

Charles University

Faculty of Science

Study programme: Applied geology



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**SURFACE WAVE ANALYSIS FOR LOCAL CONDITIONS
CHARACTERIZATION**

STANOVENÍ LOKÁLNÍCH PODMÍNEK ANALÝZOU POVRCHOVÝCH VLN

Doctoral thesis

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Prague, 2025

DECLARATION OF AUTHORSHIP

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Prague, 05.05.2025

Signature

ACKNOWLEDGMENTS

I would like to warmly thank all the people who have contributed to this dissertation, either directly or indirectly, through their technical, manual, or emotional support. Their contributions have been invaluable.

First and foremost, I would like to thank my supervisor, Jan Valenta, for his expert yet indeed friendly guidance throughout my PhD journey. His support and insights have been instrumental in shaping this work. I am also grateful to my consultant, Jiří Málek, for his inspiring ideas and continuous encouragement. A special thanks goes to Giancarlo Dal Moro, whose willingness to share his profound knowledge of surface-wave methods and for providing access to his WinMASW software.

At the Faculty of Science, Charles University, I would like to particularly thank Tomáš Fischer, Iryna Viktosenko, and especially Josef Vlček, for his endless helpfulness. I also appreciate the collaboration and support from my colleagues at the Institute of Rock Structure and Mechanics of the Czech Academy of Sciences, particularly within the Seismotectonics Department.

Finally, I owe my deepest gratitude to my family. I thank my parents, who have always encouraged my education and personal growth. Most importantly, I am profoundly grateful to my wife, Anna, for her understanding and support, particularly during the challenging early years of family life with young children, allowing me the time and space to pursue this research.

Thank you all.

Felix qui potuit rerum cognoscere causas

Fortunate is he, who is able to know the causes of things

PUBLIUS VERGILIUS MARO

ABSTRACT

The presented doctoral thesis focuses on the characterization of local site conditions within the framework of seismic hazard assessment, with a particular emphasis on surface-wave dispersion analysis and the proxy parameter V_{s30} , defined as the average shear-wave velocity of the top 30 meters of the subsurface. The study integrates theoretical insights, methodological advancements, and applied research across three main areas: (1) the physical understanding and quantification of site effects, including their role in ground motion amplification; (2) advanced surface-wave analysis techniques for near-surface shear-wave velocity profiling; and (3) the development and application of surface wave-based strategies for V_{s30} estimation and interpretation.

An extensive literature review forms the foundation of this thesis, providing a detailed exploration of the physical principles underlying local site amplification, with a particular emphasis on shear-wave velocity and related proxy parameters. Special attention is given to V_{s30} , including its methodological challenges and its debated role in engineering seismology. The thesis contributes to the ongoing international discussion regarding the suitability of V_{s30} as a standalone site-effect parameter by analyzing a unique dataset from the West Bohemia seismic network (WEBNET). In this study, V_{s30} values derived from surface-wave measurements were correlated with recorded ground motion amplification, providing new empirical evidence for its predictive capabilities. This analysis resulted in the definition of a new empirical relationship between V_{s30} and amplification, further advancing our understanding of the factors controlling site response.

Surface-wave analysis is presented through an extensive, practically oriented review, accenting both fundamental principles and real-world applications. The thesis provides a comprehensive overview of critical aspects of dispersion analysis, addressing practical challenges such as mode misidentification, resolution limitations, and the trade-offs between data quality and computational effort. In addition to illustrating basic surface-wave techniques, the work includes progressively complex case studies. The third case study, in particular, focuses on a challenging industrial site with significant ambient noise and complex subsurface conditions, demonstrating the potential for advanced joint analysis to overcome site-specific obstacles and achieve robust shear-wave velocity models. This approach highlights the critical importance of integrating multiple datasets and analysis techniques for accurate subsurface characterization in complex environments.

The final part of the thesis integrates the two main themes, demonstrating how surface-wave analysis can be effectively used for local site characterization within seismic hazard assessment. This section includes practical applications such as the site characterization of seismic stations within the WEBNET network, the creation of a V_{s30} and seismic hazard map

for Czechia, and the development of a novel rapid Vs30 estimation approach (RAVE). This newly introduced method addresses key challenges in rapid site characterization by providing a computationally efficient alternative to traditional inversion-based approaches, making it particularly valuable for large-scale surveys and preliminary hazard assessments. These studies collectively highlight the versatility of surface-wave methods for evaluating site conditions, while also contributing to the ongoing global discussion on site-effect proxies. By combining theoretical insights with practical applications, this work provides a comprehensive framework for leveraging surface-wave data to improve the accuracy and reliability of site response assessments.

ABSTRAKT

Předložená disertační práce se zaměřuje na charakterizaci lokálních podmínek v rámci hodnocení seismického hazardu, s důrazem na disperzní analýzu povrchových vln a proxy parametr Vs_{30} , který je definován jako průměrná rychlost šíření střížných vln v horních 30 metrech podloží. Práce integruje teoretické poznatky, metodologický pokrok a aplikovaný výzkum ve třech hlavních oblastech: (1) fyzikální pochopení a kvantifikace lokálních účinků zemětřesení (tzv. site effects), včetně jejich role v zesílení procházejících seismických vln; (2) pokročilé techniky analýzy povrchových vln pro profilování rychlostí střížných vln v mělkých částech podloží; a (3) vývoj a aplikace metod založených na povrchových vlnách pro odhad a interpretaci parametru Vs_{30} .

Obsáhlá literární rešerše tvoří základ práce a podrobně se věnuje fyzikálním principům zesilování lokálních podmínek, přičemž klade důraz na rychlost střížných vln a související proxy parametry. Zvláštní pozornost je věnována Vs_{30} , včetně jeho metodologických výzev a diskutované role v inženýrské seismologii. Do probíhající mezinárodní diskuze o vhodnosti Vs_{30} jako samostatného parametru pro charakterizaci lokálních podmínek přispívá analýzou unikátního datasetu ze seismické sítě WEBNET v západních Čechách, kde byly hodnoty Vs_{30} , odvozené z povrchových vln, korelovány se zaznamenaným zesílením zemětřesných pohybů, což poskytlo nové empirické důkazy pro jeho prediktivní schopnosti. Výsledkem analýzy byla v této studii definována nová empirická závislost mezi Vs_{30} a zesílením.

Analýza povrchových vln je prezentována prostřednictvím rozsáhlého, prakticky orientovaného přehledu, který zdůrazňuje jak základní principy, tak jejich aplikace. Práce poskytuje komplexní přehled klíčových aspektů disperzní analýzy, včetně praktických výzev, jako je identifikace módů, omezení rozlišení a kompromisy mezi kvalitou dat a výpočetní náročností. Vedle základních technik povrchových vln zahrnuje práce i případy postupně narůstající složitosti. Zejména třetí případová studie se zaměřuje na náročné průmyslové prostředí s významným antropogenním šumem a komplexními podložními podmínkami, čímž demonstruje potenciál pokročilých sdružených (joint) analýz k překonání specifických překážek a vytvoření robustních modelů rychlostí střížných vln. Tento přístup zdůrazňuje kritickou důležitost integrace více datasetů a analytických technik pro přesnou charakterizaci podloží v komplexních prostředích.

Závěrečná část práce propojuje obě hlavní témata a ukazuje, jak může být analýza povrchových vln efektivně využita pro lokální charakterizaci podmínek v rámci hodnocení seismického hazardu. Tato sekce zahrnuje praktické aplikace, jako je charakterizace seismických stanic sítě WEBNET, tvorba mapy Vs_{30} a seismického hazardu pro Českou

republiku a vývoj nové metody rychlého odhadu V_{s30} (RAVE). Tato nově představená metoda řeší klíčové výzvy charakterizace lokálních podmínek tím, že poskytuje výpočetně efektivní alternativu ke klasickým přístupům založeným na inverzi, což ji činí zvláště cennou pro rozsáhlé průzkumy a předběžné odhady hazardu. Tyto studie společně zdůrazňují všestrannost metod povrchových vln pro hodnocení podmínek podloží a současně přispívají k probíhající globální diskuzi o proxy parametrech pro kvantifikaci site effects. Spojením teoretických poznatků s praktickými aplikacemi tato práce poskytuje komplexní rámec pro využití dat z povrchových vln ke zlepšení přesnosti a spolehlivosti hodnocení lokálních podmínek.

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1 INTRODUCTION

1.1 Seismic Hazard Assessment

Earthquakes are among the most devastating natural phenomena, capable of causing widespread destruction, economic loss, and account for 60% of worldwide casualties associated with natural disasters (Giardini et al., 2003). Their unpredictable nature underscores the critical need for a comprehensive understanding of their causes, mechanisms, and impacts. Seismology has advanced significantly in understanding the processes that lead to earthquakes and in the last fifty years, many attempts have been made worldwide to achieve reliable, short-term earthquake prediction. Despite these efforts, no operational methodology exists today to predict major earthquakes ($M > 6$) days or hours in advance (Pulinets and Ouzounov, 2018). This inherent unpredictability has led to the development of seismic hazard assessment (SHA), which is a process that aim to quantify the likelihood of various levels of ground shaking over a given time frame. Such assessments are vital for informing urban planning, infrastructure design, and emergency preparedness, helping communities mitigate the risks posed by these sudden earthquake events.

Two principal approaches are commonly employed in seismic hazard assessment: (a) probabilistic seismic hazard analysis (PSHA) and (b) deterministic seismic hazard analysis (DSHA). Both methods rely on the same geological and seismological datasets, yet they fundamentally differ in how they define and evaluate seismic hazard. The key distinction lies in their treatment of uncertainty regarding the occurrence of future earthquake events. In PSHA, seismic hazard is quantified as the level of ground motion at or above a specified intensity associated with a specific annual probability of exceedance. This approach incorporates statistical relationships that account for all potential earthquakes, considering their probabilities of occurrence in terms of time, location, and magnitude. In contrast, DSHA estimates the maximum possible ground motion from a single earthquake or a set of events (e.g., "maximum credible earthquakes") without regard to the likelihood of their occurrence. The hazard is then assessed using deterministic calculations and simplified statistical representations of earthquake sources and ground motion (Wang, 2011; Gerstenberger et al., 2020).

Both deterministic and probabilistic approaches can yield estimates of expected ground shaking. For example, deterministic seismic hazard analysis (DSHA) may provide the 84th percentile ground motion for a specified fault rupture scenario, while probabilistic seismic hazard analysis (PSHA) allows the estimation of mean, median, or other statistical percentiles of ground motion associated with a defined probability of exceedance (PoE), such as 10% in

50 years. These methodologies can be applied independently or in a complementary fashion to support robust decision-making in seismic design and risk assessment. In contemporary practice, PSHA is more widely implemented and forms the foundation of most national and international seismic hazard models (Gerstenberger et al., 2020).

Despite ongoing debates regarding its suitability (Wang, 2011; Mulargia et al., 2017), PSHA remains the dominant framework in seismic hazard assessment and is employed in the majority of ground motion models used in seismic hazard assessments worldwide (Málek and Vackář, 2019; Ademović et al., 2021; Vackář et al., 2024).

The probabilistic seismic hazard analysis (PSHA) framework involves a sequence of interrelated steps, as outlined in Figure 1. The process begins with the compilation and evaluation of earthquake data, including historical and instrumental catalogs, to assess seismicity parameters such as recurrence rates and maximum magnitudes. This is followed by the definition of seismic source zones based on geological and tectonic features, and the selection of appropriate ground motion prediction equations (GMMs) derived from recorded strong-motion data.

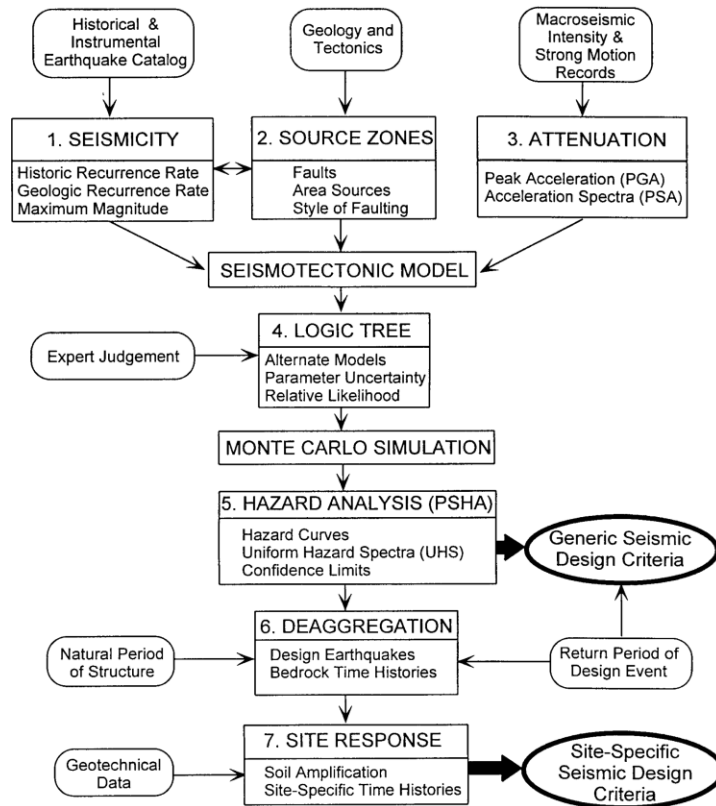


Figure 1: Flowchart illustrating the components of the probabilistic seismic hazard methodology within the framework of a seismic design criteria approach. Adapted from Thenhaus and Campbell (2003).

A logic tree framework is employed to account for model and parameter uncertainties. The seismotectonic model is then used in Monte Carlo simulations to compute hazard curves and uniform hazard spectra (UHS). Subsequent deaggregation identifies the contributions of

specific earthquake scenarios, which are then combined with site-specific geotechnical data to evaluate local soil response and amplification effects based on local geology conditions. The resulting outputs inform both generic and site-specific seismic design criteria (Thenhaus and Campbell, 2003; Vackář et al., 2024). PSHA process can be summarized as the solution of the following expression of the total probability theorem:

$$\lambda[X \geq x] \approx \sum_{\text{Sources } i} v_i \int_{M_0}^{M_{Max}} \int_{R|M} P[X \geq x | M, R] f_M(m) f_{M|R}(r | m) dr dm \quad (1.1)$$

where $\lambda[X \geq x]$ is the annual frequency at which a specified ground motion level x is exceeded at a site. In this formulation, v_i denotes the annual occurrence rate of earthquakes on seismic source i , $f_M(m)$ is probability density function of magnitude, $f_{M|R}(r | m)$ is conditional probability density function of distance given magnitude and $P[X \geq x | M, R]$ is probability that ground motion exceeds x , given magnitude M and distance R .

An essential step in the seismic hazard assessment workflow is the characterization of local site conditions, as these strongly influence ground motion at the surface. Even if the regional hazard model is accurately defined, neglecting site-specific properties can lead to under- or overestimation of seismic demand on structures. The evaluation of the characteristics of the strongest expected ground shaking for specific areas of the site under study is a fundamental tool for high-priority actions aimed at mitigating seismic risk (Caserta, 2011). In this context, the quantification of site effects represents a key component of seismic hazard modeling, particularly for engineering and microzonation applications. This thesis focuses on the evaluation of local effects using surface-wave analysis.

1.2 Objectives and Structure of the Thesis

The primary focus of this thesis is the evaluation of local site conditions in the context of seismic hazard assessment, with a particular emphasis on the application of surface-wave dispersion analysis and the widely used proxy parameter V_{s30} , the time-averaged shear-wave velocity of the top 30 meters of the subsurface. The work is motivated by the significant role that site effects play in modifying ground motion and thereby influencing the seismic risk at a given location. Despite significant advances in seismic hazard modeling, the accurate quantification of site response remains a major challenge, particularly in regions with complex geology or limited subsurface data. This thesis addresses this challenge by combining theoretical analysis, methodological development, and practical application of surface-wave techniques for the robust estimation of local seismic site parameters.

The thesis builds upon two interconnected fields: (1) the estimation and interpretation of site effects, and (2) the surface-wave methods used for near-surface shear-wave velocity profiling. These two domains are systematically integrated to demonstrate how surface-wave analysis can support reliable site characterization and improve regional and site-specific seismic hazard assessments.

The main objectives of the thesis are as follows:

- To clarify the physical basis and engineering relevance of site effects and their quantification in seismic hazard modelling.
- To review and evaluate existing methodologies for surface-wave analysis, with an emphasis on dispersion-based techniques used to estimate shear-wave velocity profiles.
- To apply and compare different surface-wave methodologies in real case studies, ranging from basic single-component analysis to advanced multicomponent joint inversions.
- To improve the understanding, estimation, and application of the V_{s30} parameter in seismic site characterization and hazard assessment, through the evaluation of its effectiveness, the development of a new estimation framework, and its integration into regional-scale models.

To fulfil these objectives, the thesis is structured into a series of chapters that guide the reader from theoretical foundations to applied examples and innovative methodologies.

Chapter 1 provides a general overview of seismic hazard assessment, emphasizing the importance of local site conditions.

Chapter 2 addresses the characterization of local site conditions, defining the phenomenon of site amplification and summarizing its physical origins, with special attention given to the role of shear-wave velocity. Key mechanisms are introduced and discussed in relation to how they affect seismic wave propagation and ground motion amplification. The chapter systematically presents the principal approaches used to evaluate site effects, highlighting their strengths, limitations, and appropriate applications within seismic hazard assessment. Various site-effect proxies are then introduced, with a particular emphasis on those based on shear-wave velocity. A detailed discussion is devoted to the parameter V_{s30} , covering its physical basis, measurement strategies, inclusion in building codes, and wide use in ground motion models. The chapter also critically examines and develops the ongoing international debate regarding the adequacy of V_{s30} as a standalone predictor of site response, referencing numerous global and regional studies that evaluate its limitations and propose complementary proxies. A substantial portion of this work contributes directly to this discussion by analyzing the relationship between V_{s30} and measured amplification at seismic stations of the West Bohemia Seismic Network (WEBNET) network based on Mazanec et al. (2024).

Chapter 3 presents a comprehensive theoretical and methodological overview of surface-wave analysis, including the physics of Rayleigh and Love wave propagation, dispersion theory, and the principles of velocity inversion. A thorough methodological review is provided, highlighting various processing workflows, inversion techniques, and joint analysis strategies. This chapter is intended as a reference resource for students and prospective users of surface-wave methods, offering a structured overview of the diverse literature related to their application in seismic site characterization. In addition to the theoretical framework, the chapter includes a discussion of practical challenges and methodological limitations, supported by real data examples that illustrate key processing steps and their sensitivity. The chapter concludes with a graduated set of case studies demonstrating increasing levels of methodological complexity. These range from basic examples aimed at introducing surface-wave integration into conventional refraction workflows, based on Mazanec and Valenta (2023) to an advanced multicomponent joint analysis from Dal Moro and Mazanec (2024), which seeks to identify the most robust analysis strategy in complex urban and industrial environments.

Chapter 4 integrates the two thematic pillars - site characterization and surface-wave analysis, through three studies:

- Section 4.1 focuses on the use of surface-wave dispersion analysis to derive shear-wave velocity profiles and characterize local site conditions of WEBNET seismic network based Mazanec et al. (2024).
- Section 4.2 contributes to regional seismic hazard assessment by calibrating and validating a topography-based Vs30 proxy model, integrating it with in-situ measurements, and implementing the results within an interactive hazard mapping platform based on Vackář et al. (2024, under review).
- Section 4.3 presents and evaluates a novel framework for rapid Vs30 estimation (RAVE), offering an efficient alternative to conventional inversion methods by exploiting statistical patterns in dispersion data based on Mazanec and Dal Moro (2025, submitted).

The thesis concludes in Chapter 5 with a summary of findings and a reflection on the implications of the work.

Chapter 6 summarizes potential future directions arising from the presented findings.

2 CHARACTERIZATION OF LOCAL SITE CONDITIONS

The actual ground shaking amplitude during an earthquake is not only controlled by the earthquake's size and distance, but also by the local site conditions. This local influence—known as site effects—can significantly modify earthquake ground motions, primarily due to the amplification caused by soft soils and unconsolidated sediments overlying stiff bedrock and has been recognized as a major cause of structural collapse (Borcherdt, 1970; Borcherdt and Gibbs, 1976; Esteva, 1988; Chávez-García and Bard, 1994; Boore, 2004; Stambouli et al., 2017). This phenomenon is commonly referred to as site response or local site amplification, describing how local ground conditions influence incoming seismic waves by altering their amplitude and frequency content. Borcherdt (1970) first documented that local San Francisco Bay (California, US) area geology caused stronger shaking on soft sediment than on bedrock during earthquakes. Even moderate earthquakes can result in markedly different levels of ground shaking at nearby locations, depending on local geology. A famous example is the 1985 Mw 8.0 Michoacan (Mexico) earthquake (Figure 2), when Mexico City, despite its considerable distance from the epicentre (approx. 350 km), experienced strong amplification of ground motion due to the presence of soft soils. Large parts of Mexico City are built on a former lakebed composed of soft, water-saturated tens of meters deep clay and volcanic sediments layers (Figure 3). These low-velocity, low-density materials have a strong impedance contrast with the underlying harder strata, leading to extreme amplification of ground shaking (Beck and Hall, 1986; Bard et al., 1988; Ohmachi et al., 1988; Singh et al., 1988; Campillo et al., 1989).

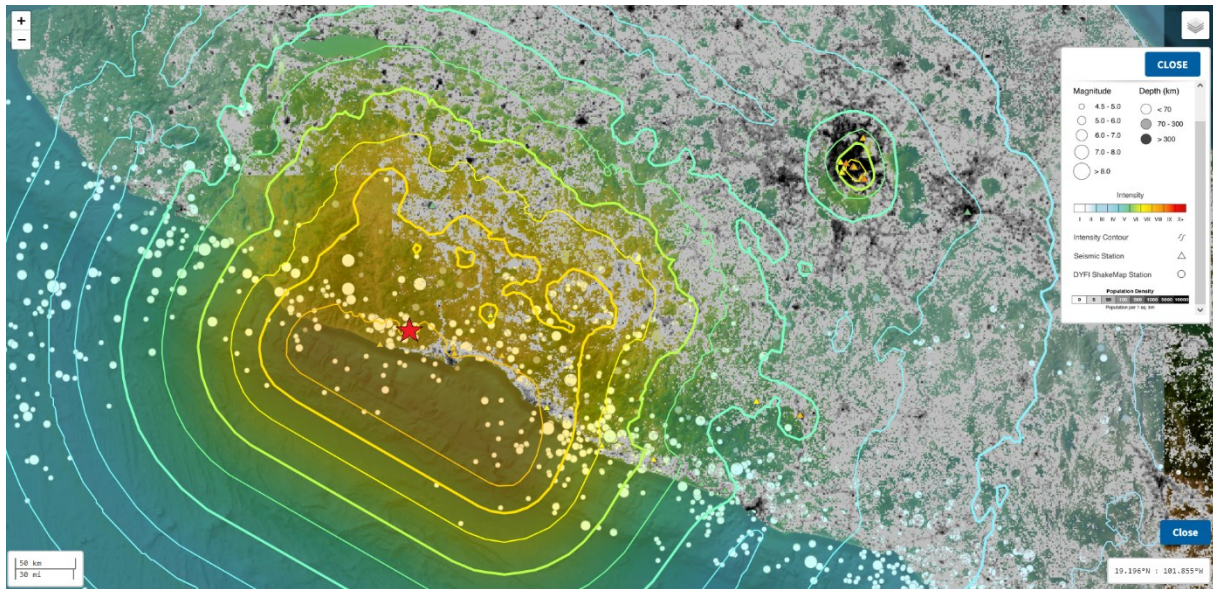


Figure 2: Shake map of the Mw 8.0 Mexico earthquake, 1985 (epicentre marked with a red star). Despite being located approximately 350 km from the epicentre near the Pacific coast, local site conditions in Mexico City caused ground motions to exceed those recorded much closer to the source (see intensity isolines), resulting in widespread destruction and thousands of fatalities. For illustration, seismicity from 1900 to 2015 is plotted. (modified from USGS, 2025).

Such observations were later supported by further studies from significant recent earthquakes, such as the 1989 Loma Prieta, 1994 Northridge, and 1995 Kobe events (Dietz and Ellsworth, 1990; Peek-Asa et al., 1998; Chen et al., 2021). Since then, numerous studies around the world have confirmed that local conditions (soil thickness, stiffness, stratigraphy, and topography) can amplify seismic shaking by several times relative to bedrock sites and therefor site characterization, i.e., the assessment of local soil conditions, has become a fundamental component of both seismic hazard studies and building code provisions (Trifunac, 1990; Ergin et al., 2004; Baise et al., 2016; Firat et al., 2016; Weatherill et al., 2020; Bergamo et al., 2023).

Site characterization involves determining the properties of the shallow subsurface (typically the upper tens of meters) that control site response. Key parameters include the thickness of soil layers, lithology, density, plasticity, and especially the seismic wave velocities. By evaluating these, one can estimate how much a given site will amplify or filter earthquake shaking. In practice, this evaluation might be part of a microzonation study (dividing an area into zones of different seismic response (Lachet et al., 1996; Fäh et al., 1997; Anbazhagan and Sitharam, 2008; Sitharam and Anbazhagan, 2008; Moscatelli et al., 2020), or part of site-specific hazard analysis for an important structure (Irsyam et al., 2007; Rodriguez-Marek et al., 2014; Ebrahimian et al., 2019; Bhusal et al., 2022).

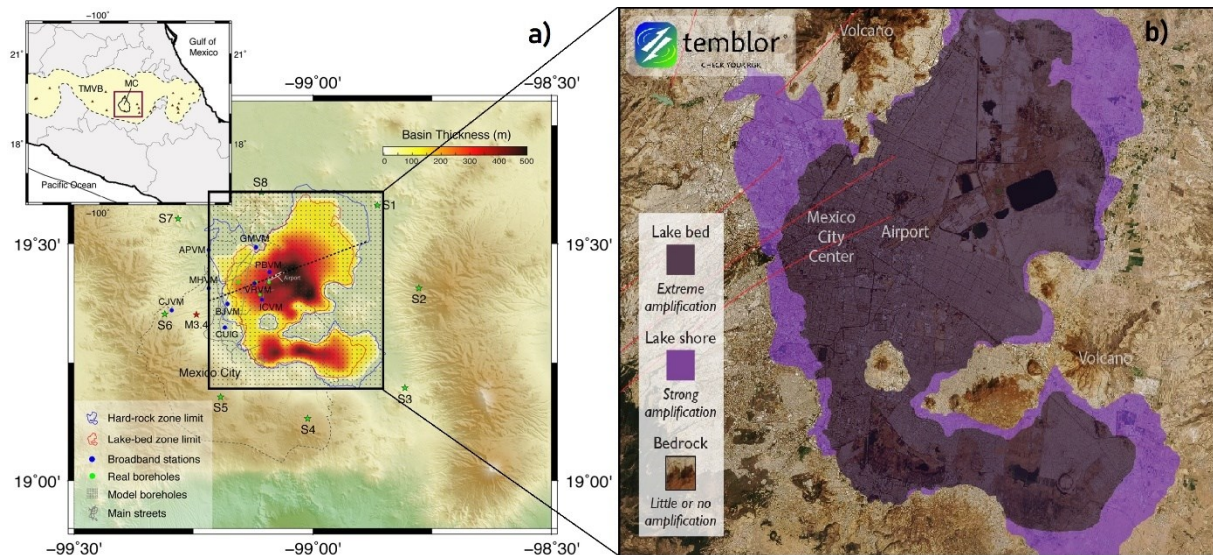


Figure 3: Mexico Cit (MC) and the Valley of Mexico: a) topographic configuration, the color scale represents basin thickness, defined by the depth to the Oligocene volcanic basement of the Trans-Mexican Volcanic Belt. Green stars indicate the locations of vertical body force applications at the surface, while the red star marks the epicentre of the M3.4 earthquake on December 1, 2014; b) Satellite image of MC with simplified lithology (shaded purple) and associated amplification effects. Compilation based on Cruz-Atienza et al. (2016) and Temblor (2017).

2.1 Basic Physical Concepts of Site Effects

Near-surface formations result from long-term processes of erosion, weathering, and sediment deposition, and they play a crucial role in the amplification and spatial variability of ground motion during earthquakes, as illustrated in Figure 4. The term *site effects* refers to the influence of local geological conditions on the modification (especially amplification) of the seismic wavefield at a recording site. Key parameters charactering a site include the shape of surface topography, the dynamic, physical, and mechanical properties of soil and rock materials and geometry of the soil stratigraphy (e.g., thickness and lateral discontinuities) (Wair et al., 2012).

Site effects in a form of amplification may occur under four main scenarios as illustrated in Figure 5. First two principal cases involve (1) a large shear-wave velocity contrast between bedrock and overlying soil (Figure 5a), and (2) resonance amplification at the fundamental site period (Figure 5b). Further details are provided in the text below. Other two amplification situation types are related to topographic features, such as convex (e.g., ridges, hills, mountains) and concave forms (e.g., valleys, basins, canyons).

