

# Investigating the subsurface in a shallow water environment using array and single-station ambient vibration techniques

A. Shynkarenko<sup>1</sup>,<sup>1</sup> A. M. Lontsi,<sup>1</sup> K. Kremer,<sup>1,2</sup> P. Bergamo,<sup>1</sup> M. Hobiger<sup>1</sup>,<sup>1</sup> M. Hallo<sup>1</sup> and D. Fäh<sup>1</sup>

<sup>1</sup>Swiss Seismological Service (SED), ETH Zürich, Sonneggstrasse 5, 8092 Zürich, Switzerland. E-mail: [a.shynkarenko@sed.ethz.ch](mailto:a.shynkarenko@sed.ethz.ch)

<sup>2</sup>Geological Institute and Oeschger Centre for Climate Change Research, University of Bern, Baltzerstrasse 1+3, 3012 Bern, Switzerland

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## SUMMARY

Single-station and array ambient vibration techniques are widely used in onshore environments, in particular to retrieve the subsurface structure and shear-wave velocity profiles. We apply these techniques offshore in Lake Lucerne (Switzerland) using single-station and array Ocean Bottom Seismometer (OBS) data. This lake has experienced tsunamigenic subaquatic slope failures in the past and still has sediment-charged slopes that might fail in the presence of a seismic or aseismic trigger. The application of traditional onshore methods offshore brings additional challenges related to the processing of recorded data. To overcome these challenges, we perform multibeam bathymetry surveys to precisely locate the OBS on the lake floor and airgun shootings to determine the orientation of the horizontal components of the seismometer and to correct the time drift of the recorder. Then we obtain surface-wave phase velocity dispersion curves of Scholte and Love waves, and Scholte wave ellipticity curves at six subaquatic slopes. After the estimation of the dispersion curves, we deal with their modal identification using mode attribution analysis. The shear-wave velocity and thickness of the sedimentary layers at the investigated slopes are inferred using a transdimensional Bayesian inversion algorithm. The resolved velocity profiles show very low shear-wave velocities in shallow lake sediments and allow us to improve the understanding of the local stratigraphy. This research contributes to the assessment of stability and tsunamigenic potential of subaquatic slopes in Lake Lucerne.

**Key words:** Tsunamis; Earthquake hazards; Seismic noise; Site effects; Surface waves and free oscillations.

## 1 INTRODUCTION

Ambient vibration techniques are non-invasive passive seismic methods based on the analysis of natural and anthropogenic seismic noise. They allow measuring the fundamental frequency of resonance, estimation of the dispersion and ellipticity curves of surface waves, and extraction of the shear-wave (*S*-wave) velocity profiles by inverting the measured curves (e.g. Tokimatsu 1995; Scherbaum *et al.* 2003; Michel *et al.* 2014; Foti *et al.* 2018; Hollender *et al.* 2018), that is especially important to understand the structure and seismic response of potentially unstable slopes (Couboulex *et al.* 2020). Unstable slopes underwater can fail and, if the volume of failed sediment is large enough, generate tsunamis. This can also happen in lakes and can represent an important natural hazard for densely populated shores. The failure of subaquatic slopes is primarily triggered by seismic impact or by excess pore water pressure generation (Schnellmann *et al.* 2006; Stegmann *et al.*

2007; Strasser *et al.* 2011; Strupler *et al.* 2018; Sammartini *et al.* 2019).

To assess the failure potential of different subaquatic slopes and to estimate the sediment volume susceptible to failure, we need to resolve the subsurface structure of the slope, its seismic response, and sediment properties; in particular, we need to infer the *S*-wave velocity depth profile at the target locations. The preference for the offshore site characterization is usually given to active seismic experiments rather than passive ones (Klein *et al.* 2005; Kugler *et al.* 2007; Exley *et al.* 2010; Socco *et al.* 2011; Vanneste *et al.* 2011; Long *et al.* 2020), although ambient vibration methods are more environment-friendly (Frisk 2003) and economical than most active seismic data acquisition techniques. We transfer the state-of-the-art experience of onshore site characterization using ambient vibration techniques (Michel *et al.* 2014; Fäh *et al.* 2016b; Hobiger *et al.* 2016, 2021) to the shallow water environment in a lake to assess the subsurface structure and seismic response of the sediments,

and to contribute to the experience of offshore passive seismic experiments.

We select Lake Lucerne in central Switzerland as a test site for our experiments, as this lake has experienced historical tsunamis in the past, for example in 1601 AD, triggered by the  $M_w$  5.9 Unterwalden earthquake (Siegenthaler *et al.* 1987; Schnellmann *et al.* 2002; Schwarz-Zanetti *et al.* 2003) and in 1687 AD, triggered by the spontaneous collapse of the Muota river delta (Hilbe & Anselmetti 2014a, b). Nowadays, the repetition of such an event would pose a significant danger for the densely populated lake shores and the shore infrastructure.

We carry out site characterization at several subaquatic slopes using ambient vibration data recorded by nine Ocean Bottom Seismometers (OBS). Waves propagating at the fluid–solid interface in the offshore environment with similar properties to Rayleigh waves, which propagate at the air–solid interface onshore, are called Scholte waves and can be analysed with the techniques usually applied for Rayleigh waves, although the forward modelling algorithm has to be adjusted to take into account the water layer (Muyzert 2007; Maraschini 2008; Nguyen *et al.* 2009; Foti *et al.* 2018; Tomar *et al.* 2018). In comparison with onshore measurements, several new challenges, such as distorted geometry of the array, unknown seismometer location and orientation, or clock drift of the recorder, arise in the offshore environment and require additional fieldwork (e.g. high-resolution multibeam bathymetry mapping of the site, active seismic experiments) and data pre-processing steps (Anderson *et al.* 1987; Duennebiele *et al.* 1987; Jurkevics 1988; Oshida *et al.* 2008; Ao *et al.* 2010; Hannemann *et al.* 2013; Gouédard *et al.* 2014; Doran & Laske 2017; Hable *et al.* 2018; Le *et al.* 2018). For the analysis of the collected data, we apply different single-station and array techniques. Single-station methods utilize 3-component recordings of one station and allow determining the fundamental frequency of the site from the spectral ratio of horizontal-to-vertical components (H/V ratio; Nakamura 1989), the Scholte wave ellipticity using RayDec (Hobiger *et al.* 2009) and the polarization parameters of the soil particle motion (Vidale 1986; Burjáněk *et al.* 2010, 2012). However, single-station techniques cannot provide the  $S$ -wave velocity structure of the subsurface without additional constraints to the subsurface model due to the non-uniqueness of this inversion problem (e.g. Hobiger *et al.* 2013). Improper constraints on the subsurface model might also bias the outcome of the analysis. At the same time, array techniques utilize the data of all array sensors together and allow the extraction of surface-wave phase velocity dispersion curves and azimuthal information for the recorded surface waves using different approaches: Three-Component High-Resolution Frequency-Wavenumber technique (3C-HRFK; Capon 1969; Poggi & Fäh 2010; Wathelet *et al.* 2018), Modified Spatial Autocorrelation (MSPAC; Aki 1957; Bettig *et al.* 2001) and Wave-field Decomposition (WaveDec; Marano *et al.* 2012). The acquired dispersion and ellipticity curves of surface waves carry information about mechanical properties of the subsurface. Hence, we can jointly invert these data for the  $S$ -wave velocity profile (Scherbaum *et al.* 2003; Wathelet *et al.* 2004; Arai & Tokimatsu 2005; Wathelet 2008; Poggi *et al.* 2012; Fazlavi & Haghshenas 2015; Ullah & Prado 2016; Gouveia *et al.* 2018). In this work, we perform a joint inversion using a newly developed algorithm for the multizonal transdimensional Bayesian inversion (Hallo *et al.* 2021).

As a final result, we get important information about the internal slope structure and sediment properties which allow us to better understand the stratigraphy and variability of the sedimentary cover for the investigated sites. Quantitative data regarding the fundamental frequency of resonance and  $S$ -wave velocity profile can be later

used for the assessment of stability and non-linear behavior of the potentially unstable areas.

## 2 REGIONAL SETTING, SITE SELECTION AND ARRAY CONFIGURATION

### 2.1 Regional setting

Lake Lucerne is a fjord-type perialpine lake in central Switzerland. It has a total area of 114 km<sup>2</sup>, maximum depth of 214 m and its surface is at a mean level of 433.6 m a.s.l. Lake Lucerne consists of seven basins filled largely by lacustrine and glaciolacustrine sediments with a maximum thickness of up to 200 m. All the basins are carved in the bedrock consisting of Swiss Molasse and Helvetic nappes. Some of the basins are separated by moraine ridges (Finckh *et al.* 1984; Hilbe *et al.* 2011; Strupler 2012). The shores of the lake are densely populated and bear a well-developed infrastructure.

Although large-scale earthquakes ( $M_w > 6$ ) are quite rare in Switzerland, they can cause far-reaching and very costly damage and lead to numerous fatalities (Fäh *et al.* 2016a; Wiemer *et al.* 2016). In Lake Lucerne, a tsunami has occurred in 1601 AD after the  $M_w$  5.9 Unterwalden earthquake (Schwarz-Zanetti *et al.* 2003). This earthquake caused numerous sublacustrine and subaerial mass movements that generated a tsunami with up to 4 m wave height followed by a seiche that lasted for several days (Siegenthaler *et al.* 1987; Schnellmann *et al.* 2006; Hilbe & Anselmetti 2014a).

### 2.2 Site selection and available data sets

We carried out site characterization measurements at six selected locations: Chindli (CIA), Ennetbürgen (ENA and ENB), Muota (MUA), Nase (NAA) and Weggis (WED; Fig. 1a, Table 1). These sites represent sublacustrine slopes of different geomorphological settings (delta, hemipelagic lateral slopes and moraine ridges; Hilbe & Anselmetti 2014a). Several previously acquired data sets were used to define the location of the arrays at the selected sites:

- (i) A high-resolution bathymetry map (Fig. 1a; Hilbe & Anselmetti 2014a) was used to define the lake floor gradient and the sites large enough to accommodate the designed array size.
- (ii) Reflection seismic profiles (Fig. 1a) collected with a single-channel 3.5 kHz pinger source, ‘centipede-sparker’ with 150–1500 Hz central frequency and 1 in<sup>3</sup> airgun with 120–1600 Hz central frequency (Schnellmann *et al.* 2005; Hilbe *et al.* 2011; Strupler 2012) were used to estimate the thickness of shallow (*ca.* first 2–10 m) slope sediments (Fig. 1b).
- (iii) The mass movement history of the lake (Schnellmann *et al.* 2002; Strasser *et al.* 2011; Hilbe & Anselmetti 2014a, b) was analysed to select potentially unstable slopes.
- (iv) Coring and Cone Penetration Testing were performed at target areas to check that the lake sediments can bear the OBS weight (Strasser *et al.* 2007; Shynkarenko *et al.* 2018).

Based on this information and on-site gained experience, we selected the most suitable locations for our investigations as the sites that have not failed recently, have low-to-moderate slope gradient (usually between 0° and 15°), relatively homogeneous sediment cover and enough space for array deployments, taking into account possible instrument drifts (from our experience up to 30 m) while descending in the water column in a free-fall mode.

