to the maximum of the direct P-waves as well as from the direct S-wave arrivals onwards. f) Seismic network (gray dots) and stations along Line L2 with sufficient data quality (black dots), as described in Section 3.1. The earthquake shown in panels a), b), and e) is marked by a large red star, while light red stars indicate epicenters of events included in panel d).

for which take-off angles differ significantly from those of the direct waves. Thus, even if we align direct P-wave polarities, those of reflected phases may still be inconsistent across events. To assess and correct for this effect, we analyze common-station gathers (CSGs), which compile waveforms from multiple events recorded at the same station. For each event in the dataset, we perform clustering based on nearby events within a 500-meter epicentral radius. For instance, the CSG shown in Figure 4c includes traces from the main event displayed in panels a) and b), along with events occurring within this 500-meter radius. High waveform coherency within the CSG indicates that reflected phases in the P-wave coda exhibit consistent polarity across events. Overall, waveforms within the clusters show similar characteristics, and we observe no significant polarity reversals among reflected phases. Figure 5 compares two earthquake wavefields close to the IDDP-1 with synthetic data generated using SPECFEM2D (Sec-

tion 3.2). In the synthetics, the P-P reflection from the boundary at 2.4 km depth appears at approximately 0.6 s, consistent with the expected traveltimes as confirmed via raytracing (blue dashed lines). Although the real data exhibit more scattering, clear coherent phases are registered with comparable arrival times. Note the wavefield complexity and lack of coherency in the real-data S-wave coda, as discussed in Section 3.1.

4.2.4 Muting of phases

To suppress imaging artifacts caused by interfering wavefronts, we mute all energy from the arrival time of the direct S-wave onward by zeroing amplitudes. Similarly, we mute all energy prior to the direct P-wave arrival. Thus, in the resulting dataset, only the P-wave codas are retained (Figure 4e). The rationale and implications of this processing step are discussed in detail in Section 6.1.