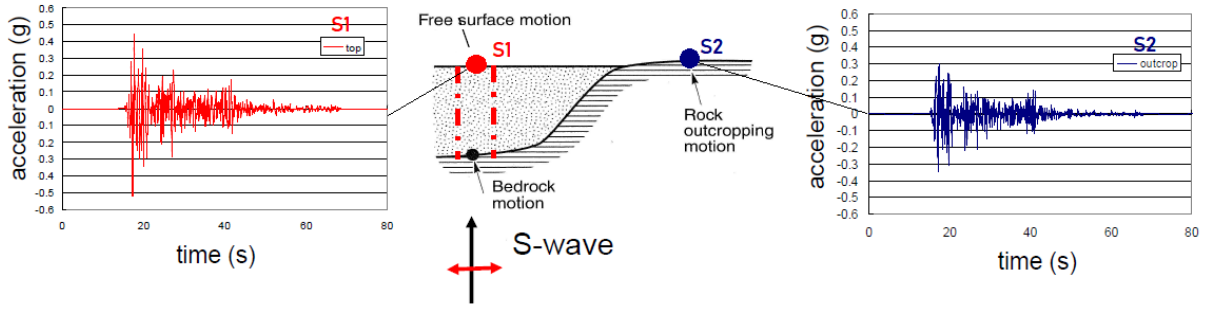


Figure 4: Simplified illustration of site effects. An S-wave propagates upward from deeper layers, with surface station S1 (red) located on a sedimentary basin and station S2 (dark blue) on bedrock, resulting in different ground motion responses. Modified after Foti (2012).

Amplification can also occur in soft soils when thick sedimentary layers fill narrow bedrock depressions (such as sedimentary basin). In such settings, (3) geometric focusing effects (Figure 5c), may develop, in which seismic waves are refracted and reflected in two or three dimensions, leading to focusing of energy (Bard and Bouchon, 1985). Furthermore, (4) seismic waves approaching the edges of buried bedrock valleys can generate surface waves (Figure 5d) that constructively interfere within the basin, leading to unusually high horizontal and vertical ground motion amplitudes (Lomnitz et al., 1999). For convex topographies, amplification tends to be greater at the ridge than at the base, while in concave settings, amplification is often more pronounced along the sides than at the centre (Pitilakis, 2004; Hunter et al., 2010). Burjáněk et al. (2014) analyzed several sites with pronounced topography and concluded that the strong amplification observed at topographically pronounced sites is primarily governed by subsurface structure, not surface geometry. Pure topographic effects may only emerge at hard rock sites where subsurface amplification is minimal.

The dynamic stiffness of geologic materials is strongly dependent on strain level and is directly related to shear-wave velocity (V_s), making V_s a key parameter for assessing site effects. The shear-wave velocity reflects the dynamic mechanical behavior of soils and rocks and is commonly used to estimate the small-strain shear modulus (G_{max}), as expressed by the relation:

$$G_{max} = \rho \cdot V_s^2 \quad (2.1)$$

where ρ denotes the mass density of the soil. This relationship illustrates how variations in V_s reflect differences in stiffness and seismic impedance, which are fundamental to understanding site amplification phenomena (Shearer and Orcutt, 1987; Foti et al., 2014).

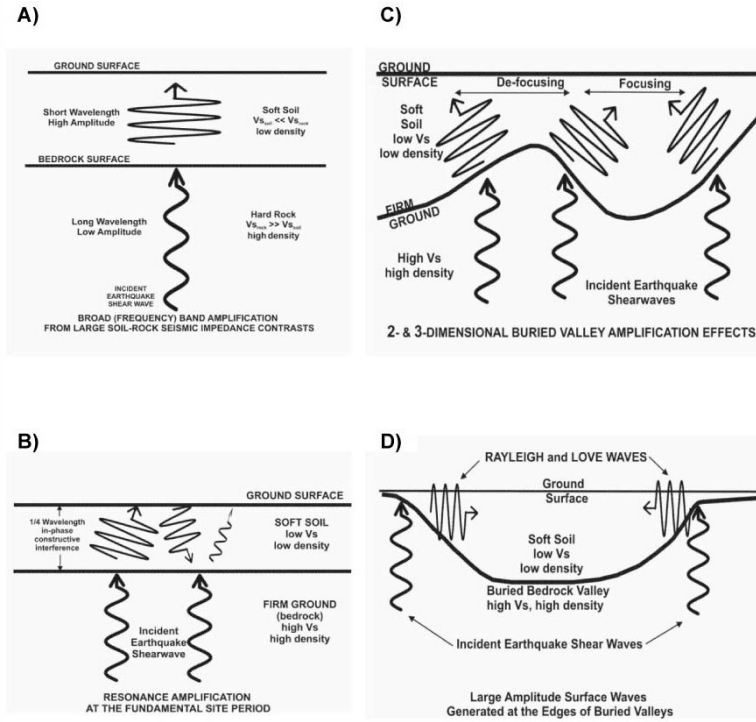


Figure 5: Graphic illustrations of key amplification effects associated with soft soil overlying firm ground or bedrock: a) broad-band amplification, b) resonance amplification, c) 2D & 3D valley focusing effects, d) surface wave generation at the edges of buried valleys (from Hunter et al. 2010).

The amplitude of earthquake ground motion is primarily governed by two key physical properties: seismic impedance, which quantifies the resistance to particle motion, and material damping, which reflects the energy dissipation within the medium (Aki and Richards, 1980).

2.1.1 Impedance contrast effects

Among the site effects, one of the most critical is the amplification of ground motion due to contrasts in subsurface seismic impedance, such as those observed during the 1985 Mexico City earthquake (Seed et al., 1988; Chávez-García, 2007). In the simplest case of horizontally polarized shear waves (SH), seismic impedance (Z_{SH}) is defined as the product of the medium's density (ρ), shear-wave velocity (V_s), and the cosine of the angle of incidence (θ):

$$Z_{SH} = \rho \cdot V_s \cdot \cos(\theta) \quad (2.2)$$

The angle of incidence θ is typically small near the Earth's surface, allowing the cosine term to be approximated as unity. When earthquake seismic waves travel upward from hard bedrock into softer surface layers, the contrast in density and stiffness causes impedance contrast effects. At a boundary where a fast, dense rock underlies a slower, soft soil, part of the seismic energy is reflected downward and part is transmitted upward into the soft layer, with a higher amplitude to satisfy energy conservation in the lower-impedance medium (Figure 6).

The impedance contrast (Z) between two can be expressed in simplified form as:

$$Z = \frac{\rho_2 \cdot V_{s2}}{\rho_1 \cdot V_{s1}} \quad (2.3)$$

where ρ_1 and ρ_2 are the densities of the upper and lower layer, respectively, and V_{s1} and V_{s2} are the corresponding shear-wave velocities. In principle, as shear waves slow down in softer material, their amplitude increase. A large impedance contrast (for example, a ten-fold drop in shear-wave velocity from rock to soil) can lead to significant amplification of ground motion in the softer layer (Baise et al., 2016). As an example, Khareshi Banab et al. (2012) investigated seismic site amplification in Ottawa, Canada, where the subsurface is predominantly composed of soft clay sediments with low V_s ~150 m/s, overlying a high-velocity bedrock ~2700 m/s. This configuration results in an impedance contrast of at least 18.

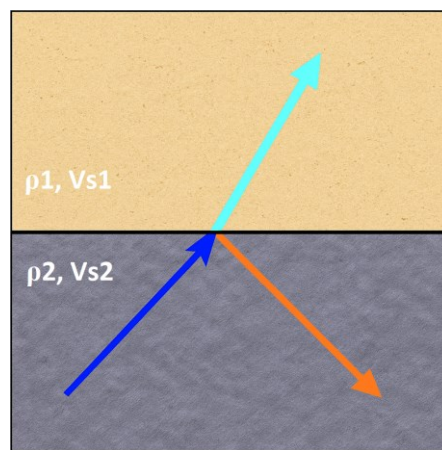


Figure 6: Acoustic impedance contrast illustration for two layers characterized by ρ and V_s , where $\rho_2 > \rho_1$ and $V_{s2} > V_{s1}$. Seismic wave is propagating upwards from bedrock (dark blue line), interacts at the layers boundary resulting in reflected (red) and transmitted wave (light blue) to near-surface soft sediments.

G_{max} and V_s are primarily functions of soil density, void ratio, and effective stress, with secondary influences including soil type, age, depositional environment, cementation, and stress history (Hardin and Drnevich, 1972a, 1972b). Table 1 summarizes the effect of increasing various parameters on V_s and G_{max} .

Table 1: Effect of various factors on G_{max} and V_s . Adapted from Wair et al. (2012).

Increasing Factor/Parameter	Influence on G_{max} and V_s
Confining Pressure or Overburden Stress	Increases ↑
Void Ratio	Decreases ↓
Geologic Age	Increases ↑
Cementation	Increases ↑
Overconsolidation Ratio (OCR)	Increases ↑
Strain Rate or Frequency of Cyclic Loading	Increases ↑

2.1.2 Resonance amplification

In the case of horizontally layered (1D) structures, wave trapping primarily affects vertically propagating body waves within the near-surface layers. However, when the subsurface exhibits lateral variations in sediment thickness, resulting in 2D or 3D structures, surface waves are also influenced as they resonate along the interface between sediment and bedrock. In both scenarios, the effect is most pronounced when the reflected and incident waves become phase-aligned, resulting in constructive interference and resonance. Resonance is a frequency-dependent phenomenon governed by the geometric and mechanical properties of the subsurface, including density, P and S-wave velocities, and material damping. This frequency dependent amplification means that site effects can be particularly dangerous for structures whose natural vibration period coincides with the site's resonant frequency, as occurred in Mexico City (where many multi-story buildings with ~2 s period were severely damaged due to resonance with the 0.5 Hz ground motion). In a horizontally layered (1D) structure, resonance arises from vertical interference of body waves and typically manifests as a single, well-defined amplitude peak (Figure 7). In contrast, 2D and 3D configurations, such as basins or irregular topographies, lead to more complex resonance patterns, as they involve both body and surface waves (Pitilakis, 2004; Kramer and Stewart, 2024).

A simple qualitative and quantitative assessment of site effects is often expressed in terms of the maximum amplification factor (A_{max}) and the resonant frequencies (f_0), including both fundamental and higher modes of propagation. The fundamental resonant frequency in a horizontally layered media can be approximated with the knowledge of Vs and thickness of the sedimentary layer (H):

$$f_0 = \frac{V_s}{4H} \quad (2.4)$$

f_0 usually ranges from as low as 0.2 Hz in the case of very thick or soft deposits, up to 10 Hz or more for shallow, stiff layers or weathered rock. The amplitude of the fundamental peak is primarily influenced by the impedance contrast between surface sediments and underlying bedrock, as well as the material damping (D) (Park and Hashash, 2004a). The characteristics of the incoming wavefield (e.g., wave type, incidence angle, and distance to the source) play a secondary role. For the simplest 1D case discussed above, the maximum amplification (A_{max}) at the fundamental frequency can be approximated by the following formula:

$$A_{max} = \frac{2}{\frac{1}{Z} + \frac{\pi \cdot D_1}{2}} \quad (2.5)$$

where D_1 is the sediments damping. In case of very small damping ($\leq 2-3\%$), the maximum amplification can approximate double the impedance contrast. A high damping layer ($\geq 10\%$) may also attenuate the incoming motion, reducing amplification and potentially resulting in

de-amplification, where the ground surface motion is less intense than the input motion (Pitilakis, 2004; Chávez-García, 2007; Kramer and Stewart, 2024).

Figure 7 illustrates the resonance effect in the frequency domain, focusing on a typical case of a soft sediments layer overlying hard rock with an impedance contrast of $Z = 5$. A 100-meter-thick layer produces amplification peaks around 0.5, 1.5, 2.5 Hz and higher, whereas a 50-meter-thick layer shifts the peaks to approximately 1.0 and 3.0 Hz. Amplitude of these peaks decreases with increasing frequency due to inelastic attenuation or damping (Pitilakis, 2004).

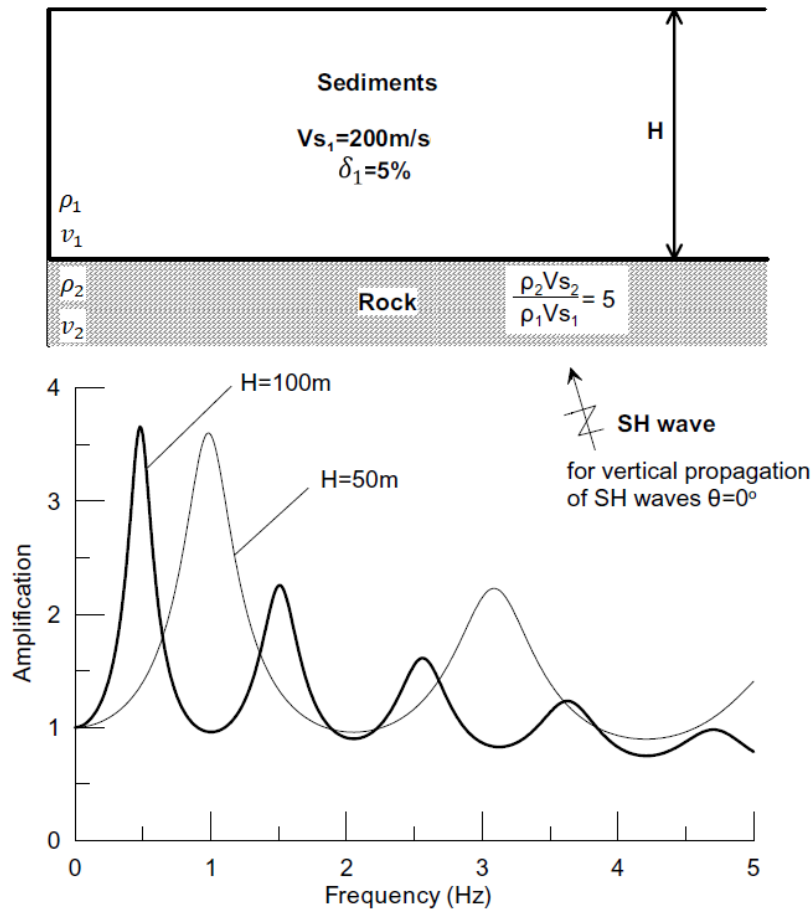


Figure 7: Simplified 1D model (upper part) illustrating site amplification (lower part) for sedimentary layers of varying thicknesses 50 and 100 m. Modified after Pitilakis (2004).

2.1.3 Non-linear Soil Behaviour and Attenuation

While amplification effects due to impedance contrast, resonance, and topographic focusing are typically analyzed under the assumption of linear site response, it is important to recognize that such simplification may not hold under strong ground motions. In reality, near-surface soils often exhibit non-linear behavior (usually at large strain during strong motion), where both shear stiffness and damping properties vary with strain amplitude. This non-linear site response can significantly alter the amplitude, frequency content, and duration of seismic shaking, particularly in soft soil deposits (Yoshida and Iai, 1998; Arslan and Siyahi, 2006). One

notable consequence of this behavior is the limitation of amplification during peak ground motion - a phenomenon observed in some strong motion recordings. For example, during the 1994 Northridge earthquake, deep soft soils exhibited less high-frequency amplification than expected, largely due to non-linear damping effects (Peek-Asa et al., 1998). Therefore, a comprehensive understanding of site effects must also account for these strain-dependent changes in dynamic soil properties (Parolai, 2012).

Importantly, non-linear behavior also affects seismic attenuation, as represented by the quality factor (Q_A), see Figure 8, which describes the rate of energy dissipation in the medium (Kramer and Stewart, 2024). In the context of soil dynamics and ground motion, site attenuation is understood to counteract amplification effects. Very soft or saturated soils can absorb high-frequency energy (due to material damping and non-linear behaviour), acting as a filter. The quality factor is defined as:

$$Q_A = 2\pi \frac{E^{max}}{\Delta E^{diss}} \quad (2.6)$$

where ΔE^{diss} is the ratio of energy dissipated in the geomaterial during one cycle of harmonic loading and E^{max} is the maximum stored energy per unit volume of geomaterial during the same cycle of loading. Under strong shaking, the effective Q_A typically decreases, reflecting increased damping and energy loss. Using the Q_A material damping ratio D_A can be estimated by the relation (Parolai et al., 2022):

$$D_A = \frac{1}{2Q_A} \quad (2.7)$$

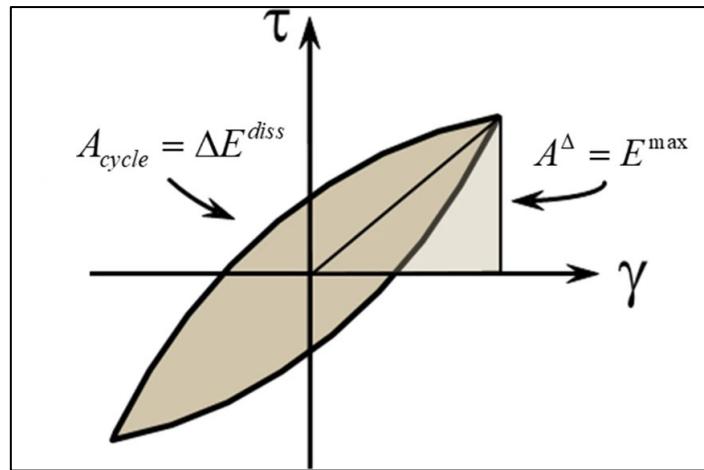


Figure 8: Hysteretic loop of a cyclic stress (τ) – strain (γ) curve illustrating the definition of the quality factor Q_A . The light brown elliptical hysteresis loop indicates the dissipated energy ΔE^{diss} . The coloured triangle represents the maximum strain energy E^{max} . Adopted from Parolai et al. (2022).

Understanding the main physical basis of site effects: the impedance contrast and the resonance allows scientists and engineers to anticipate which areas are likely to experience stronger shaking and to account for that in seismic hazard models. Accurately accounting for non-linear and attenuative behaviors is crucial for realistic ground motion prediction, particularly in regions with soft, unconsolidated sediments. However, under moderate shaking, assuming largely linear soil response, the amplification due to elasticity dominates. As shown above, shear-wave velocity plays a fundamental role in controlling how seismic energy propagates through the near-surface and is therefore considered a key factor in accurately evaluating site response (Stokoe, 1994; Tokimatsu and Asaka, 1998; Wald and Mori, 2000). It is also widely used for site characterization (Borcherdt, 1994; Park et al., 1999; Derras et al., 2017; Hollender, 2019) as will be demonstrated in the following sections.

2.2 Site Effects Estimation

In the previous chapter it was shown that site effects have a great impact on seismic ground motion and could thus cause increased damage to structures. Estimating site effects involves identifying and quantifying how local subsurface conditions modify incoming seismic waves. The key objective is to determine the degree and frequency-dependence of ground motion amplification, which is primarily influenced by parameters such as V_s , Z , f_0 , and D . These factors govern the amplitude, duration, and spectral content of seismic shaking at a given site. To characterize these effects, various measurable quantities are analyzed - seismic response parameters, including spectral ratios, response spectra, transfer functions, or amplification factors (Figure 9 and Figure 10) - typically expressed as acceleration, velocity, or displacement (Mohraz, 1976; Seed and Murarka, 1976; Seed et al., 1976; Lermo and Chávez-García, 1993; Foti, 2004; Michel et al., 2014). Among the seismic response parameters characterizing the intensity of ground motion, response spectra are the most commonly used in engineering practice (Stambouli et al., 2017).

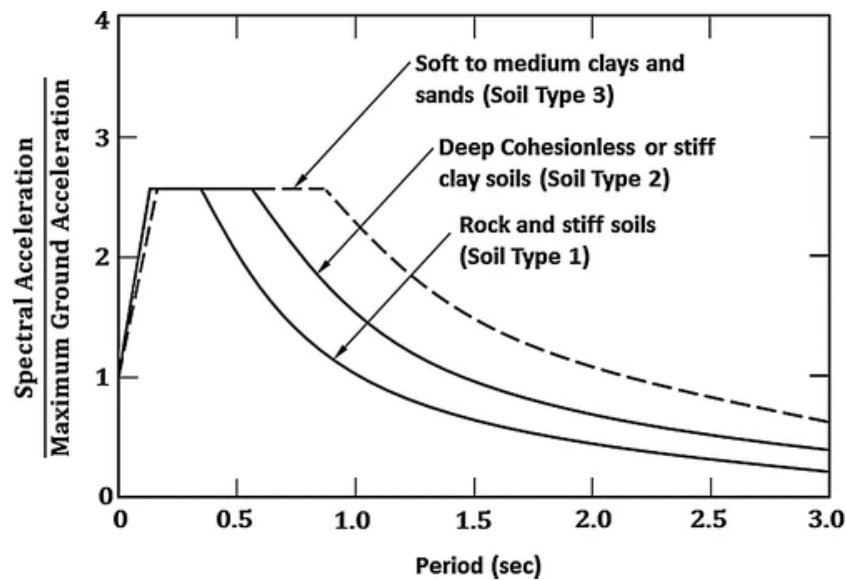


Figure 9: Based on a statistical analysis of over 100 recordings from California Seed and Murarka (1976) and Seed et al. (1976), developed normalized design spectral shapes for 5% damping across different soil conditions, which continue to serve as a key reference in many seismic design codes worldwide. Adapted from UBC (1997).

2.2.1 Amplification factors

In practice, amplification is most often quantified by comparing the response at a site of interest to that at a reference site or input motion, leading to the definition of amplification factors. These factors are typically expressed as the ratio of peak or spectral amplitudes: such as peak ground acceleration (PGA), peak ground velocity (PGV), or spectral acceleration (SA), between the site and a (Joyner and Fumal, 1984; Campbell, 1997; Wald et al., 1999; Baker and Jayaram, 2008; Zhang et al., 2022). Amplification factors are coefficients used to scale the reference ground motion according to local site conditions, accounting for the expected increase in shaking due to soft soils or other geotechnical characteristics. They represent a practical means to include site response effects in engineering design without the need for detailed site-specific analyses (Borcherdt, 1997, 2002; Dobry et al., 2000; Stewart et al., 2003). Various amplification factors are employed in seismic design, including (1) F_a and F_v (Figure 10 used in NEHRP), which correspond to short and mid-period site coefficients (Borcherdt, 1994); (2) S factor (used in Eurocode 8), defining spectral shape modifications based on soil class (CEN, 2005); and (3) empirical corrections incorporated in GMPs, typically as functions of V_s (Chiou and Youngs, 2008; Boore et al., 2014).

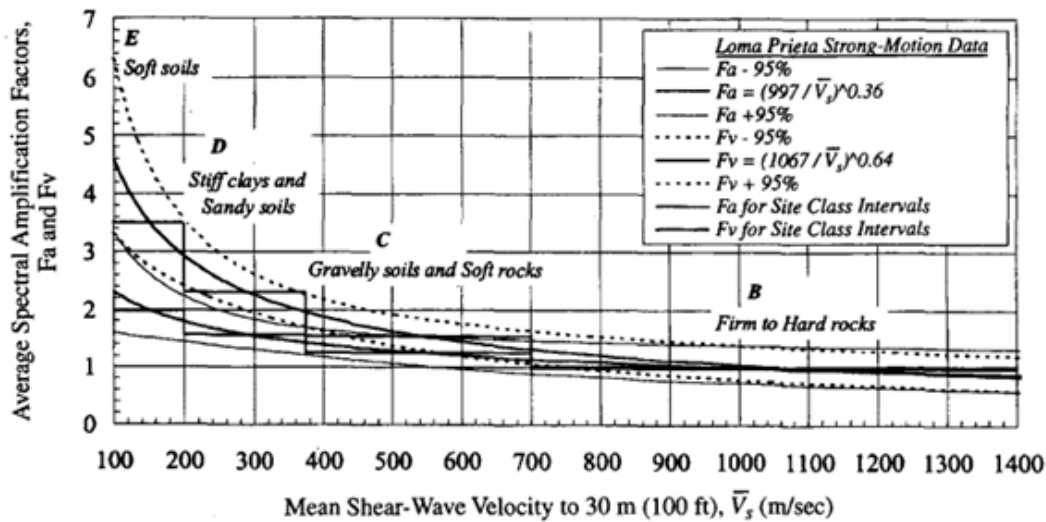


Figure 10: Short- and mid-period amplification factors (F_a , F_v) relative to Firm-to-Hard Rock (Site Class B) plotted as a continuous function of V_{s30} , based on regression from the Loma Prieta earthquake using spectral ratios of nearby soil and rock recordings. Shaded areas indicate 95% confidence intervals and ± 2 standard errors. Predicted amplification factors for simplified site classes (relative to Class B) are also shown, following Borcherdt (1994). Adapted from Dobry et al. (2000).

Site amplification factors are often derived from empirical analyses of earthquake recordings (Borcherdt, 1994, 2002). An illustrative example of such an approach is given below. Correlation dependences between amplification coefficients of short and mid-period Fa , Fv and velocity V_{s30} are presented as follows:

$$Fa = (V_{ref}/V)^{m_a}; Fv = (V_{ref}/V)^{m_v} \quad (2.8)$$

where V_{ref} is the base velocity (e.g. 850, 1050 m/s); V refers to V_{s30} ; m_a and m_v are empirical coefficients that depend on the level of seismic motion, determined according to the corresponding base acceleration (Borcherdt, 1994). Alternatively, site amplification factors can also be obtained through theoretical or numerical modeling, by simulating wave propagation through layered soil profiles to estimate amplification relative to bedrock motion (Field and Jacob, 1995).

2.2.2 Methods for Site Effects Estimation

In the context of seismic hazard assessment, the quantification of site effects through seismic response parameters for local ground motion amplification estimation can be evaluated through:

- 1) empirical methods that directly observe amplification,
- 2) theoretical methods that model wave propagation using analytical transfer functions,
- 3) geotechnical methods that estimate subsurface properties,
- 4) geophysical methods that estimate subsurface properties.

2.2.2.1 Direct empirical methods

Direct empirical approaches commonly utilize methods such as Standard Spectral Ratio (SSR, Borcherdt, 1970; Field and Jacob, 1995; Bonilla et al., 1997; Michel et al., 2014; Poggi et al., 2017) or Horizontal-to-Vertical Spectral Ratio (HVSr, (Nakamura, 1989; Fäh et al., 2001; Havenith, 2004), see Figure 11, analyzing the frequency-dependent spectral ratios of seismic signals to identify site-specific amplification characteristics related to local subsurface conditions. These methods rely on recorded ground motions to compare spectral amplitudes either between horizontal and vertical components at a single site or between a target site and a reference rock site, allowing for the estimation of resonance frequencies and amplification peaks associated with shallow geological structures (Bard, 1999).

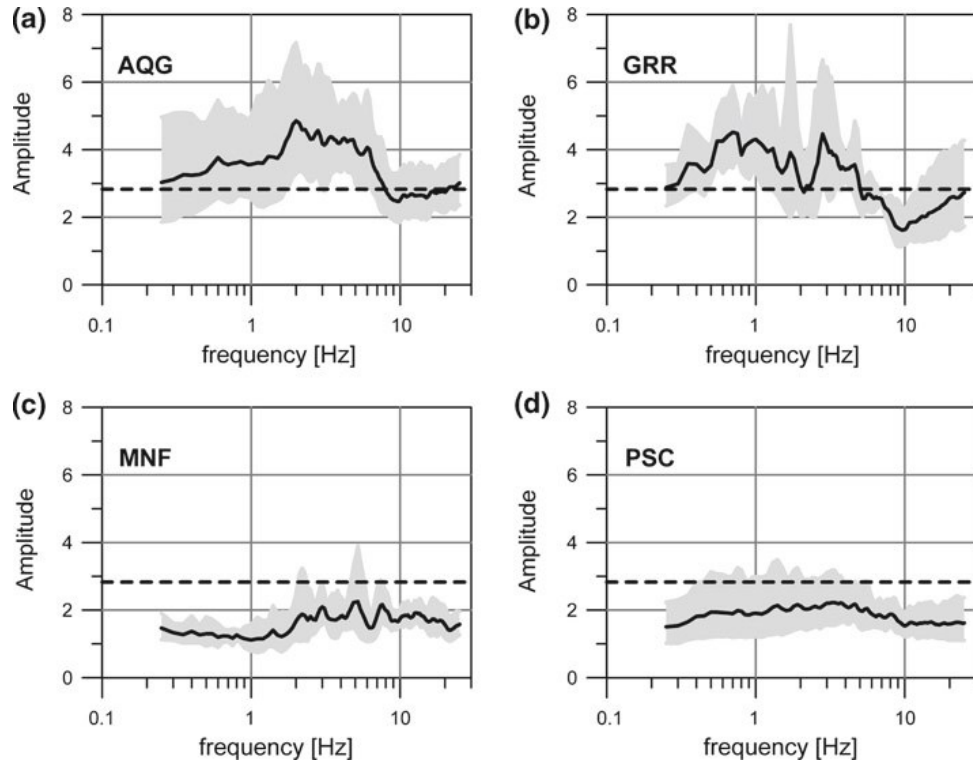


Figure 11: Illustrative examples of stations exhibiting either broad-band or flat spectral ratios. Top panels (left and right) show cases of broad-band amplification, while the bottom panels display flat HVSR curves. Stations PSC d) and MNF c), which exhibit flat responses across the full frequency range, are installed on carbonate bedrock. In contrast, stations AQQ a) and GRR b) are located on highly fractured and weathered rock, respectively. The dashed line represents the threshold used to distinguish between rock and soil sites. Adapted from Luzi et al. (2011).