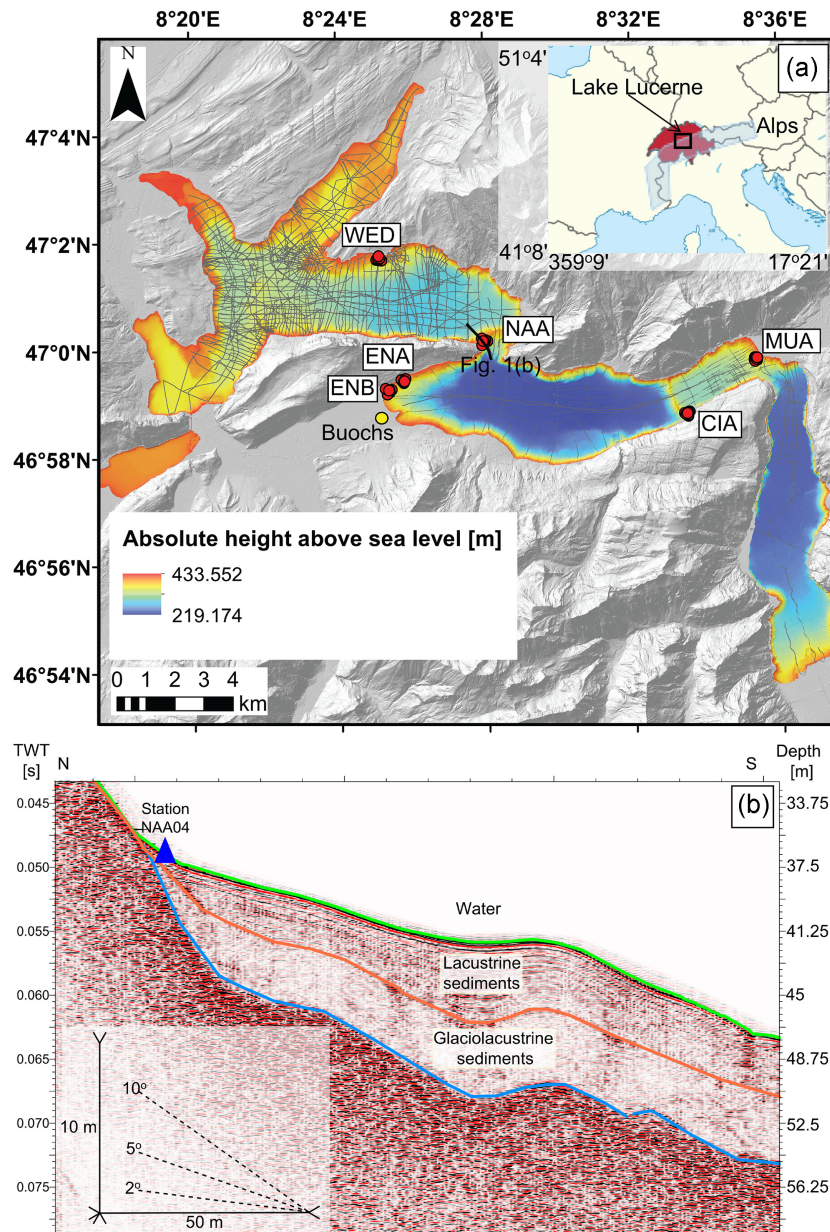


**Figure 1.** (a) Bathymetric map of Lake Lucerne (Source: Federal Office of Topography; Hilbe & Anselmetti 2014b) showing the location of the six study sites (CIA, ENA, ENB, MUA, NAA and WED) and the position of the OBS stations (red circles). Grey lines represent existing reflection seismic profiles, the black line refers to the profile in (b). The coordinates are given in WGS84 coordinates; (b) Example of a reflection seismic profile close to array NAA in depth and two-way traveltime (TWT): interpretation of the data acquired by Schnellmann *et al.* (2002).

**Table 1.** Information about analysed arrays.

| Array and Site name | Geomorphological type of the slope | Geometry and # of the stations | Duration of the measurement [days] | Water depth [m] | Slope angle below the stations [°] | Min and Max interstation distance [m] |
|---------------------|------------------------------------|--------------------------------|------------------------------------|-----------------|------------------------------------|---------------------------------------|
| CIA Chindli         | Moraine ridge                      | Spiral, 9 stations             | 3                                  | 47.3–61.1       | 7.0–26.5                           | 11.3–177.6                            |
| ENA Ennetbürgen     | Lateral hemipelagic                | Spiral, 9 stations             | 2                                  | 30.9–41.4       | 2.0–10.8                           | 14.9–152.5                            |
| ENB Ennetbürgen     | Deltaic                            | Spiral, 9 stations             | 53                                 | 19.2–49.0       | 4.5–11                             | 20.6–218.0                            |
| MUA Muota           | Deltaic                            | Circular, 8 stations           | 5                                  | 38.9–61.6       | 8.5–10.8                           | 32.6–127.9                            |
| NAA Nase            | Moraine ridge                      | Spiral, 9 stations             | 1                                  | 36.3–66.4       | 2.0–14.8                           | 28.6–232.2                            |
| WED Weggis          | Lateral hemipelagic                | Spiral, 9 stations             | 2                                  | 22.5–47.9       | 4.3–9.8                            | 16.0–192.1                            |

### 2.3 Array configuration

After the site selection, the array geometry was planned depending on the expected subsurface structure, target array response and the number of available instruments (Maranò *et al.* 2014; Kennett *et al.* 2015; Foti *et al.* 2018). In total, we were operating nine OBSs: eight with LOBSTER frame and one with NAMMU frame. Except for the frame, all other components of the OBS are the same: Trillium Compact 120s broad-band seismometer, 32-bit data logger '6D6', hydrophone, titanium pressure tube with a battery pack and electronics, and titanium releaser KUMQuat.

At each site, an array measurement using eight or nine OBS stations (which can be also analysed as eight or nine independent single stations) with an acquisition time between 1 day and 2 months was performed. The difference in the arrays' acquisition times is due to the limited duration and number of field campaigns.

For the 9-station arrays, we used a spiral-arm geometry (three circles with three stations in each; the radius of the following circle is 2.2–2.5 times larger than the previous one; the stations of the following circle are rotated by 25–35° with respect to the previous one). In one case (array MUA), we operated only eight OBSs and used a circular geometry (one central station and seven stations on the circle; Table 1). Due to the previously mentioned drift of the OBS in the water column while descending to the lake floor, the final geometries of the arrays are distorted from the planned ones (Fig. 2 and Section 2 of the Electronic Supplement). The layouts of the arrays and the lake floor morphology below them are shown in Fig. 2.

## 3 METHODS AND DATA PROCESSING

For data processing, we apply different single-station and array ambient vibration techniques. Due to the specifics of the experiment, we also developed data pre-processing tools required for the array data analysis and carried out a mode attribution analysis before the inversion. For the inversion, we take advantage of a recently developed tool for transdimensional inversion (Hallo *et al.* 2021). In the following, we provide the details of the techniques applied.

### 3.1 Single-station methods

Single-station methods focus on the ground-motion polarization of ambient vibrations at a site. The acquired spectral properties allow us to compare different sites in terms of spatial variability in resonance frequency and sediment thickness. The following single-station techniques are applied to our data:

(i) *H/V spectral ratio* (e.g. Nakamura 1989; Fäh *et al.* 2001, 2009; Poggi *et al.* 2012)—the horizontal-to-vertical spectral ratio allows the identification of the fundamental frequency of resonance of the site and the presence of strong *S*-wave velocity contrasts in the soil profile. The *H/V* curves were estimated using a continuous wavelet transform to produce an accurate time-frequency representation of the noise signal. All three components of particle motion are treated separately, horizontal components are combined through the vector summation. To increase the robustness of the estimate, we divided the recording into 60 s time windows and averaged the resulting *H/V* curves over 6-hour-long time series.

(ii) *RayDec* (Hobiger *et al.* 2009)—this technique allows us to obtain the Scholte wave ellipticity curve using the random decrement technique (Asmussen 1997): for every positive zero-crossing on the vertical component of a filtered signal, the signal of all three

components is stacked, shifting the horizontal components by a quarter of the period corresponding to the current frequency. The two horizontal components are projected in the direction maximizing the correlation with the vertical component. All stacked signals on the vertical and horizontal components are summed and the Rayleigh/Scholte wave ellipticity is calculated as the *H/V* ratio of the resulting signals. The 2-hour signal was cut into 10 windows, which were analysed separately; the resulting curves were averaged.

(iii) *The polarization analysis* (Burjáněk *et al.* 2010, 2012)—this time-frequency analysis determines the polarization ellipsoid of the particle motion: the ellipticity as described in Burjáněk *et al.* (2010) is defined as a parameter ranging from 0 (linear particle motion) to 1 (circular particle motion), the dip is the inclination of the horizontal polarization ellipse, and the strike indicates the horizontal direction of the major axis of this ellipse. In this way, 2-D resonances can be identified. The sensor orientation has to be corrected before the polarization analysis (see Section 3.2.1). Two-hour-long recordings were used for the polarization analysis.

Properties of the particle motion as acquired by the single-station techniques (*H/V* spectral ratio or Rayleigh wave ellipticity curve) can be used for inversion on the *S*-wave velocity profile below the station (Fäh *et al.* 2003; Hobiger *et al.* 2013; Berbellini *et al.* 2019). Nevertheless, such an inversion requires additional constraints as it is an inherently non-unique inverse problem.

### 3.2 Array methods

#### 3.2.1 Array data pre-processing

Array methods provide the possibility to determine the propagation parameters of seismic waves. During the array data processing, we can extract the Love and the Scholte wave phase velocity dispersion curves, but due to the specifics of the offshore experiment, additional pre-processing steps are required before the proper array analysis. Passive seismic arrays require (i) accurate instrument location, (ii) correct time stamps of the data and (iii) knowledge of the seismometer orientation. However, when applying passive array techniques in a lake environment using OBS, the following issues arise: (i) final position and orientation of the horizontal components of the seismometer are unknown as the instruments are deployed in a free-fall mode and can drift within the water column; (ii) time drift of the OBS internal clock due to the missing connection to the GPS underwater. To resolve these issues, we applied the following techniques to all stations:

(i) To localize the OBS on the lake floor, a multibeam bathymetry mapping survey was performed at the array site using a Kongsberg EM2040 multibeam echo sounder in a 1° by 1° beam-width configuration with 300 kHz operating frequency. The calculated bathymetry raster has a lateral and vertical resolution of 0.5 and 0.1 m, respectively (Hilbe *et al.* 2011; Hilbe & Anselmetti 2014a), and provides detailed morphological information, allowing the identification of objects such as OBS on the ground. In addition to the depth information, the amplitude of the backscatter intensity was also measured, which is controlled by the signal attenuation in the water column, the local relief of the lake floor and the contrast of acoustic impedance between water and sediment/objects on the lake floor (Beyer *et al.* 2007), simplifying the identification of the OBS. Moreover, during the deployment and recovery of each station, we measured its approximate position with the differential GPS (dGPS) to avoid possible misinterpretations of the multibeam result. The combined

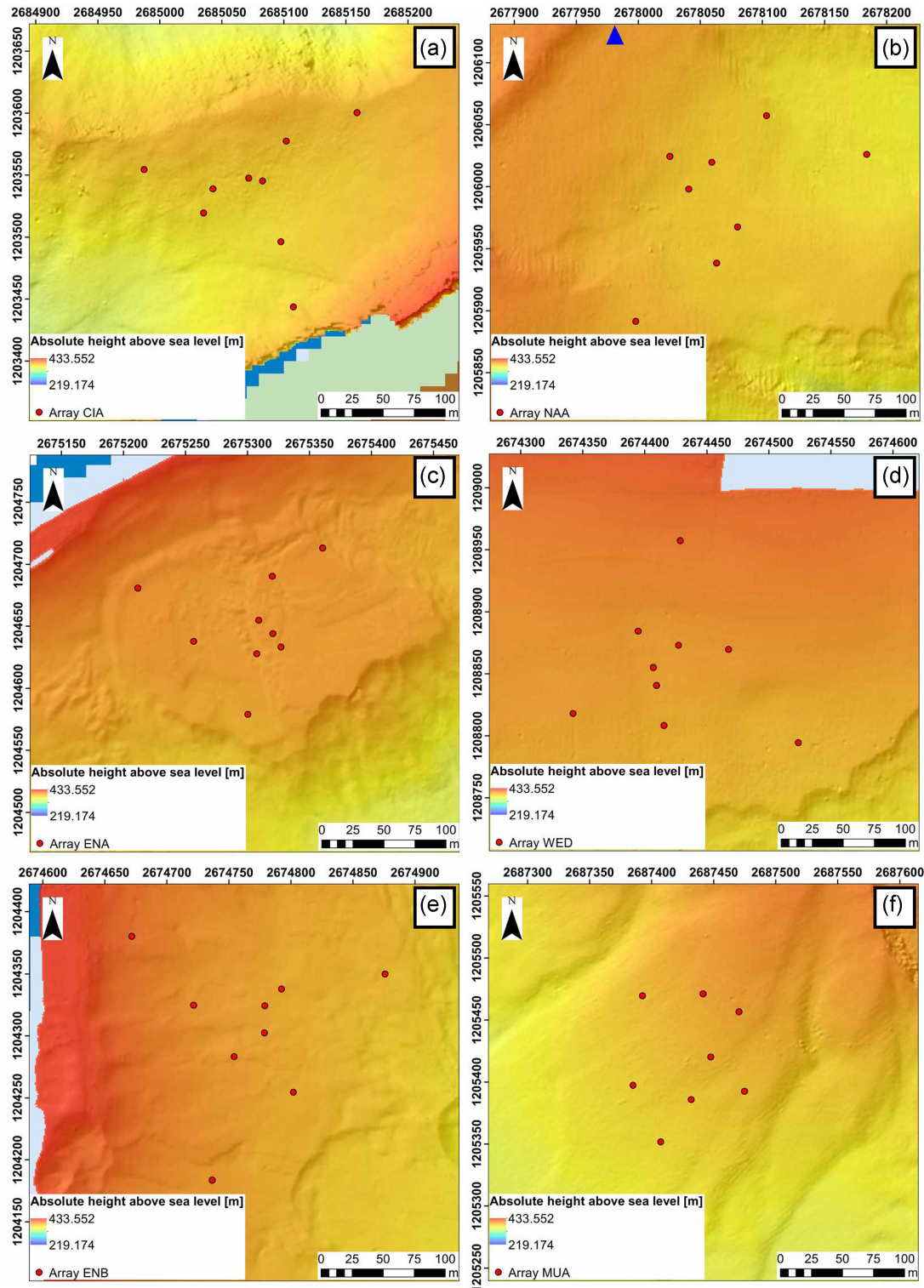


**Figure 2.** Maps of the analysed arrays (see Table 1): (a) Chindli CIA; (b) Nase NAA, blue triangle represents the station NAA04; (c) Ennetbürgen ENA; (d) Weggis WED; (e) Ennetbürgen ENB; (f) Muota MUA. The coordinates are given in the Swiss Coordinate System LV95 with units of 1 m.