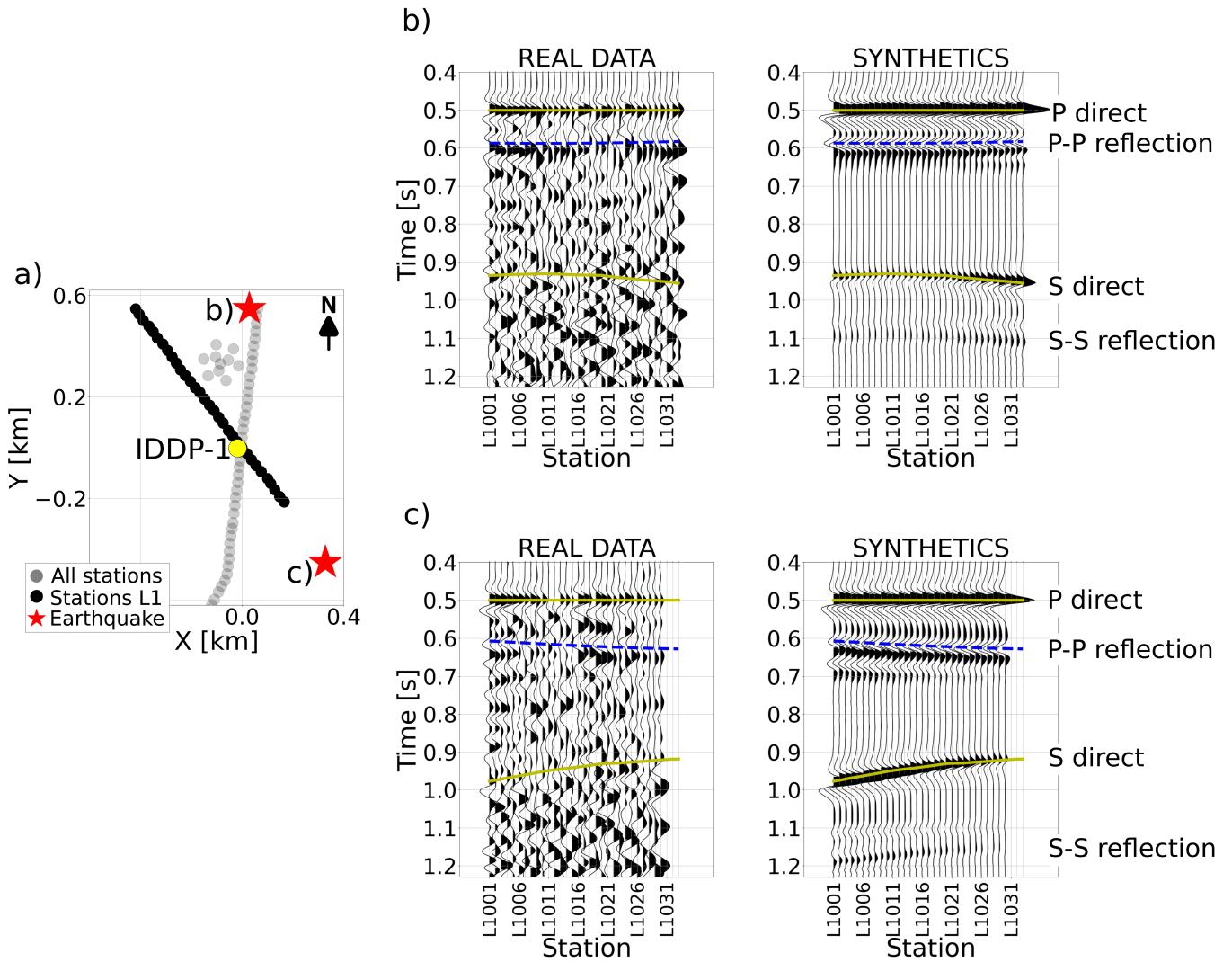


Figure 5 a) Location of the two earthquakes shown in b) and c) relative to the seismic network and the IDDP-1 borehole. b) Comparison between real and synthetic data for an event north of the IDDP-1 after application of the pre-processing, except for the muting of phases (Section 4.2). A coherent arrival is observed in the real data shortly after 0.6 s, consistent with the predicted traveltime of a P–P reflection from 2.4 km depth, as computed using a modified 1D velocity model (Section 3.1, Figure 2e). c) Same as in b) but for an event south-east of the IDDP-1.

4.2.5 Designing the grid

To implement the migration method, the computational domain is discretized into a three-dimensional grid covering the study area in the horizontal (X , Y) and vertical (depth, Z) dimensions (Section 4.1). The vertical grid spacing, dz , controls the depth sampling and determines the time step used for extracting amplitudes from seismic traces. We set $dz = 20$ m. Assuming the lowest P-wave velocity $v_{P_{min}} = 2600$ m/s in the velocity model, this corresponds to a maximum time increment of $dt = dz/v_{P_{min}} = 0.0076$ s. This ensures that the shortest period included in the analysis, 0.04 s (corresponding to 25 Hz), is sampled more than five times per cycle, providing sufficiently dense temporal sampling and minimizing the risk of cycle skipping. The horizontal grid spacing, dx and dy , is primarily constrained by the fold, defined as the number of reflection points within each grid cell. A coarser spacing increases computational efficiency and results in a higher and more uniform fold, but a finer grid provides higher

spatial resolution. Figure 6 shows the fold distributions at depths of 2.1 km, 2.5 km, and 3.0 km for three different horizontal grid spacings (50 m, 70 m, and 100 m). The grid is rotated by -7° to be aligned with station line L2, maximizing the fold along cross-sections parallel to L2 (e.g., Quiros et al., 2017; Kim et al., 2020). At 2.1 km, the coverage of reflection points is sparse due to the limited number of earthquakes that occur above this depth (Figure 1). This restricts our ability to image shallow structures. The reflection coverage improves significantly at 2.5 km and 3.0 km, resulting in more uniformly resolved structures at these depths. Similar to the vertical grid spacing, the horizontal grid also depends on the seismic wavelength. To conform with the Nyquist theorem, the apparent wavelength across the array must be sampled at least twice to avoid spatial aliasing. In our dataset, direct waves and waves within the P-wave coda arrive at steep incidence angles, resulting in high apparent velocities and long apparent wavelengths across the array (Figures 2 and 4). This is primarily due to the small aperture of the station lines (approximately 1 km for L1