Perron et al. (2018) presented a clear overview of SSR methods with emphasis on ambient noise vibrations methods, referred here as SSRn, see Figure 12, which offer an attractive solution for the easy and rapid computation of site responses with sufficient spatial resolution, particularly in regions of low seismicity (and could be categorized as geophysical method). SSRn represents the spectral ratio of simultaneously recorded seismic noise at a target site and a rock reference station, analogous to the earthquake-based SSR method. Perron et al. (2018) also introduced the hybrid standard spectral ratio (SSRh) approach, which seeks to enhance the SSRn method by incorporating an intermediate station within the basin, for which the SSR is already known Figure 12.

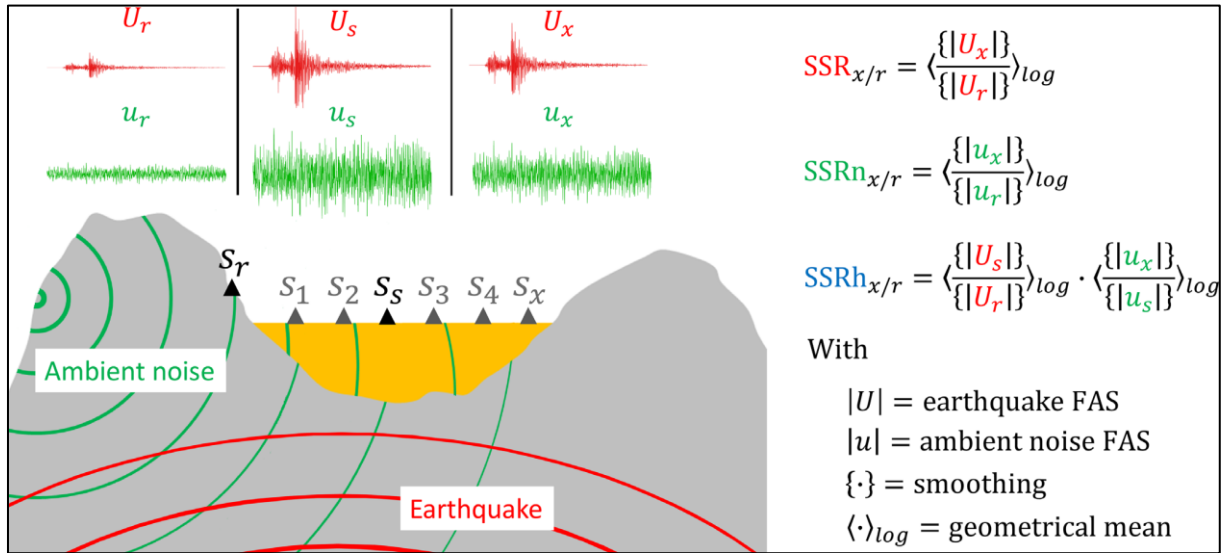


Figure 12: Illustration of Standard Spectral Ratio methods: from earthquake data – SSR, from ambient noise data – SSRn or using hybrid method – SSRh in the case of sedimentary basin (yellow) compared to the bedrock (gray). The two permanent stations (s_r and s_s) are shown in black, while the temporary network stations (s_1 to s_x) are displayed in gray. Right side of the figure refers to the fundamental principles of discussed SSR methods. FAS refers to the Fourier Amplitude Spectrum. After Perron et al. (2018).

2.2.2.2 Theoretical methods

Theoretical methods for evaluating site effects rely on modeling seismic wave propagation through layered media using analytical transfer functions (TFs, (Knopoff, 1964; Telford and Birks, 2009; Garcia-Suarez et al., 2022)). These functions describe how incoming seismic energy is modified by the subsurface structure, particularly due to impedance contrasts between layers of different stiffness and density. In the simplest 1D case, the transfer function can be derived using the matrix method or Haskell-Thomson transfer matrix (Thomson, 1950; Haskell, 1990), assuming vertically incident SH or SV waves and linear, elastic material behavior (Kramer and Stewart, 2024). The transfer function expresses the frequency-dependent amplification of motion at the surface relative to a reference depth, typically bedrock (Mikami et al., 2008; Petrosino et al., 2008).

2.2.2.3 Geotechnical methods

Geotechnical methods are typically based on in-situ or laboratory measurements of soil properties. such as the Standard Penetration Test (SPT, Cetin et al., 2004; Anbazhagan et al., 2013), Cone Penetration Test (CPT, (Schneider et al., 2001; Wang et al., 2016), or undrained shear strength (Rémai, 2013; Cao and Wang, 2014) and are often used to infer site conditions indirectly (Figure 14). An extensive review of previous studies between V_s and SPT N-value can be found in Wair et al. (2012). Although these methods do not provide amplification values directly, they offer valuable input for classification systems (e.g., Eurocode 8, CEN, 2005) or numerical models, and are especially practical where geophysical data are unavailable. This category also includes borehole drilling (Olsson et al., 1988; Mayne et al., 2002) and in-situ downhole or cross-hole seismic approaches (Figure 13), that provide direct estimation of shear-wave velocity profiles in the borehole(s) vicinity (Hunter et al., 2002; Angioni et al., 2003; Du and Pan, 2016).

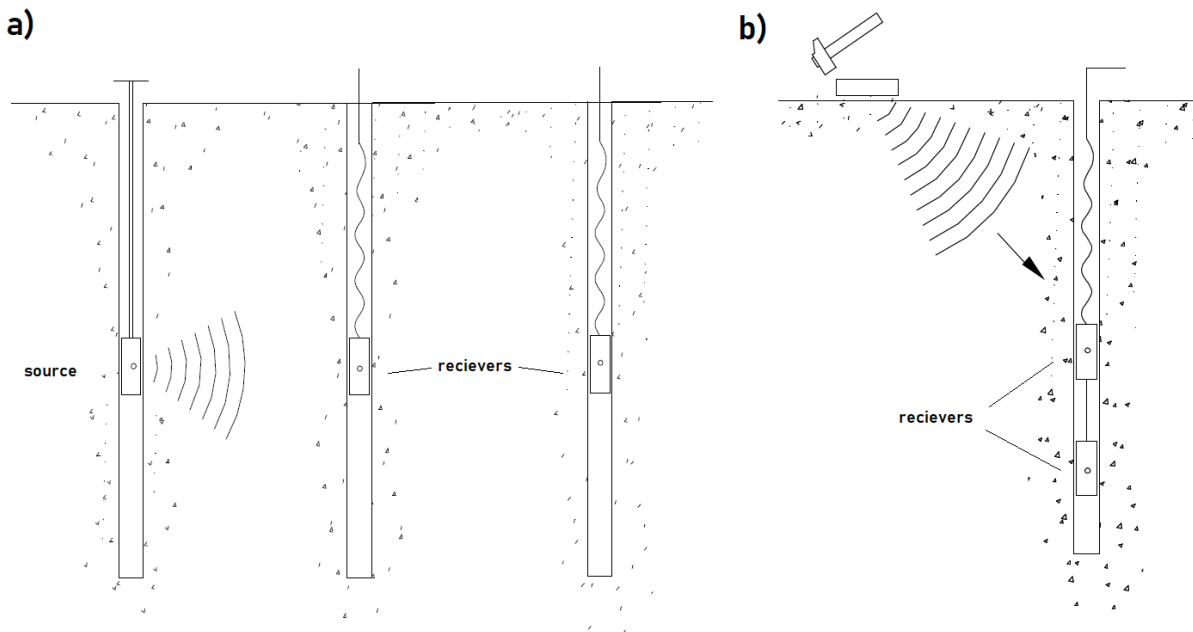


Figure 13: Schematic illustration showing a) the cross-hole method, where the source is placed in one borehole and receivers are located in neighbouring boreholes; and (b) the downhole technique, in which receivers are placed in a borehole and the source is located at the surface. Modified after Foti (2012).

2.2.2.4 Geophysical methods

Geophysical site characterization has become increasingly popular because it provides a relatively fast, non-invasive way to determine key subsurface properties, especially seismic wave velocities (Soupios et al., 2005). The most prevalent geophysical methods for site characterization are those that provides shear-wave velocity (V_s) depth profiles. Geophysical methods include seismic refraction surveys (which can measure P-wave or S-wave velocity or at least layer thickness, (Azwin et al., 2013; Hunter et al., 2022), downhole/crosshole logging

as mentioned (which overlap geophysical and direct categories) and surface wave methods. Using surface wave analysis provides a dispersion curve that is sensitive to the V_s structure of the subsurface and can be inverted into a V_s model. The most common methods include active Multi-channel Analysis of Surface Waves (MASW, Park et al., 1999), and passive techniques such as Spatial Autocorrelation (SPAC, Aki, 1957), Extended Spatial Autocorrelation (ESAC, Okada and Suto, 2003), Refraction Microtremor (ReMi, Louie, 2001), and HVSr (already mentioned among empirical methods). Surface wave methods represent an essential part of this work and are discussed in detail in chapter 3.

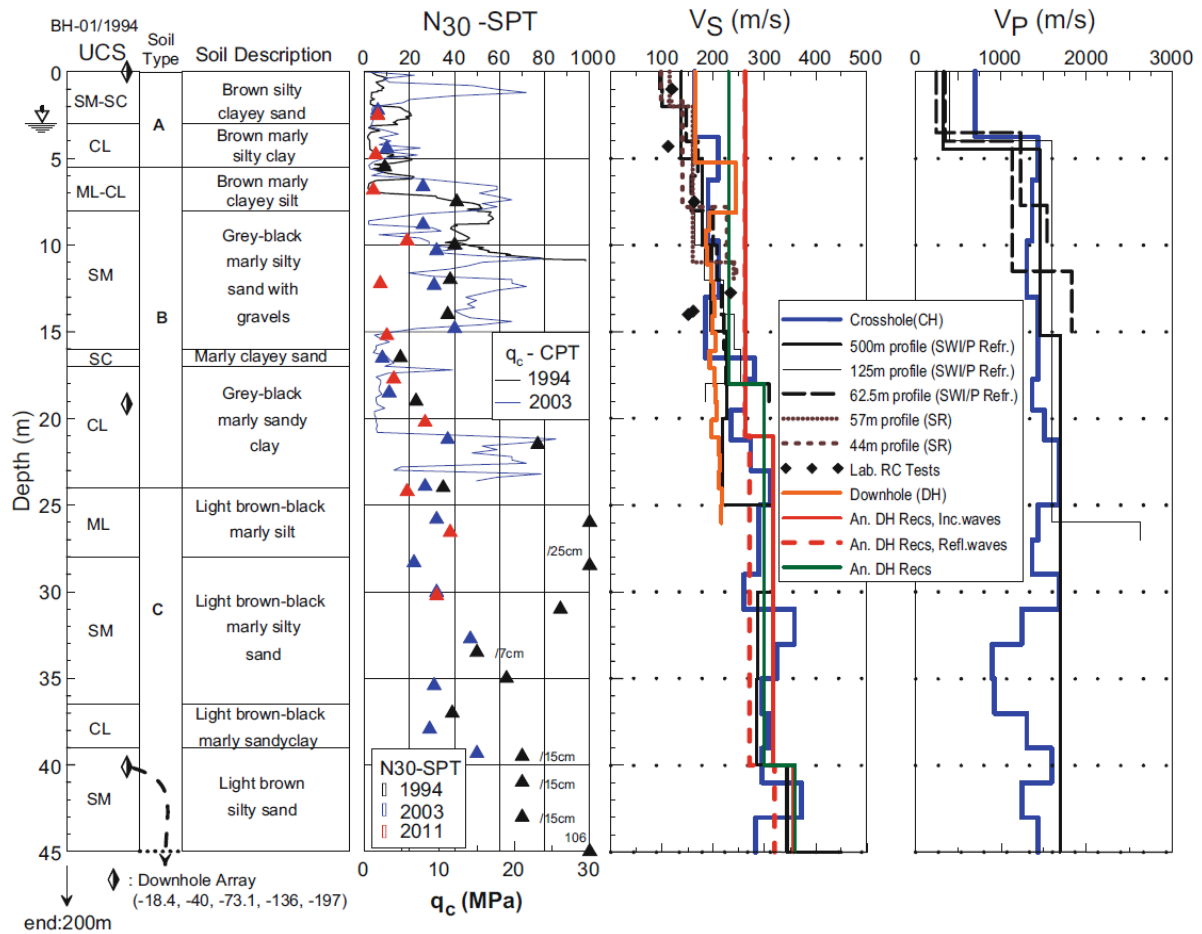


Figure 14: Borehole and CPT logs, along with SPT values, and shear- and compressional-wave velocities (V_s and V_p) derived from multiple methods, are available for the upper 40 meters at the Euroseistest DH array (TST) site. Based on Pitilakis et al. (1999) and Raptakis et al. (2000).

Accurate evaluation of site amplification requires detailed investigation of subsurface structures and the dynamic behavior of the target soils. However, as noted by Shingaki et al. (2018), even when high-quality subsurface data are available at a specific site, extending this level of detail to the surrounding area typically proves challenging. Consequently, while precise approaches are valuable at the local scale, they may not be practical for large-scale or regional seismic assessments.

2.3 Site Classification in Seismic Design

Following the identification, description and evaluation of site effects through various methods discussed in previous chapters, seismic design codes incorporate this information into standardized site classifications for structural design. These classifications, such as building codes - Eurocode 8 (EC8, Part 1 in EN 1998-1:2004 and updated version EN 1998-1-1:2024, (CEN, 2005) in Europe, see Table 2, or the National Earthquake Hazards Reduction Program (NEHRP, BSSC, 1988) provisions in the United States have been developed. These seismic design standards define soil categories based on specific site proxy parameters that serve as simplified indicators of local ground conditions. Such proxies, predominantly based on shear-wave velocity metrics, are used to assign amplification factors and to shape the design spectra for various site classes (Borcherdt and Glassmoyer, 1992; Pitilakis et al., 2012).

Table 2: Site classes according to EC8-1 (CEN, 2005).

Ground type	Description of stratigraphic profile	Parameters		
		$V_{s,30}$ (m/s)	N_{SPT}	c_u (kPa)
A	Rock or other rock-like geological formation, including at most 5 m of weaker material at the surface	>800	–	–
B	Deposits of very dense sand, gravel, or very stiff clay, at least several tens of meters in thickness, characterized by a gradual increase of mechanical properties with depth	360–800	>50	>250
C	Deep deposits of dense or medium dense sand, gravel or stiff clay with thickness from several tens to many hundreds of meters	180–360	15–50	70–250
D	Deposits of loose-to-medium cohesionless soil (with or without some soft cohesive layers), or of predominantly soft-to-firm cohesive soil	<180	<15	<70
E	A soil profile consisting of a surface alluvium layer with V_s values of type C or D and thickness varying between about 5 m and 20 m, underlain by stiffer material with V_s > 800 m/s			

2.3.1 Seismic Site Proxies

Given the limitations of detailed site-specific investigations, site proxies offer a practical means of characterizing local ground conditions for seismic design and hazard assessment, with most of them being directly or indirectly linked to shear-wave velocity (V_s). A brief summary of the most commonly used proxies is provided below, see Table 3, while the following chapter focuses in detail on the most widely applied parameter - V_{s30} .

Table 3: Summary of the most common site proxies.

Proxy	Description	Reference
$V_{s_{bedrock}}$	Average V_s down to bedrock	Héloïse et al. (2012)
V_{s30}	Average V_s over top 30 m	Borcherdt (1994)
f_0	Fundamental resonance frequency (often from HVSR)	Luzi et al. (2011)
H800	Depth to V_s > 800 m/s layer	Derras et al. (2017)

IS_1/IS_0	Impedance ratio (surface vs. base layer)	Shingaki et al. (2018)
C_v	Velocity contrast ($V_{S_{bedrock}}/V_{S_{surface}}$)	Stambouli et al. (2017)
Depth to bedrock	Total thickness of sediments	Power et al. (2008)

2.3.1.1 Depth to bedrock $H_{bedrock}$

Depth to bedrock is a commonly used proxy parameter that represents the depth from the surface to the first significantly stiffer geological layer, typically associated with a strong impedance contrast. It can be expressed as the sum of the thickness of sedimentary layers $H_{bedrock}$ above the underlying rock formation:

$$H_{bedrock} = \sum_{i=1}^n H_i \quad (2.9)$$

It is often estimated through seismic surveys, borehole data, or indirectly inferred from the fundamental resonance frequency (Day et al., 2008; Lee and Trifunac, 2010; Luzi et al., 2011; Garofalo et al., 2016a; Stambouli et al., 2017). Another possibility is to define the bedrock (or basement) with certain V_s value, this approach represents e.g. Steidl (2000), who parametrized a possible deep basin site effect by the depth to $V_s = 2.5$ km/sec iso-surface.

2.3.1.2 $V_{S_{bedrock}}$

The average shear wave velocity up to bedrock ($V_{S_{bedrock}}$) is representing the average seismic velocity of the soil column down to the depth where engineering or seismic bedrock is encountered, typically defined by a threshold V_s (e.g., 800 m/s or 1 000 m/s, (H  lo  se et al., 2012; Garofalo et al., 2016a; Stambouli et al., 2017; Anbazhagan et al., 2018)).

2.3.1.3 Fundamental frequency f_0

The fundamental frequency of a site, commonly determined via horizontal-to-vertical spectral ratios (HVSr, see Chapter 2.2.2), reflects the depth and stiffness contrast between soft sediments and underlying bedrock and is explained in detail in Chapter 2.1.2. It has shown superior correlation with observed amplifications in many contexts, particularly in areas with strong impedance contrasts (Nakamura, 1989; Chatelain et al., 2008; Guillier et al., 2008; Luzi et al., 2011; Burj  nek et al., 2014; Zhu et al., 2020).

2.3.1.4 H_{800}

Suggested as a practical depth marker, H_{800} indicates the depth at which the velocity profile exceeds $V_s = 800$ m/s. This threshold aligns with typical definitions of engineering bedrock and

offers a physically intuitive measure linked to the thickness of the amplifying layer (Derras et al., 2017; Paolucci et al., 2021; Mazanec et al., 2024).

2.3.1.5 Shear wave Impedance Ratio

This parameter captures the contrast in impedance ($\rho \cdot Vs$) between the uppermost and underlaying layers – IS_1/IS_2 (as described in Chapter 2.1.1). It directly relates to the reflection and transmission of seismic energy and has proven especially effective in estimating short-period amplification, where surface-layer properties dominate the response (Shingaki et al., 2018).

2.3.1.6 Velocity contrast C_v

The velocity contrast (C_v) between bedrock ($V_{bedrock}$) and surface materials (V_1):

$$C_v = \frac{V_{bedrock}}{V_1} \quad (2.10)$$

is perhaps the most physically relevant proxy for site amplification. It characterizes the impedance divergence that governs wave trapping and amplification. However, its practical use is limited due to the difficulty in acquiring deep and robust velocity data at most sites (Stambouli et al., 2017).

2.3.1.7 V_{s30}

Introduced by Borchardt (1994), V_{s30} is defined as the average S-wave velocities over the uppermost 30 meters:

$$V_{s30} = \frac{30}{\sum_{i=1}^m \frac{H_i}{v_i}} \quad (2.11)$$

where H_i and v_i denote the thickness and the shear wave velocity for the i -th formation or layer, in a total of m . It is the most widely adopted proxy - used in ground motion prediction equations (GMPEs), and integrated into design codes such as NEHRP, Eurocode 8, and IBC. As this parameter represents one of the central topics of this thesis, it is examined in detail in the next section.

2.4 Vs30 as the Most Common Proxy

The following chapter provides a detailed examination of the Vs30 parameter, including its origin, the most common methods of its determination, relevant case studies, and discussion regarding its relationship to seismic amplification.

A commonly used parameter for site classification is the average shear-wave velocity within the uppermost z meters of the subsurface, denoted as V_{Sz} . In the case of a horizontally layered medium composed of layers overlying a half-space, V_{Sz} is defined by the following expression:

$$V_{Sz} = \frac{z}{\sum_{i=1}^m \frac{H_i}{v_i}} \quad (2.12)$$

where v_i and H_i represent the shear-wave velocity and thickness of the i -th layer, respectively, and m denotes the number of layers extending to depth z . The thickness of the deepest layer is adjusted to match the target depth z :

$$z = \sum_{i=1}^m H_i \quad (2.13)$$

As described in Shingaki et al. (2018), the concept of V_{Sz} is historically based on the observation that the shear-wave velocity of near-surface layers significantly influences site amplification effects (e.g., Borchardt and Gibbs, 1976; Joyner and Fumal, 1984; Shearer and Orcutt, 1987). Borchardt (1978) reported a negative correlation between site amplification and the average shear-wave velocity within the uppermost approximately 30 meters, based on data from the San Francisco Bay area. Joyner and Fumal (1984), had indicated that average velocity within certain layers correlates more effectively with seismic wave amplification than merely considering the depth to bedrock. They suggested using the average velocity to a depth corresponding to a quarter of the seismic wavelength for a given frequency.

Later, Borchardt and Glassmoyer (1992) demonstrated a strong correlation between peak ground motion parameters - such as peak ground acceleration (PGA) - and shear-wave velocity within the uppermost 30 meters, using strong-motion recordings from the 1989 Loma Prieta earthquake (Dietz and Ellsworth, 1990). Similar findings were reported for Japanese sites by Midorikawa (1980) and Tokimatsu et al. (1994), who confirmed a comparable correlation based on national databases and earthquake records.

During the preparation of the U.S. National Earthquake Hazards Reduction Program (NEHRP, BSSC, 1988) provision, Borchardt (1994) proposed adopting average shear-wave velocity to a depth of 30 meters (Vs30) as a definitive quantitative indicator for site characterization (site parameter). His recommendation was based on extensive measurements of velocity profiles (boreholes) and comparisons with data on ground motion amplification in California.

In building codes such as NEHRP, the use of a fixed reference depth was adopted as a more practical alternative to wavelength-based approaches. A depth of 30 meters, approximately equivalent to 100 feet (a standard unit in U.S. geotechnical practice), was chosen, as it corresponds to the typical reach of geotechnical boreholes and aligns well with engineering practice. More importantly, the upper ~30 meters of soil and rock have a dominant influence on the amplification of seismic ground motions. Softer sediments within this zone can significantly amplify shaking, whereas materials at greater depths often transition into stiffer bedrock.

Table 4: Vs30-based site classifications adopted in seismic codes and standards in Europe, Spain, and the United States. From Pinzón et al. (2019).

Eurocode 8 (Europe) [3]			NCSE-02 (Spain) [4]			ASCE 7-16 (USA) [5]		
Soil Type	Description	Vs30 (m/s)	Soil Type	Description	Vs30 (m/s)	Soil Type	Description	Vs30 (m/s)
A	Rock or other rock-like geological formation, including at most 5 m of weaker material at the surface.	Vs > 800	I	Compact rock, very dense cemented or granular soil	Vs > 750	A	Hard rock	Vs > 1500
B	Deposits of very dense sand, gravel, or very stiff clay, at least several tens of meters in thickness, characterized by a gradual increase of mechanical properties with depth.	360 < Vs ≤ 800	II	Very fractured rock, dense or cohesive hard granular soils.	400 < Vs ≤ 750	B	Rock	750 < Vs ≤ 1 500
C	Deep deposits of dense or medium dense sand, gravel, or stiff clay with thickness from several tens to many hundreds of meters.	180 < Vs ≤ 360	III	Granular soil of medium compactness or cohesive soil of firm consistency to very firm.	200 < Vs ≤ 400	C	Very dense soil and soft rock	360 < Vs ≤ 750
D	Deposits of loose-to-medium cohesionless soil or of predominantly soft-to-firm cohesive soil.	Vs < 180	IV	Loose granular soil or soft cohesive soil.	Vs < 200	D	Stiff soil profile	180 < Vs ≤ 360
E	A soil profile consisting of a surface alluvium layer with vs values of type C or D and thickness varying between about 5 m and 20 m, underlain by stiffer material with vs > 800 m/s.	-	-	-	-	E	Soft soil profile	Vs < 180
S1	Deposits consisting of or containing a layer at least 10 m thick of soft clays/silts with a high plasticity index (PI > 40) and high water content.	Vs < 100	-	-	-	F	Other. See ASCE 7-16 Table 20.3-1	-
S2	Deposits of liquefiable soils, sensitive clays or any other soil profile not included in types A–E or S1	-	-	-	-	-	-	-

The parameter Vs30 has become a widely adopted proxy in various seismic and geotechnical applications. It serves as the basis for site classification in seismic design codes such as NEHRP, IBC, ASCE 7, and Eurocode 8 (Table 4), where it is used to assign site categories and associated amplification factors for response spectra (BSSC, 1988; Borchardt, 1994; Dobry et al., 2000; UBC 1997; CEN, 2005). Vs30 is also extensively applied to estimate site amplification effects in both empirical and theoretical studies (Joyner and Fumal, 1984; Borchardt and Glassmoyer, 1992; Castellaro et al., 2008; Mazanec et al., 2024). In strong-motion seismology, it is used as a standard input parameter in ground motion prediction equations (GMPes) for estimating peak ground acceleration and spectral ordinates (Kanno et al., 2006; Power et al., 2008; Abrahamson et al., 2014). Furthermore, Vs30 is employed in microzonation and national hazard mapping projects, providing a spatially continuous proxy for near-surface stiffness (Wald and Allen, 2007; Pitilakis et al., 2013; Yong et al., 2016). Finally, Vs30 remains a practical and accessible parameter for geotechnical engineers in routine seismic site characterization and foundation design (UBC 1997).

2.4.1 Methods for Vs30 Estimation

The Vs30 parameter is widely regarded as a cost-effective alternative compared to more expensive geotechnical characterization methods. Its accurate determination is essential for reliable application in building codes and seismic hazard assessments (Albarelo and Gargani, 2010). A variety of approaches are available for estimating Vs30, several of which have already been briefly mentioned in Chapter 2.2.2. In this chapter, selected methods specifically targeting the measurement of shear-wave velocity (Vs) and the estimation of Vs30 are discussed in greater detail.

2.4.1.1 *Direct Estimation Using Invasive Methods*

An important method for deriving Vs30, especially in engineering practice, involves geotechnical boreholes combined with in-situ or laboratory measurements. Invasive techniques such as downhole and crosshole seismic testing directly measure shear-wave velocities at various depths. Alternatively, correlations between geotechnical parameters and Vs are often employed. Standard Penetration Test (SPT) or Cone Penetration Test (CPT) results can be empirically related to shear-wave velocities through established regression models (Cadet and Duval, 2009). These methods were discussed in Chapter 2.2.2.3. While such empirical relations provide a practical tool, they introduce additional uncertainty, particularly in heterogeneous or unusual soil conditions.

Although borehole-based methods offer a high-resolution and detailed characterization of the subsurface, their application is limited by the high cost and logistical demands of drilling, especially for surveys requiring dense spatial coverage (Albarelo and Gargani, 2010). Borehole data take into account rock properties only in the vicinity of the borehole and can therefore be significantly affected by the location of the borehole. As a result, in regional-scale mapping or in areas with limited resources, borehole data are often supplemented or replaced by geophysical or proxy-based approaches. Moreover, when sufficient investigation depth is achieved, surface-wave methods provide Vs30 estimates comparable to those from invasive tests, with solution non-uniqueness posing no significant concern even under simplified approaches (Foti et al., 2009).

2.4.1.2 *Direct Estimation from Geophysical data*

A common way of Vs30 determination is achieved through direct non-invasive measurement of shear-wave velocity depth profiles. Seismic refraction represents possible approach, relying on the travel times of refracted seismic waves to estimate subsurface velocities (Hunter et al., 2022).

Other alternative involves surface-wave-based methods, particularly the analysis of Rayleigh and Love wave dispersion. Techniques such as Multichannel Analysis of Surface Waves (MASW) or array-based approaches exploit the dispersive nature of surface waves, whose phase velocity varies with wavelength, to reconstruct shear-wave velocity profiles (Park et al.,

1999; Dal Moro, 2014). Surface wave methods are discussed in detail in Chapter 3.3. The fundamental relationship comes from the fact that surface waves sample the subsurface down to approximately one wavelength, and their phase velocity is intimately linked to the shear modulus profile. By targeting wavelengths corresponding to depths of ~30 m, Vs30 can be estimated with high reliability (Foti et al., 2018).