analysis of the multibeam and dGPS data provides an unambiguous location of the OBSs on the lake floor with a precision of *ca.* 0.5 m. (ii) To correct the time stamps of the recorded data, we (a) either performed airgun shootings on top of the array (using a Bolt 1 in<sup>3</sup> airgun filled with compressed air at 7–8 MPa pressure) and detected

the arrival of the airgun-triggered pressure waves on the OBS hydrophone or (b) used the skew time (deviation between the GPS time and instrument time) of the internal OBS clock measured after OBS recovery. (a) The usage of the airgun signals for the time correction is based on the fact that we know the source and receiver posi-

tions and exact shooting time, thus we can calculate the theoretical source–receiver traveltime and compare it with the measured one. Such an analysis of each airgun shot and an optimization process to minimize the root mean square error allows the determination of the skew time for each station. (b) Using the skew time means a shift of the data time stamps for the last 2 hours of the recording (used for the array analysis) for the measured skew time value of the OBS clock, assuming it to be constant during that time window. The airgun option is more complex, but it allows us to perform time corrections at any time point of the recording when there are available airgun shots, while the skew option is only applicable to the last few hours of the recording due to the possible non-linearity of the clock drift (it depends on the clock aging, meteorological conditions etc.).

(iii) The correct orientation of the horizontal components of the OBS is crucial for the proper separation of Love and Scholte waves and for any other kind of analysis which needs an exact orientation of sensors (e.g. polarization analysis). To properly orient the horizontal components of the OBS, we also used airgun shootings with a wide azimuth coverage around the measuring instruments and estimated the back-azimuth of the recorded airgun signals. This approach is based on the knowledge that an airgun shot should produce a pure compression-wave (*P*-wave) signal. In order to estimate the orientation of the OBS horizontal components, we were looking for the rotation angle which maximizes the signal energy on the radial component, parallel to the direction of the airgun impulse propagation, and minimizes the signal energy on the transverse component. The polarization of the airgun shot is estimated according to Jurkevics (1988). The data are treated in 3-D and, as the airgun shot point is located on the water surface and the sensor at the lake floor, the 180° ambiguity of the orientation is resolved by identifying the upgoing direction as the one pointing at the airgun. To refine and check these results, we performed a cross-correlation of the horizontal components of different instruments at low frequencies (usually in the range 0.2–0.5 Hz). An additional constraint for the orientation of the horizontal components is the slope gradient: the internal levelling system of the OBS is constructed in a way that if the slope gradient is more than 5°, the east component is oriented along the slope. Similarly to the airgun signal, it is possible to use earthquake signals to correct the time stamps or orientation of the seismometer components, provided that enough seismic events were recorded during the operation time of the OBS. However, this was not the case for our experiment.

As mentioned before, in this study, we used two different types of OBS, which have the same electronic and measuring components, but different size and aspect ratio between vertical and horizontal dimensions of the frame. The base of the LOBSTER is larger than its height and is stable while descending and settling on the lake floor, providing almost no chance for instrument overturning. At the same time, the height of the NAMMU is larger than its horizontal dimensions (bell-like shape) and poses a risk for overturning during the descent or settling. This would lead to the failure of the levelling system and recording useless data. To overcome this issue, we developed a tilt- and orientation-tracking device to determine the tilt and the direction of maximum tilt of the NAMMU immediately after the deployment (more details can be found in Section 1 of the Electronic Supplement).

Together with the above-mentioned steps and to facilitate the outcome of the array analysis, we checked the thickness of the shallow (*ca.* first 2–10 m) soft sediment cover estimated from the interpreted reflection seismic profiles and the variability of the *H/V*

and Scholte wave ellipticity curves for the array stations. In some arrays, the responses of a few stations (the shape of the curves and frequency of the peaks) are significantly different from the others. Their exclusion from the array analysis allowed a better resolution of the phase velocity dispersion curves of surface waves. One example of such a case is station NAA04 of array NAA. While NAA04 was sitting on top of a thin sediment layer close to the moraine outcropping (Figs 1b and 2b, the location of NAA04 is marked by the blue triangle), the rest of the array was sitting on a thicker homogeneous sediment cover. This difference in the thickness of underlying sediments is also reflected in the shape of the *H/V* and ellipticity curves: for NAA04 they are significantly different from the curves for the other stations of this array (Figs 3b and d). Thus, station NAA04 was excluded from the array analysis.

### 3.2.2 Array methods

After the pre-processing of the array data, we used 2-hour-long recordings of all array stations and analysed them with the following methods to derive the phase velocity dispersion curves of Scholte and Love waves:

(i) *3C-HRFK* (Fäh *et al.* 2008; Poggi & Fäh 2010)—3-Component High-Resolution Frequency-Wavenumber technique, which allows the estimation of Love and Scholte wave dispersion curves and the Scholte wave ellipticity curve using high-resolution beamforming applied to the three-component seismic recordings. The data of the vertical component is processed as in Capon (1969), while the analysis of the horizontal components is angle-dependent, meaning that before processing each time window with Capon's method, they are projected in different angles with a step of 5° and the correct azimuth is identified in the direction of maximum coherence of the signals.

(ii) *WaveDec* (Maranò *et al.* 2012)—Wavefield Decomposition technique, which allows estimating Love and Scholte wave dispersion curves and the Scholte wave ellipticity curve using a maximum likelihood approach. The contribution of different surface waves is estimated for each time window and the parametrization of this method allows a mixture of Bayesian Information Criterion and Maximum Likelihood approaches.

(iii) *MSPAC* (Bettig *et al.* 2001)—Modified Spatial Autocorrelation technique, applied to the recordings of the vertical component of seismic sensors to extract the Scholte wave dispersion curve by computing the spatial autocorrelation coefficient, based on the SPAC technique (Aki 1957). Station pairs are grouped in rings based on the interstation distance, where for each ring an average cross-correlation function which should correspond to a Bessel function is computed. The derived Bessel functions are then inverted to define the Scholte wave dispersion curves.

These methods have different theoretical backgrounds and differ also in the performance depending on the composition of the recorded wavefield. Thus, it is beneficial to apply all of them to extract as much useful information as possible from the data. As a result of the application of these array techniques, we can extract phase velocity dispersion curves for Scholte and Love waves bounded by the array resolution limits (Wathelet *et al.* 2008).

### 3.2.3 Mode attribution analysis

Due to a relative lack of *a priori* information on the geological structure of the lake subsurface and the possible presence of lateral

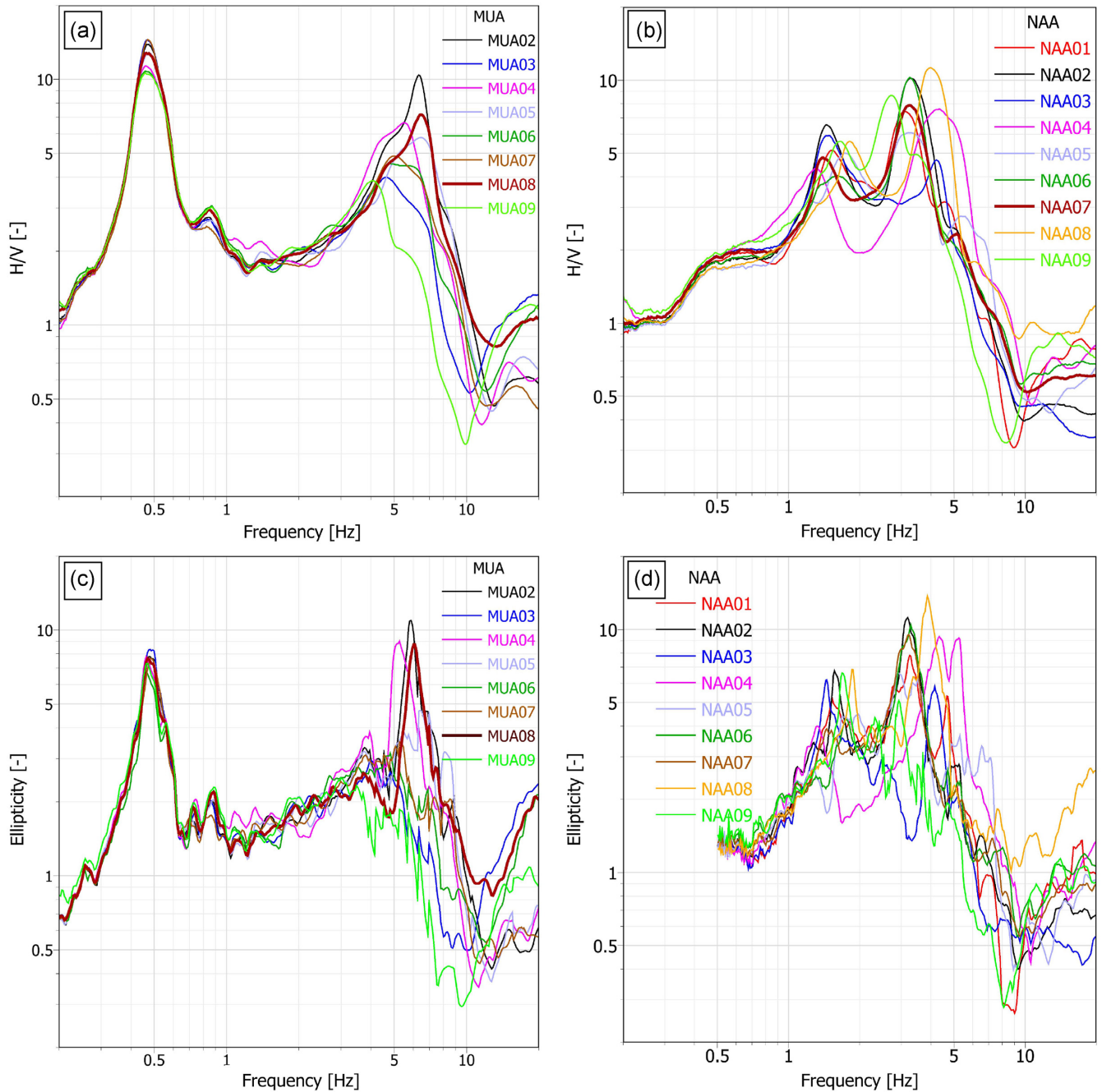


**Figure 3.** (a and b) H/V spectral ratios for the array (a) MUA and (b) NAA; (c and d) Scholte wave ellipticity curves for the array (c) MUA and (d) NAA.

heterogeneities, the attribution of the retrieved dispersion branches to defined modes of Scholte or Love waves is not trivial (Fazlavi & Haghshenas 2015; Chieppa *et al.* 2020). Moreover, we might observe shadows of strong Love waves in the Scholte wave analysis and vice versa; such shadows need to be identified and excluded from the further analysis. Hence, before the actual dispersion curve inversion we carry out a modal assignment procedure, inspired by that of Maraschini & Foti (2010) and similar to that of Abdel Moati *et al.* (2013); see also Bergamo *et al.* (2018) for an application to Rayleigh waves.

The modal assignment procedure follows this workflow:

- (i) All the picked dispersion curves (retrieved by applying different methods) are collected and subdivided according to the surface wave type, i.e. Scholte or Love.
- (ii) The data points which, according to expert judgement, belong to a particular propagation mode, are manually selected and given a defined mode number; the other data points are left without *a priori* modal assignment. In the current work, this *a priori* modal attribution involved only data points assigned to the fundamental mode. We attributed the phase velocity estimates belonging to the slowest dispersion branch to the fundamental mode, possibly identified by more than one processing method and, for Scholte

wave processed with 3C-HRFK analysis, at both radial and vertical components.