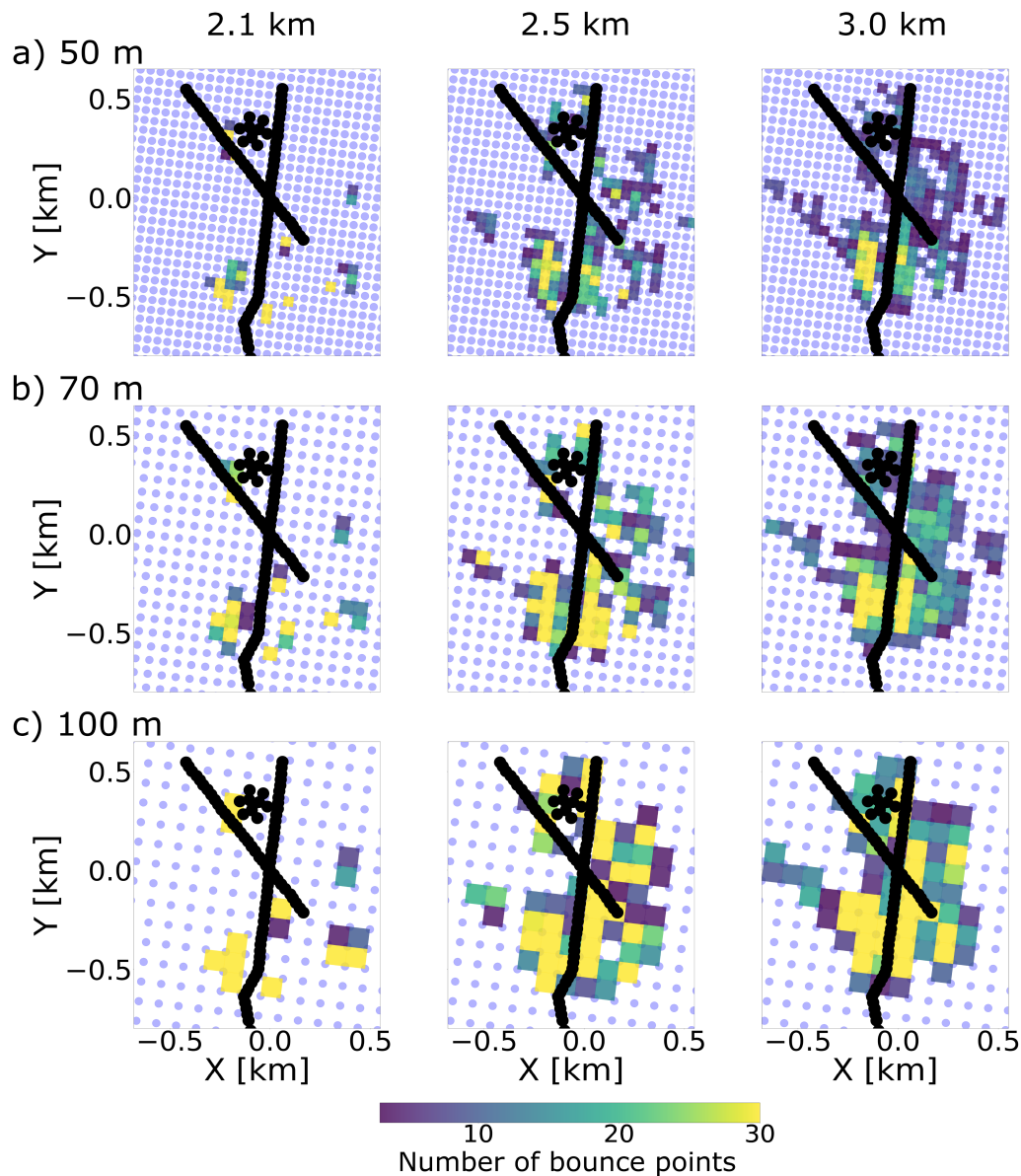


Figure 6 Distribution of reflection points (fold) in our study area for three considered horizontal grid cell sizes (50 m, 70 m and 100 m) in a), b) and c), respectively. The fold is shown for different depths (2.1 km, 2.5 km and 3.0 km). Only fold values larger than 3 (reflection points per grid cell) are displayed.

and 2 km for L2) combined with earthquake sources located directly beneath the arrays at depths greater than 2 km. For this reason, we do not expect spatial aliasing to occur, even for a relatively coarse horizontal spacing (100 m). After testing various horizontal grid spacings, we select $dx = dy = 70$ m as an effective compromise between image resolution and computational cost.

4.3 Post-processing

The migration procedure yields a 3D volume representing subsurface reflectivity across the study area. This volume can be visualized using vertical (depth) cross-sections. Figure 7b presents a South–North (S–N) trending cross-section located approximately 210 m east of the IDDP-1 borehole, providing an image of the crustal structure as a function of depth and Y-coordinate. The

cross-section reveals complex patterns that may reflect significant geological heterogeneity. Despite this complexity, a localized zone of increased reflectivity is observed below 2.3 km depth, extending over a segment roughly 600 m long in the northern part of the section. As the final processing step in our workflow, we apply a coherency-enhancing filter. This filter is conceptually similar to the beamforming method described in Section 3.1, but operates in the image domain. For each grid cell in the 2D-cross-sections, the filter performs a grid search over a plausible range of slopes within a predefined aperture in the horizontal (X) and vertical (Z) dimensions, centered on the target grid cell. For each slope value, semblance is computed as a measure of coherency. The maximum semblance value found during the slope search is used to scale the amplitude at each grid cell. In our implementation, we use a hori-

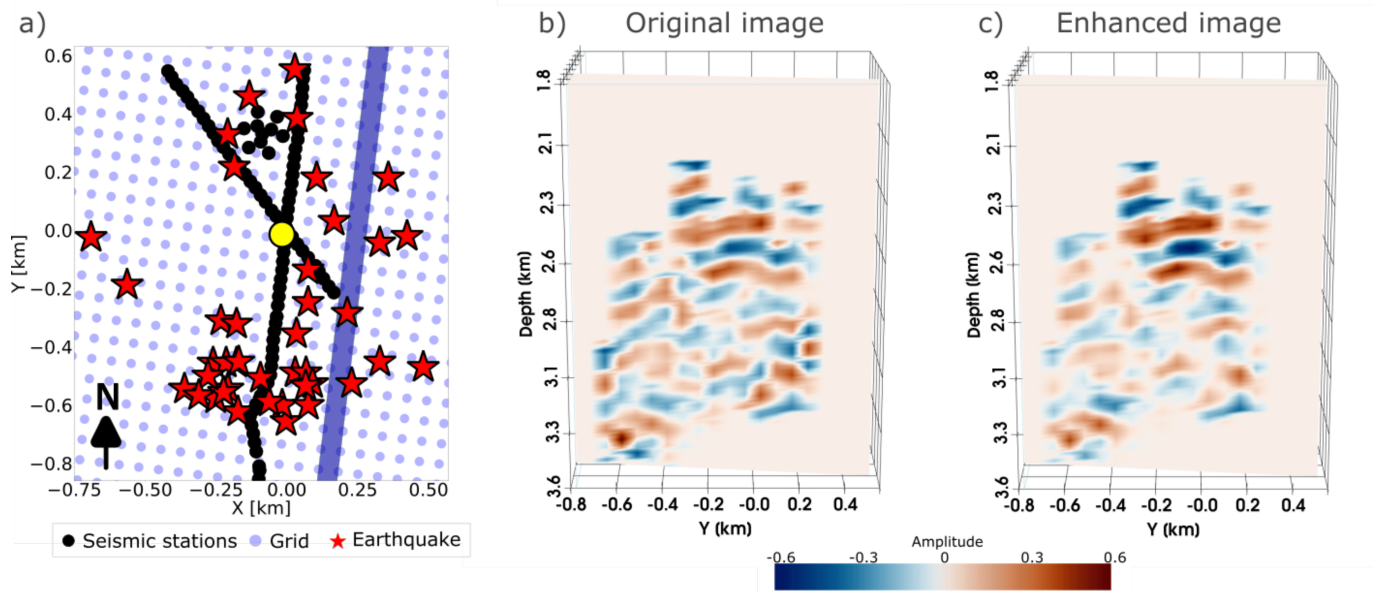


Figure 7 Illustration of the effect of the coherency filter applied in the image domain. a) Study area and grid. The cross-section shown in b) and c) is indicated with purple color and the yellow dot marks the IDDP-1 borehole. b) and c) Original image and enhanced image after application of the filter. Laterally discontinuous structures are suppressed in the enhanced cross-section.