Surface wave dispersion curves currently represent one of the most widely applied non-invasive geophysical methods in both geotechnical and seismic engineering practice (Castellaro, 2016). The analysis of surface waves has been widely employed in numerous studies aimed at Vs30 mapping, as it offers a practical and efficient means of characterizing shallow subsurface conditions over large areas. Kanlı et al. (2006) were using MASW for Vs30 mapping and soil classification in Turkey, (Rahman et al., 2019) estimated Vs30 in 151 sites in Dhaka, capital city of Bangladesh using MASW as one of the methods. Gaytan et al. (2020) applied MASW in combination with refraction microtremor in 33 sites in Guadalajara, Mexico to define the soil classification in terms of Vs30, the bedrock depth and the fundamental period. Sairam et al. (2019) carried out MASW surveys at 500 sites in combination with PS-logging (100 sites) for site condition map. Hollender (2019) characterized 33 sites from the French permanent accelerometric network using MASW and ambient vibration arrays. Odum et al. (2013) presented shallow VS data and determined Vs30 for 27 sites in Puerto Rico, where they used multimethod acquisition approach containing MASW. Buckreis et al. (2024) has evaluated site response for 36 seismic stations in Sacramento-San Joaquin Delta region of California using HVSR method. Prajapati et al. (2024) determined Vs30 of soil in Bhaktapur district in the eastern part of the Kathmandu Basin at 73 locations using microtremor array measurements. Janusz et al. (2025) were seeking site-specific amplification information in soft sedimentary basin in Lucerne area (Switzerland) using ambient vibrations methods. Mazanec et al. (2024) obtained Vs30 and H800 at 17 West Bohemia Seismic Network sites using MASW, see Chapter 4.1. Dovbnich et al. (2022) combined Vs30 mapping based on surface wave analysis with topography-based Vs30 estimation at the Litoměřice geothermal site (Figure 15).

In dispersion analysis, Vs30 is typically calculated from the final inverted shear-wave velocity model using the standard averaging approach, as defined by Eq. 2.11. Recognizing the challenges associated with surface-wave dispersion inversion, particularly the risk of non-uniqueness of dispersion curve inversion interpretation (Zhang and Chan, 2003; Cercato, 2009; Dal Moro, 2014), as discussed in Chapter 3.2.2, several studies have proposed simplified approaches focused on using the phase velocity of the fundamental mode Rayleigh wave at a selected wavelength as a proxy for Vs30 (Brown et al., 2000; Konno and Kataoka, 2000; Martin and Diehl, 2004). Martin and Diehl (2004) proposed estimating Vs30 as 1.045 times the Rayleigh-wave phase velocity at a wavelength of 40 meters. In a related effort to improved Vs30 estimation, Albarello and Gargani (2010) introduced a method based on the direct interpretation of the effective Rayleigh-wave dispersion curve $V_E(\lambda)$.

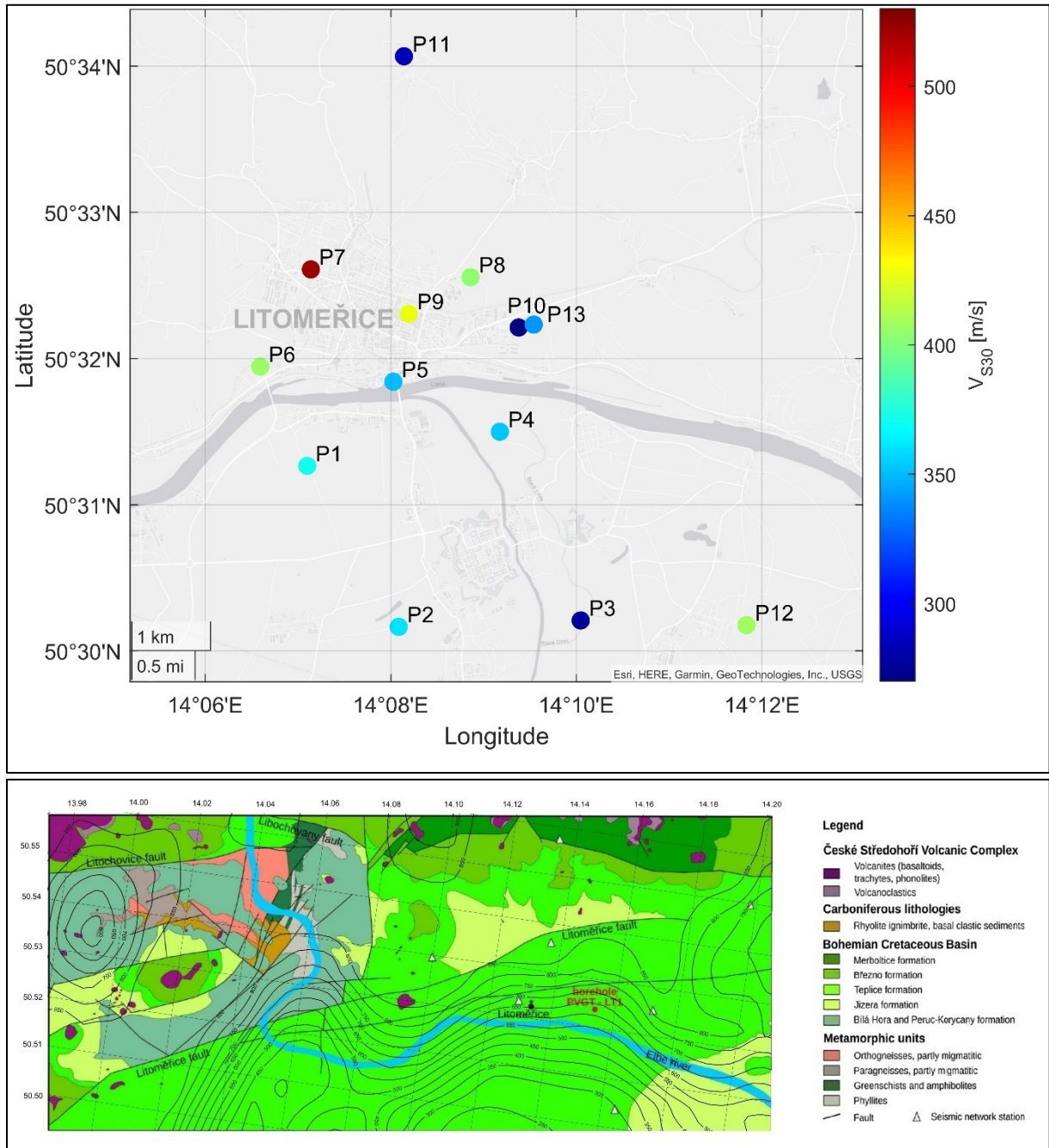


Figure 15: Example of V_{s30} mapping at the Litoměřice geothermal site, showing V_{s30} estimated from surface wave analysis (upper panel) and from topography-based methods. Adapted from Dvořáček et al. (2022).

In this thesis new method for simplified V_{s30} estimation is introduced. In the Chapter 4.3 Rapid V_{s30} Estimation (RAVE) is discussed as a novel approach that derives V_{s30} directly from observed dispersion data without inversion. Two complementary methods are introduced: RAVE-A, which relates V_{s30} to the area under the dispersion curve, and RAVE-SF, which uses the phase velocity at a single frequency for estimation (Mazanec and Dal Moro, 2025).

2.4.1.3 Indirect Estimation

In situations where direct measurements are not feasible, Vs30 may be inferred indirectly from proxy parameters. Among the most widely used proxies is the topographic slope, based on the observation that steeper terrains typically correspond to rock outcrops with higher Vs30 values, whereas flat regions tend to be composed of soft sediments with lower Vs30 (Figure 16, Wald and Allen, 2007). By calibrating empirical relationships between slope and Vs30, it is possible to map seismic site conditions globally, however with considerable non-uniqueness. Chapter 4.2 of this thesis introduce the Vs30 map for the Czech Republic Vackář et al. (2024) and regional Vs30 map of Litoměřice geothermal site from Dovbnich et al. (2022) in Figure 15.

Other possible proxies for Vs30 estimation are surface geology (Stewart et al., 2003; Kottke et al., 2012), terrain categorization (Yong et al., 2012), hybrid slope geology (Thompson and Silva, 2013), and P-wave seismogram (Kim et al., 2016; Zalachoris et al., 2017; Kang et al., 2020).

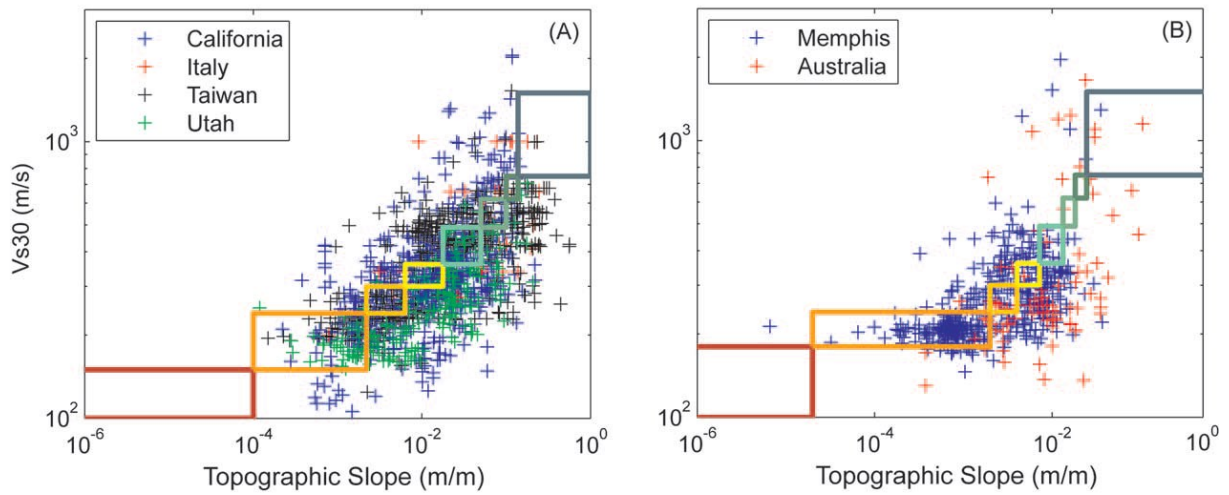


Figure 16: Relationship between measured Vs30 and topographic slope for (a) active tectonic regions and (b) stable continental regions. Color-coded polygons indicate Vs30, and slope intervals related to NEHRP site classes (B-E), for details see Wald and Allen (2007) from where the figure is adopted.

2.4.2 Ongoing Discussions on the Relevance of Vs30

As showed in previous chapters, Vs30 offers a practical, standardized, and easily measurable parameter linked to site effects. Empirical studies, such as those conducted in California, shown in Figure 10, demonstrated systematic differences in ground-motion amplification between Vs30 based site classes, validating its predictive utility (Borcherdt, 1994; Hartzell et al., 2001).

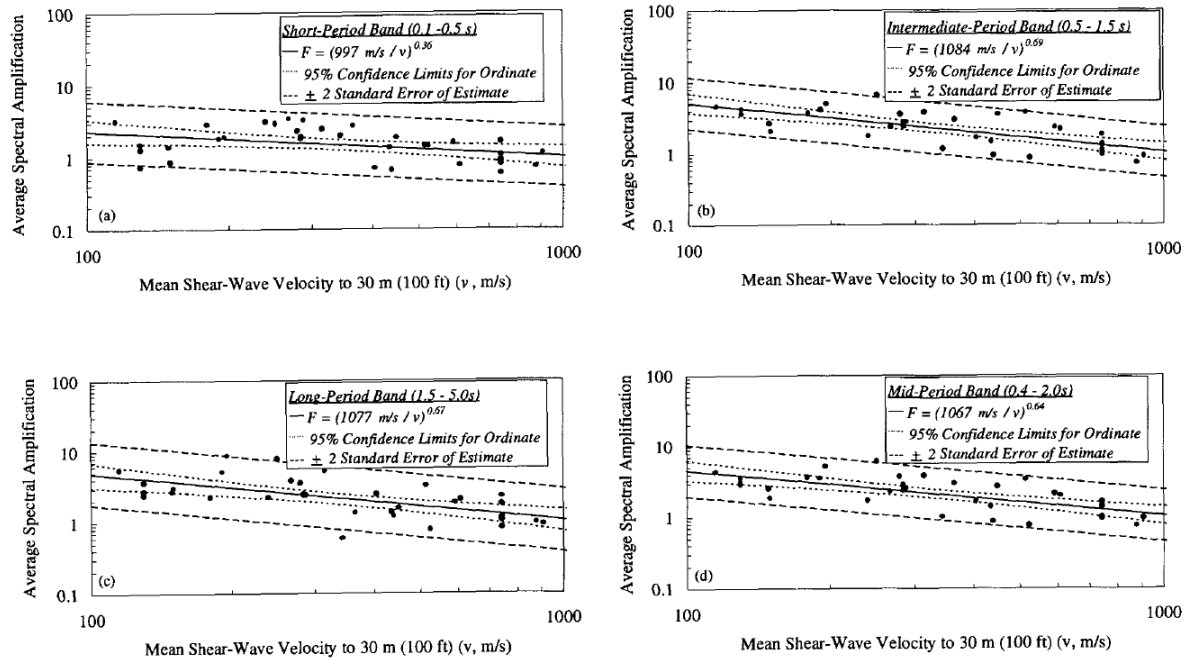


Figure 17: Average horizontal amplification factors with regression curves and confidence intervals derived from free-field strong-motion recordings of the 1989 Loma Prieta earthquake. Results are shown for a) short (0.1–0.5 s), b) intermediate (0.5–1.5 s), c) long (1.5–5.0 s), and d) mid-period (0.4–2.0 s) bands. Shaded areas indicate 95% confidence intervals for the true regression line and ± 2 standard errors of estimate. Adapted from Borcherdt (1994).

Several studies have demonstrated a consistent correlation between Vs30 and site response, as illustrated in Figure 17 and Figure 18 (Borcherdt, 1994; Hartzell et al., 2001; Roser and Gosar, 2010; Rastogi et al., 2011; Sairam et al., 2019). Borcherdt (2012) emphasized, that correlations between Vs30 and strong-motion spectral amplification, as well as average Vs at other than 30 m depths, have established Vs30 as a reliable single parameter for characterizing site response for simplified building codes, GMPE, ShakeMap, and seismic hazard mapping.

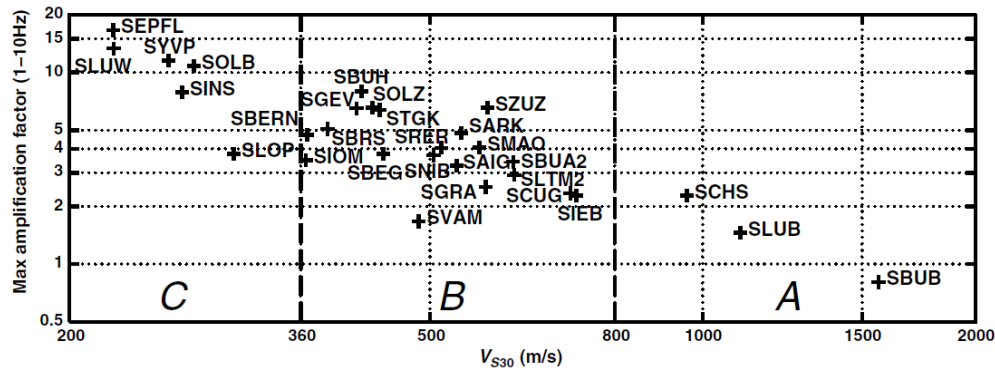


Figure 18: Michele (2014) demonstrated a clear correlation between V_{s30} and amplification within the Swiss Strong Motion Network (SSMNet) site characterization. However, the study also highlighted the presence of 2D/3D site effects, which are not adequately captured by a single parameter such as V_{s30} . Adapted from Michel et al. (2014).

Its practicality and strength were later supported by works such as Wald and Allen (2007) and Boore et al. (2011), who emphasized that V_{s30} , despite its simplicity, captured essential characteristics of site conditions relevant for seismic hazard assessments. Consequently, V_{s30} remains the main parameter of site amplification adjustments in GMPE and seismic design codes globally (Foti et al., 2018).

However, several studies have questioned the adequacy of V_{s30} as a sole proxy for capturing the complexity of site effects, Figure 19 (Steidl, 2000; Wald and Mori, 2000; Stewart et al., 2003; Boore, 2004; Park and Hashash, 2004; Di Giacomo et al., 2005; Lee and Trifunac, 2010; Farrugia et al., 2021). Ghanbari et al. (2018) cautioned against relying solely on V_{s30} , noting that it may oversimplify subsurface conditions. Anbazhagan et al. (2011) emphasized that amplification is influenced by more factors than just shear-wave velocity alone. Further complications were observed in shallow bedrock sites, where using V_{s30} can lead to underestimation of seismic amplification. Anbazhagan et al. (2018) observed that V_{s30} based classifications overestimate shear-wave velocity for very shallow rock conditions and consequently underestimate amplification.

Steidl (2000) observed a weak correlation between site class and amplification, proposing that depth to basement (defined by $V_s = 2500$ m/s) could serve as a more effective predictor of ground motion amplification. Similarly, Boore et al. (2011) examined correlations between V_{s30} and velocities averaged over shallower and deeper intervals, noting that for soft soil sites, V_{s30} may not adequately capture deeper structural effects influencing long-period ground motions.

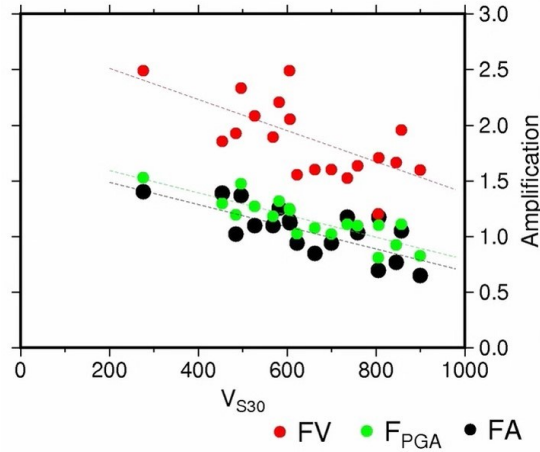


Figure 19: Graphs illustrating the relationship between amplification factors and V_{s30} for the the Maltese islands. All three amplification factors generally decrease with increasing V_{s30} , though significant scatter in FV values is observed. Farrugia et al. (2021) concluded that this highlights the influence of the full Vs profile on site amplification and the limitations of classifying sites based solely on the upper 30 meters. Adapted from Farrugia et al. (2021).

Both studies recommended using the average shear-wave velocity to bedrock ($V_{s_{bedrock}}$) as a more representative parameter, as it better captures the geological features responsible for amplification. However, as highlighted by Foti et al. (2018) and Garofalo et al. (2016b) while near-surface velocities can be reliably estimated using surface-wave methods, the determination of $V_{s_{bedrock}}$ carries significantly higher uncertainty.

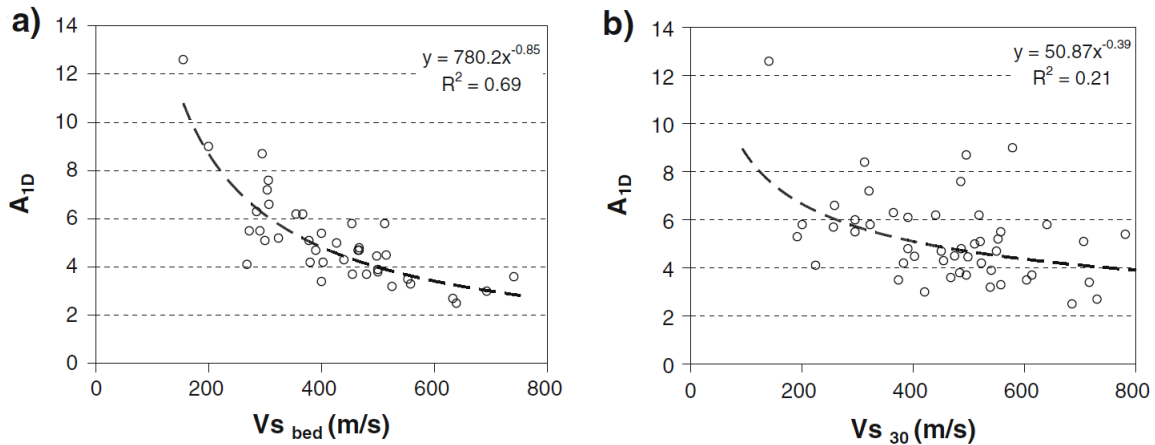


Figure 20: Correlation between average shear-wave velocity and amplification at the fundamental frequency (A_{1D}) based on 1D modeling assuming linear soil behavior: a) average shear-wave velocity to bedrock; b) average shear-wave velocity within the uppermost 30 meters V_{s30} . The average shear-wave velocity to bedrock ($V_{s_{bedrock}}$) shows the strongest correlation with amplification, closely linked to the impedance ratio. Luzi et al. (2011) concluded that as illustrated in a) 1D modelling reveals an inverse power-law relationship with an exponent of -0.85 . However, due to frequent lack of bedrock depth data, V_{s30} is often used as a proxy b), though the fit is weaker. This highlights the advantage of characterizing the full profile down to bedrock to better capture site response. Adapted Luzi et al. (2011).

Some studies proposed enhancing the site classification process by using alternative and complementary proxies. E.g., Stambouli et al. (2017), Luzi et al. (2011), Figure 20 and Derras et al. (2017) showed that combining V_{s30} with other parameters, like the site's fundamental

frequency (f_0) or the depth to a shear-wave velocity of 800 m/s (H800), significantly improves the prediction of amplification. Stambouli et al. (2017) reported that adding f_0 reduced the variance in amplification predictions by at least 60%. Zhu et al. (2020) confirmed that f_0 and depth to bedrock can outperform Vs30 in modeling linear site response, particularly at lower frequencies. Shingaki et al. (2018) concluded that S-wave impedance (I_{s1}/I_{s0}) is effective for use in site amplification evaluations because I_{s1} only requires a physical parameter of the uppermost surface layer. Gallipoli and Mucciarelli (2009) also suggested that the average velocity to 10 meters depth (V_{s10}) could serve as a similarly efficient predictor under certain conditions. Paolucci et al. (2021) indicated that for the new draft of Part 1 Eurocode 8 will likely integrate H800 and VsH (average Vs to a depth of H, where H is conditional on H800) than Vs30. Zhu et al. (2020) analyzed 7 single proxies and 11 proxy-pairs using surface-borehole recordings from Kik-net database, concluding that f_0 is the best single-performing proxy and (f_0 , Vs30) is the best-performing proxy pair, see Figure 20.

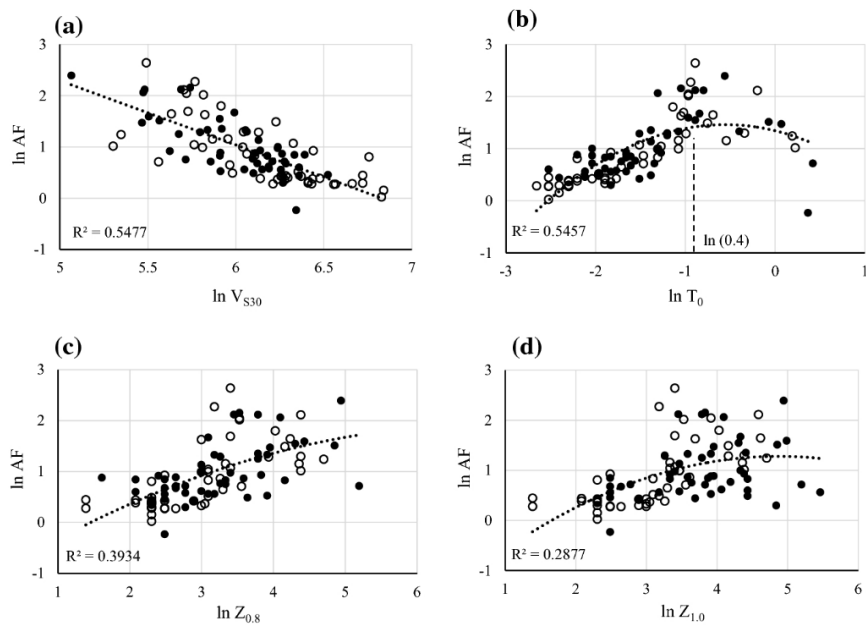


Figure 21: Site amplification as an observed amplification factor (AF) at period $T=0.4s$ plotted against (a) Vs30, (b) f_0 , (c) $Z_{0.8}$, and (d) Z_1 in log-log scale. Dots indicate sites with shear-wave velocity at borehole (V_{shole}) in the range $800 < V_{shole} < 1800$ m/s, while circles represent sites with $1800 < V_{shole} < 3300$ m/s. The dotted line shows the regression fit to all data points characterized by the coefficient of determination R^2 . Adapted from Zhu et al. (2020).

Regional studies further emphasize the limitations of global Vs30 applications. Nambiar et al. (2022) found that Vs30 correlations vary significantly across different geologies in the UAE, advising against uncritical application of global models without regional calibration.

However, one of the strongest objections to Vs30 as an amplification proxy came from Castellaro et al. (2008), who reanalysed the empirical bases of the correlation between Vs30 and amplification from Borchardt (1994) and claimed that seismic amplification is too complex to be related only to the Vs30 (Figure 22). Nevertheless, Borchardt (2012) replied that Castellaro et al. (2008) they used incorrect assumptions leading to a misapplication of the regression model. One of the most serious concerns raised by Castellaro et al. (2008) is the large scatter of

empirical data used by Borchardt (1994) to derive the regression models indicating that sites with identical V_{s30} can exhibit very different seismic responses. Ghanbari et al. (2018) illustrated this issue by presenting three distinct soil profiles with identical V_{s30} values but different amplification characteristics and structural responses, underscoring the risks of oversimplifying site classification. Since one possible source of such scatter could be data quality, the study by Mazanec et al. (2024) focused on a dataset recorded by an instrumentally homogeneous and spatially dense seismic network and seek for the relationship between V_{s30} and amplification.

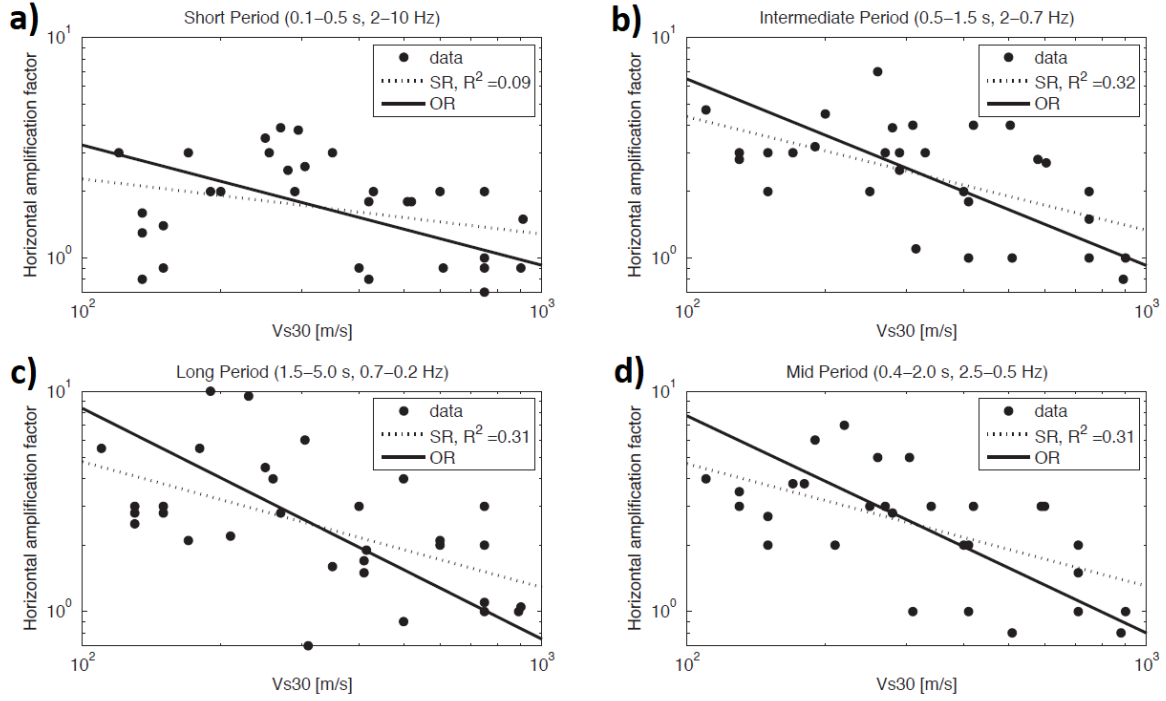


Figure 22: Horizontal amplification factor versus V_{s30} across various period bands: a) short (0.1–0.5 s), b) intermediate (0.5–1.5 s), c) long (1.5–5.0 s), and d) mid-period (0.4–2.0 s) for the 1989 Loma Prieta earthquake data from Borchardt (1994). Standard linear regression (SR, dashed line) with its coefficient of determination (R^2), and orthogonal regression (OR, solid line) are shown. Adapted Castellaro et al. (2008).