(iii) We model the subsurface as a stack of six homogeneous layers, overlain by the water column and overlying a half-space. This number of layers provides a flexible parametrization without introducing an excessive number of unknowns. We define wide but reasonable boundaries for the geophysical parameters of these layers ( $v_s$ ,  $v_p$  and thickness; the density is attributed *a priori*). The approximate depth range which they cover is defined by half of the maximum available wavelength (Socco & Strobbia 2004). This depth interval is divided into six layers; the central value of the thickness range of each layer increases logarithmically with depth. The  $S$ -wave velocity boundaries are defined adopting a tentative  $v_s$  based on the approximate conversion of the identified fundamental mode to a  $v_s$ -depth profile (e.g. Abbis 1981). The corresponding  $v_p$  boundaries are defined by imposing Poisson's ratio ranges compatible with water-saturated formations (0.4–0.5) for the layers closer to the lake bed, and lower Poisson's ratios (0.2–0.35) for the deeper formations.

(iv) A set of joint  $v_s$ – $v_p$  profiles is obtained by randomly extracting values for the geophysical parameters within the pre-set boundaries. The number of generated velocity models is between 250 000 and 750 000. For each model, the synthetic Love and Scholte wave dispersion curves are computed. The modes from the fundamental to the 4th higher mode are simulated.

(v) The collation between the experimental and each synthetic dispersion curve is summarized through the computation of a root-mean-square error ( $RMSE$ ):

$$RMSE = \sqrt{\frac{1}{n} \sum_{i=1}^n \left( \frac{d_i - s_i}{\sigma_i} \right)^2}, \quad (1)$$

where  $d_i$  is the  $i$ th data point (out of  $n$ ) of the experimental dispersion curves with its associated uncertainty  $\sigma_i$ , and  $s_i$  is its counterpart from the synthetic curve. For the phase velocities  $d_i$  attributed *a priori* to a given propagation mode, the counterpart  $s_i$  is the phase velocity from the corresponding synthetic mode at the same frequency. For the unassigned phase velocities  $d_i$ , the counterpart  $s_i$  is the closest (in terms of phase velocity) data point having the same frequency:  $d_i$  is therefore attributed to the mode having the closest phase velocity at the same frequency. If the distance between the unassigned experimental data point and the closest synthetic value exceeds a threshold (set to 20 %), the experimental point is identified as an outlier and discarded from the  $RMSE$  computation.

(vi) The pairs of synthetic curves (Love and Scholte) that achieve the lowest misfit, are assumed to propose the most plausible modal attribution. We visually collate the 20 best performing synthetic curves with the experimental data; if they propose more than one alternative for mode numbering (as in Maraschini & Foti 2010), we additionally compute the synthetic  $H/V_{\text{microtremor}}$  curves for them (forward modelling of Lontsi et al. 2019). We then compare these curves with the experimental  $H/V$  from the stations located at the array centre. We finally select the modal attribution corresponding to the  $H/V$  curves with a better match with the experimental data, focusing on the frequency band covered by surface wave dispersion data.

This mode identification procedure does not constitute the actual inversion process, but a preliminary operation to determine the most likely mode numbering for Love and/or Scholte wave dispersion curves. The obtained modal assignments are then adopted in the following phase of the actual inversion.

### 3.3 Inversion of dispersion and Scholte wave ellipticity curves

An inversion for the subsurface structure from dispersion and/or ellipticity curves of surface waves is a non-linear inverse problem (e.g. Parolai et al. 2006; Hobiger et al. 2013). In particular, Fäh et al. (2003) proposed a non-linear inversion of the  $H/V$  ratio, Wathelet et al. (2004) developed an optimization inversion of dispersion curves of surface waves, Arai & Tokimatsu (2005) proposed a joint iterative approach to the inversion of both types of data, and Wathelet (2008) developed a joint inversion of both types of data utilizing the neighbourhood algorithm. More recently, Li et al. (2012) applied a non-linear Bayesian inversion of multimodal Scholte wave dispersion curves.

In this work, we invert the dispersion and ellipticity curves jointly. This allows us to better constrain the final model in terms of  $S$ -wave velocity and layer thickness (e.g. Fazlavi & Haghshenas 2015; Ullah & Prado 2016; Gouveia et al. 2018). For this non-linear inversion, we use the multizonal transdimensional Bayesian inversion (Hallo et al. 2021) implemented in the Neowet Coral software (more information on the offshore method approbation can be found in Section 8 of the Electronic Supplement). A drawback of the widely used standard optimization inversion techniques is a significant inherent non-uniqueness as different velocity models may provide a similar fit to the observed data. The used inversion approach utilizes the Bayesian probabilistic framework (Tarantola 2005; Bodin et al. 2012) and treats the number of layers in the model as an unknown parameter of the inverse problem itself. The measured data and model parameters are related through multidimensional conditional probability density functions (PDF). The thickness of the layers is parametrized on the logarithmic scale in order to reflect the decrease of the information available in measured data with increasing depth. As the number of layers is not constrained *a priori*, the dimension of the model space may vary. This feature allows the utilization of a self-adapting mechanism, where models consist of as few layers as possible, but not less than required by the observed data (Hallo & Gallovič 2020). For the exploration of such a model space, it utilizes a reversible-jump Markov-chain Monte Carlo algorithm (Hastings 1970; Green 1995; Sambridge et al. 2006) enhanced by the parallel tempering of Markov chains (Sambridge 2014). To suppress a possible dependency on the initial model and intrachain correlations, Neowet uses multiple parallel chains with independent and random initial models. For the forward modelling of Love wave dispersion curves, it utilizes the geopsy package 'gpd' (Wathelet et al. 2020). The forward modelling of the Scholte wave dispersion and ellipticity curves is performed using the Matlab codes developed at the Geophysics Laboratory of Politecnico di Torino (Maraschini 2008; Socco et al. 2011). To summarize, an important advantage of this inversion method is that it suppresses the dependency on the initial model and does not require a prescribed number of layers or their thickness ranges due to the utilized self-adopting data-driven mechanism.

The main result of the inversion is represented by the posterior PDF on the model space. However, to interpret the inversion result and compare it with other investigations, a single representative subsurface model is usually required. To comply with the standard optimization inversion approaches, we derive a Maximum Likelihood (ML) model which has the lowest misfit between observed and modelled data. The ML model can be misleading due to overfitting of generally uncertain data, so we derive also the Maximum *a Posteriori* (MAP) model which is representative in terms of the Bayesian statistic. In particular, our MAP model is closest to the maximum

posterior marginal probability of  $v_s$  in terms of the L1-norm (Hallo *et al.* 2021). This model together with the uncertainty estimate from the posterior PDF is then used for geological interpretation of the subsurface structure.

### 3.4 Strategy for the geological interpretation of the inversion results

To provide a geological interpretation for the derived  $S$ -wave velocity profiles of the subsurface and to attribute a lithological unit to the range of  $v_s$ , we use the following strategy: (i) check the overall range of velocities and the thicknesses of the identified layers, (ii) analyse the velocity variations along the profile and (iii) compare the obtained data with the previously available knowledge from reflection seismic profiles, Cone Penetration Testing, sediment cores, drilling reports and literature data.

The main lithological units of the subaquatic slopes previously defined in Lake Lucerne are lacustrine, glaciolacustrine and glacial sediments/moraine (Strasser *et al.* 2007; Strupler 2012; Hilbe & Anselmetti 2014a). The lacustrine sediments are described as homogeneous to mottled silty clays with abundant organic material (deposited in a Holocene lake); the glaciolacustrine sediments consist of thinly laminated (mm to sub-mm-scale) clays and clayey silts (deposited in a cold proglacial lake); the glacial sediments that are often tilted and deformed are composed of thinly-laminated clayey silts to silty clays with sparse (<20 %) sand and medium-sized gravels (deposited in direct ice-vicinity or even ice-contact in front or below the glacier). For the study sites in a deltaic setting (MUA and ENB), we use additional lithological units and divide the lacustrine succession into two sublithologies: fine-grained sediments composed of silty clays with the presence of sandy layers and probably related to a comparably distal deltaic sedimentation (deposited as distal deltaic sediments, quite similar to the normal lacustrine sediments) and highly variable sediment sequences, consisting of the coarse- to the fine-grained material probably related to the proximal deltaic sedimentation and mostly delivered by the inflowing river (proximal deltaic sediments). From the geological map and previous investigations (Pfiffner *et al.* 1990; Greber *et al.* 1994; Hilbe *et al.* 2011; Pfiffner 2018), the bedrock at the investigated sites consists either of Cretaceous limestone/marl intercalation from the Helvetic nappes (at NAA, ENA, ENB, CIA and MUA) or of Neogene clastic Swiss molasse deposits composed of silty, sandy or conglomerate material (at WED).

The lithological units described can be identified on the inverted  $S$ -wave velocity profiles. For example, for the lacustrine sediments, we expect relatively low  $v_s$  values (below  $100 \text{ m s}^{-1}$ ; Caiti *et al.* 1994; Carr *et al.* 1998; Odum *et al.* 2004; Kugler *et al.* 2007; Vanneste *et al.* 2011); for the glaciolacustrine sediments, we expect relatively stable  $v_s$  values as we know that their shear strength is also quite stable (Strasser *et al.* 2007; Strupler *et al.* 2017) and shear strength can be related to  $v_s$  (Long *et al.* 2009; Qureshi *et al.* 2013; L'Heureux & Long 2017); for the glacial and proximal deltaic sediments, the range of  $v_s$  might be quite wide and variable (Carr *et al.* 1998; Odum *et al.* 2004; Kugler *et al.* 2007; Hunter *et al.* 2016; Tong *et al.* 2018). The  $S$ -wave velocity of the bedrock strongly depends on the weathering and glacially-induced deformations and can be quite variable.

## 4 RESULTS AND DISCUSSION

Using different data processing techniques, we obtained  $H/V$  and Scholte wave ellipticity curves as well as the parameters of the

particle motion polarization for every single station of the arrays. The array analysis was targeting the extraction of Scholte and Love wave dispersion curves for the investigated sites. In the following, we present results for selected stations and arrays; results for the other stations/arrays can be found in Sections 3–7 of the Electronic Supplement.

### 4.1 $H/V$ spectral ratio estimation

Figs 3(a) and (b) represent the  $H/V$  curves determined for each station of two different arrays: MUA (deltaic slope) and NAA (moraine ridge slope). The main outcome of the  $H/V$  analysis is the fundamental frequency of resonance  $f_0$  and the relative intensity of the velocity contrasts related to the peaks in the  $H/V$  curves (stronger velocity contrasts should produce a stronger peak). Moreover, the variability of the resonance frequency between the stations of the same array indicates the variability of sediment thickness (assuming no lateral variability of the  $S$ -wave velocity). The presence of several peaks in the  $H/V$  curves indicates the presence of several strong velocity contrasts at depth.

For the array MUA,  $f_0$  is around 0.47 Hz and the  $H/V$  spectral ratio at  $f_0$  has an amplitude between 10 and 10.5. This is very similar across all the stations of the array, indicating a homogeneous subsurface structure. The second peak between 4 and 6.5 Hz is more variable in terms of frequency and amplitude. As it corresponds to the shallower velocity contrast, we explain its variability by the differences in sediment thickness and composition in the upper part of the profile (as we are close to the delta of the Muota, the content of clay, silt and sand in the upper part of the profile can significantly change over a short distance).

For the array NAA, we also observe two peaks, but in this case, they are very close and both of them vary inside the array.  $f_0$  is between 1.3 and 1.8 Hz, while the second peak ranges from 2.7 to 4.6 Hz. The variability of the peak frequency and intensity is again explained by the changes of sediment thickness, which are also observed on the reflection seismic profiles.

The other arrays (except WED) usually also have two peaks in the  $H/V$  curves in the frequency range between 0.5 and 6 Hz; the  $H/V$  amplitude of the second peak ( $\sim 10$ ) is higher than the amplitude of the first one ( $\sim 5$ –6). For array WED, we observe a single wide and relatively flat peak.

### 4.2 Scholte wave ellipticity

Figs 3(c) and (d) represent the examples of the ellipticity curves for the arrays MUA and NAA. Here again, we observe a more homogeneous appearance of the curves for the array MUA and a more heterogeneous picture for the array NAA. The most central station of MUA is MUA08, and for NAA it is NAA07.