zontal aperture of 210 meters (3 traces) and a vertical aperture of 70 depth samples, corresponding to 140 meters in depth. With the filter applied, the reflector below 2.3 km depth stands out more clearly in the image (Figure 7). The procedure enhances the continuity of flat-lying and gently dipping reflectors, while suppressing laterally incoherent energy. We note that this enhancement may come at the expense of true steeply dipping features, such as faults. Therefore, the filtered images should be interpreted with caution.

5 Results

5.1 A reflector at 2.4 km depth

Figure 8 presents selected cross-sections from the 3D imaging results. Panels b) and c) show vertical slices in the X - Z (E-W trending) and Y - Z (S-N trending) planes, while panels d) and e) provide side views of the crustal structure. A prominent and predominantly flat reflector is imaged at approximately 2.4 km depth. In the E-W trending section located about 200 m south of the IDDP-1, the reflector extends laterally from approximately 300 m west to 350 m east of the IDDP-1 borehole. In the S-N oriented section at around 140 m east of IDDP-1, the same reflector is visible from approximately 200 m south to the borehole position. Figure 8g and h show that this reflector is also visible on S-N oriented cross-sections at 70 m and 250 m east of the borehole. Outside this depth range, the images show rather complex and scattered reflectivity, even though some other laterally coherent structures are present, for example at depths below 2.5 km in Figure 8c.

5.2 Variability

Results vary across the study area (Figure S1). Images are complex, with partially coherent energy distributed across a range of depths. Some cross-sections show increased reflectivity near 2.4 km, but this feature is not consistently resolved across all considered cross-sections. The irregular source-receiver geometry limits its resolution in certain areas and affects both the detectability and continuity of reflectors, including the one around 2.4 km. For example, in the westernmost S-N trending cross-section in Figure S1a), reflectors are imaged at approximately 2.2 km and 2.8 km depth at $\approx$ 400 m south and north of IDDP-1. However, these features are not resolved in between due to sparse ray coverage in that region.

6 Discussion

6.1 Importance of direct P-wave alignment and muting of phases

To ensure the accurate extraction of reflection travel-times, we aligned the direct P-wave arrivals in the seismic traces with theoretical traveltimes computed with a 1D velocity model (Section 4.2.2). This alignment corrects for static traveltime shifts caused by discrepancies between the 1D model and the true subsurface velocity structure, as well as by uncertainties in earthquake hypocenter locations. At Krafla, we found that the 1D model systematically overestimates seismic velocities (Figures 2 and 4a), resulting in predicted P-wave arrivals that are earlier than those observed. Thus, without the alignment, the migration algorithm would start extracting amplitudes prior to the direct P-wave arrival and misinterpret the direct P-wave as a shallow sub-