2.5 Does Vs30 reflect seismic amplification?

Observations from the West Bohemia Seismic Network

In the previous sections various site proxies were introduced, with a particular focus on Vs30, which remains the most widely used parameter globally. However, as demonstrated in the preceding chapter, its reliability is subject to ongoing scientific debate. The current chapter, based primarily on Mazanec et al. (2024), contributes to this discussion by exploring the relationship between site amplification and Vs30 estimated from in situ geophysical data, using high-quality earthquake recordings from the instrumentally homogeneous and spatially dense West Bohemia Seismic Network (WEBNET).

The development of the WEBNET seismic network began in response to the significant earthquake swarm of 1985/1986 in West Bohemia/Vogtland region (Figure 71) on the Czech German border (Horálek and Vavryčuk, 1985; Horálek et al., 2000). In terms of seismicity, this region is characterized by the frequent occurrence of weak earthquake swarm events, mostly of magnitudes $M_L \leq 3.5$ (Institute of Geophysics, 1991; Horálek et al., 2000; Fischer et al., 2010, 2014). The predominant part of the present seismic activity in the region is roughly defined by the area between 49.9° to 51.0°N and 12.0° to 12.8°E and is characteristically spatially and temporarily clustered (Fischer and Bachura, 2014). Today, the network spans over 900 km² and comprises 25 permanent and temporary stations, providing dense spatial and azimuthal coverage of the focal area, particularly around the active Nový Kostel fault zone (Fischer et al., 2010). The stations are equipped with broadband, three-component weak-motion sensors, predominantly Guralp CMG-3ESPC. Core station parameters are detailed in Fischer et al. (2010), with updated data and earthquake catalogues available from the Geophysical Institute of the CAS.

Using a well-established network such as WEBNET provides the advantage of high-quality, consistently calibrated recordings, ensuring reliable amplitude measurements. However, due to the typical placement of stations on hard bedrock for optimal sensor-ground coupling, the dataset may lack sites with very low Vs30 values. For study in Mazanec et al. (2024), 17 representative and accessible WEBNET stations were selected, as shown in Figure 71.

Shear-wave velocity profiles were derived at 17 WEBNET sites using surface wave analysis, providing both Vs30 and H800 values. The Vs30 values were subsequently compared with (i) H800 and (ii) bedrock densities obtained for the respective sites. This part of the study (Mazanec et al., 2024) is further discussed in Chapter 4.1, while the remainder of this chapter focuses on the amplification analysis. To investigate the relationship between Vs30 and seismic amplification, recordings from WEBNET stations of two moderate earthquakes: Mw 6.4 in Croatia and Mw 4.2 in Austria, were analyzed (Figure 23). The analysis was performed across four types of seismic waves and nine frequency bands, and a regression model was introduced to characterize the amplification–Vs30 relationship.

2.5.1 Data analysis

To assess site response and examine the relationship between Vs30 and seismic amplification, Vs30 values estimated in situ for WEBNET stations were correlated with ground motion amplitudes from two recent earthquakes recorded by the network (Figure 23):

- (1) Mw 6.4 Petrinja, Croatia (45.42°N, 16.26°E), December 29, 2020 (Ganas et al., 2021),
- (2) Mw 4.2 southeast of Vienna, Austria (47.76°N, 16.21°E), April 30, 2021.

The second event is for simplicity hereinafter referred to as the “Vienna” earthquake. These events were selected because: (i) their epicentral distances (~300 km and ~600 km) greatly exceed inter-station distances (<30 km), minimizing azimuthal and distance-related effects; (ii) earthquakes arriving from the southeast produce significantly higher ground motion amplitudes in the Bohemian Massif (Vavryčuk et al., 2004; Málek et al., 2005, 2017); and (iii) strong regional earthquakes (Mw > 4.0) are relatively rare.

Both earthquakes were clearly recorded with a high noise-to-signal ratio at all WEBNET stations, except for the Vienna event, which was not captured at the VAC station due to a temporary technical issue. Example seismograms for both events are shown in Figure 24.

The analysis of earthquake recordings involved first the amplitude spectra (Figure 24) computation following the approach of Borchardt and Gibbs (1976). Next, spectral amplitudes were summed within nine frequency bands: 0–10 Hz, 0–0.5 Hz, 0.5–1 Hz, 0.5–2 Hz, 1–2 Hz, 1–3 Hz, 2–4 Hz, 4–8 Hz, and 5–10 Hz selected as representative for site amplification studies (Borchardt, 1970; Harmsen, 1997; Castellaro et al., 2008).



Figure 23: Map of Central and Southern Europe showing the locations of the two analyzed earthquakes (marked by red stars) - Mw 4.2 near Vienna (~300 km and azimuth ~133° from WEBNET) and Mw 6.4 near Petrinja (~600 km from WEBNET with azimuth ~150°) - in relation to the WEBNET seismic network (red square). Adapted from Mazanec et al. (2024).

The summed amplitudes in each frequency band were taken as proxies for amplification. To allow interstation comparison, these amplification values were normalized to the reference station JOC (amplification ratio = 1), resulting in the so-called *relative amplification ratio*. To investigate the influence of different seismic phases, four waveform categories were analyzed: (i) full seismogram, (ii) P-wave group, (iii) S-wave group, and (iv) surface-wave window. The relative amplification ratios for each category were then plotted as a function of Vs30 across the frequency bands.

Finally, based on previous studies (Midorikawa, 1987; Borchardt, 1994; Gallipoli and Mucciarelli, 2009) and trends observed in the WEBNET dataset, the relationship between Vs30 and seismic amplification was described using a simple empirical regression model:

$$A = A_0 + \frac{k}{V_{s30}} \quad (2.14)$$

where A denotes the amplification amplitude, k is the slope of decline, V_{s30} is the average shear-wave velocity in the uppermost 30 m and A_0 corresponds to the amplitudes for high V_{s30} . The model's fit to the observed data is evaluated using the coefficient of determination (R^2).

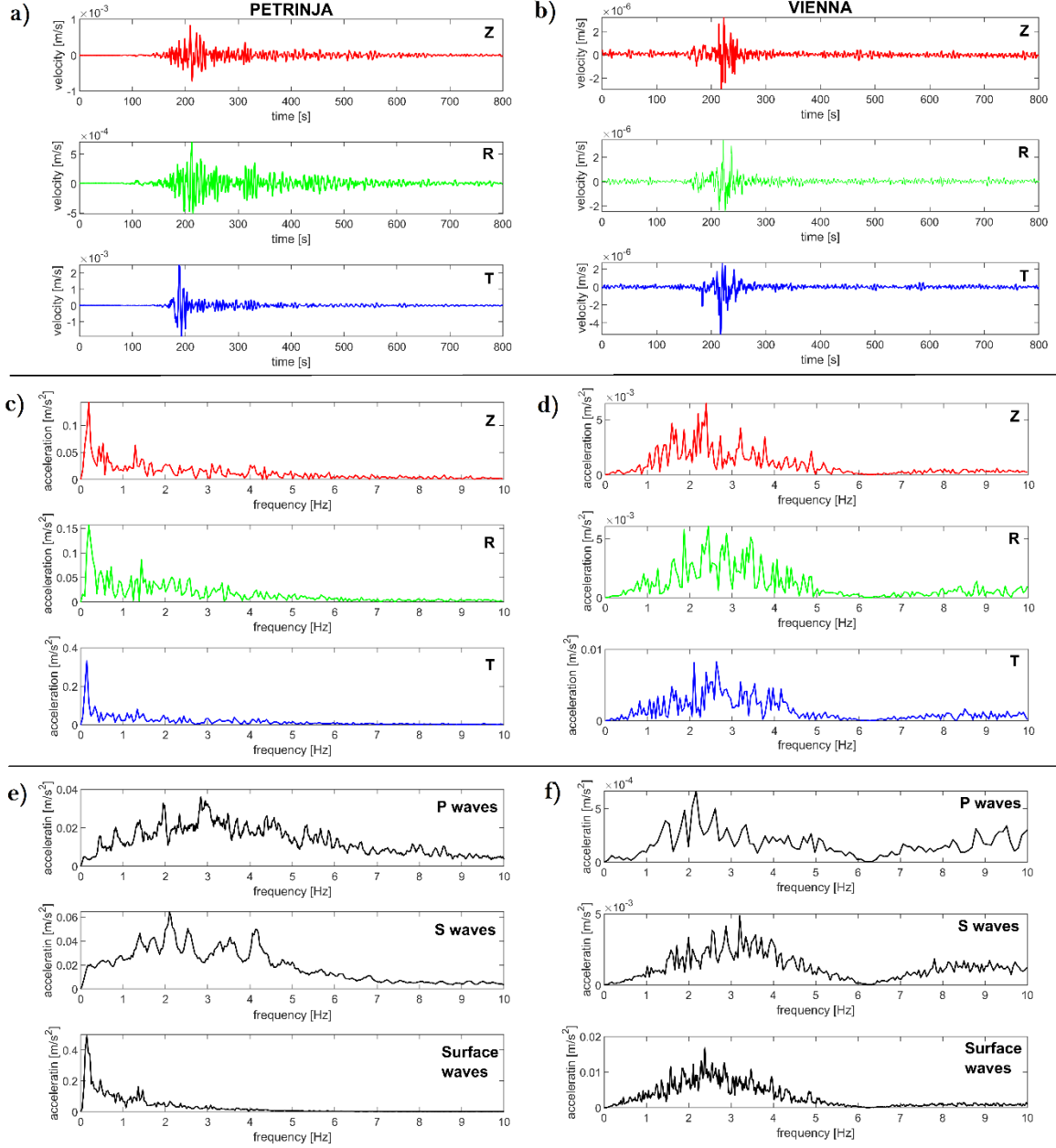


Figure 24: Example of earthquake recordings from the BUBD station. The left column shows data from the Mw 4.2 Vienna event, and the right column from the Mw 6.4 Petrinja event. Panels a) and b) display seismograms (three components: Z, R, and T); panels c) and d) show the corresponding acceleration spectra; and panels e) and f) present acceleration spectra for P-wave, S-wave, and surface-wave segments. Note that amplitude ranges differ due to the varying magnitudes and epicentral distances of the two events. Adapted from Mazanec et al. (2024).

2.5.2 Results on Correlation Between Amplification and Vs30

For each of the nine frequency windows, the relative amplification ratio from two WEBNET recorded earthquakes was plotted against Vs30 across four waveform categories (i-iv), resulting in 36 correlation plots. A simple empirical model (defined by equation 2.14) was fitted to each dataset to describe the relationship between site amplification and Vs30. Figure 25 displays six selected plots with the highest regression coefficients (R^2), three for each of the two events. While the strength of the correlation varies across frequency windows and waveform categories, consistent trends are identifiable and further discussed in the following section.

Figure 26 summarizes the R^2 values from the regression analysis for all four waveform categories across the nine frequency bands. The weakest correlations occur in low-frequency bands (<1 Hz), while the strongest correlations are observed between 1 and 4 Hz, particularly in the 1–3 Hz window, where R^2 values typically range from 0.5 to 0.6 depending on waveform type. Correlation decreases again above 4 Hz. Statistically, the full seismogram category (i) exhibits the most consistent correlations, whereas the P-wave category (ii) shows the lowest overall R^2 values.

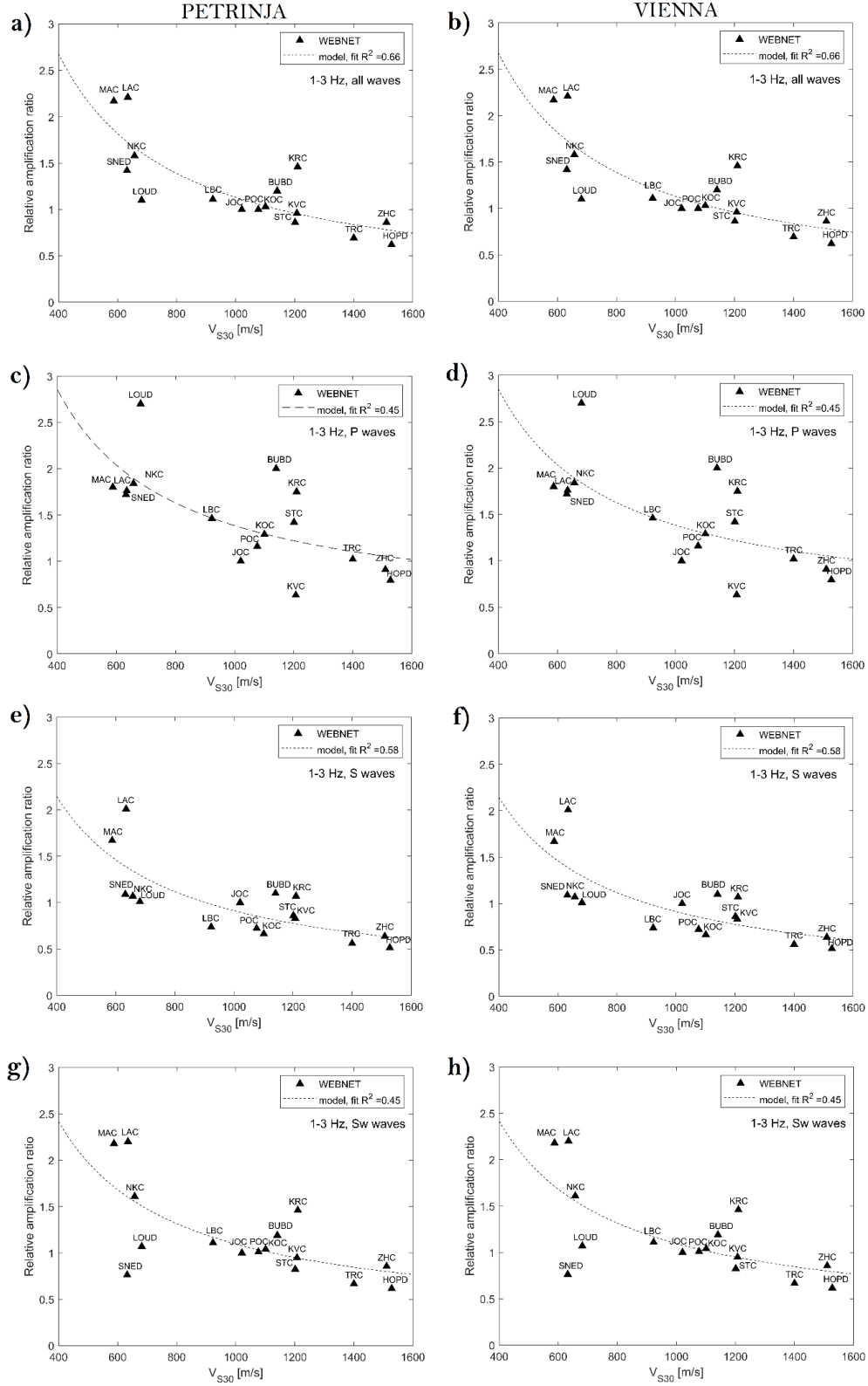


Figure 25: Example scatter plots illustrating the relationship between amplification and V_{S30} . Seismic stations are shown as black triangles labeled with station codes. The dashed line represents the regression model defined by Equation 2, with the corresponding regression coefficient R^2 indicated. The left column displays four correlation plots for the Petrinja earthquake, and the right column shows the corresponding plots for the Vienna earthquake, all within the 1–3 Hz frequency window: (a, b) full waveform, (c, d) P waves, (e, f) S waves, and (g, h) surface waves. Adapted from Mazanec et al. (2024).

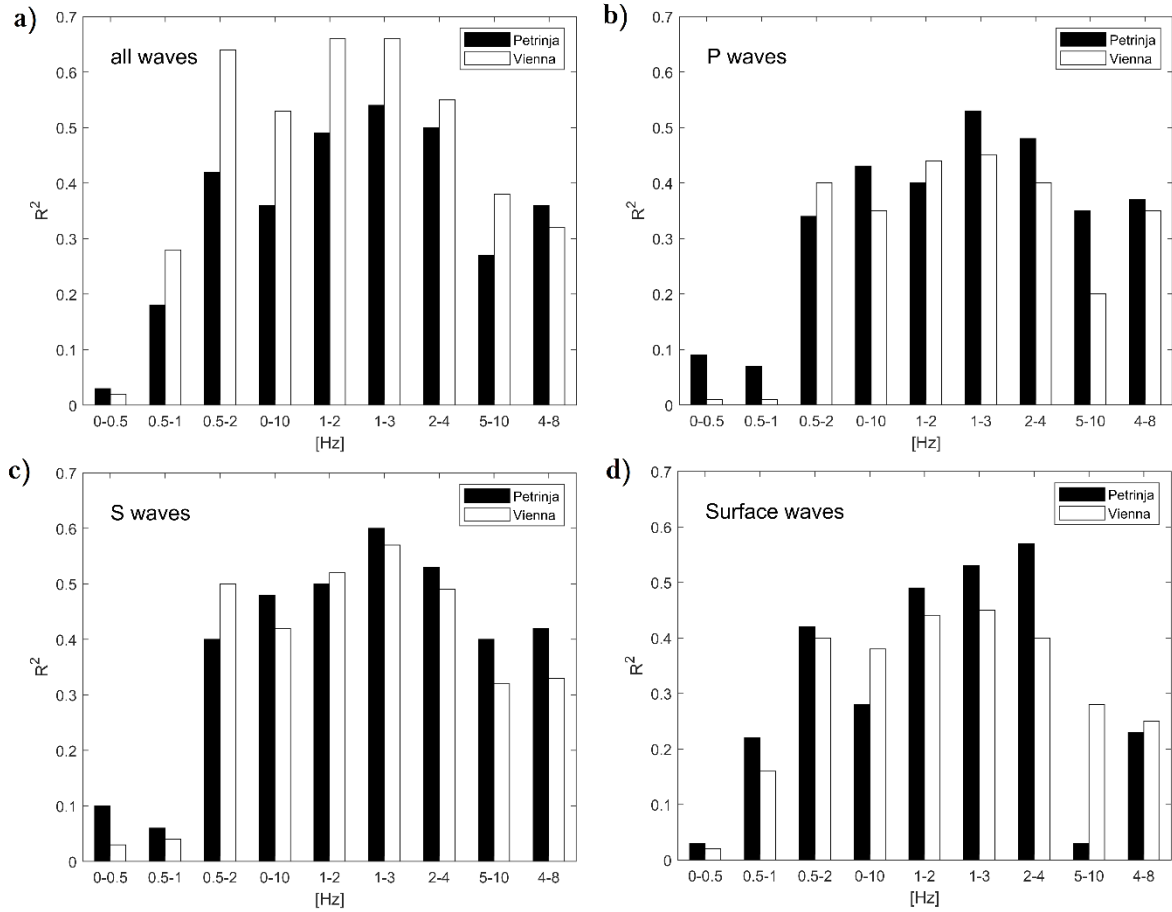


Figure 26: Bar charts showing the regression coefficient R^2 for the Petrinja (black bars) and Vienna (white bars) datasets across nine frequency windows and four waveform categories: (a) all wave types, (b) P waves, (c) S waves, and (d) surface waves. Adapted from Mazanec et al. (2024).

2.5.3 Discussion and Conclusion

Figure 25 indicates a clear relationship between amplification and Vs30 within the analyzed dataset, although this relationship is frequency dependent. Sites classified as EC8 class A, with high average shear-wave velocities, exhibit significantly lower amplification compared to sites with lower Vs30 values (EC8 class B). This trend aligns with the general understanding that seismic amplification tends to increase with decreasing site stiffness (Sairam et al., 2019). Three distinct zones can be distinguished:

- (1) a high-velocity zone ($V_{s30} > \sim 1300$ m/s) including stations HOPD, ZHC, and TRC, where amplification is the lowest - less than half the maximum observed across the dataset,
- (2) a mid-velocity zone ($V_{s30} \sim 680\text{--}1300$ m/s), where amplification shows minimal variation with velocity,
- (3) a low-velocity zone ($V_{s30} < \sim 680$ m/s), primarily represented by sites LAC, MAC, and NKC, where the lowest velocities are recorded and amplification reaches the dataset's maximum, typically at least twice that of the high-velocity zone.

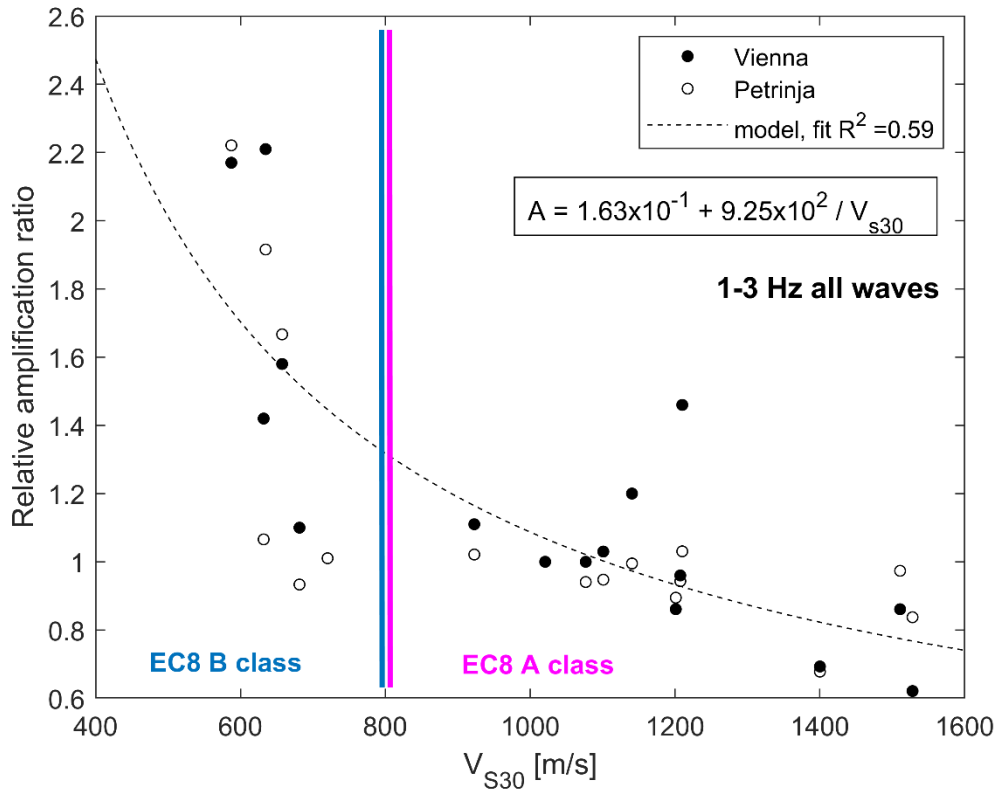


Figure 27: Combined scatter plot showing the relationship between amplification and V_{s30} for both the Vienna (black dots) and Petrinja (white dots) datasets. The dashed line represents the regression model defined by Equation 2, with corresponding model parameters. The goodness of fit is indicated by the regression coefficient R^2 . The plot illustrates the correlation within the 1–3 Hz frequency window for all wave types. Eurocode 8 soil classes A and B are marked with magenta and blue lines, respectively, indicating the velocity boundary at 800 m/s. Adapted from Mazanec et al. (2024).

As illustrated in Figure 26, the regression coefficient R^2 demonstrates a clear frequency dependence, with the strongest correlations observed in the 1–4 Hz range. This corresponds with the energy distribution in the acceleration spectra (Figure 24), confirming that the 1–4 Hz band is statistically optimal for capturing the V_{s30} –amplification relationship, consistent with previous studies: Borchardt and Gibbs (1976) showed that amplitude variations between 0.25–3.0 Hz correlate with site conditions and remain constant up to 4.0 Hz. This also aligns with the mid-period amplification factor F_v (0.5–2.5 Hz), used in seismic design codes (BSSC, 1988; Borchardt, 1994). Similar observations were made by Gallipoli and Mucciarelli (2009), who reported HVSr peaks between 0.5–1.5 s (0.6–2 Hz), and by Harmsen (1997) who found significant V_{s30} –amplification correlations in the 0.5–1.5 Hz and 2–6 Hz bands.

Figure 27 shows a joint scatter plot of amplification versus V_{s30} for the 1–3 Hz frequency window (category i), based on 33 observations from both Petrinja and Vienna earthquakes, fitted with a single regression model (Eq. 2.14). The model yields an R^2 of 0.59 and confirms the trend of decreasing amplification with increasing V_{s30} . However, significant data scatter within EC8 class B (600–750 m/s) is visible. A possible explanation might be a presence of shallow low-velocity layers (~5 m), as seen at high-amplification sites LAC and MAC. Such

shallow features can significantly influence amplification yet remain undetected by standard proxies like V_{s30} . This finding aligns with observations by Anbazhagan et al. (2018). Despite this complexity, the applied hyperbolic regression model in Figure 27 offers a useful means of capturing the overall amplification– V_{s30} relationship, including sharp increases due to near-surface velocity contrasts—an effect not adequately addressed by conventional classifications such as EC8 or NEHRP.

However, the analyzed dataset is known to be influenced, and possibly biased, by several distinct characteristics, such as (i) limited number of observables (17 sites) and (ii) lack of low-velocity stations ($V_{s30} < 500$ m/s), which is typical for high-quality seismic networks optimized for detecting weak events (Di Giulio et al., 2021). However, thanks to the high quality and instrumental consistency of the WEBNET network, local amplification effects remain observable, and the data scatter is lower than in Borchardt (1994) dataset—probably, one of the reasons why Castellaro et al. (2008) criticized the use of V_{s30} as a site proxy.

Previous studies have demonstrated that characterizing local site response is inherently complex and often yields non-unique solutions (Boore, 2004; Castellaro et al., 2008). In this context, concerns raised by various authors regarding the limitations of V_{s30} are well justified, prompting continued efforts to identify alternative or complementary site proxies (Park and Hashash, 2004b; Castellaro et al., 2008; Gallipoli and Mucciarelli, 2009; Lee and Trifunac, 2010; Pitilakis et al., 2013). Derras et al. (2017) evaluated multiple site-condition proxies to identify optimal combinations for characterizing local site effects. One of their key recommendations was the joint use of V_{s30} and the H_{800} parameter. This approach was further supported by Paolucci et al. (2021), who proposed H_{800} and $V_{s,H}$ as new site classification criteria within the Eurocode 8 framework. These findings are consistent with the results of the present study. Correlation of H_{800} with relative amplification, shown in Figure 28a, has clear linear trend and the four stations exhibiting the highest amplification and lowest V_{s30} values (LAC, MAC, NKC, and SNED) also show notably high H_{800} values (>20 m), with H_{800} generally decreasing as V_{s30} increases, as shown in the correlation of these two parameters on Figure 28b. A limitation of the H_{800} parameter at this particular study is that, for 7 of the analyzed stations, the shear-wave velocity of 800 m/s is already reached at the surface (notably, the stations are installed in boreholes directly at the bedrock interface). However, this issue is not expected to be common under standard surface conditions. However, in contrast to the regression-based model developed here, neither H_{800} nor V_{s30} alone captures the influence of near-surface low-velocity layers on amplification, as observed at the LAC and MAC sites.

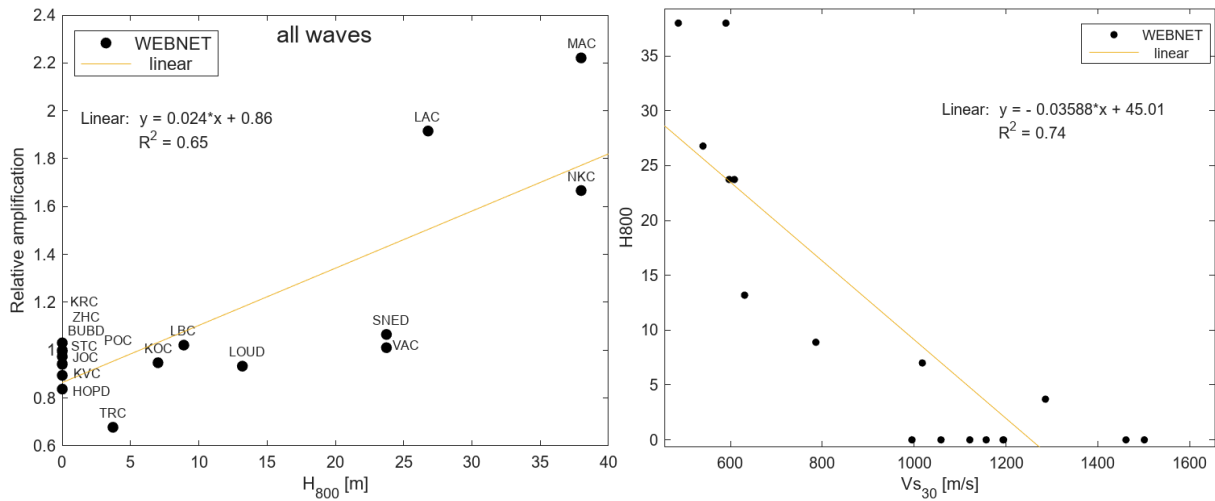


Figure 28: a) Scatter plot illustrating the relationship between amplification and H₈₀₀ with linear regression (Petrinja dataset, 1–3 Hz window, all waves); b) correlation between Vs₃₀ vs. H₈₀₀ for WEBNET sites with linear regression.