### 4.3 Polarization analysis

Figs 4(a) and (b) show an example of the polarization analysis for the central stations of the arrays MUA and NAA, respectively. The polarization ellipticity curves show consistency with the  $H/V$  and Scholte wave ellipticity curves: the peaks in the latter ones correspond to low values of the polarization ellipticity, indicating a linear polarization. The dip is usually between  $0^\circ$  and  $\pm 10^\circ$ , indicating a flat installation of the sensor on the ground. The results for the different stations of the arrays are usually very similar and there are no signs of possible 2-D polarization effects on the strike

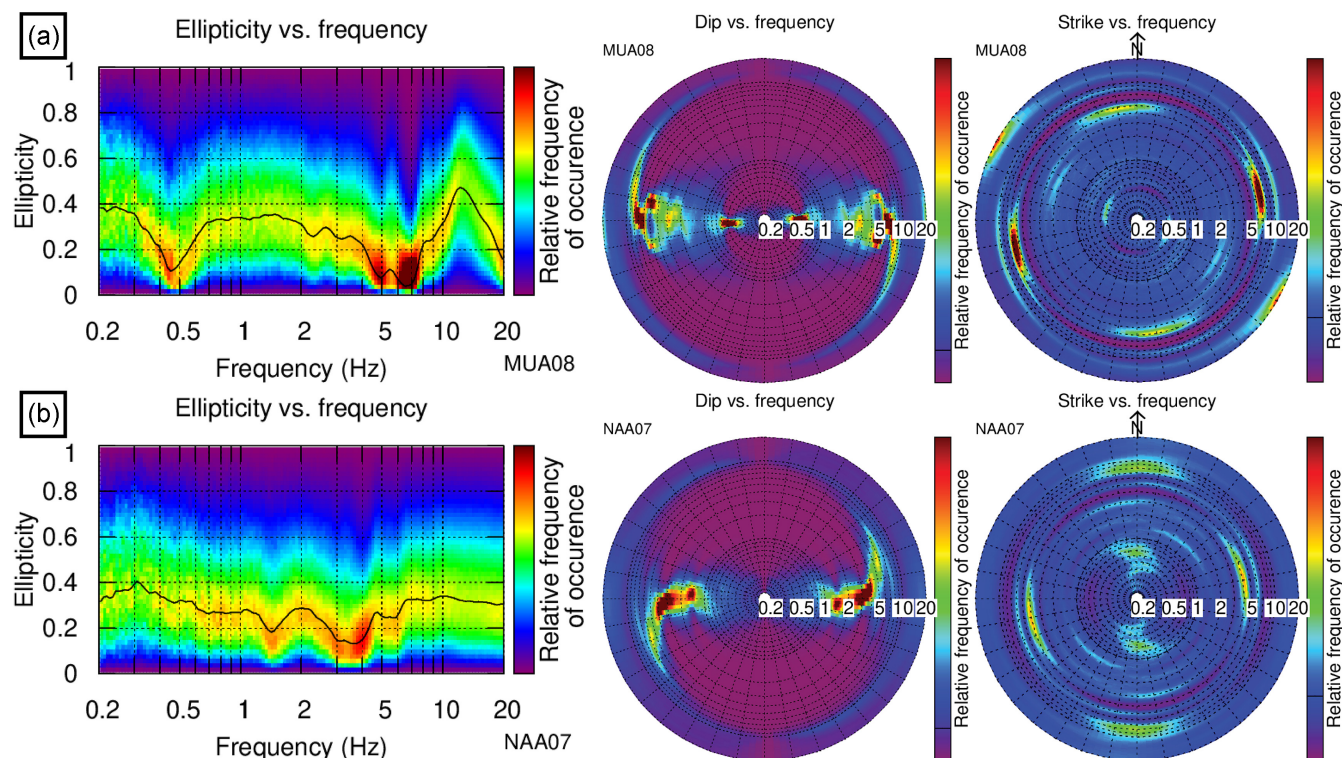


**Figure 4.** An example of polarization analysis results for the central stations MUA08 and NAA07 of the arrays (a) MUA and (b) NAA, respectively.

plots. The results for different arrays vary due to the site-specific topography and subsurface structure.

#### 4.4 Scholte and Love wave dispersion curves

By applying different array processing techniques (3C-HRFK, MSPAC and WaveDec), we could extract dispersion curves for Scholte and Love waves, except for array WED, where we observe only a Scholte wave dispersion curve, and the application of the MSPAC technique, which provided Scholte wave dispersion curves only for some arrays.

Fig. 5 represents the overlap of dispersion curves for the arrays MUA and NAA detected using the different methods. The outcome of each processing technique can be found in Section 6 of the Electronic Supplement. We did not extract reliable Scholte wave ellipticity curves from the array data, so they are not presented. The theoretical array resolution limits were calculated according to Wathelet (2008).

The dispersion curves, derived by different techniques, have different quality, resolution and extent. For example, MSPAC is limited to the vertical component of the 3-component recording and allows the derivation of the dispersion curves only for three arrays (ENA, NAA and MUA) out of six. However, for the array MUA, the dispersion curve could be retrieved to lower frequencies than observed with the other two techniques. WaveDec was successful for all arrays, but in some cases the resolution of the frequency-wavenumber plots was not very good and the delineation of the dispersion curves was not straightforward. Among these three techniques, the 3C-HRFK provided the most robust result, although the modal attribution of the defined dispersion curves was still not very clear and required post-processing represented by the mode attribution analysis. The combination of the results of all three techniques provides

the most robust estimation of surface wave dispersion curves and an increased frequency extent.

Overall, the extracted dispersion curves cover a limited frequency range between 0.5 and 4 Hz. The irregularities of the dispersion curves can be mainly related to: (i) difficulties with the data processing, artefacts from the dispersion curve picking and pre-processing steps which have to be applied to the data (e.g. time correction or horizontal component re-orientation) and might affect the processing results and (ii) the heterogeneity of the shallow subsurface structure below the array stations. In the best case, the dispersion curves cover the frequency range from the fundamental peak of resonance to the onset of the left flank of the second  $H/V$  peak at a higher frequency. High variability of the second  $H/V$  peak within the array indicates lateral variability of the shallow structure so that fundamental mode surface waves change their properties inside the array and are difficult to analyse with array methods that assume a plane wave field.

For some arrays, the dispersion curves obtained with different techniques overlap quite well and the mode identification is straightforward (e.g. MUA and NAA), while for the other arrays the mode attribution is not clear: we identified a few possible mode attributions, which had to be verified with the mode attribution analysis (see Section 4.5). Additionally, if we compare dispersion curves for the sites which belong to the same geomorphological types of the slopes, they are quite similar: for the moraine ridge slopes (CIA and NAA) we observe a more pronounced change of velocity and a narrower frequency range for the dispersion curve; deltaic slopes (ENB and MUA) have the flattest and most continuous dispersion curves that correspond to the massive amount of sediments which can be found there; hemipelagic lateral slopes (WED and ENA) have the lowest surface wave velocities in the upper part of the profile (especially WED).

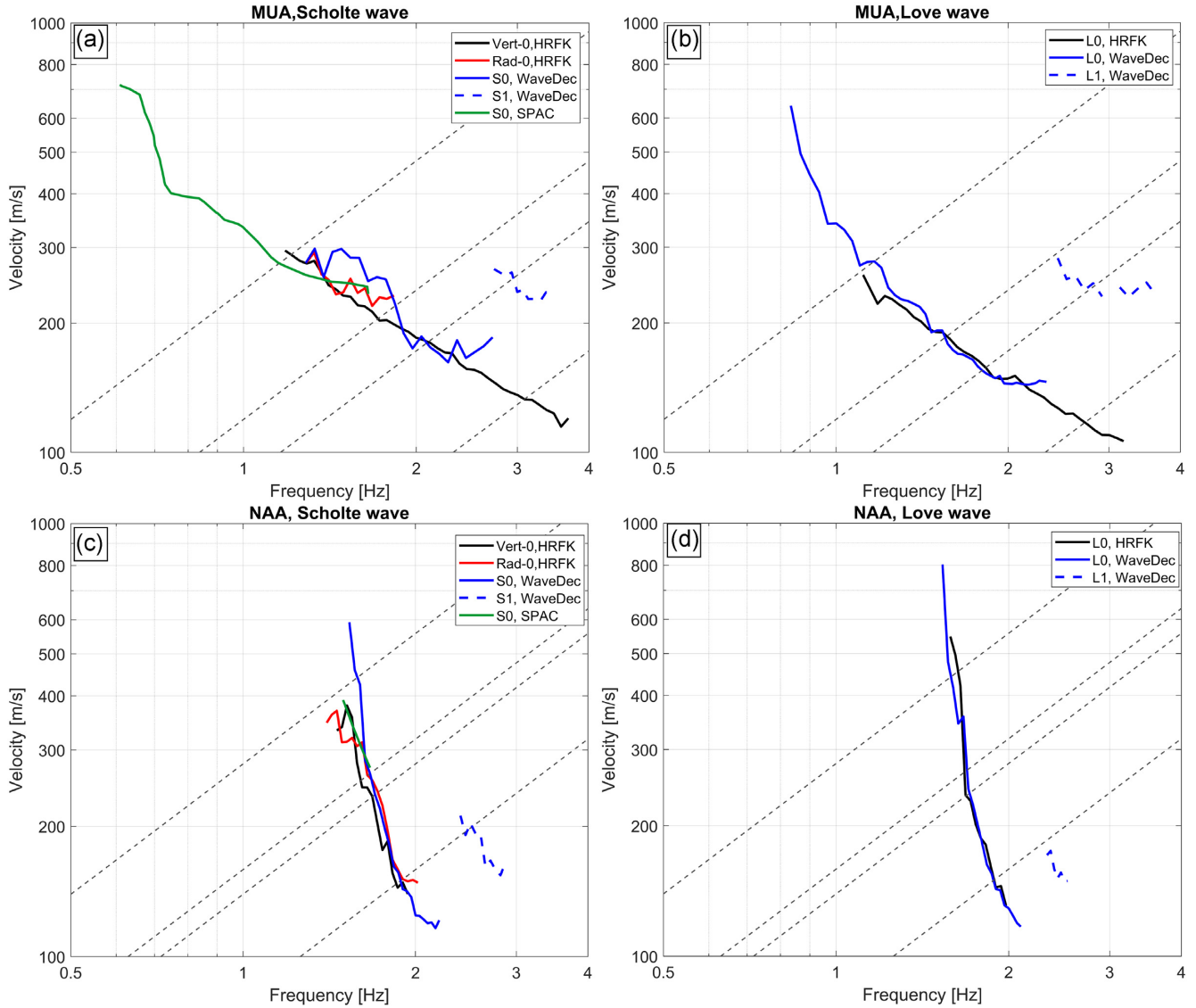


**Figure 5.** Scholte and Love wave dispersion curves obtained for (a and b) array MUA and (c and d) array NAA using 3C-HRHK, MSPAC and WaveDec techniques. S/L–Scholte/Love wave; Vert/Rad–vertical/radial component; 0 or 1: fundamental mode or first higher mode, respectively; HRHK–dispersion curves obtained using 3C-HRHK; WaveDec–dispersion curves obtained using WaveDec; SPAC–dispersion curves obtained using MSPAC. The grey dashed lines represent the theoretical array resolution limits.

#### 4.5 Mode attribution analysis

For three sites, the assignment of the modes was straightforward (MUA, NAA and WED). The other sites (CIA, ENA and ENB) required a mode attribution analysis before the inversion. We present an example of the mode attribution analysis for array ENB in Fig. 6; for array MUA (Fig. 7), the attribution of the first higher mode is verified.

For arrays ENB and MUA, we tested 750 000 and 250 000 different models, respectively. For both arrays, we used all dispersion curves identified during the data processing (Figs 6a, b and 7a, b); selected segments were attributed *a priori* to the fundamental mode (black squares in Figs 6c and 7c). The resulting pairs of Love/Scholte synthetic curves that best approach the experimental data are grouped into three (ENB, Fig. 6c) or two (MUA, Fig. 7c) different families (marked by different colours); each group proposes a different modal assignment. To choose the most probable assignment, we compared the *H/V* spectral ratio from a station in

the centre of the array (Fäh *et al.* 2009) with the synthetic *H/V* curves (Lontsi *et al.* 2019) derived from the velocity models corresponding to the best-fitting curves from Figs 6(d) and 7(d). The synthetic *H/V* curves follow consistent patterns for each family of models. For ENB (Fig. 6d), the synthetic *H/V* curves from the family of models ‘C’ (red lines in Fig. 6d) showed a better match with the experimental curve, particularly at low frequencies; hence the mode numbering proposed by this family of models was finally adopted (Fig. 6c). For MUA, the family of models whose synthetic *H/V* curves best approached the experimental curve was the family named ‘A’ (yellow lines in Fig. 7d); the corresponding modal assignment was selected for the inversion (Fig. 7c).

