S-wave arrivals for which the associated P-wave is also detected, but the S-wave was not automatically removed during the 3 s de-clustering process. Manual inspection of signals also reveals missed events, particularly during the most active periods (i.e., after 8 August, Fig. S6).

4 Earthquake location and local magnitude estimates

We initially locate and calculate local magnitudes for high-SNR detected local earthquakes by combining DAS and seismic station data. We calculate SNR on offshore cable channels (4000–6800) considering a noise window from 1–5 s before the detection time and a signal window from 0.5–4.5 s after the detection time (Fig. 5). We compute the average Root-Mean-Square (RMS) values for noise and signal on traces band-pass filtered between 5–20 Hz, using a trimmed mean that removes the highest and lowest 10 % of values (trim fraction 0.1).

We then calculate SNR as follows:

$$\mathrm{SNR} = 20\log_{10}(\mathrm{RMS}_{\mathrm{signal}}/\mathrm{RMS}_{\mathrm{noise}}). \quad (1)$$

We consider local earthquakes (following the criteria outlined above) with a SNR $>$ 12 dB for hypocentral location and magnitude estimation, resulting in a subset of 456 events out of the 5734 detected local earthquakes. We manually include a M_L 3.4 earthquake on 8 August 2024 at 19:18 (event id: kef2196) despite its low SNR of 5.3 dB, as it represents the largest earthquake recorded northwest of Kefalonia during our study period. The low SNR results from the origin time being $\sim$ 12 s after a nearby M_L 3.0 (event id: kef2195) whose coda wave contaminates the window containing the primary phase arrival of the M_L 3.4 earthquake. Thus, our initial dataset consists of 457 earthquakes for location. The restrictive SNR threshold reflects the goal of building a high-quality reference catalog of locations and magnitudes. We perform automatic P- and S-wave arrival picking on DAS data using PhaseNet-DAS (Zhu et al., 2023) for each of the 457 earthquakes. We define 20 s windows for PhaseNet-DAS that start 6 s before the detection time and end 14 s after it (same as in Fig. 5). We follow the preprocessing used in the PhaseNet-DAS implementation (https://ai4eps.github.io/EQNet/phasenet_das/, last access: 26 May 2026), including conversion of strain to strain rate via temporal differentiation. We apply an additional band-pass filter between 5–40 Hz as part of our preprocessing. We manually inspect all picking results and observe good quality P- and S-phase picks (Fig. S7).

To reduce the number of DAS channels for earthquake location and achieve a number comparable to the seismic stations, we subdivide the cable into 12 segments (Fig. 1c) using a custom geometric clustering algorithm. The algorithm uses Density-Based Spatial Clustering of Applications with Noise (DBSCAN, Ester et al., 1996) to group channels exclusively by their spatial coordinates. It automatically instantiates a new segment boundary whenever the azimuthal angle variation exceeds a specified threshold. We allow a maximum azimuthal variation of 60°, applied an azimuth smoothing window of 50 channels, and enforce a minimum constraint of 100 channels per segment. We assign median P- and S-wave arrival times that we estimate in a prior step with PhaseNet-DAS applied to the entire cable to the median channel for each of the 12 segments. Finally, we download waveforms from the seven stations closest to the cable (Figs. 1, S3) and manually pick P- and S-wave arrival for each of the 297 earthquakes in Snuffler (Heimann et al., 2017) when data quality permits.

4.1 Earthquake location

We consider events with a P- or an S-wave arrival identified by Phasenet-DAS on at least 10 of the 12 cable segments and with a clear P- or S-wave arrival at a minimum of three seismic stations for earthquake location. The minimum required number of segments and stations assures adequate azimuthal coverage for the events in the cluster offshore northwest of Kefalonia. We compute earthquake locations with NLLoc (Lomax et al., 2000) using arrival times from both DAS and seismic station data and a local 1D velocity model (Haslinger et al., 1999). We revised the original velocity model by removing the second layer at 0.5 km depth (Vp = 5.47 km s^{-1}), maintaining lower velocities from 0 to 2 km depth (Table S1) to reflect the presence of unconsolidated sediments beneath the offshore cable segment. The removal of the second layer from the original velocity model reduces the number of very shallow earthquakes (0–1 km) and improves depth resolution slightly. We calculate static phase arrival time residuals at the seismic stations and DAS channels used for location and include them in the final NLLoc run. Incorporating these residuals partially accounts for velocity model heterogeneities and mitigates the limitations of a 1D model. We note, however, that static station corrections are influenced by the spatial distribution of seismicity and are most appropriate for the region with the highest event density for this data set. We successfully locate 356 of the 457 local earthquakes (Fig. 6). The remaining events were either recorded at only two seismic stations or lacked a median phase arrival time on at least 10 cable segments, which inhibited robust hypocentral solutions. We obtain average semi-major and semi-minor horizontal ellipse error axis values of 1.5 ± 0.5 and 0.8 ± 0.4 km, respectively, and vertical errors of 1.9 ± 0.6 km (Fig. S8).

4.2 Earthquake magnitude estimate

We estimate M_L using seismic station data following the empirical relation of Hutton and Boore (1987) because the relation is also employed by NOA for M_L calculations (Melis and Konstantinou, 2006). We remove instrument response and convert waveforms to Wood-Anderson equivalent ampli-