Following the discussion on the suitability of Vs₃₀ as a proxy for seismic amplification, it can be concluded that for the WEBNET sites with shear-wave velocity lower than ~ 680 m/s the amplification is significantly higher than for the sites with higher Vs₃₀. Above this threshold (soft to hard rock), site conditions have only a minor influence on amplification. However, considerable scatter remains within the EC8 class B range, which Vs₃₀ or other standard proxies fail to fully capture. The relationship is well described by the regression model (Eq. 2.14), applied across nine frequency windows and four wave type categories. The best fit (highest R²) was observed in the 1–3 Hz range, particularly for the full waveform category. The approach shown in Figure 27 is proposed as a general tool to estimate amplification from Vs₃₀ directly, without introducing new parameters. Unlike categorical classifications, the regression (Eq. 2.14) provides a continuous amplification ratio useful for engineering applications.

3 SURFACE WAVE ANALYSIS

Surface wave analysis are non-invasive geophysical techniques that can estimate the subsurface shear-wave velocity profile by solving the inverse problem of Rayleigh or Love wave propagation. Due to their ability to be easily generated and detected at the Earth's surface, surface waves are an effective means of seismic site characterization (Foti et al., 2018).

Surface wave analysis is a classical research field in seismology since 1920s, especially for crustal studies (Gutenberg, 1924; Evinson et al., 1959; Capon, 1969; Knopoff and Panza, 1977; Kolínský and Brokešová, 2007; Novotný et al., 2013; Prodehl et al., 2013; Málek et al., 2019). Although early studies (e.g., Gabriels et al., 1987; Jongmans & Demanet, 1993) demonstrated the potential of surface waves for near-surface characterization, they have historically been regarded as noise ('ground roll') in this field. However, seismic sources near the surface typically generate longitudinal, shear, and surface waves, with surface waves carrying a dominant portion of the total wave energy (Miller et al., 1955). Therefore, compared to traditional methods based on bulk wave analysis, surface-based wave techniques provide a more intuitive and effective detection approach (Chen et al., 2021).

Surface waves are particularly useful for site characterization due to their strong sensitivity to the mechanical and elastic properties of the stratified subsurface, such as shear modulus. Their propagation velocity depends primarily on the shear wave velocity (V_s) of the medium, making them a powerful tool for assessing the stiffness and rigidity of geological layers (Stokoe, 1994; Park et al., 1999). Unlike body waves, which propagate through the entire volume of a medium, surface waves are confined to the uppermost layers, where most of the variations in mechanical properties occur (Nazarian, 1984; Park et al., 1999). This confinement allows for high-resolution profiling of shear wave velocity, a key parameter in evaluating seismic site response and soil behavior under dynamic loading (Tokimatsu, 1997).

Additionally, surface waves exhibit frequency-dependent dispersion, meaning that different wavelengths penetrate to different depths. This specific characteristic enables the reconstruction of V_s -depth profiles by analyzing the dispersion curve, providing insights into the vertical variations of stiffness and elasticity (Xia et al., 1999; Socco and Strobbia, 2004). As surface waves propagate along a free surface, it is relatively easy to measure their motion. Furthermore, surface waves are less affected by small-scale heterogeneities and local velocity anomalies compared to body waves, leading to more stable and reliable estimates of subsurface mechanical properties (Foti et al., 2018). These advantages make surface wave methods effective for geotechnical assessments, seismic hazard studies, and the classification of site conditions in engineering applications (Garofalo et al., 2016b, 2016a).

3.1 Fundamental Characteristics of Surface Wave propagation

Surface waves are one of the seismic waves categories that propagate along the Earth's free surface, resulting from the complex interaction between body waves and the mechanical boundary condition imposed by the surface. Surface waves are generated when seismic energy encounters a free surface boundary, such as the Earth's surface. They can be excited by both natural sources, including earthquakes, volcanic processes, and landslides, and anthropogenic sources such as controlled explosions mechanical impacts or ambient vibrations (Lay and Wallace, 1995).

There are two main types of surface waves: Rayleigh (Rayleigh, 1887) and Love (Love, 1911) waves. Rayleigh waves involve both vertical and radial particle motion in the propagation plane and are characterized by retrograde elliptical paths near the surface (Figure 32). Love waves, by contrast, involve purely horizontal transverse particle motion perpendicular to the direction of propagation and exist only when a low-velocity layer overlies a higher-velocity material (Figure 30).

Surface waves exhibit a set of distinctive physical and mathematical properties that differentiate them from body waves and render them particularly useful for probing near-surface structures. Unlike body waves, their geometrical confinement to a near-surface results in energy concentration. Miller et al. (1955) noted that normal load at the surface generates stress waves in the form of longitudinal waves, shear waves, and surface waves, which account for approximately 7%, 26%, and 67% of the total wave energy, respectively. Thus, surface waves carry about two-thirds of the total wave energy. Surface waves typically exhibit larger amplitudes compared to body waves at longer periods. They often dominate the seismograms (Figure 29) in the period range 10-200 s, making them particularly significant in seismological studies of Earth's crust and upper mantle structure, as well as in engineering seismology for assessing ground motion and structural response during earthquakes (Aki and Richards, 1980).

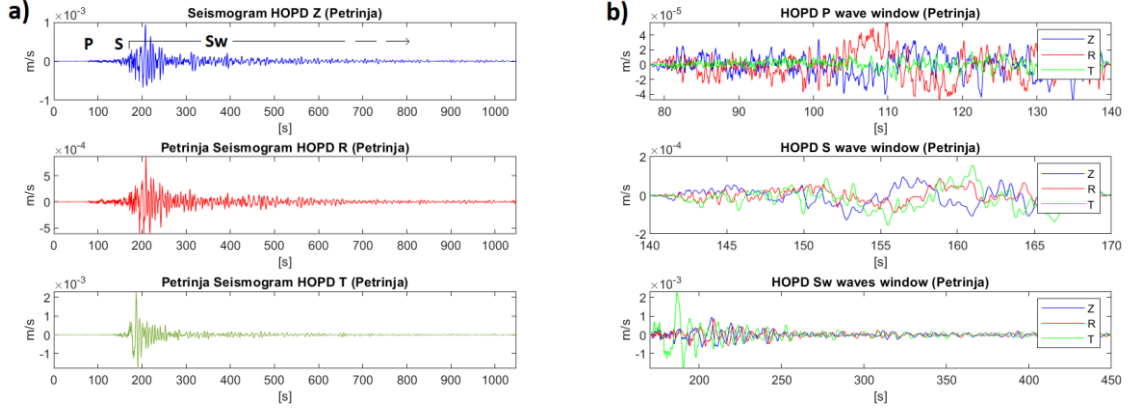


Figure 29: Seismograms illustrating the dominance of surface waves. a) Z, R, and T components (often called Z, N-S, E-W) of the Mw 6.4 Petrinja earthquake recorded at the HOPD station (WEBNET), with labels indicating the arrivals of P, S, and surface waves (Sw); b) the same seismograms segmented into P, S, and Sw time windows, showing all three components. Adapted from Mazanec et al. (2024).

Due to the near-surface confinement surface waves exhibit more rapid attenuation of amplitude with depth than body waves. This also leads to their characteristic dispersive behavior in heterogeneous media, as their phase and group velocities become frequency-dependent due to varying sensitivities to subsurface elastic properties at different wavelengths. The depth of influence, or penetration depth z , is commonly approximated by the wavelength λ , where lower frequencies (longer wavelengths) sample deeper structures, and higher frequencies (shorter wavelengths) are sensitive to shallower layers (Foti et al., 2014).

The following sections outline key characteristics of surface waves. As a representative example, the propagation of Love waves, chosen for their relative mathematical simplicity, will be briefly derived to illustrate the concept of geometric dispersion. A simplified approach for Rayleigh wave propagation will also be outlined, highlighting the fundamental differences introduced by P-SV wave coupling.

3.1.1 Love Waves

Love waves (Lw), Figure 30 are surface waves that result from the constructive interference of horizontally polarized shear waves (SH-waves) trapped in a low-velocity surface layer overlying a higher-velocity half-space (Shearer, 2019).

The following section outlines the general approach and key steps in deriving the propagation velocity of Love waves in a vertically heterogeneous half-space, based on Aki and Richards (1980), Panza et al. (2001) and Shearer (2019).

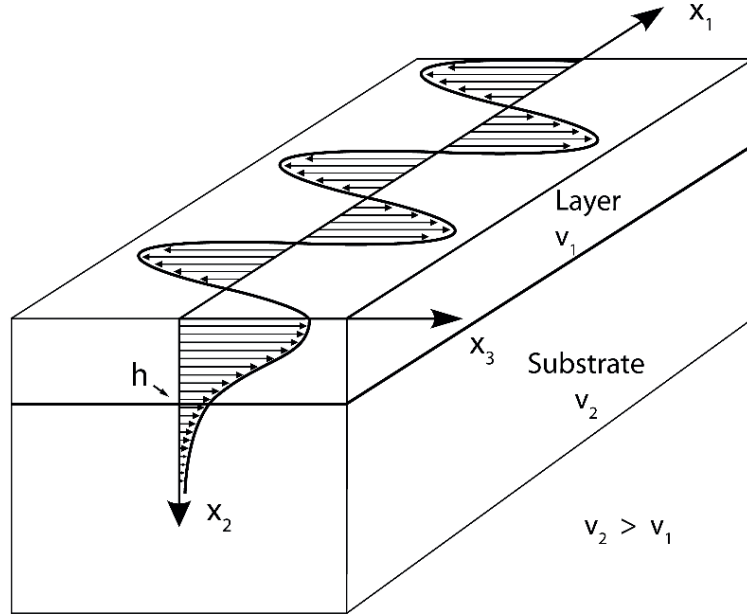


Figure 30: Illustration of a Love wave propagation in isotropic elastic vertically heterogeneous half-space. An elastic surface layer of thickness h and a shear velocity v_1 is rigidly connected to the underlying semi-infinite substrate with $v_2 > v_1$. Adapted from Kiełczyński (2018).

First, several assumptions need to be set. In the considered half space ($z \geq 0$), the elastic properties, specifically the shear modulus $\mu(z) = \rho(z)V_s(z)^2$, and the density $\rho(z)$ are functions of depth z . The free surface is defined as $z = 0$. Love waves propagate in the horizontal x -direction, with particle motion (v) only in the transverse horizontal (y) direction. The displacement vector (u) is defined:

$$u = (0, v(x, z, t), 0) \quad (3.1)$$

The equation of motion for SH-waves in an elastic medium is given by:

$$\rho(z) \frac{\partial^2 v}{\partial t^2} = \frac{\partial}{\partial x} \left(\mu(z) \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial z} \left(\mu(z) \frac{\partial v}{\partial z} \right) \quad (3.2)$$

When assuming a harmonic wave solution propagating in the x -direction with angular frequency (ω) and wavenumber (k):

$$v(x, z, t) = \beta(z) e^{i(kx - \omega t)} \quad (3.3)$$

where $\beta(z)$ is the amplitude of the wave as a function of depth. The phase velocity (c) of the wave is $c = \frac{\omega}{k}$.

Next steps involve: 1) the assumed harmonic wave solution (3.3) is substituted into the equation of motion (3.2), 2) the partial derivatives with respect to x and z , (where $\mu(z)$ is independent of x) are performed; and 3) substituting $V_s(z)^2 = \frac{\mu(z)}{\rho(z)}$. Which leads to an differential equation for the depth-dependent amplitude $\beta(z)$:

$$\frac{\partial^2 \beta(z)}{\partial z^2} + \frac{1}{\mu(z)} \frac{\partial \mu(z)}{\partial z} \frac{\partial \beta(z)}{\partial z} + k^2 \left(\frac{c^2}{Vs(z)^2} - 1 \right) \beta(z) = 0 \quad (3.4)$$

This equation governs how the wave amplitude changes with depth, considering the varying material properties and the wave's frequency and wavenumber.

To obtain non-trivial solutions for $\beta(z)$ and consequently determine the possible values of the phase velocity $c(\omega)$, appropriate boundary conditions need to be applied: 1) Free surface condition ($z = 0$): zero shear stress: $\frac{\partial v(z)}{\partial z} \Big|_{z=0} = 0$ (no force acting on the surface in the y -direction); 2) radiation condition: $z \rightarrow \infty$, wave amplitude does not increase with depth, therefore the wave energy is either trapped near the surface or decays with depth. Solving the differential equation (3.4) for $\beta(z)$ under these boundary conditions is generally a complex task and depend on the specific forms of $\mu(z)$ and $\rho(z)$. For a homogeneous layer over a homogeneous half-space the equation has this form:

$$\tan \left(Hk \sqrt{\frac{c^2}{Vs_1^2} - 1} \right) = \frac{\mu_2 \sqrt{1 - \frac{c^2}{Vs_2^2}}}{\mu_1 \sqrt{\frac{c^2}{Vs_2^2} - 1}} \quad (3.5)$$

which is the dispersion equation for Love wave, where H is the thickness of the layer, Vs_1 and μ_1 are the shear wave velocity and shear modulus of the layer, and Vs_2 and μ_2 are parameters of the half-space. This equation implicitly defines the phase velocity c as a function of frequency ω (since $k = \frac{\omega}{c}$), showing dispersion properties of Love wave in a dispersive medium (such as vertically heterogeneous half-space). This phenomenon arises because different frequency components of the wave interact differently with the layered structure of the medium: higher frequencies, having shorter wavelengths, are more influenced by the properties of shallow layers, whereas lower frequencies sample deeper parts.

The presence of the tangent function dependent on c , implies that for a given frequency ω , there exists more than one solution for the phase velocity c that satisfies the equation. Each of these solutions corresponds to a distinct mode of propagation for Love waves. These modes are differentiated by their vertical amplitude structure (the dependence of $\beta(z)$ on depth) and the extent to which the wave energy is restricted within the surface layer (Aki and Richards, 1980; Sheriff and Geldart, 1995).

The argument of the tangent function, $Hk \sqrt{\frac{c^2}{Vs_1^2} - 1}$, determines how many times the vertical component of the wavelength within the surface layer fits into the thickness H of the surface layer.

Different modes correspond to a different number of parts of the wavelengths that fit into the layer and are subject of the constructive interference, which is a condition for the existence of Love waves.

The fundamental mode exhibits the lowest frequency (for a given phase velocity) or the highest phase velocity (for a given frequency). Its vertical amplitude typically displays a single maximum within the surface layer and decays with increasing depth.

For higher frequencies (or lower phase velocities), additional solutions to the dispersion equation exist, corresponding to higher modes (overtones). These modes have more complex vertical amplitude structures. The energy of higher modes may be less concentrated in the surface layer, allowing them to penetrate to greater depths (Panza et al., 2001; Foti et al., 2014). Since multiple possible phase velocities c exist for Love waves at each frequency ω (each corresponding to a different mode), this dependence can be plotted for each mode. These graphical representations are termed dispersion curves.

The dispersion curve of the fundamental mode typically originates from frequency close to zero, and its phase velocity varies with increasing frequency. In the low-frequency limit, the phase velocity approaches the SH-wave velocity of the half-space (V_{s2}), whereas in the high-frequency limit, it approaches the SH-wave velocity of the surface layer (V_{s1}).

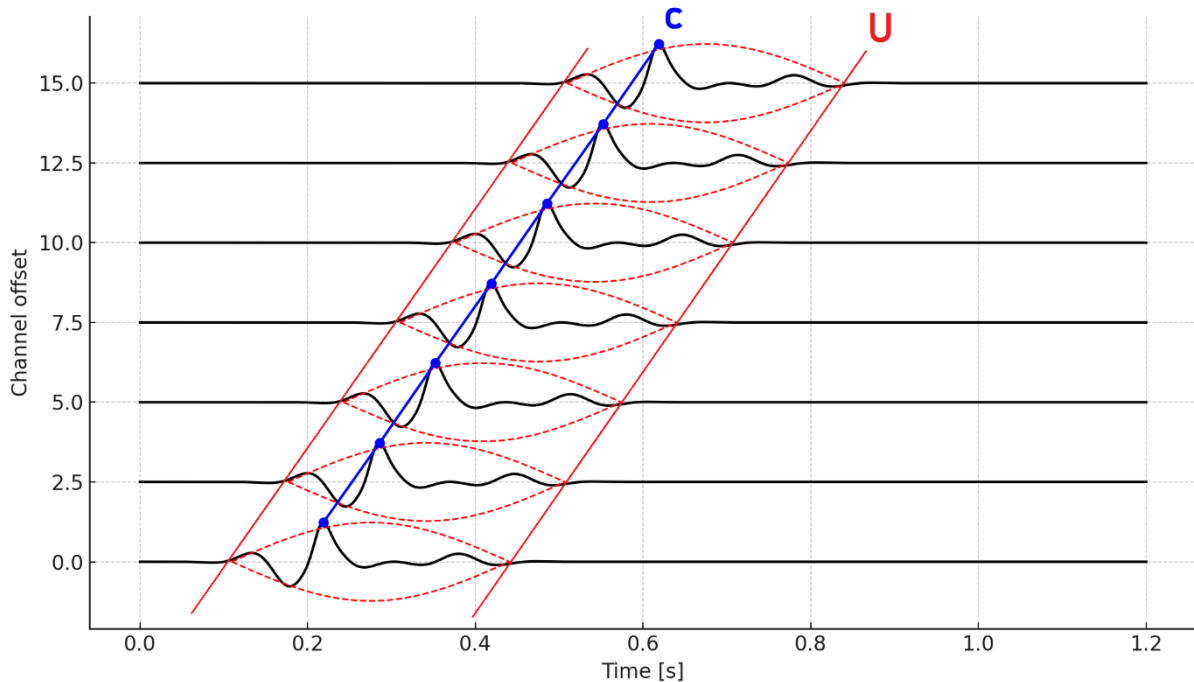


Figure 31: Illustration of group (U , red lines) and phase (c , blue line) velocities on synthetic multichannel record.

The dispersion curves of higher modes develop only above certain cut-off frequencies. Below this frequency, the corresponding higher mode does not exist as a propagating wave. Above the cut-off frequency, the phase velocity of this mode also depends on frequency, forming its own dispersion curve. These curves are generally situated below the curve of the fundamental mode. The shape of the dispersion curves is governed by the parameters of the medium, specifically the thickness of the surface layer H and the elastic properties of both layers (V_{s1} , μ_1 , V_{s2} , μ_2). Analyzing the shape of dispersion curves obtained from seismic data allows for the inference of subsurface structural information (Dal Moro, 2014).

The group velocity U , which represents the velocity at which the energy of a wave packet propagates (Figure 31), is given by:

$$U = \frac{\partial k}{\partial \omega} \quad (3.6)$$

For a vertically heterogeneous half-space, the dispersion relation and the resulting phase and group velocities will depend on the specific nature of the heterogeneity (i.e., how $\mu(z)$ and $\rho(z)$ vary with depth).

3.1.2 Rayleigh waves

In 1887, Lord Rayleigh showed that elastic bodies with free surfaces admit additional solutions to the equations (later named after him as Rayleigh waves) which propagate along the surface and involve motion predominantly confined to the near-surface region (Rayleigh, 1887; Lay and Wallace, 1995). Rayleigh waves (Rw) motion is retrograde elliptical in the vertical plane parallel to the direction of propagation, meaning that particles move both vertically and horizontally, and the sense of the ellipse is counterclockwise as the wave propagates forward (Figure 32). Their motion results from the coupling of compressional (P) and vertically polarized shear (SV) waves, leading to more complex behavior compared to Love waves.

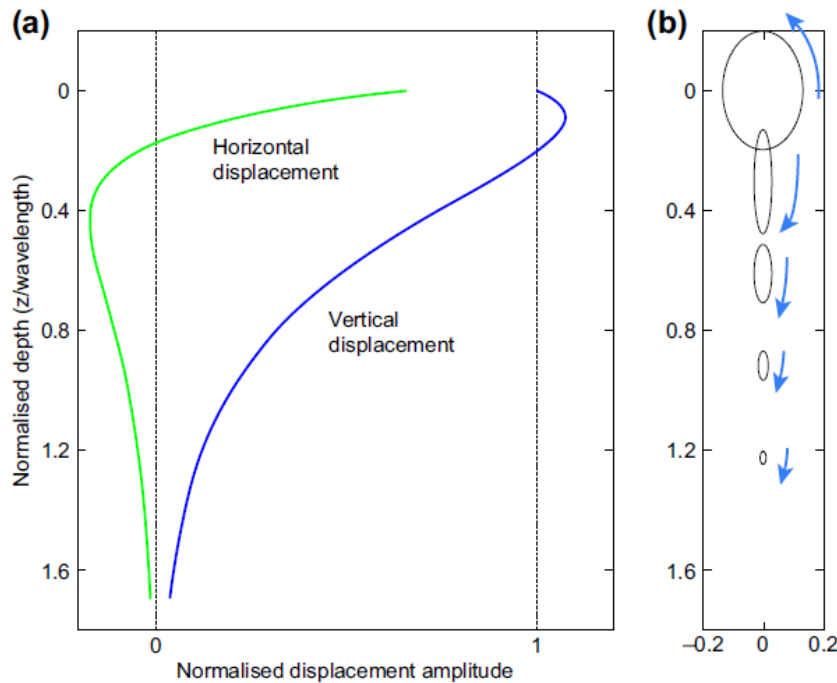


Figure 32: (a) Displacement amplitude of a Rayleigh wave (both vertical – blue curve and radial – green curve components) as a function of depth in an example isotropic medium. (b) Elliptical particle motion at various depths, with each ellipse representing the trajectory of a particle centered at its equilibrium position. The direction of rotation differs with depth; the surface particle (top ellipse) rotates oppositely to those at greater depths, as shown by the blue arrows. Adapted from Gedge and Hill (2012).

Deriving the dispersion relation for Rayleigh waves in a vertically heterogeneous half-space follows a similar approach to that used for Love waves in the previous chapter. However, the fundamental differences lie in the nature of the wave motion and the resulting equations and boundary conditions. For illustration purposes, only selected steps of the derivation are presented here, following (Aki and Richards, (1980), Panza et al. (2001) and Shearer (2019).

For Rayleigh waves, the full elastic wave equation in two dimensions (e.g., the $x - z$ plane, where x is the direction of propagation and z is depth) needs to consider, as the displacement has both horizontal (u) and vertical (w) components:

$$\rho \frac{\partial^2 u}{\partial t^2} = \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xz}}{\partial z} \quad (3.7)$$

$$\rho \frac{\partial^2 w}{\partial t^2} = \frac{\partial \sigma_{zx}}{\partial x} + \frac{\partial \sigma_{zz}}{\partial z} \quad (3.8)$$

where σ_{ij} are the stress components, related to the strains through Hooke's Law involving the depth-dependent Lamé parameters $\lambda(z)$ and $\mu(z)$. This contrasts with the single equation for the y -component of displacement used for SH waves in case of Love waves.

When assuming a harmonic wave solution propagating in the x -direction with $x - z$ plane oscillation:

$$u(x, z, t) = U(z)e^{i(kx - \omega t)} \quad (3.9)$$

$$w(x, z, t) = W(z)e^{i(kx - \omega t)} \quad (3.10)$$

where $V(z)$ and $W(z)$ are the depth-dependent horizontal and vertical amplitudes, respectively. This is more complex than the single displacement component $\beta(z)$ for Love waves.

Boundary Conditions: At the free surface ($z = 0$), both the normal stress (σ_{zz}) and the shear stress (σ_{xz}) must be zero:

$$\sigma_{zz}|_{z=0} = (\lambda + 2\mu) \frac{\partial w}{\partial z} + \lambda \frac{\partial u}{\partial x} \Big|_{z=0} = 0 \quad (3.11)$$

$$\sigma_{xz}|_{z=0} = (\lambda + 2\mu) \frac{\partial w}{\partial z} + \lambda \frac{\partial u}{\partial x} \Big|_{z=0} = 0 \quad (3.12)$$

This is different from the single free surface condition ($\frac{dV}{dz} \Big|_{z=0} = 0$) for Love waves (see the previous chapter). Next, a radiation condition ($U(z), W(z) \rightarrow 0$ as $z \rightarrow \infty$) is applied to depend on the wave energy to near-surface.

Substituting the wave solutions into the equations of motion and using the stress-strain relationships will result in a system of two coupled second-order ordinary differential equations for $U(z)$ and $W(z)$ with depth-dependent coefficients involving $\lambda(z)$ and $\mu(z)$. For a homogeneous half-space, solving this system under the free surface boundary conditions and the radiation condition leads to the Rayleigh wave dispersion equation:

$$\left(2 - \frac{c^2}{V_s^2}\right)^2 - 4\sqrt{1 - \frac{c^2}{V_p^2}}\sqrt{1 - \frac{c^2}{V_s^2}} = 0 \quad (3.13)$$

where c is the Rayleigh wave phase velocity, $V_p = \sqrt{\frac{\lambda+2\mu}{\rho}}$ is the P-wave velocity, and $V_s = \sqrt{\frac{\mu}{\rho}}$ is the S-wave velocity. This equation is a relationship between the Rayleigh wave phase velocity c and the material properties (V_p and V_s). For a homogeneous, isotropic elastic half-space, this equation provides a single, non-dispersive phase velocity c_R , which is slightly less than the shear wave velocity V_s (typically $\sim 0.92V_s$).

However, in a vertically heterogeneous half-space (analogous to the Love wave case), where $\lambda(z)$ and $\mu(z)$ vary with depth, the coefficients in the differential equations for $U(z)$ and $W(z)$ become functions of depth. Solving the boundary value problem then leads to a dispersion relation $D(\omega, c) = 0$, implying that the Rayleigh wave phase velocity c becomes dependent on the frequency ω . Similar to Love waves, this dispersion arises because different frequencies sample the depth varying velocity structure to different points. Lower frequencies with longer wavelengths penetrate deeper and are influenced by the material properties at greater depths, while higher frequencies with shorter wavelengths are more sensitive to the near-surface layers (Aki and Richards, 1980; Foti et al., 2014).

In a heterogeneous medium, Rayleigh wave modes of propagation are observed: a fundamental mode and higher modes. Each mode can be represented by its own dispersion curve, plotting the phase velocity as a function of frequency. The fundamental mode typically has the strongest amplitude. Higher modes exist above certain cut-off frequencies and exhibit more complex depth-dependent amplitude structures with nodes. The analysis of Rayleigh wave dispersion curves obtained from seismological data is a powerful tool for inferring the near-surface velocity structure of the Earth, providing information about layering, gradients in material properties, and other subsurface features. This is analogous to how Love wave dispersion is used to constrain the shear wave velocity structure, as discussed in the previous chapter. However, Rayleigh waves provide additional constraints due to their sensitivity to both compressional and shear wave velocities (Dal Moro, 2014; Foti et al., 2014).

3.2 Surface Wave Dispersion Analysis

This chapter aims to provide an overview of the most relevant surface wave methods in near-surface applied geophysics, with a primary focus on approaches based on multichannel phase velocity analysis. It reviews the entire process of surface wave analysis, highlighting key aspects, challenges, and advantages of selected approaches through case studies and examples. The discussion is based on personal experience with different datasets and methodologies (e.g., Dovbnich et al., 2022; Mazanec and Valenta, 2023; Dal Moro and Mazanec, 2024; Mazanec et al., 2024; Vackář et al., 2024; Mazanec and Dal Moro, 2025) as well as a broader review.

The essential point to emphasize is that surface-wave analysis can be performed using various techniques, either active (with a controlled source) or passive (relying on ambient noise vibrations). The primary goal is usually to investigate surface-wave propagation to derive phase or group velocity, which is then used to estimate the vertical V_s profile. This chapter does not aim to provide an exhaustive overview of all methods but rather focuses on the most relevant ones, highlighting specific aspects, challenges, and advantages of selected approaches through case studies and examples.