#### 4.6 Inversion

The purpose of the inversion was to infer the *S*-wave velocity profile (supplemented with an uncertainty estimate) of the site assuming

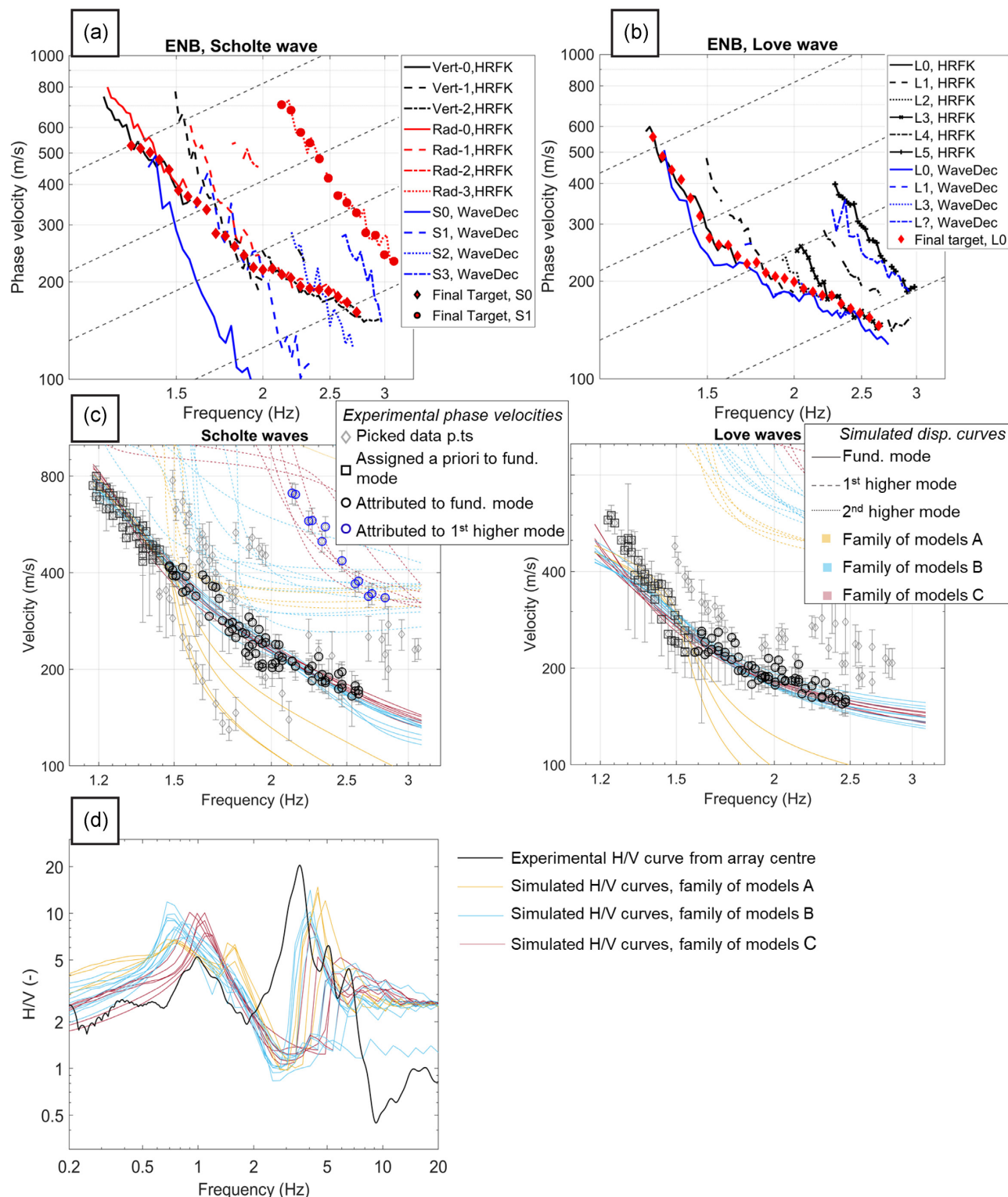


**Figure 6.** Mode attribution analysis for the array ENB. (a) Scholte and (b) Love wave dispersion curves obtained using different techniques (numbers identify different dispersive branches without referring to the mode). The red diamonds and circles represent the targets for the inversion, as identified from the mode attribution analysis (Figs 6c and 6d). S/L–Scholte/Love wave; Vert/Rad–vertical/radial component; 0, 1, ...  $n$  – fundamental, first or  $n$ th higher mode. The grey dashed lines represent the theoretical array resolution limits. (c) Collation between the experimental dispersion data (grey diamonds) and 20 best-fitting synthetic multimodal dispersion curves (continuous lines represent fundamental modes, dashed and dotted lines represent first and second higher modes, respectively). (d) Comparison of the  $H/V$  curve from a station at the centre of the array (black line) and the simulated  $H/V$  curves corresponding to the models in Fig. 6(c).

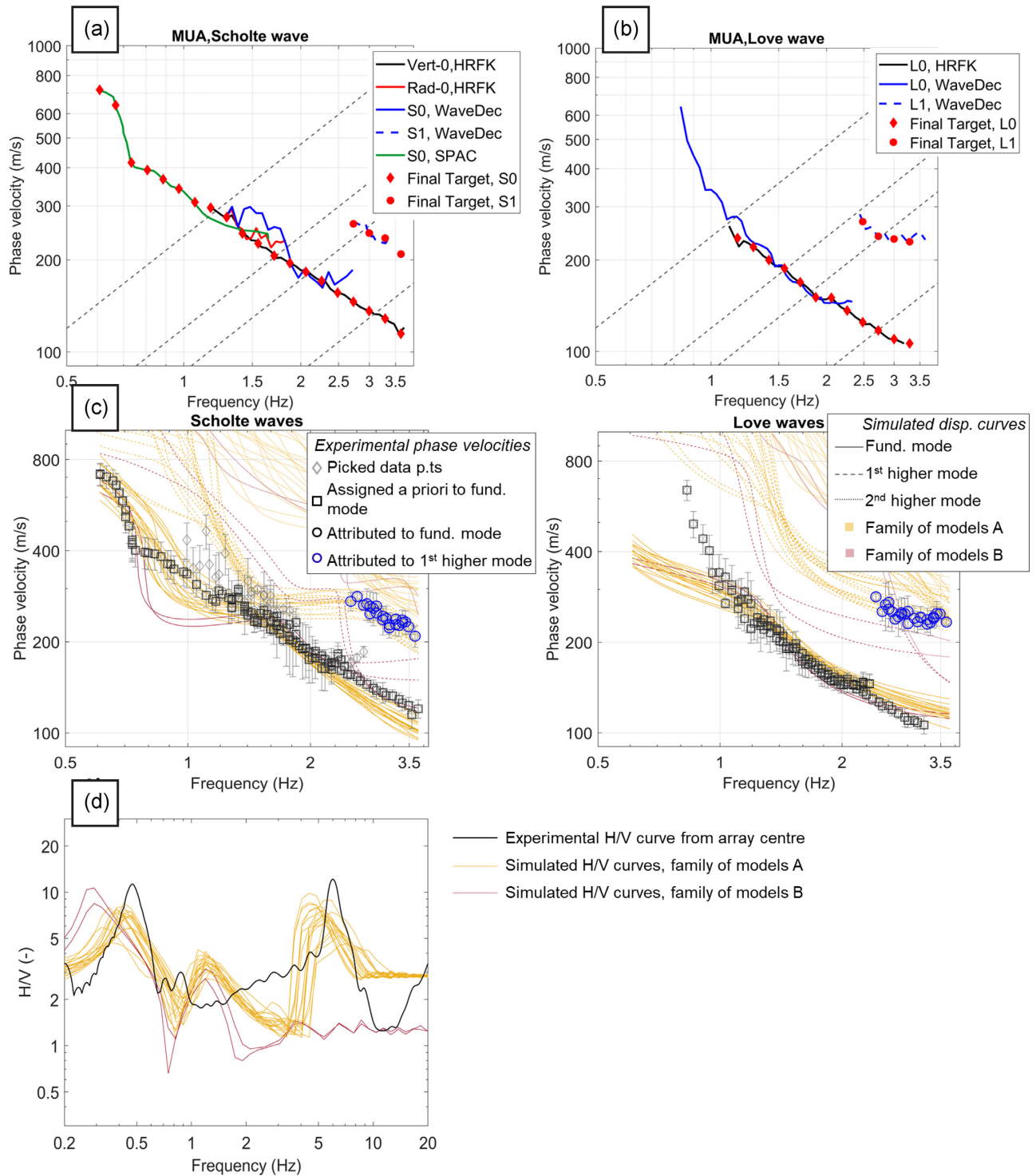


**Figure 7.** Mode attribution analysis for the array MUA. Figure layout is the same as in Fig. 6.

a horizontally layered 1-D subsurface structure. Additionally, the  $P$ -wave velocity and associated Poisson ratio were retrieved, but due to the limited data quality, these parameters were poorly resolved and are not presented in this section (see Section 9 of the Electronic Supplement). The input for the inversion included the selected dispersion curves after the mode attribution analysis and parts of the Scholte wave ellipticity from the analysis of the recordings of the central array station obtained using RayDec. Each input data curve

was supplemented with the uncertainty estimate (see Fig. 8). Due to the limited frequency range of the acquired dispersion curves, the resolution and maximum resolved depth of the inferred  $S$ -wave velocity profiles was also limited. Adding the Scholte wave ellipticity curves increased the resolution of the depth of the layer interfaces with strong velocity contrasts (e.g. between the sediments and the bedrock). The Scholte wave ellipticity derived using RayDec was always attributed to the fundamental mode.

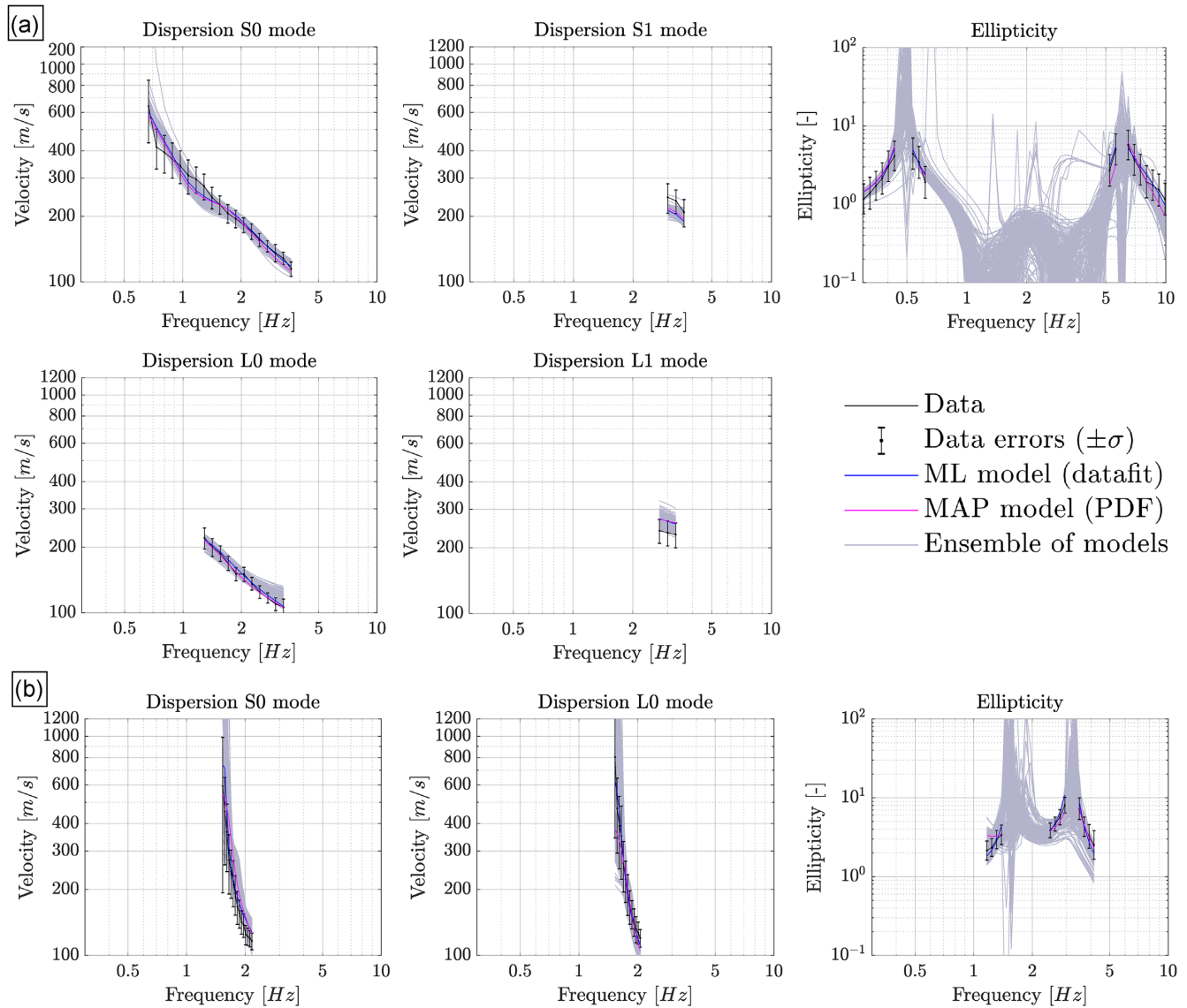


**Figure 8.** The fit of the synthetic data to the experimental data (dispersion curves of a fundamental and first higher mode of Scholte and Love waves and ellipticity of the fundamental mode of Scholte wave) for the array (a) MUA and (b) NAA: the grey curves represent 200 random models from the ensemble of solutions to show the scatter of possible inverse problem solutions with different misfit (i.e. posterior predictive distribution); the black points with the data errors represents the experimental data; the blue and purple curves are the ML and MAP models, respectively.