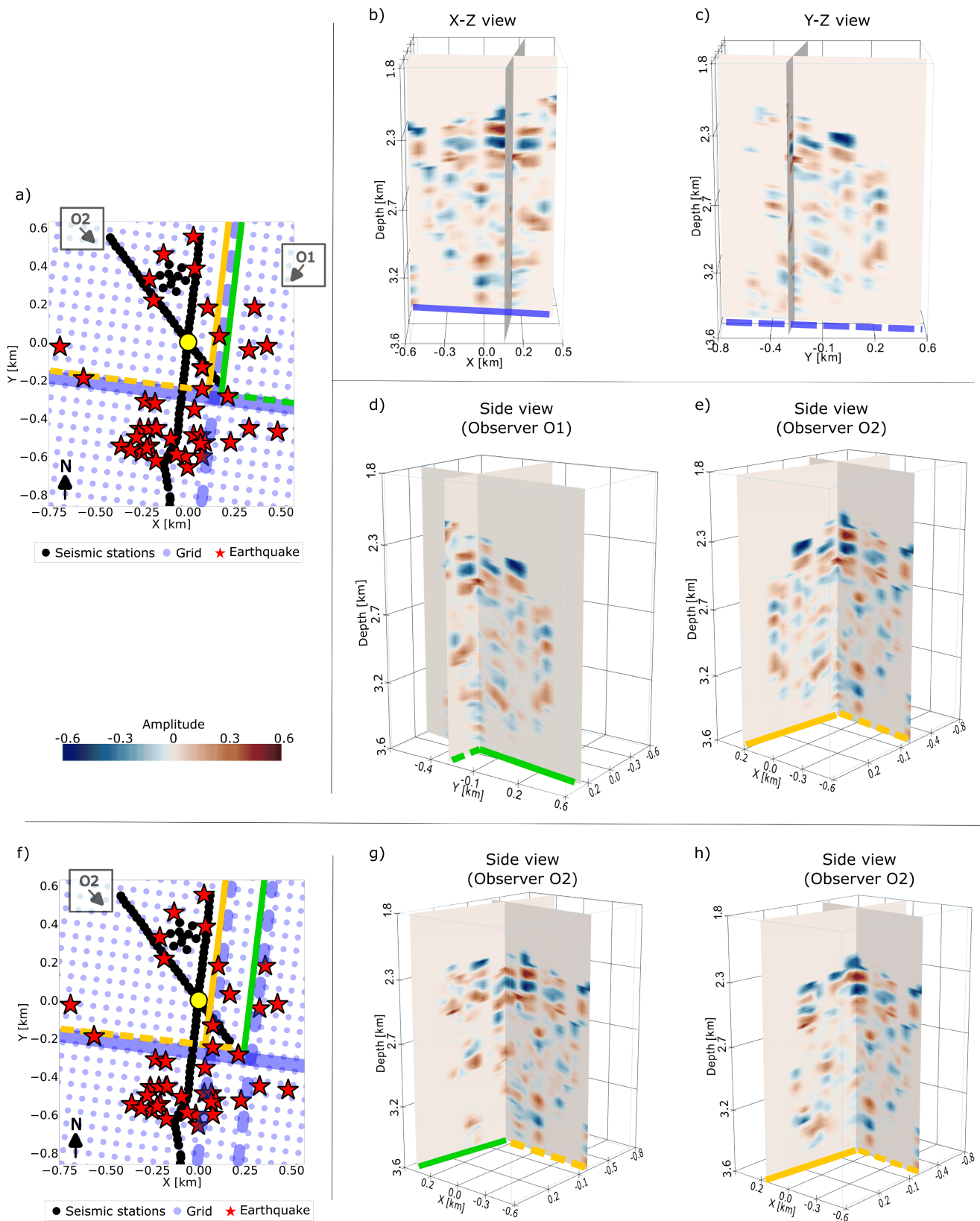


Figure 8 Results. a) Study setup and grid points. The locations of two intersecting cross-sections, one E-W trending (continuous purple line) and one S-N trending (dashed purple line) are indicated. Yellow and green colors illustrate the areas shown in the images in b-e). b) and c) Vertical ($X-Z$ and $Y-Z$) cross-sections, illustrating a reflector at 2.4 km depth. d) and e) Side views on the intersecting planes and the reflector at 2.4 km depth. The observer viewpoints for these panels are indicated on the map by arrows (O1, O2). f) Location of cross-sections shown in g) and h). The yellow dot in a) and f) marks the location of the IDDP-1 borehole.

surface reflection, creating an artifact in the migration images. In addition, we muted direct S-waves, which might seem counterintuitive given its lower (apparent) velocity relative to P–P reflections. In the common-image point (CIP) gathers, which combine traces from multiple events, amplitudes are stacked based on pre-computed reflection traveltimes (Figure 3d). Ideally, any phase with moveout differing from that of the P–P reflection should therefore interfere destructively and be suppressed. However, tests with synthetic data show that direct S-waves can produce spurious reflectors in the imaging results if not muted. This is demonstrated in Figure 9, which shows two selected cross-sections based on unmuted synthetic and real data. In the synthetics, direct S-waves appear as spurious, dipping reflectors at about 3.2 km depth. In real data, scattering and noise may suppress these arrivals, but artifacts remain visible. For example, in Figure 9b, the signature of the direct S-wave can be clearly seen, while the effect seems less pronounced in Figure 9d. These artifacts result from steep incidence angles and high apparent velocities caused by the limited array aperture and earthquakes occurring just underneath the stations. As a result, different phases have similar apparent velocities, preventing full destructive interference during stacking. For similar reasons, we muted the S-wave coda. With increasing lapse time, the wavefield becomes progressively complex due to overlapping arrivals from converted phases, multiples, and scattered waves. Because of similar moveouts of different phases, distinguishing true P–P reflections becomes increasingly unreliable at later times. For instance, the images based on synthetics in Figures 9a and c clearly show artifacts below 3.5 km depth due to P–S and S–S reflections. It remains unclear whether the reflectors observed around 4 km depth in Figure 9b and d represent actual subsurface features or spurious signals resulting from phase conversions or multiples.

Given these limitations, we restricted our analysis to the P-wave coda, targeting a limited depth range (from > 2.1 km to about 3 km). As noted by Quiros et al. (2017), rotating the data from the ZEN system to the LQT (P, SV, and SH) coordinate system may help to separate the P- and S-wave energy, potentially improving the geological interpretation. However, since most of our stations are single-component seismic nodes, this rotation is only feasible for the northernmost stations on our profiles (14 stations on L1 and 21 stations on L2, respectively), spatially constraining the area that could be imaged. Still, given the limitations encountered here, future work could include the implementation of this rotation for the available 3C-stations. Additionally, denoising techniques such as a resonance removal filter (e.g., Maass et al., 2025) could be applied to reduce site effects and recover useful information from the S-wave coda.