Active methods rely on controlled sources (impact or impulsive) to generate surface waves, typically allowing for higher-frequency content and, consequently, a more detailed investigation of shallow subsurface structures. In contrast, passive methods utilize natural or artificial ambient vibrations (Aki, 1957), also referred seismic microtremor noise, which predominantly exhibit lower frequencies, as illustrated in Recent research and near-surface applications predominantly rely on Rayleigh and Love-wave phase-velocity dispersion for estimating S-wave velocity, with only a few studies utilizing group-velocity dispersion (e.g., Long and Kocaoglu, 2001; Pedersen et al., 2003; Tillmann, 2005; Liner et al., 2009; Luo et al., 2011; Dal Moro and Mazanec, 2024). This preference for phase velocity analysis in applied geophysics likely arises from the fact that most surface-wave processing techniques focus on dispersive energy imaging in the phase velocity domain, as it directly corresponds to a specific wavelength, making it easier to link with subsurface properties, particularly the V_s profile (McMechan and Yedlin, 1981; Yilmaz and Doherty, 1987; Park et al., 1998; Song and Gu, 2007).

Table 5: >1 Hz for anthropogenic noise and < 1 Hz for coastal and meteorological sources (Bonnefoy-Claudet et al., 2006; Hayashi et al., 2022). The difference in frequency content directly influences the depth range that can be effectively explored, with active methods generally suited for near-surface studies and passive approaches providing insights into deeper subsurface conditions.

Recent research and near-surface applications predominantly rely on Rayleigh and Love-wave phase-velocity dispersion for estimating S-wave velocity, with only a few studies utilizing group-velocity dispersion (e.g., Long and Kocaoglu, 2001; Pedersen et al., 2003; Tillmann, 2005; Liner et al., 2009; Luo et al., 2011; Dal Moro and Mazanec, 2024). This preference for phase velocity analysis in applied geophysics likely arises from the fact that most surface-wave processing techniques focus on dispersive energy imaging in the phase velocity domain, as it directly corresponds to a specific wavelength, making it easier to link with subsurface properties, particularly the V_s profile (McMechan and Yedlin, 1981; Yilmaz and Doherty, 1987; Park et al., 1998; Song and Gu, 2007).

Table 5: Overview of ambient noise sources as a function of frequency according to various studies. Adapted from Bonnefoy-Claudet et al. (2006).

	Gutenberg (1958)	Asten (1978), Asten and Henstridge (1984)
Ocean waves impacting coastal areas	0.05–0.1 Hz	0.5–1.2 Hz
Large scale meteorological perturbations	0.1–0.25 Hz	0.16–0.5 Hz
Oceanic cyclones	0.3–1 Hz	0.5–3 Hz
Local scale meteorological conditions	1.4–5 Hz	
Volcanic tremor	2–10 Hz	
Antropogenic	1–100 Hz	1.4–30 Hz

In contrast, group velocity depends on the dispersion characteristics across a broader spectral range, making its interpretation more complex as it is influenced by energy distribution over multiple modes. Even though group velocity analysis is classic approach in seismology (Ritzwoller and Levshin, 1998; Levshin et al., 1999; Cotte and Laske, 2002; Kolínský and Brokešová, 2007; Kolínský et al., 2016), the limited adoption of group velocity analysis for near surface applications may be also attributed to a lack of cross-disciplinary exchange between applied geophysicists and seismologists (Dal Moro and Mazanec, 2024).

Although individual approaches to surface wave dispersion analysis differ, multichannel analysis using active or passive data generally consists of the following steps (Figure 33):

- (i) Multichannel field data acquisition
- (ii) Dispersion analysis, in which the acquired t-x (time-distance) records are transformed into a phase velocity spectrum and an experimental dispersion curve is estimated
- (iii) Inversion analysis, where the experimental curve is inverted to obtain the VS profile.

However, these steps are independent, meaning that data can be acquired using e.g., MASW but subsequently analyzed by various methods to obtain velocity spectra (Dal Moro, 2014).

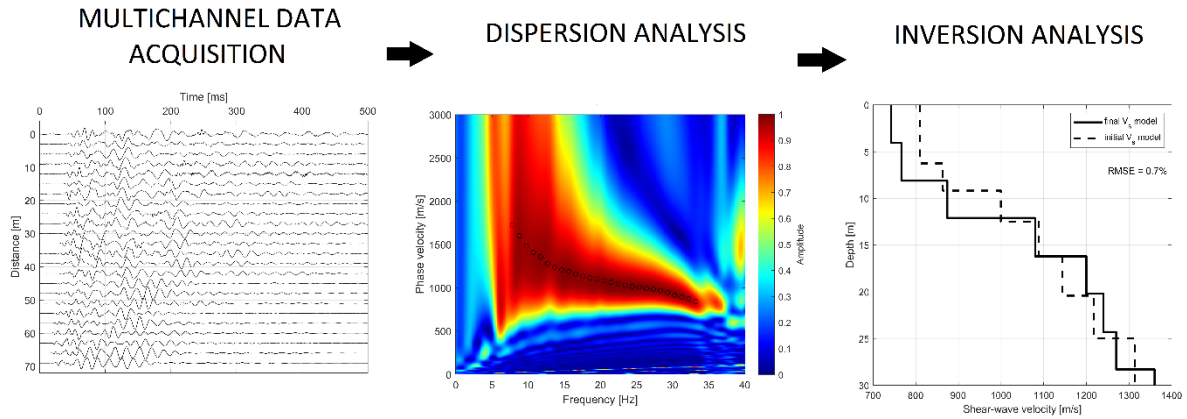


Figure 33: Dispersion analysis workflow: on the left - multichannel shot gather record (source position 81 m); in the middle - Rayleigh wave dispersion image with picked dispersion curve (black circles); and on the right - final (bold black) and initial (black dashed line) Vs models with estimated RMSE. Adapted from Mazanec et al. (2024).

Data Acquisition

This chapter reviews and discusses recommended field procedures and acquisition parameters for multichannel analysis (Figure 34). While most key parameters allow some flexibility, the following guidelines provide an initial orientation based on personal experience and several independent studies. Comprehensive background information can be found, for example, in the following studies: (Park et al., 1999; Socco and Strobbia, 2004; Park and Carnevale, 2010; Dal Moro, 2014, 2020a; Foti et al., 2014, 2018; Ólafsdóttir, 2016; Hayashi et al., 2022; Mazanec and Valenta, 2023; Dal Moro and Mazanec, 2024). A proper understanding of the geological context, initial test measurements, and adjustments to acquisition settings are essential for reliable data collection.

The active MASW approach is often preferred for the shallow applications such as site characterization, particularly due to:

- (1) its ability to control the source offset (Figure 38). Since the sources generating microtremors are typically unknown, determining their propagation direction and accurately estimating the phase velocities of the surface waves present in them can be more challenging than in active-source analysis (Hayashi et al., 2022),
- (2) its principal simplicity, logistical, instrumental and data compatibility with seismic refraction, and the availability of processing software (Sarkar et al., 2021; Mazanec and Valenta, 2023).



Figure 34: Illustrative setup of a Geometrics seismic array, including horizontal, vertical, and 3C geophones, seismic cables, Geodes, sledgehammers, steel plates, trigger cables, a notebook, GNSS receiver, and additional accessories. This configuration can be used for seismic refraction, reflection, and most types of surface wave analysis.

In the MASW approach the most commonly analyzed frequency range is 5–50 Hz (Park and Carnevale, 2010), typically corresponding to an exploration depth of ~0–30 m. However, the actual penetration depth depends on acquisition parameters and the subsurface properties. To achieve greater depth penetration, ambient noise vibration analysis is preferred. Passive methods – generally referred as microtremor array measurements (MAM) approach enables the recording of lower frequency components (<5 Hz; Park et al., 2005), thereby extending the investigation depth.

The optimal approach, as discussed later in Chapter 3.2.2, combines both passive and active methods, extending the analyzable frequency range and number of observables, reducing the non-uniqueness of the solution (e.g., Dal Moro, 2020; Dal Moro and Mazanec, 2024).

3.2.1.1 Seismic receivers

Both active and passive data acquisition is performed using a multichannel array equipped with low-frequency receivers. These receivers primarily differ in:

- (1) the number of components (typically one or three),
- (2) their orientation (vertical or horizontal),
- (3) frequency range,
- (4) cable/wireless connection.

Due to its availability and applicability to refraction seismics the most common active data approach utilizes vertical 1C (one-component) geophones to record the vertical component of Rayleigh waves. To capture the radial component of Rayleigh waves, horizontally oriented geophones aligned with the array are required. For Love wave acquisition, horizontal geophones oriented perpendicular to the array are necessary. The most effective approach is to use three-component (3C) geophones, allowing simultaneous acquisition of all components and thus both Love and Rayleigh waves.

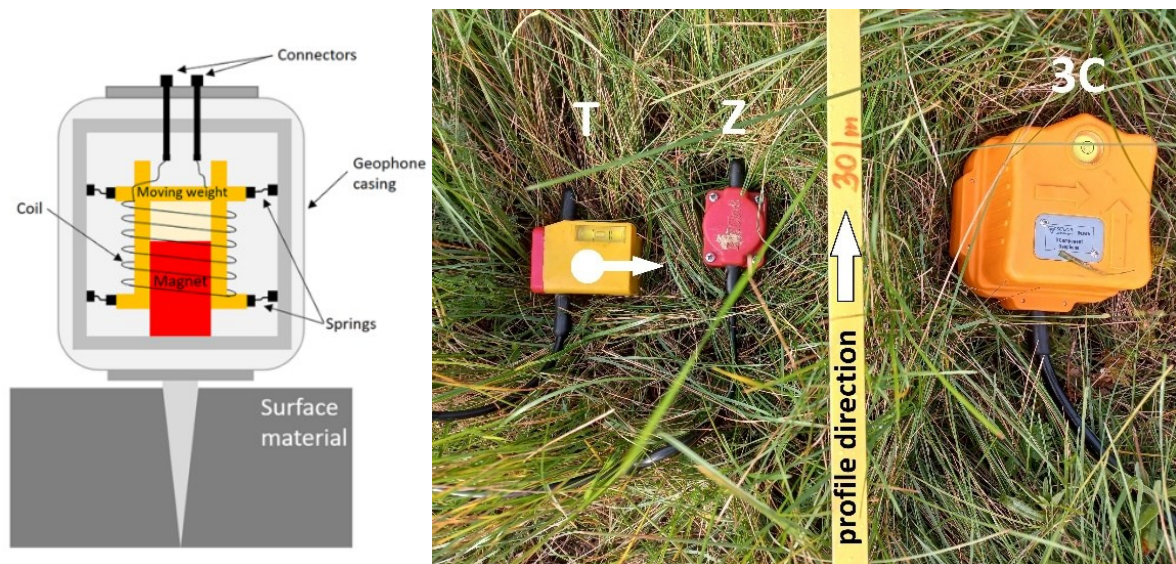


Figure 35: Schematic diagram of geophone body (left panel). A geophone operates on the principle of electromagnetic induction, where ground motion causes relative movement between a suspended coil and a fixed magnet. This relative displacement induces a voltage in the coil proportional to the velocity of ground motion, enabling the detection and recording of seismic waves. Adapted from Apeh et al. (2024); right panel: field picture of 1C horizontal (T component, yellow) and vertical (Z, red) geophones along with the 3C receiver (orange).

Geophones are typically characterized by their natural frequency (eigenfrequency), which represents the frequency at which the internal mass oscillates naturally and below which the sensitivity decreases (Figure 36). For surface-wave studies, the low frequency geophones (4.5 Hz or lower) are generally recommended to capture the long-wavelength, deep penetrating components of the wavefield, see Table 6. However, as demonstrated by Mazanec and Valenta (2023) and in Chapter 3.4.1, geophones can still detect signals at frequencies lower than their eigenfrequency. Additionally, Dal Moro (2014) noted that a commonly used rule of thumb suggests a practical “visibility limit” at approximately half the nominal eigenfrequency.

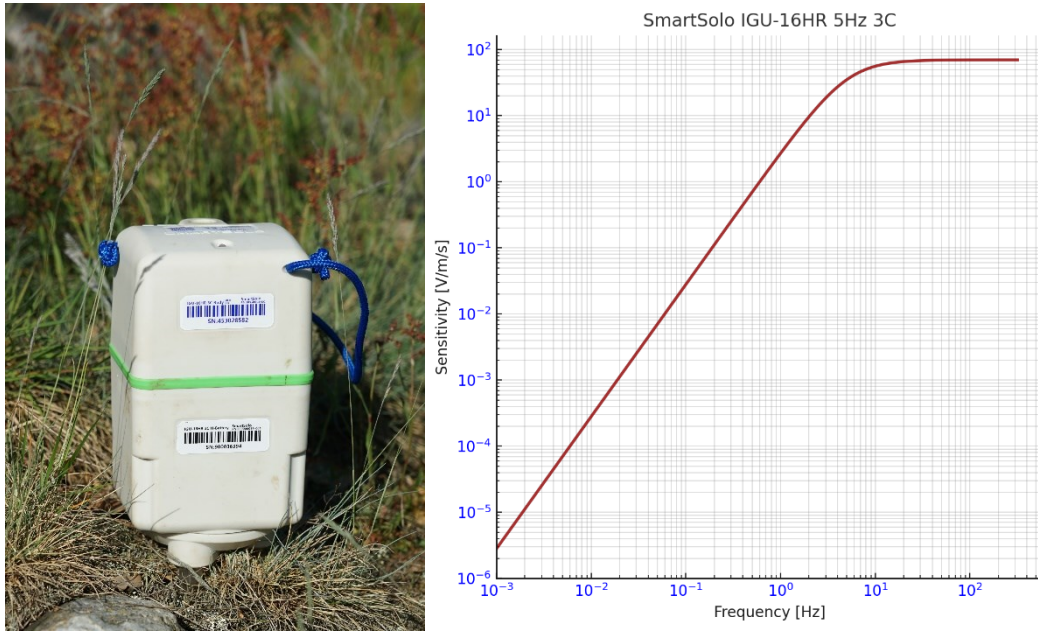


Figure 36: On the left: all-in-one SmartSolo IGU-16HR 5 Hz 3C as an example of a seismic node; b) its response curve, illustrating sensitivity across different frequencies.

A variety of acquisition systems are available for surface-wave surveys, ranging from conventional cabled arrays to advanced wireless units and broadband installations. The selection of a suitable system depends on the survey objectives (e.g. active or passive acquisition), target depth, required resolution, and logistical conditions in the field.

Cabled multi-channel systems are widely used for both active and passive measurements. These systems typically employ single component geophones (commonly 4.5 Hz) connected via analog or digital cables to a centralized recorder. They provide high data fidelity and precise timing synchronization, making them suitable for controlled-source methods such as MASW. In passive surveys, cabled arrays are advantageous for short-aperture configurations where portability is less critical. Triaxial (3C) geophones in cabled configurations allow for multicomponent acquisition, enabling the recording of both Rayleigh and Love waves, and facilitating joint inversion strategies.

Table 6: Typical data acquisition systems and investigation depths for unconsolidated or weakly consolidated sediments. Adapted from Foti et al. (2018) and Hayashi et al. (2022).

<i>Acquisition System</i>	<i>Sensor Type</i>	<i>Natural Frequency</i>	<i>Effective Frequency Range</i>	<i>Approx. Maximum Investigation Depth</i>
Multi-channel wired system/nodes	Geophone	≥ 4.5 Hz	> 1 Hz	~ 100 m
Wireless nodal system	Geophone (low frequency)	1–2 Hz	> 0.2 Hz	~ 1000 m
Wireless or broadband	Broadband velocity sensor	≥ 5 s (≈ 0.2 Hz)	> 0.1 Hz	~ 2000 m

In contrast, nodal all-in-one systems (Figure 36) have gained importance in both research and applied geophysics due to their ease of deployment and flexibility. These self-contained units integrate three-component sensors, GPS timing, onboard digitizers, and battery power, allowing fully autonomous, cable-free operation. They are particularly well suited for passive ambient vibration studies and are increasingly used to supplement active-source campaigns. Many nodal systems are equipped with low-frequency geophones (e.g. 2 Hz), enabling easy deployment of large arrays (Hayashi et al., 2022).

For applications requiring high sensitivity and long-period recording - such as deep basin characterization and teleseismic analysis, broadband sensors are employed. These instruments offer flat velocity response over a wide frequency range (e.g., 0.1–50 Hz or more) and are primarily associated with passive methods.

3.2.1.2 Array geometry

The geometry of a seismic array plays a critical role in the successful acquisition and interpretation of surface-wave data. The spatial configuration of receivers directly influences the frequency content, resolution, and azimuthal coverage of the recorded wavefield. While both active and passive methods rely on the dispersion of surface waves to infer subsurface shear-wave velocity profiles, their requirements for array geometry differ significantly due to the nature of the source and wavefield.

For active-source data, a linear array (Figure 37) is typically employed, as it facilitates controlled generation and acquisition of surface waves propagating perpendicular to the array (Park et al., 1999). This configuration is particularly effective in methods such as MASW, where the source-receiver geometry ensures directional wavefields and simplifies dispersion analysis. The linear geometry simplifies data processing steps such as the generation of dispersion images and velocity spectrum analysis.

In contrast to the directional and controlled nature of active source wavefields, passive surveys rely on ambient seismic noise, which is typically stochastic, multi-directional, and unpolarized. Therefore, nonlinear (or two-dimensional) geometries such as circular, triangular, or L-shaped arrays (Figure 37), are preferred to ensure adequate azimuthal averaging (Aki, 1957; Okada and Suto, 2003; Hayashi et al., 2022). However, linear arrays can still be used in passive surveys if the dominant wavefield direction is known, but they may introduce directional biases (Foti et al., 2018). The choice of array geometry should therefore be carefully considered based on the expected wavefield characteristics and the study's objectives.

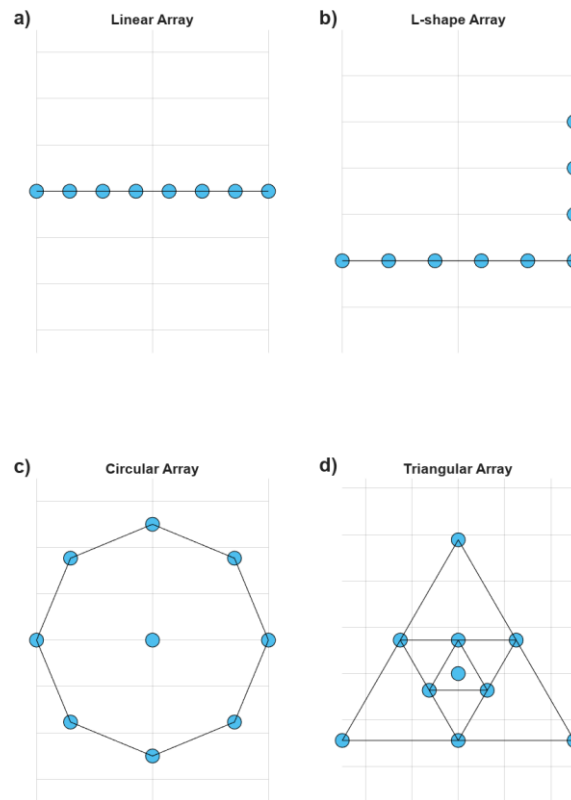


Figure 37: Illustration of the most common array geometries in surface-wave array acquisition.

From a practical perspective, as highlighted in Dal Moro and Mazanec (2024) linear arrays offer a distinct advantage in the analysis of the transverse (T) component of the wavefield, as they allow for the straightforward acquisition of Love-wave dispersion curves using standard horizontal component geophones. Such arrays enable the use of both active and passive seismic data (Dal Moro, 2014) for T-component analysis. In contrast, for 2D array geometries, the extraction of Love-wave dispersion typically requires three-component (3C) sensors, as the transverse component must be computed between sensor pairs with known azimuthal orientation (Okada and Suto, 2003).

3.2.1.3 Active data acquisition parameters

The maximum depth of investigation (Z_{max}) depends on site conditions, the eigenfrequency of the geophones, and the type of seismic source used. It is primarily determined by the maximum analyzable wavelength of surface waves λ_{max} (Figure 38) following a commonly adopted empirical criterion (Park and Carnevale, 2010):

$$Z_{max} = \frac{\lambda_{max}}{2} \quad (3.14)$$

and by the length of the geophone array (D), which is directly related to the longest possible analysed wavelength λ_{max} :

$$D = \lambda_{max} \quad (3.15)$$

and therefor, the Z_{max} should be considered not larger than half of the geophone array:

$$Z_{max} = \frac{D}{2} \quad (3.16)$$

However, some sources (e.g., Hayashi et al., 2022) suggest a practical relationship in which the λ_{max} is approximately 2 to 4 times the D .

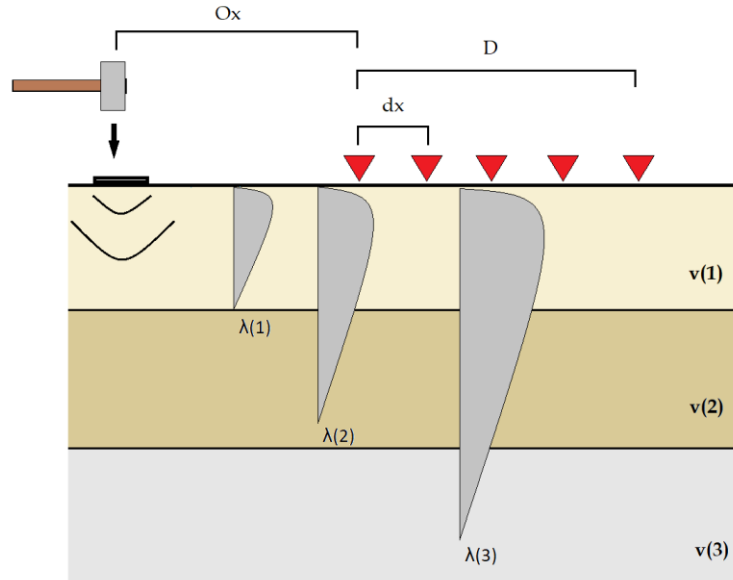


Figure 38: Multichannel acquisition scheme showing active seismic source (a sledgehammer), geophones (red triangles), source offset (Ox), length of the receiver spread (D) and receiver spacing (dx). Different wavelengths ($\lambda(x)$) of Rayleigh waves propagate from the source through a layered medium (where each layer has a different velocity $v(x)$) with a different depth of penetration. Adapted from Mazanec and Valenta (2023).

After deploying the geophone array, it is recommended to perform preliminary evaluation of the lowest frequency of surface waves recorded to verify the achievement of the required depth penetration (Dal Moro, 2014).

In active multichannel acquisition, the profile length is also influenced by the energy of the seismic source and its offset (O_x , Figure 38), as surface waves generated by the source can become attenuated and contaminated by body waves below the noise level. These so-called far-field effects are typically observed in the high-frequency segment of the seismic record. Initially, they appear in the far-offset traces but become noticeable in near-offset traces as well as the frequency increases. Additionally, large values of O_x and D increase the risk of higher-mode dominance and reduce the noise-to-signal ratio (NSR) of the fundamental mode.

Near-field effects occur when a seismic source is positioned too close to the receiver array, causing significant signal distortion and resulting in altered waveforms. This phenomenon is often observed in the low-frequency part of multichannel swept-frequency records as a lack of linear phase coherence. To mitigate these effects, the source offset must be sufficiently large to ensure the recorded signal accurately represents the propagating wavefield (Roy and Jakka, 2017).

Near-field and far-field effects can be reduced by optimizing data acquisition parameters and selecting an appropriate test configuration. Based on acquired experience, the most effective approach has proven to be performing multiple shots with varying offsets. This practice allows for optimizing the recorded wavefield, mitigating near-field and far-field effects, can reveal lateral variations and improving the reliability of the dispersion analysis. To suppress incoherent noise, it is recommended to perform multiple shots at each offset position separately, followed by stacking (i.e., summation of traces from multiple shots), such as Non-linear stacking (Vilhelm et al., 2022).

In active multichannel acquisition, surface waves are typically generated by striking an impact plate (metallic or non-metallic) with a 5–10 kg sledgehammer. A heavier seismic source capable of delivering greater impact energy can generate lower-frequency surface waves with longer wavelengths, potentially extending the depth of investigation.

Receiver spacing (d_x) is related to the shortest wavelength (λ_{min}) and therefore determines the shallowest resolvable depth of investigation (Z_{min}):

$$Z_{min} = k d_x \quad (0.3 \leq k \leq 1) \quad (3.17)$$

In practice, receiver spacing is often related to the total length of the receiver spread, as standard geophones are typically connected to a seismic cable that generally allows a maximum spacing of 5 m for a standard 24-channel configuration (the aforementioned does not apply to autonomous receivers). However, Dal Moro (2014) demonstrates that the number of channels is not a critical parameter and should be maintained between 12 and 24, see Figure 39. Consequently, a lower number of geophones can potentially allow for a larger receiver

spacing. However, it is not beneficial to increase the number of geophones without lengthening the receiver spread (Park et al., 2001).

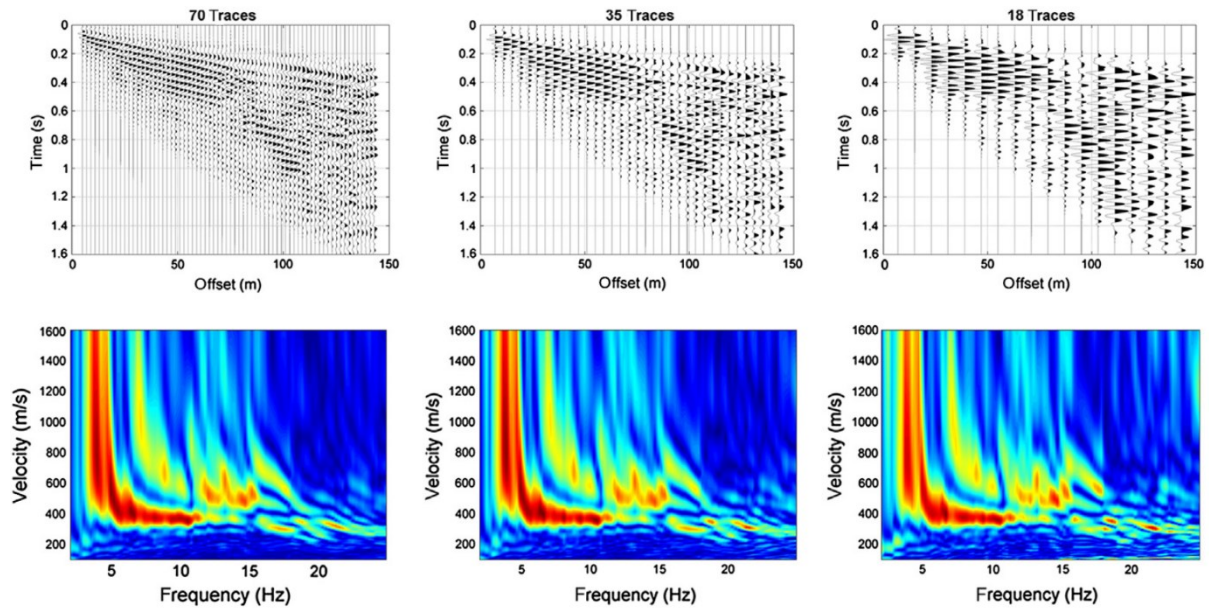


Figure 39: Illustration of the influence of geophone number on the resulting dispersion velocity spectra in Multichannel Analysis of Surface Waves (MASW). On the left column, the original 70 trace dataset and the corresponding phase-velocity spectrum; in the middle, the data after removing half of the original traces (35); on the right, the data after further removing half of the traces (18) showing no significant difference in the velocity spectra. Adapted from Dal Moro (2014).

The typical recording duration (T) ranges from 1 to 2 seconds. It is essential to capture the complete surface wave signal across all channels. Figure 40 illustrates an example of insufficient recording duration (1 s) for active MASW analysis of Love waves. For near-surface geological investigations, a sampling interval of 1ms (sampling rate 1000 Hz) is generally sufficient. Its role is determined by the Nyquist-Shannon sampling theorem: for the sampling frequency (f_s) of a time series, the maximum analyzable frequency is $f_s/2$ ms (for a sampling interval of 1 ms, the corresponding Nyquist frequency is 500 Hz). Consequently, to accurately analyze dispersion up to a given frequency, the data must be sampled at a rate of at least twice that frequency.