The resultant posterior PDF of the subsurface can be used for the construction of the posterior marginal PDFs of  $S$ -wave velocity,  $P$ -wave velocity, and associated Poisson ratio. These posterior marginal PDFs can be visualized by 2-D histograms with the probability of having a certain velocity at a certain depth shown by colour. They are also supplemented with representative ML and MAP models. In Figs 8 and 9 we present the inversion examples for the arrays MUA and NAA: the fit between the experimental and modelled data and the posterior marginal PDF on  $v_s$  with the MAP and ML models, respectively. The grey-shaded areas on the velocity profiles in Fig. 9 mark the depth with no resolution; in the unresolved part of the profile, the prior PDF is stronger than the evidence in the observed data, and as a result, we observe random values of seismic wave velocities at this depth. Based on the analysis of the single-station data (Figs 3a and c), we conclude that array MUA at the Muota delta represents a site with laterally homogeneous subsurface structure.

The dispersion curves extracted for this array (Figs 5a and b) are very consistent between the different processing methods. For the inversion at MUA, we used dispersion curves for the fundamental and first higher modes of Scholte and Love waves, and the flanks of the peaks of the ellipticity curve for the most central array station (Fig. 8a). The ensemble of the solutions sampling the posterior PDF for this inversion (Fig. 9a) includes 172 000 models, while in total the inversion with parallel tempering explored more than 3 600 000 models (each inversion node had 20 chains on various temperatures).

For site NAA, the dispersion curve segments are shorter (Figs 5c and d) and therefore the uncertainties are larger than for MUA. For the inversion, we used dispersion curves for the fundamental mode of Scholte and Love waves, and the flanks of the peaks of the ellipticity curve for the most central array station (Fig. 8b). The MAP and ML models start to diverge below 27 m depth (Fig. 9b). The

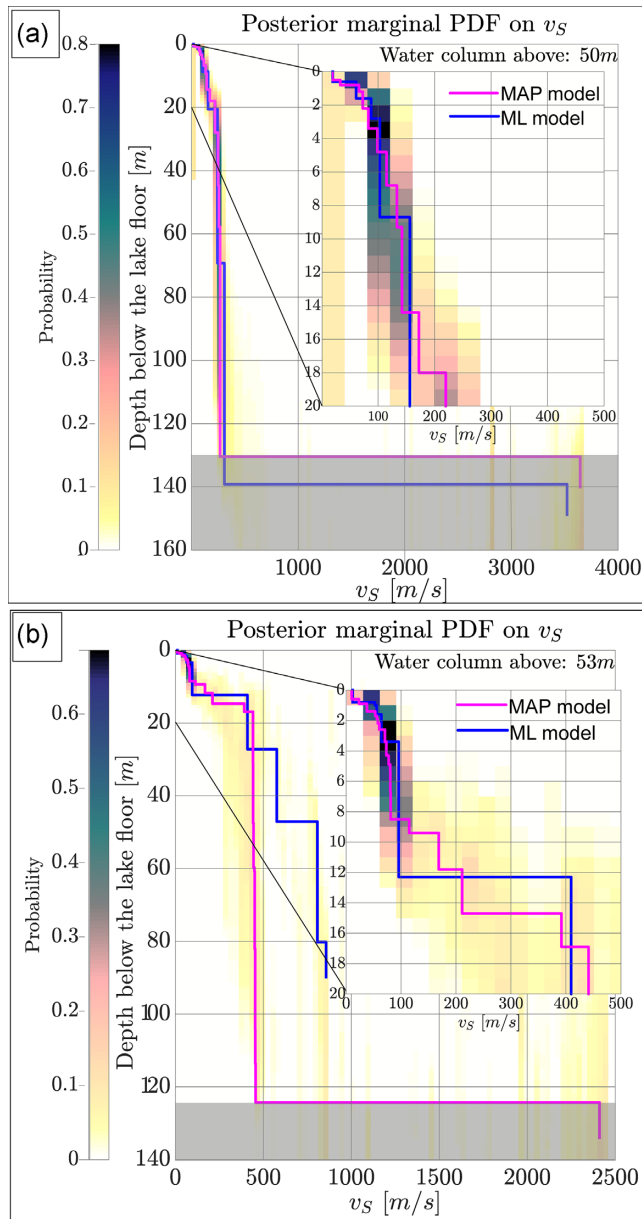


**Figure 9.** *S*-wave velocity profile at site (a) MUA and (b) NAA and zoom to the upper 20 m. The colour represents the probability of the posterior marginal PDF; the blue and purple curves are the ML and MAP models, respectively. The shaded area marks the part of the profile with random values of velocity where we don't have resolution.

ensemble of the solutions sampling the posterior PDF for this inversion includes 67400 models, while in total the inversion explored more than 1 570 000 models.

For array MUA, the MAP and ML models are very similar, which means that the solution is very robust. For array NAA, we see a decent fit of the MAP and ML models in the upper 27 m, but below they are quite different, meaning that the ML model corresponds to a single local minimum of the misfit but is not representative in terms of posterior PDF for the complete ensemble of solutions. Also note that the uncertainty (shape of the posterior PDF) increases significantly below this depth. For this reason, we always rely on the MAP model and consider the uncertainty when interpreting the inversion results. The fit between the experimental and modelled data is quite good and the velocity structure is resolved down to

approximately 130 m depth for MUA and 125 m depth for NAA. Below these depths, we do not have resolution due to the limited frequency range of the dispersion curves (implied by very uncertain posterior PDF). For MUA, the  $v_{S30} - ML$  (where  $v_{S30}$  is the average *S*-wave velocity in the topmost 30 m) is equal to  $130 \text{ m s}^{-1}$  and  $v_{S30} - MAP$  is  $130 \text{ m s}^{-1}$ , while for NAA, the  $v_{S30} - ML$  is equal to  $125 \text{ m s}^{-1}$  and  $v_{S30} - MAP$  is  $121 \text{ m s}^{-1}$ .

In the following, we provide a lithological interpretation of the velocity profiles obtained at each site.

#### 4.7 Geological interpretation of the *S*-wave velocity profiles

Following our strategy for the geological interpretation of the inversion results, we assigned the lithological units to the derived *S*-wave velocity profiles. The typical lithological units met in Lake Lucerne (i.e. lacustrine, glaciolacustrine, glacial and proximal/distal deltaic sediments) were identified on the inverted *S*-wave velocity profiles (Figs 10–12). Although the derived MAP and ML  $v_s$  models are very similar, we used the MAP models for the geological interpretation as they are representative in terms of the PDF. Table 2 provides a summary of the possible ranges of *S*-wave velocities for different lithological units (as observed in our data and confirmed by previous studies mentioned in Section 3.4).

The maximum resolved depth of the *S*-wave velocity profiles is represented by the dashed horizontal lines. Below this maximum depth, our measurements do not constrain the inversion and we can only see the random values from prior parameter distribution, but not the real properties of the investigated site. Thus, the bedrock depth is not resolved in our  $v_s$  profiles. Table 3 summarizes the information about the thickness of different lithological units at different sites, as revealed by the inversion results. Our interpretation of the *S*-wave velocity profiles is also compared with existing data from seismic reflection, sediment core and drill core data. This comparison, as described in the following paragraphs, shows that ambient vibration techniques allow us to extract reliable information on the subsurface structure in offshore environments and can thus be used in areas where conventional methods (e.g. reflection seismic) are not successful.

At CIA, the interpreted  $v_s$  profile has 2.2 m of lacustrine and 9.8 m glaciolacustrine sediments. These thicknesses compare well with the results of two borehole drillings in a distance of 350 and 450 m from the array which recorded around 4 m of lacustrine deposits and around 9 m of glaciolacustrine sediments on top of the moraine (Freimoser & Gautschi 1991).

At NAA, the estimated thickness of the lacustrine sediments is 4.3 m. This result fits with the stratigraphy of the sediment core 4WS11-MO6, retrieved in 2011 by Strupler (2012) at the array site NAA, where the thickness of the lacustrine sediments is 4.1 m. This is also in line with the interpretation of reflection seismic data at this location, which gives the thickness of the lacustrine sediment cover between 4 and 8 m.

At WED, the total thickness of lacustrine and glaciolacustrine sediments is estimated to be 3.5 m (it is not possible to differentiate them, probably due to the very small thickness of glaciolacustrine material). The sediment core 4WS05-K6 retrieved in 2005 by Strasser *et al.* (2007) at ca. 600 m distance from array WED, covers the same sediment sequence, with 2.5 m and 1.2 m thickness of lacustrine and glaciolacustrine sediments, respectively. The total thickness of both units, retrieved from the reflection seismic profiles below the array, is in the range 3.1–8.3 m. Our thickness estimate is a

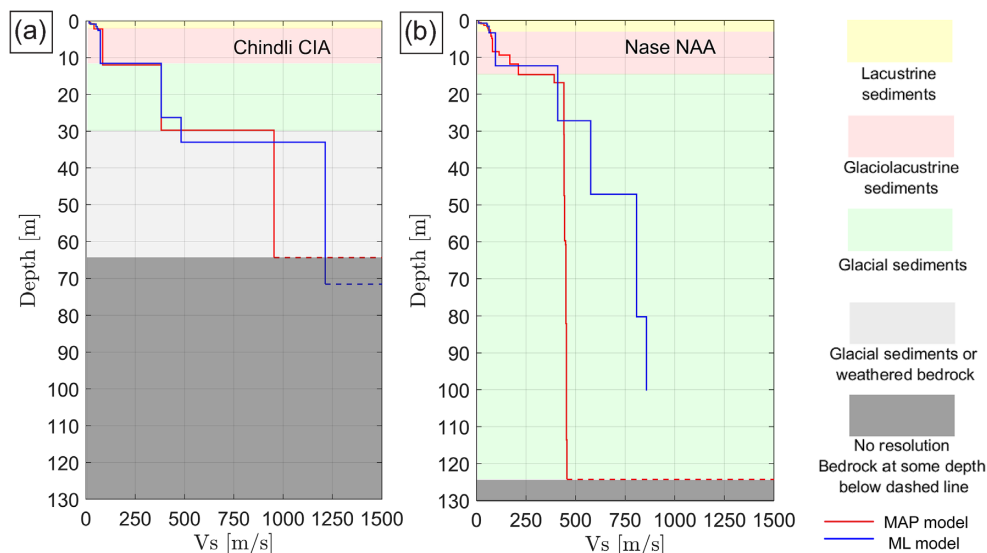


**Figure 10.** MAP and ML  $S$ -wave velocity profiles (red and blue curves, respectively) with assigned lithological units for (a) Chindli CIA and (b) Nase NAA.

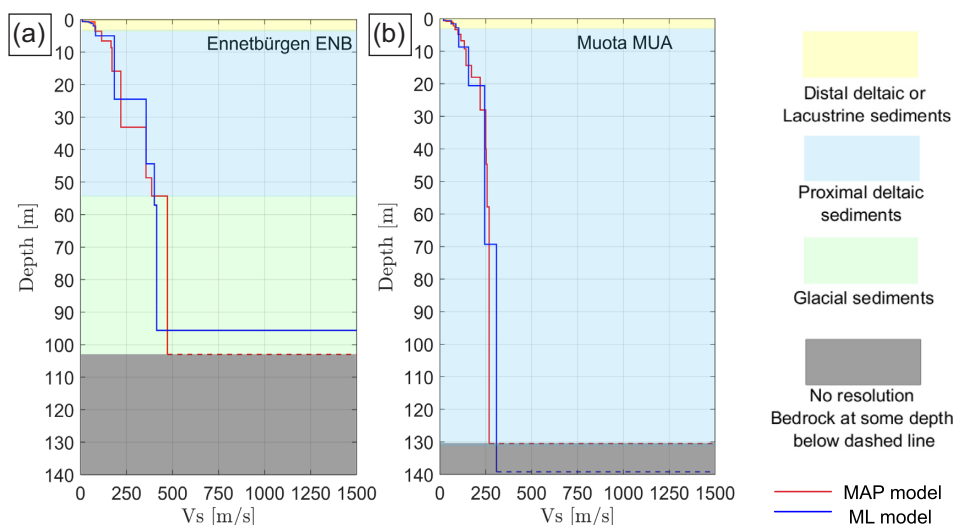


**Figure 11.** MAP and ML  $S$ -wave velocity profiles (red and blue curves, respectively) with assigned lithological units for (a) Ennetbürgen ENB and (b) Muota MUA.

bit smaller than in the previously available data due to the variability of the sediment thickness below the array and considerable distance from the array to the coring location. The results of an available deep onshore drilling allow us to describe the bedrock below the array as sea molasse consisting mostly of conglomerate sandstones and conglomerates (Greber *et al.* 1994).