6.2 Possible factors causing image degradation

While the most prominent feature observed in this study is a coherent reflector at 2.4 km depth beneath the Krafla caldera, it is not consistently imaged across all

cross-sections (Figure S1). This inconsistency is partly due to limited ray coverage in some areas and an irregular source–receiver distribution, as previously discussed. Spatial illumination could be improved by using a more complete earthquake catalogue. Although we rely on manually reviewed event locations, ensuring a high level of reliability in phase picks, smaller events may not have been fully captured. In future work, advanced detection and location techniques—such as template matching (e.g., Gibbons and Ringdal, 2006; Li et al., 2020) and Deep Learning-based methods (e.g., Dokht et al., 2019)—could be used to recover additional events. This could fill data gaps and increase the signal-to-noise ratio during the stacking process, improving the detectability of weak or discontinuous reflectors. Beyond resolution limitations, scattered images which lack clear and laterally coherent reflectors may result from strong lateral geological heterogeneity, complex reflector topography, or sharp lateral variations in seismic impedance, leading to inconsistent reflection amplitudes. Such heterogeneities are expected at Krafla, as volcanic environments are known for their pronounced structural complexity (e.g., O'Brien and Bean, 2009; Paulatto et al., 2022).

Discontinuous or scattered reflectors may also arise from uncertainties in earthquake locations. In this study, horizontal as well as depth uncertainties vary between 200 m and 500 m. Mislocated events lead to the incorrect assignment of reflection points to grid cells and therefore degrade image quality. Discrepancies between the 1D velocity model used for traveltime calculations and the true velocity structure can introduce further errors. Although the alignment of the direct P-wave to theoretical traveltimes compensates for static shifts at each station, this correction applies only to the direct arrivals. Reflected P–P-waves follow different raypaths, and the divergence between direct and reflected paths increases with reflector depth (Figure 3). A more accurate approach would involve a 3D velocity model, for example constrained with seismic tomography, to better estimate both traveltimes and bounce points of reflections. However, limited ray coverage below 2.1 km around IDDP-1 restricts the ability of traveltime tomography to resolve deeper structures. That said, the main reflector identified at 2.4 km lies just beneath the earthquake cluster, so P–P reflections from these depths follow paths similar to those of the direct P-waves. Given the limited resolution of tomography and taking horizontal smoothing within the Fresnel zone into account, lateral velocity differences along these paths might not be resolvable. We therefore speculate that, at least for shallow reflectors, degradation of the image quality is more likely caused by earthquake mislocations than by inaccuracies in the velocity model; that is, by errors in the estimated hypocenter positions (both lateral and depth), which lead to mispositioned inferred reflection points and therefore a distortion or smearing of coherent structures in the final image.

Finally, incoherent scattering and site effects from within the alternating basaltic lava flows and hyaloclastites that make up the shallow subsurface at Krafla, may also degrade the image. Scattered waves as well as