At ENB, the maximum resolved depth is around 103 m. The upper 3.6 m consist of distal deltaic sediments, followed by 50.7 m of proximal deltaic sediments. To identify the deeper lithological units in the sediment profile, the result for this site can be compared with the result of an onshore ambient vibration survey carried out at Buochs (Fig. 1 a; Hobiger *et al.* 2017, 2021). Both sites should have the same material source for the proximal deltaic and subjacent sediments. Fig. 13 shows the MAP  $v_s$  profile for array ENB and the best six profiles obtained for different parametrizations of the inversion onshore with the  $v_s$  of the upper 20 m measured using a Seismic Cone Penetration Testing. Below 35 m, the  $S$ -wave velocities for both onshore and offshore sites are quite similar, allowing us to use the  $v_s$  profile obtained for Buochs for site ENB below 103 m depth.

Therefore, the bedrock for ENB may be below 300 m, but the depth resolution of the onshore inversions is not constraining the bedrock depth reliably. The material between 103 m depth and the bedrock can be provisionally interpreted as a succession of deltaic deposits and/or glacial deposits and/or weathered marl/limestones from the Helvetic nappes.

At ENA, we resolved 4.6 m of lacustrine sediments, followed by 10.1 m of glaciolacustrine sediments and at least 130 m of glacial deposits. There is no previously available information about the subsurface structure at this site. Based on the comparison of the  $v_s$  profile at ENA with the ones at WED and NAA, ENA, which was defined as a hemipelagic slope before the analysis, is actually closer to the moraine ridge slope and probably can be considered as an intermediate case between the hemipelagic and moraine ridge slope.

At MUA, the upper 3.4 m of the subsurface consist of distal deltaic sediments, which are followed by at least 125 m of proximal deltaic sediments. Although there is also no previously available data about the structure of the subsurface at MUA, we expect this

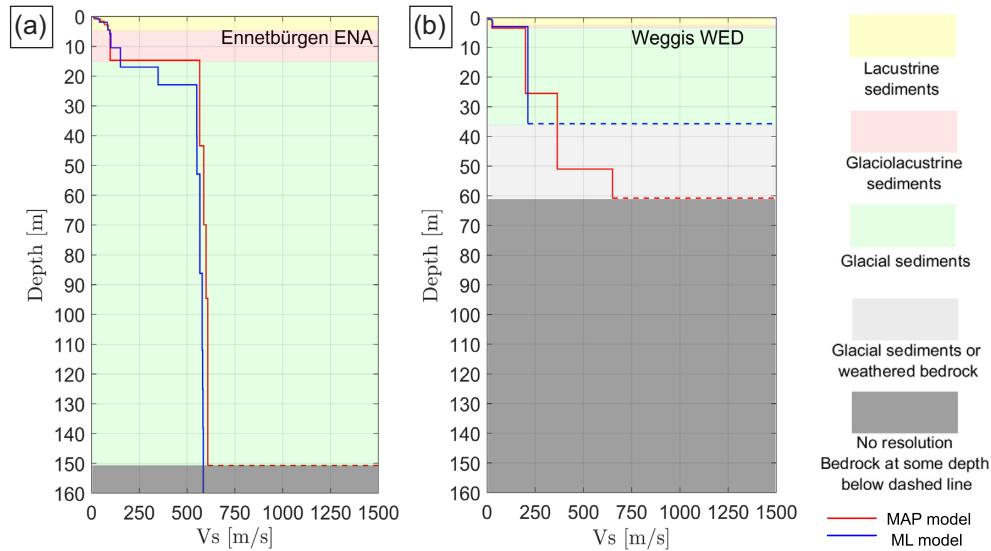


**Figure 12.** MAP and ML  $S$ -wave velocity profiles (red and blue curves, respectively) with assigned lithological units for (a) Ennetbürgen ENA and (b) Weggis WED.

**Table 2.** Range of  $S$ -wave velocity values for different lithological units.

| Lithological unit                                     | $v_s$ range [m/s] |
|-------------------------------------------------------|-------------------|
| Lacustrine sediments (incl. Distal deltaic sediments) | <100              |
| Glaciolacustrine sediments                            | 50-250            |
| Glacial sediments, moraine                            | 200-1000          |
| Proximal deltaic sediments                            | 100-250           |

**Table 3.** Thickness of the lithological units observed at our study sites.

| Lithological unit                                           | Thickness [m]  |                    |                    |              |             |               |
|-------------------------------------------------------------|----------------|--------------------|--------------------|--------------|-------------|---------------|
|                                                             | Chindli<br>CIA | Ennetbürgen<br>ENA | Ennetbürgen<br>ENB | Muota<br>MUA | Nase<br>NAA | Weggis<br>WED |
| Lacustrine sediments<br>(incl. Distal deltaic<br>sediments) | 2.2            | 4.6                | 3.6                | 3.4          | 4.3         | 3.5           |
| Glaciolacustrine<br>sediments                               | 9.8            | 10.1               | n/a                | n/a          | 10.4        | ?             |
| Glacial sediments,<br>moraine                               | $\geq 18$      | $\geq 130$         | $\geq 48.7$        | ?            | $\geq 110$  | $\geq 30$     |
| Proximal deltaic<br>sediments                               | n/a            | n/a                | 50.7               | $\geq 125$   | n/a         | n/a           |

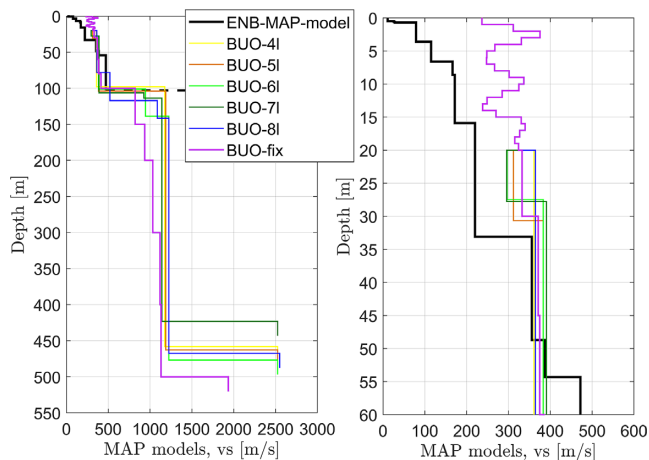


**Figure 13.** Comparison of the MAP S-wave velocity profile for array ENB (bold black line) and the best ML models for Buochs (Hobiger *et al.* 2017, left) and zoom to the upper 60 m (right). The black dashed line on the left figure represents the maximum resolved depth at ENB.

site to be similar to the other deltaic site, ENB. Thus, we assume that the proximal deltaic deposits at MUA should be followed by the glacial sediments, but we cannot clarify this due to the limited data resolution of the retrieved  $v_s$  profile.

To summarize, the presented S-wave velocity profiles and interpreted sediment units fit the previously available coring and drilling data (Freimoser & Gautschi 1991; Greber *et al.* 1994; Strasser *et al.* 2007; Strupler 2012) and provide new information about the deeper part of the subsurface structure (in some cases down to 150 m). Especially the latter is of high importance for the deltaic sites because this work represents the first successful attempt to extract quantitative information on the sediment thickness of the deltaic settings of Lake Lucerne. Moreover, the obtained data match the expectations from the  $H/V$  analysis: for example,  $f_0$  for array MUA is lower than for array ENB, indicating a thicker package of the sediments, while the amplitude of the peak at  $f_0$  for MUA is higher than for ENB, which means a stronger velocity contrast. In the resulting profiles for these arrays, the proximal deltaic deposits at MUA are thicker than at ENB, while their S-wave velocity at MUA is lower than at ENB, clearly explaining the differences in  $f_0$  and amplitude of the  $H/V$  peak between these two arrays. With further investigation, we will utilize the obtained information about the subsurface structure and sediment properties as input data for the analysis and modelling of seismic response and stability of subaquatic slopes in the lake environment. The  $v_s$  profiles already show that the shallowest part of the subsurface has very low  $v_s$  and strong velocity contrasts are present at different depth levels. This information points towards the possibly relatively high (nonlinear) seismic amplification (Celebi *et al.* 1987; Choi & Stewart 2005; Gautam & Chamlagain 2016) and subsequent earthquake-triggered mass movements at the locations investigated.

## 5 CONCLUSION

In this work, we present the results of the shallow water site characterization surveys in Lake Lucerne for potentially unstable subaquatic slopes. We deployed nine Ocean Bottom Seismometers at different locations in the lake and, in total, performed six successful array measurements with the subsequent application of single-station and array data processing techniques. The single-station

data allowed us to derive the fundamental resonance frequencies and Scholte wave ellipticity curves, and to qualitatively assess the variations in the subsurface structure below the array. Before processing the array data, we established a workflow to address the challenges related to the imprecise OBS location, unknown seismometer orientation and shifted data time stamps. This workflow involved carrying out additional field experiments and development of data pre-processing tools. To estimate the dispersion curves of surface waves, we applied three different processing techniques: 3C-HRFG, WaveDec and SPAC. Each of them allowed us to resolve the dispersion curves of Scholte and Love waves over relatively short frequency bands at the locations investigated. Due to the lateral variability of the thickness of the sedimentary cover, the picking of dispersion curves was not straightforward for most of the sites, but the combination of all techniques improved the resolution and extent of the dispersion curves. The most robust results in terms of dispersion curve delineation and continuity for all arrays were obtained from the 3C-HRFG method. The unclear mode attribution of the estimated dispersion curves was addressed by using the mode attribution analysis before the actual inversion. The single-station data analysis and previously available data sets (e.g. reflection seismic profiles) were also taken into account during the array data processing and some stations were excluded from the analysis. After these steps, we performed a joint inversion of the dispersion and ellipticity curves to infer the S-wave velocity profiles at all examined sites. The target ellipticity curve was selected as the ellipticity curve from the most central station of the array. For the inversion, we used the newly developed codes for the transdimensional Bayesian inversion which allowed us to obtain also representative models of the subsurface and an uncertainty estimate (by means of the posterior PDF).

For each of the obtained  $v_s$  profiles, we provided a geological interpretation to differentiate between the available lithological units and compare the depth profile at different geomorphological types of the slopes, i.e. lateral hemipelagic slope, deltaic slope and moraine ridge slope. A comparison of the newly acquired data with previously available information showed a good match and allowed a better understanding of the stratigraphy of the investigated sites. Moreover, our results showed a possibility to use non-destructive seismic methods to resolve the structure of the subaquatic slopes. This is especially beneficial at sites where common imaging methods (e.g. reflection seismic) are not working, as for example experienced for the deltaic slopes at MUA and ENB. Our data represents the first information on the thickness of the sediment cover at these sites. The other alternative to image the subsurface at such locations could be drilling, but it is neither cheaper nor less complicated than the ambient vibration survey. Overall, the experiments we performed using passive seismic techniques allowed us to resolve the general structure of the slopes, in some cases down to more than 150 m depth. The shortcoming of our results is a limited resolution in the very shallow and very deep parts of the subsurface profile, caused by the limited frequency extent of the experimental data. Consequently, we were not able to resolve the bedrock depth or the very thin sediment layers which compose the shallowest part of the soil profile.

Our investigations were carried out at the potentially unstable subaquatic slopes and the obtained results will contribute to the assessment of their seismic response and tsunamigenic potential. In particular, the resolved lithological and S-wave velocity profiles provide an important information related to the subsurface structure and properties of the failure-prone sediments. Additionally,

for each of the investigated slopes we observed very low (below 50–100 m s<sup>-1</sup>) S-wave velocities in the near-surface part that indicates a high probability of significant seismic amplification in these sediments.

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## DATA AVAILABILITY

Data available on request from the authors. OBS data will be made publicly available after the end of the project in 2022.

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