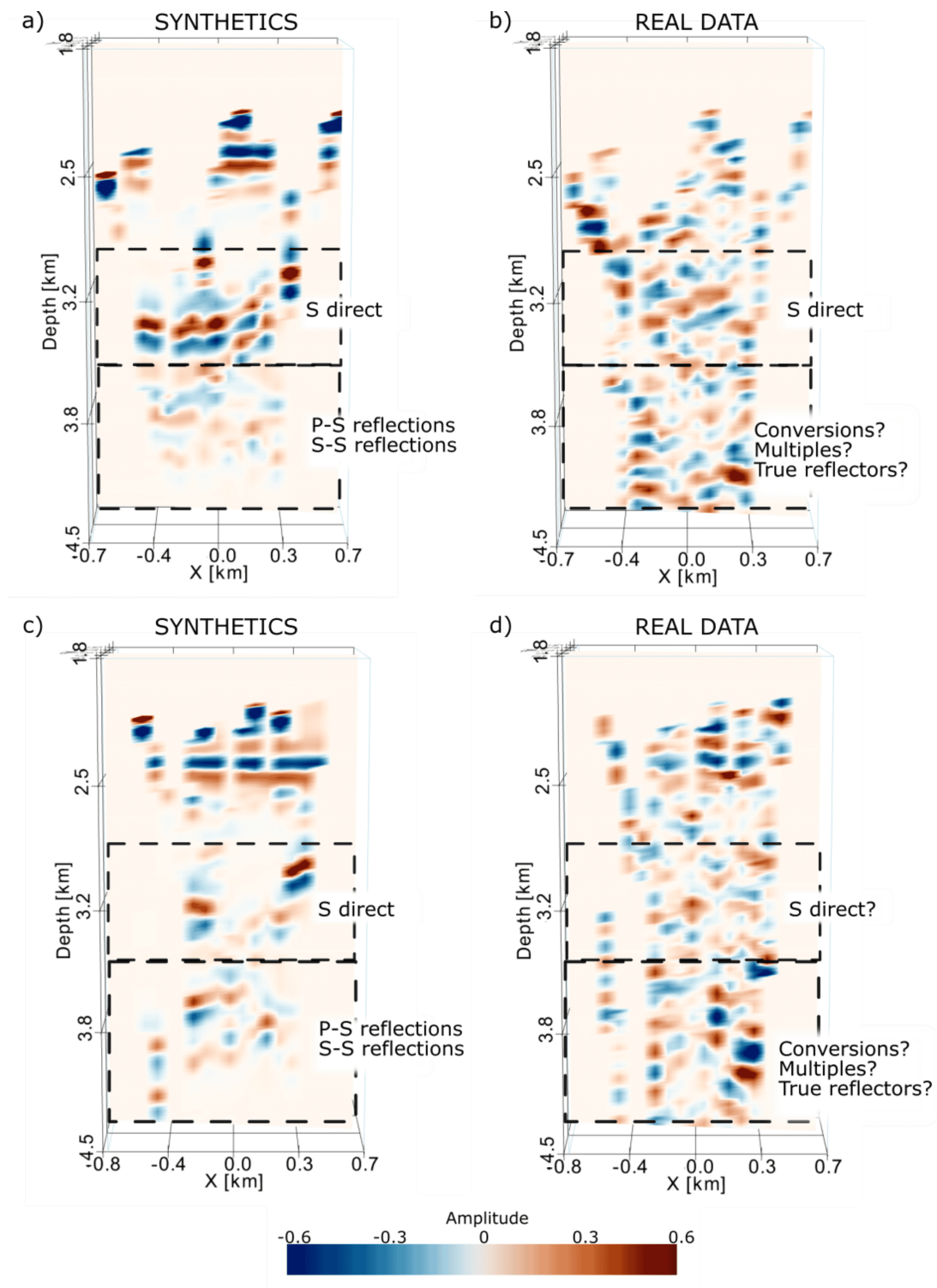


Figure 9 Two examples demonstrating imaging artifacts due to direct S-waves and conversions within the S-wave codas. a) and b) Images based on synthetic and real data for the same cross-sections as in the central panel in Figure S1b. In the synthetics, the direct S-waves create a dipping spurious reflector at around 3.2 km depth. Conversions further introduce artifacts at greater depths. In real data, contamination due to the direct S-wave can be clearly seen, and reflectors appear at larger depths, possibly due to conversions within the S-wave coda. c) and d) Same as in a) and b) but for the cross-section from Figure 8b.

near-surface reverberations may interfere with reflections from depth, reducing their coherency and modulating the amplitude of P-P reflections. NNE-SSW trending faults that traverse the Krafla area (Section 1.1)

further distort seismic wavefields. As with the S-wave coda discussed in the previous section, the P-wave coda may also contain artifacts resulting from conversions (e.g., S-P) and site effects. As a result, we interpret

our images with caution, placing greater emphasis on the shallow subsurface where contamination from converted or scattered phases is expected to be less significant and the imaging results are more reliable.

6.3 Using envelopes instead of raw amplitudes

In this study, we stacked raw amplitudes, applying polarity corrections where necessary based on visual inspection within earthquake clusters to account for varying source mechanisms. A simpler alternative would be to use only positive amplitudes—such as envelopes, absolute values, or squared amplitudes—thereby avoiding the need for polarity correction. Envelope stacking also smooths the waveforms, effectively acting as a low-pass filter. This reduces the sensitivity of the migration method to the velocity model. However, the disadvantage of using this approach is that no destructive interference is achieved. As demonstrated in Figure S2, applying the migration approach to envelopes of seismic traces yields poor imaging results and fails to highlight reflectors at Krafla. This is likely due to strong scattering and complex coda arrivals that are not suppressed anymore through destructive interference during the stacking. We conclude that, while envelope-based approaches eliminate the need to correct for polarities and reduce the sensitivity of the method to the velocity model, they may not be optimal in highly heterogeneous volcanic environments like Krafla, where noise cancellation is deemed important to reliably map out reflectivity.

6.4 Geological interpretation

Our results reveal a prominent reflector at approximately 2.4 km depth near the IDDP-1 borehole. A similar feature was identified by Kim et al. (2020), who applied the RVSP technique to local earthquakes recorded during a 2014 field campaign. They imaged a predominantly flat reflector at the same depth both southwest and southeast of IDDP-1, although resolution was limited toward the north due to the uneven distribution of earthquakes. Kim et al. (2020) interpreted this reflector as the top of the magma pocket intersected at the IDDP-1 borehole at a depth of 2.1 km. The 300 m depth discrepancy between the reflector and the actual melt encounter was attributed to uncertainties in earthquake locations and the velocity model. For example, given the event depth uncertainties of 200 m to 500 m in our dataset, a systematic upward shift of 200 m in earthquake locations would reposition our imaged reflector to ≈ 2.2 km, which is very close to the magma pocket. Similarly, overestimated velocities in the employed 1D velocity model would bias reflector depths downward, as also demonstrated by Kim et al. (2020).

The close agreement in geometry, depth, and location suggests that our study and that of Kim et al. (2020) image the same subsurface structure. If this reflector indeed marks the top of the magma reservoir, it would indicate a predominantly flat melt body detectable in all directions relative to the IDDP-1, potentially representing a sill. In addition to the findings of Kim et al.

(2020), this interpretation is further supported by Reiser et al. (2020), who resolved a reflector at ≈ 2.1 km depth beneath borehole K-26 (approximately 800 m south of IDDP-1) using active-source data, and by Maass et al. (2025), who identified possible multiple reflections from the magma pocket both up to 170 m south and 500 m north of the borehole. Seismicity at Krafla dips south-eastward from ≈ 2.1 km depth at IDDP-1 to approximately 2.6 km and occurs along the upper boundary of a low resistivity and high v_P/v_S zone, which has been interpreted as a potential melt accumulation zone (Lee et al., 2019; Glück et al., 2025). In this context, the reflector could delineate the top of a broader body containing (partial) melt beneath the seismogenic zone (Figure 1). The idea of the magma pocket residing immediately beneath the base of local seismicity has also been proposed in earlier work (e.g., Schuler et al., 2016). Interestingly, magma was also encountered through drilling in borehole KJ-39, 2.5 km south-east of IDDP-1, at 2.6 km depth, corresponding to the base of local seismicity in this area.

An alternative explanation, however, is that the reflector at 2.4 km depth marks a local brittle-ductile transition (BDT) within the Krafla geothermal system without the involvement of magmatic fluids, or—taking uncertainties of earthquake locations into account—simply another structure within the Krafla geothermal system (e.g., a deeper magmatic sill). To further constrain the characteristics of the reflector, polarity analysis could be of value. A change in polarity would be expected if the interface corresponds to a magma–rock boundary. As part of the data pre-processing, corrections were applied to ensure consistent polarities of direct P-waves, and to verify that the polarity of reflected phases were consistent across multiple events (Section 4.2.2). However, a qualitative interpretation of reflected phase polarities also relies on knowledge of the earthquake source mechanisms, which remain unaccounted for in this study. Furthermore, applying our migration workflow to the dense earthquake cluster in the southeastern part of our study area (Figure 1b) could help refine the geometry of the reflector and provide a more robust basis for the geological interpretation. We also note that the apparent lateral continuity of the reflector at 2.4 km depth may be exaggerated. Due to the finite frequency content of the seismic waves, a network comprising small melt lenses could appear as a continuous feature within the lateral resolution limit defined by the Fresnel zone.

In addition to the reflector imaged at 2.4 km depth, results show a complex network of subsurface features, including dipping and only partially coherent reflectors (Figure S1). This complexity may reflect a high degree of structural and geological heterogeneity beneath Krafla volcano. Alternatively, some of the observed features may be imaging artifacts resulting from methodological limitations, as discussed in detail in Section 6.2.

7 Conclusions

We proposed a workflow for earthquake migration seismic imaging in complex geological settings. Applied

to P–P reflections from local earthquakes at Krafla volcano, NE Iceland, it yields new images of the subsurface structure surrounding the IDDP-1 borehole. Synthetic data generated using SPECFEM2D served as a proof of concept and played a critical role in benchmarking our processing workflow. These simulations also helped identify key challenges associated with passive-source reflection imaging at Krafla, primarily linked to the sometimes unfavorable earthquake-station geometry. The main workflow-related insights and limitations encountered in this study, along with implications for future work, are summarized below:

- Alignment of the direct P-waves with the velocity model is essential to avoid migration artifacts caused by the direct P-wave arrivals.
- Direct S-waves, although slower than P–P reflections, do not fully cancel during stacking due to the earthquake–station geometry. If not muted, they can produce spurious reflectors. Similarly, the use of the S-wave coda increases the risk of interference from scattered and converted phases, hindering the reliable identification of subsurface reflectors.
- Phase discrimination based on apparent velocity could be improved using a more favorable array geometry with a larger aperture, as also noted by Quiros et al. (2017). We propose that synthetic resolution tests could be used in future studies to design optimal station configurations that minimize phase interference and distinguish true subsurface structures from migration artifacts, improving geological interpretation.
- Further limitations arise from the geological heterogeneity of the medium, inducing complex seismic wavefields and scattered waves that interfere with reflections, modulating their amplitude and phase. As the migration technique presented here requires accurate traveltimes picking, this contributes to the degradation of migrated images.
- Envelopes of seismic traces are useful waveform attributes as they reduce the sensitivity of the workflow to the velocity model and phase polarities. However, at Krafla, envelope-based stacking produced poor imaging results, likely due to scattering and wavefield complexity. This approach may therefore be more effective in less heterogeneous environments.

Our analysis reveals a prominent reflector at approximately 2.4 km depth near the IDDP-1 borehole, where magma was unexpectedly encountered during drilling. This feature was detected over an area spanning at least 650 m × 300 m. Considering the uncertainty in earthquake locations, it could represent the upper boundary of the IDDP-1 magma pocket. In that case, the magma pocket would represent a broader structure just below the main seismicity cluster at IDDP-1. This interpretation is supported by previously published studies and by evidence from a second borehole at Krafla,

where magma was encountered just beneath the base of seismicity at 2.6 km depth. However, alternative interpretations cannot be excluded; the reflector may also correspond to a local brittle–ductile transition zone or another structural feature within the volcanic system. Extending the imaging workflow to a larger area will help to better constrain its geometry and refine the geological interpretation. Moreover, the high station density of the seismic network enables the application of interferometric techniques to microearthquake data, in which seismic wavefields are redatumed from surface receivers to earthquake hypocenters using cross-correlation approaches (e.g., Curtis et al., 2009; Chamarczuk et al., 2022). Such redatuming can reduce the effects of complex near-surface structure and thus represents another promising direction for future work. Overall, this study demonstrates that seismic migration - which is more commonly used in active-source studies - of high frequency (5 - 25 Hz) data from local earthquakes can reveal valuable fine-scale structural information in geologically complex and strongly scattering environments.

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Data and code availability

This research has made extensive use of ObsPy (Beyreuther et al., 2010) and the PyVista Python package (Sullivan et al., 2023). Raytracing was performed with the open-source MATLAB toolbox CREWES (Margrave, 2000). The Python-based seismic migration code developed for this study is freely available on Git (https://git.dias.ie/maass/earthquake_migration_imaging). The repository includes three main subfunctions—binning, data processing, and data visualization—along with the seismic data from the 48 earthquakes analyzed in this work. These data are provided at a sampling rate of 200 Hz and are accompanied by complete metadata for the seismic stations and earthquake events. The provided earthquake catalogue is based on two independent catalogues published in Guðnason et al. (2023) and Glück et al. (2025).

Competing interests

The authors declare no competing interests.

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