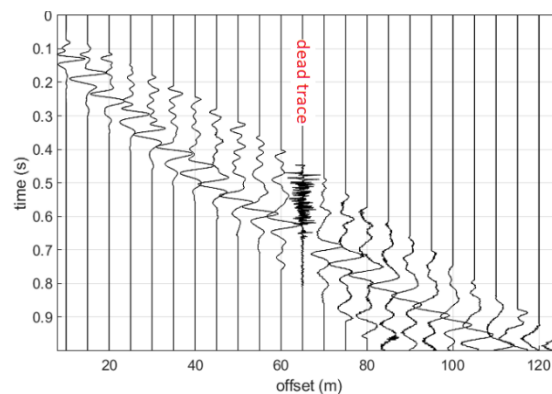


Figure 40: Real-data example illustrating insufficient recording duration from the soft soil site in Litoměřice, Czechia. In this case, the 1 s recording window failed to capture the complete propagation of Love waves across the entire profile. Note the dead trace at the 65 m offset, indicating a technical malfunction or faulty connection.

3.2.1.4 Ambient vibrations acquisition parameters

The acquisition parameters used in passive surveys are in many respects identical or closely analogous to those employed in active-source data acquisition, as discussed above. However, fundamental differences arise due to the inherently distinct nature of the seismic sources involved, while active surveys rely on controlled, point-source energy input with defined geometry, passive surveys are governed by uncontrolled, spatially distributed sources of ambient noise, resulting in different requirements for array configuration, acquisition duration, and signal processing strategies. A more detailed review can be found in Bonnefoy-Claudet et al. (2006), Hayashi et al. (2022), Foti et al. (2018) and Dal Moro (2014), who discuss the implications of source characteristics on array design, data quality, and the passive surface wave analysis.

Proper deployment of sensors during passive ambient vibration measurements is critical to ensure data quality and accurate site characterization. As illustrated in Figure 41, sensors should be installed directly on or slightly below the ground surface in locations that reflect in situ conditions. Good coupling between the geophone and the ground is essential, which often requires removing vegetation and minimalizing external influences (Foti et al., 2018; Molnar et al., 2022). Environmental factors such as wind, rain, and temperature fluctuations must be mitigated using appropriate shielding, and avoiding measurements during hostile weather is recommended to reduce unwanted noise (Chatelain et al., 2008; Molnar et al., 2022).

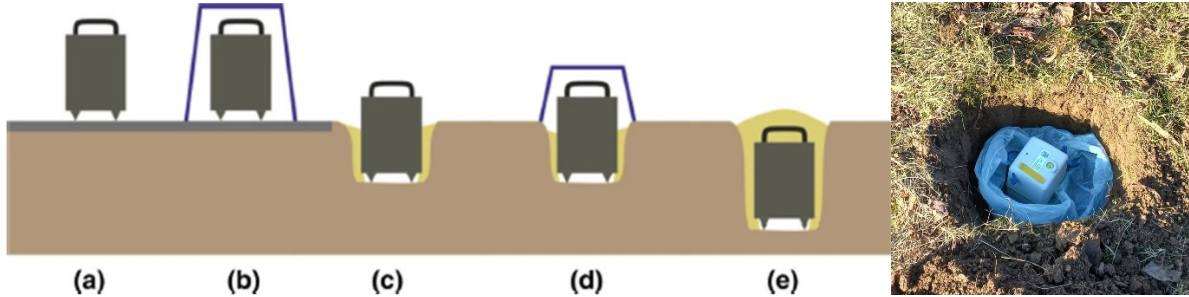


Figure 41: On the left - various seismometer deployment configurations for microtremor measurements, ordered from least (a) to most suitable (e) for high-quality data acquisition. Due to the typically short recording duration, setups with infill deployment (c, e) are uncommon; most measurements are conducted using configurations a, b, or d without infill. Adapted from Foti et al. (2018). The photo on the right shows an example of configuration (e) prior to being backfilled.

As in active-source acquisition, the maximum investigation depth is typically considered to be between one-half and one-quarter of the maximum resolvable wavelength ($\lambda_{\max}$). The minimum receiver spacing determines the shortest wavelengths that can be reliably sampled. In pure passive surveys, where low-frequency content dominates and sensor spacing is often large, high-frequency (short-wavelength) information may be underrepresented or aliased, limiting near-surface resolution. To overcome this, high-resolution data from active sources or mini-arrays such as Miniature Array Analysis of Microtremors (MAAM, Cho et al., 2006) are often integrated into the survey design (Foti et al., 2018).

Recording duration must be adapted to the desired frequency range (Table 7, Hayashi et al. 2022). In very general terms, for large arrays (>100 m), 30–60 minutes is typical. For deep surveys (>1 km) or low-noise sites, 2–4 hours may be necessary. Sampling frequency is typically set to 100–200 Hz for passive data, ensuring coverage of both low and mid-frequency content (Dal Moro, 2014; Foti et al., 2018; Hayashi et al., 2022).

Table 7: Relation between recording length and minimal target frequency according to Hayashi et al. (2022) and Dal Moro (2014).

Target Frequency	Minimum Window (s)	Minimum Recording (min)
0.2 Hz	50	30–60
1 Hz	10	10–20
>5 Hz	5	2–5

Recording ambient vibrations involves certain risks associated with the uncertainty in the location, distribution, and characteristics of the source. SPAC and similar methods assume ambient waves behave as plane waves propagating across the array. This assumption breaks down if the noise sources are too close or inside the array (e.g., traffic, industry), or sources are strongly directional or non-stationary. To improve isotropy and low-frequency content, recordings are often done at night, especially in urban areas (Hayashi et al., 2022). Coherency and ambient wavefield quality check in the field immediately after acquisition is essential.

3.2.2 Dispersion Analysis

Once the multichannel waveform field data are recorded and pre-processed (including e.g., stacking), the next step is the surface wave dispersion analysis. A key objective of dispersion analysis is to derive information on the geometric dispersion of surface waves from full multichannel waveform field records. The results are represented as dispersion curves, illustrating the relationship between surface wave phase or group velocity and frequency or wavelength (Foti et al., 2014). The following sections will focus on phase velocity analysis, as it is the most commonly used approach. A brief discussion of group velocity analysis is provided in Chapter 3.3.1.3. The processing of dispersion analysis consists of two main steps. First, a dispersion image (Figure 33b) is generated from the acquired seismic records (Figure 33) using a wavefield transformation method. In the second step, the most representative experimental dispersion curve(s) (Figure 33c) are extracted from the dispersion image.

3.2.2.1 Dispersion Image

Various surface wave analysis methods employ different dispersion analysis techniques, with the Fourier transform being fundamental for decomposing the time history of vertical particle motion into its frequency components. This decomposition allows for the isolation of

individual frequency components, which are then analyzed to determine phase velocity. Many of these analysis methods are either formally identical or linearly dependent, yielding identical results when applied to synthetic surface-wave data. Therefore, comparisons should generally focus on aspects such as sensitivity to noise, stability, robustness, ability to extract and distinguish different events, as well as the effectiveness of filtering and weighting the information (Socco and Strobba, 2004). The specific approach depends on the acquisition geometry and the number of receivers (Foti et al., 2014).

The most common transform methods for determining phase velocities include the (a) frequency - wavenumber (f - k) transform, which applies a double Fourier transform to the field data (Yilmaz and Doherty, 1987); (b) the intercept time - slowness (τ - p) transform (McMechan and Yedlin, 1981), which combines a slant-stack transformation of recorded wavefield with the summation of power spectra for each frequency; and (c) the phase shift transform (Park et al., 1998) , which integrates elements of both the slant-stack and f - k methods (Pelekis and Athanasopoulos, 2011).

3.2.2.1.1 Frequency–Wavenumber ($f - k$) transform

The $f - k$ transform (Nolet and Panza, 1976; Yilmaz and Doherty, 1987) is a natural widely recognized technique for analyzing surface waves as it corresponds to the physical nature of their propagation. It transforms the wavefield $U(t, x)$ - recorded data from time-space, into the frequency–wavenumber $F(f, k)$ domain using a 2D Fourier transform:

$$F(f, k) = \iint_{-\infty}^{+\infty} U(x, t) e^{-2\pi i(ft+kx)} dx dt \quad (3.18)$$

In an ideal case, surface wave energy is concentrated along specific lines in the f - k domain, reflecting their dispersion characteristics. However, real data exhibit energy spreading due to observation effects, and noise, both random and coherent, can distort the results. To enhance the identification of energy maxima, preprocessing steps such as muting, filtering, and windowing in the t - x domain are applied. Zero padding is also used to improve spectral resolution, and different windows help distinguish between ripples and higher modes. Once a clear image of energy density is obtained, an automatic search for maxima is performed. The phase velocity (c) is then determined using the relation $c = \frac{f}{k}$, by identifying the wavenumber with maximum spectral energy at each frequency as shown in Figure 42 (Gabriels et al., 1987; Foti, 2000; Socco and Strobba, 2004).

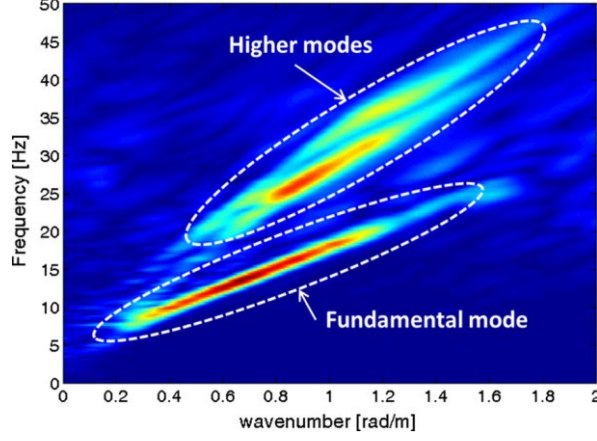


Figure 42: Example of a frequency–wavenumber (f – k) spectrum on the real data from the Mirandola site (InterPACIFIC project), showing the dispersion curves of the fundamental mode and two higher modes visible as red and yellow high amplitudes. Adapted from Foti et al. (2018).

The f – k transform is a fast and reliable method, particularly effective when the wavefield is dominated by a single direction of propagation and signal-to-noise ratio is high. However, it assumes a plane wavefront and may suffer in near-field conditions or when the spatial sampling is insufficient, leading to aliasing and model misfit (Rahimi et al., 2021).

3.2.2.1.2 Slant stack (τ – p) transform

The slant stack, also known as the τ – p transform (McMechan and Yedlin, 1981), is a widely used method for surface-wave analysis that enables the construction of the velocity (or dispersion) spectrum by transforming seismic data from the time–space (t – x) domain into the intercept time–slowness (τ – p) domain. In this context, τ denotes the intercept time, i.e., the time at which a seismic event would be recorded at zero offset, while p represents the slowness (the reciprocal of phase velocity), describing the apparent propagation rate of the wavefront across the receiver array.

The fundamental principle underlying this transform is the linear relationship between travel time t , offset x , slowness p , and intercept time τ :

$$t = \tau + px \quad (3.19)$$

This transformation is achieved in two steps. First, the recorded wavefield $U(x, t)$ is slant stacked, i.e., integrated along lines of constant slowness, thereby mapping the data into the τ – p domain. The objective of this operation is to concentrate coherent seismic energy along trajectories corresponding to specific slowness values, which enhances the identification of propagating wave modes:

$$f(\tau, p) = \int_{-\infty}^{+\infty} U(x, \tau + px) dx \quad (3.20)$$

This transformation effectively decomposes the wavefield into its plane-wave components, allowing wave arrivals that travel at different apparent velocities to be separated based on

their intercept times and slowness. This separation facilitates the identification of distinct surface-wave modes.

In the second step, a 1D Fourier transform is applied with respect to τ , resulting in a representation in the frequency–slowness domain $F(f, p)$:

$$F(f, p) = \int_{-\infty}^{+\infty} f(\tau, p) e^{-i2\pi f\tau} d\tau \quad (3.21)$$

The resulting function $F(f, p)$ expresses the amplitude of energy associated with each frequency–slowness pair and can be interpreted to extract the dispersion spectrum, i.e., the relationship between phase velocity and frequency (McMechan and Yedlin, 1981; Stoffa et al., 1981).

The $\tau - p$ transform is generally robust in homogeneous or quasi-homogeneous wavefields and is capable of resolving multiple surface-wave propagation modes under favorable conditions. However, it has known limitations: it is sensitive to velocity reversals (where deeper low-velocity layers are present), which can hinder the correct reconstruction of deeper portions of the profile. Additionally, it may offer limited resolution for higher modes or in shallow, laterally variable stratigraphy (Rahimi et al., 2021).

3.2.2.1.3 Phase shift

The phase shift method (Park et al., 1998) is a widely adopted transformation technique introduced for Multichannel analysis of surface waves. It is based on determining and compensating for the phase differences between seismic signals recorded at different offsets, enabling the estimation of phase velocity as a function of frequency. Unlike the $f - k$ and $\tau - p$ transforms, the phase shift method isolates the phase content of the signal and is particularly robust against amplitude variations and spatial aliasing (Dal Moro et al., 2003; Socco et al., 2010).

The process begins by decomposing the recorded wavefield $f(t, x)$ through a 1D Fourier transform in time, yielding the frequency–offset domain:

$$U(\omega, x) = \int_{-\infty}^{+\infty} f(t, x) e^{i\omega t} dt \quad (3.22)$$

Angular frequency ω is used here because the method operates directly on the phase of the wavefield, and phase shift is expressed as a function of ωt .

Each frequency component is then normalized in amplitude, both in frequency and offset, to ensure that the resulting phase spectrum remains unaffected by signal strength or near-field effects. This is crucial, as the dispersion characteristics are encoded in the phase, not the amplitude.

Next, a range of trial phase velocities c_{test} is defined. For each ω and c_{test} , a phase compensation term is applied to align the frequency components across all receiver positions:

$$U(\omega, x) = A(\omega, x) \cdot e^{\frac{-i\omega x}{c_{test}}} \quad (3.23)$$

The phase-aligned traces are then summed over all receivers:

$$c(f, c_{test}) = \sum_{i=1}^N \frac{U(f, x)}{|U(f, x)|} \cdot e^{-i2\pi f x / c_{test}} dx \quad (3.24)$$

This summation process is repeated for all frequencies and all trial velocities. The result is a matrix of correlation values (or stacking amplitudes), where high values indicate constructive phase alignment at that specific frequency–velocity pair.

Plotting this matrix yields the final output - the velocity dispersion spectrum, which objectively represents phase velocity as a function of frequency. It is commonly visualized as a color-coded image where warmer colours indicate higher coherence or stacking amplitude. This spectrum is a mathematically derived result, independent of subjective interpretation (Dal Moro, 2014; Ólafsdóttir, 2016).

Among available transformation techniques, the phase shift method is considered highly effective, especially for retrieving the fundamental mode, even with sparse or irregular receiver spacing. Unlike the $f - k$ transform, the phase shift method can tolerate non-uniform sensor geometry and is less sensitive to aliasing artifacts, which, if present, remain visually distinguishable and do not critically hinder interpretation (Dal Moro et al., 2003). However, Rahimi et al. (2021) conducted extensive testing of transformation techniques in MASW under various site conditions and identified limitations in most of them. They concluded that the best practice is to use the composite dispersion approach, which integrates all transformation methods, or at minimum, to apply at least two different transformation techniques.

3.2.2.2 Dispersion curve extraction

The velocity dispersion spectrum contains the complete signal content - comprised of kinematic and energy information of the wavefield. Different modes of surface wave propagation (harmonics) are recognized here by their frequency content and characteristic phase velocity at each frequency. The second step of the dispersion analysis is to extract from the dispersion spectra the most representative dispersion curve(s) (Figure 33). This extracted curve is called an experimental dispersion curve and is an input data to the next (iii) inversion analysis step. Typical dispersion curves exhibit smooth and continuous variations in phase velocity with frequency, without abrupt slope changes or discontinuities. Visual inspection of the experimental dispersion curve can provide valuable insights into the expected shear wave

velocity profile (Foti et al., 2018). Based on the approach used for extracting dispersion curves during data processing, conventional methods can be categorized into fundamental mode (M0) approach, multimodal-curve methods (the last two types are often grouped as modal-curve methods), and effective dispersion curve techniques.

In a normally dispersive medium, where S-wave velocity increases monotonically with depth, the dispersion spectrum typically exhibits energy concentrated along a smooth and continuous “ridge”. In this ridge phase velocity smoothly and continuously decreases with the increasing frequency and represents the fundamental mode, which predominates in the wavefield of such a medium (Figure 43). In contrast, a discontinuity or flat segment in the ridge may suggest the presence of a velocity inversion (inverse layering), such as a softer layer beneath a stiffer one. The experimental fundamental-mode dispersion curve is obtained by selecting spectral maxima along this ridge, revealing a dispersion trend (Figure 43), where velocity decreases with frequency (Dou and Ajo-Franklin, 2014). This approach is for its simplicity the most common (Socco et al., 2010; Olafsdottir et al., 2018; Mazanec and Valenta, 2023; Mazanec et al., 2024).

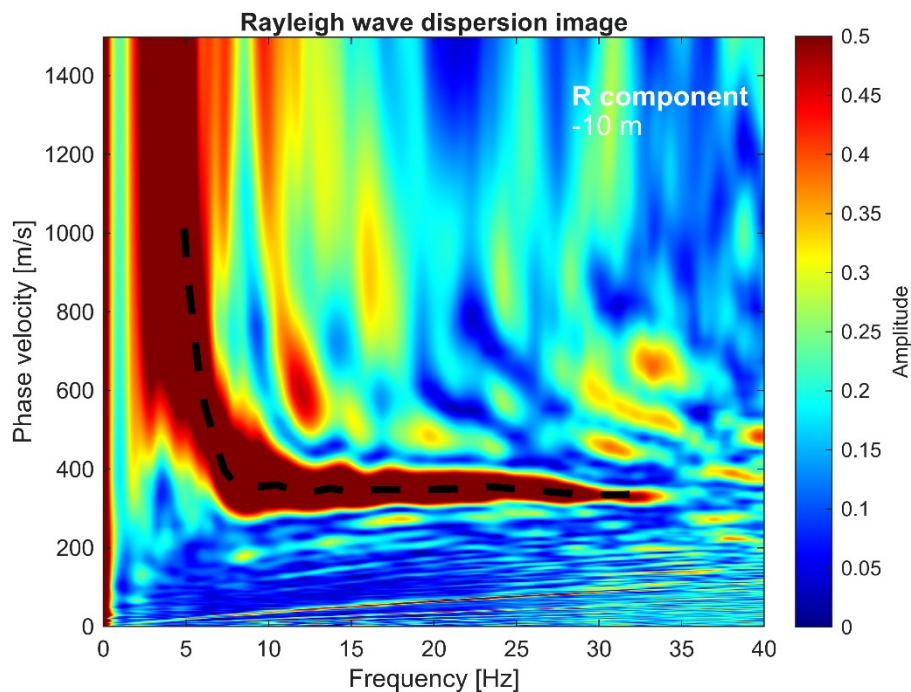


Figure 43: Example of Rayleigh wave dispersion spectra (radial component) with picked fundamental mode (black dashed line).

The number of dispersion modes in the dispersion spectra depends on the complexity of the ground condition (Zhang and Chan, 2003). The multi-modal approach considers the presence of higher modes alongside the fundamental mode (Figure 44), allowing for a more detailed characterization of the subsurface. Extracting higher modes (M1, M2, etc.), which can provide valuable information about deeper layers or velocity inversions (Socco et al., 2010), typically involves identifying multiple spectral ridges in the dispersion image, see Figure 44 and Figure

45, with each ridge corresponding to a distinct propagation mode (Gabriels et al., 1987; Ryden et al., 2004; Song and Gu, 2007; Karray and Lefebvre, 2009).

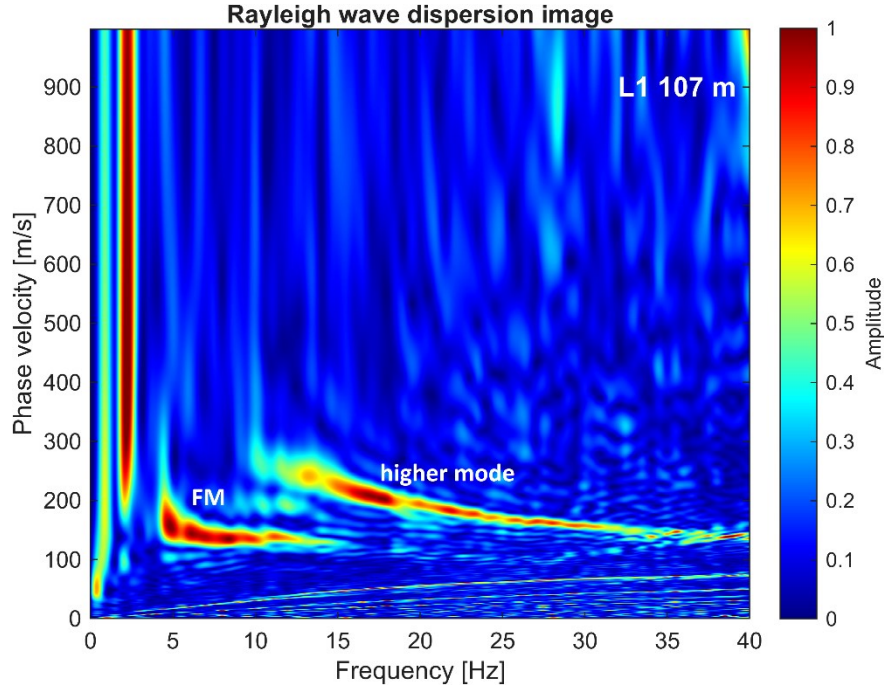


Figure 44: The multimodal approach takes into account presence of higher-modes of propagation, here clear fundamental mode (FM) and dominant higher mode (vertical component of Rayleigh wave).

The effective (or apparent) dispersion curve is a pragmatic representation derived from surface wave data, integrating energy contributions from multiple propagation modes (mode superposition), as mathematically formulated by Lai et al. (2014), Lunedei and Albarello (2009), Tokimatsu et al. (1992) and Ikeda et al. (2012). This curve represents a smoothed or dominant trend extracted from a complex dispersion image potentially containing overlapping modes, noise, or interference. It typically tracks the highest-amplitude energy ridge in the frequency–velocity domain, often aligning with the fundamental mode but subject to influence from contributions from higher modes or mixed propagation effects (Dal Moro, 2014; Foti et al., 2018). The method bypasses mode identification and accounts for acquisition and processing effects; however, it does not fully exploit the energy distribution within the dispersion spectra and instead relies mainly on the kinematic properties of surface-wave datasets (Dou and Ajo-Franklin, 2014).

For calculating the effective dispersion curve $\hat{V}(\omega)$, the following equations from Lunedei and Albarello (2009) can be used:

First the average power $P_n(\omega)$ for each n -th wave propagation mode at the angular frequency ω is computed:

$$P_n(\omega) = \int M [|A_n(\omega; x, y)|^2] dx dy \quad (3.25)$$

where $A_n(\omega; x, y)$ is the spectral amplitude of the n -th wave mode at position (x, y) . M denotes the mean value operator. Next the effective dispersion curve $\hat{V}(\omega)$ as a function of ω can be obtain:

$$\hat{V}(\omega) \equiv \frac{\omega \xi}{\arccos \left[\sum_n \frac{P_n(\omega)}{P(\omega)} \cos \left(\frac{\omega}{c_n(\omega)} \xi \right) \right]} \quad (3.26)$$

where ξ is the distance between geophones, $c_n(\omega)$ is the phase velocity of the n -th wave mode at frequency ω and $P \equiv \sum_n P_n$ is the total average power of all modes.

The effective dispersion-curve method obviates the necessity for explicit mode identification by combining spectral maxima at each frequency into a single curve (Figure 45), which is then used as the target for inversion (e.g., O'Neill et al., 2004). Application of effective dispersion curve can be seen in (Zhang and Chan, 2003; Albarello and Gargani, 2010; Hayashi and Craig, 2017; Foti et al., 2018; Dal Moro, 2020a; Hayashi et al., 2022; Dal Moro and Mazanec, 2024), where is applied to the passive data and briefly discussed.

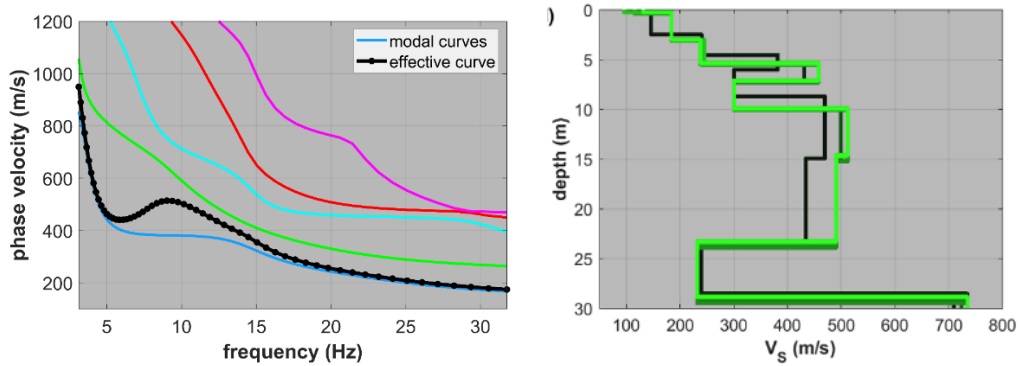


Figure 45: a) comparison between the modal (solid lines) and effective curve (dotted line) for the green model shown in b). Adapted from Dal Moro and Mazanec (2024).

There are also ways how the experimental dispersion curve is replaced by other types of data. These include the full velocity spectrum (FVS) or full-waveform analyses (FWA). These methods a

The Full Velocity Spectrum (FVS) approach (Dal Moro, 2014) preserves the entire dispersion image rather than extracting a single dispersion curve, offering a more comprehensive representation of modal energy distribution (Foti et al., 2018). This method is particularly advantageous in cases involving higher modes or complex subsurface structures, as it eliminates the need for explicit mode identification (Dal Moro, 2014). Instead of interpreting dispersion curves, FVS analyzes the full velocity spectrum through an inversion process that minimizes the misfit between observed and synthetic spectra. Synthetic traces are computed for an initial subsurface model, followed by the calculation of their velocity spectra and comparison with field data. The inversion iteratively refines the model to achieve the best match between observed and synthetic spectra. FVS shares conceptual similarities with full-

waveform inversion but focuses specifically on the wavefield's dispersion characteristics, see Figure 46.

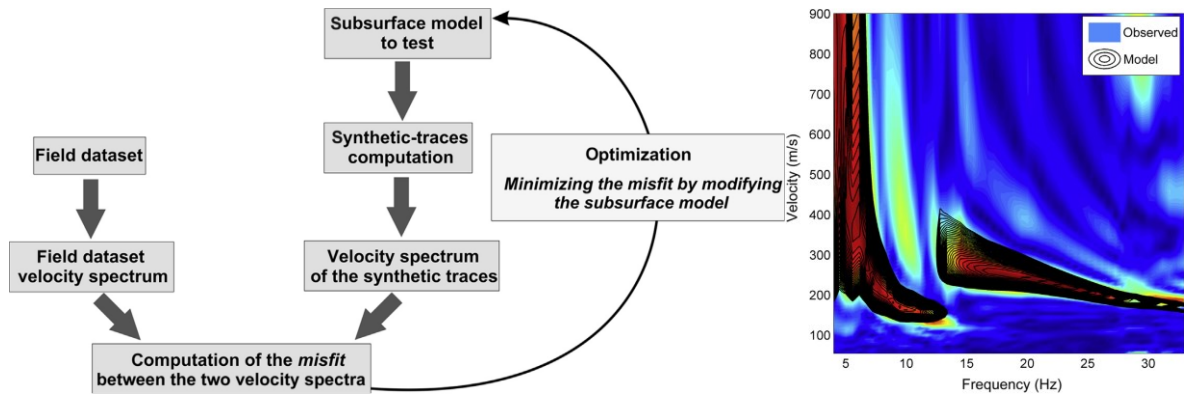


Figure 46: Full Velocity Spectrum (FVS) inversion—schematic representation of the implementation process (left panel). The right panel shows an example of FVS inversion, where the background color indicates the phase-velocity spectrum of the observed field data, and the overlaid black contour lines represent the modeled spectrum of the selected velocity model. Adapted from Dal Moro (2014).

Unlike FVS, which focuses on velocity spectra, full-wavefield inversion (Dou and Ajo-Franklin, 2014) incorporates both phase and amplitude information, allowing for a more detailed reconstruction of subsurface properties. This technique iteratively compares observed wavefields with synthetic ones generated through forward modeling, refining the velocity model to minimize waveform misfit. The approach is particularly effective in environments with complex wave propagation, such as permafrost regions, where higher modes and scattered energy play a significant role. However, similar to other full-waveform techniques, it demands precise source characterization and significant computational resources to achieve robust results.

3.2.2.3 Few remarks regarding the dispersion analysis

Dispersion analysis presents several challenges and potential risks that must be carefully considered during the analysis. First, although the dispersion spectrum is an objective, mathematically derived entity, the experimental dispersion curve represents an interpretation of that spectrum. As such, unless a clearly defined mathematical procedure is used for its extraction, it remains a subjective interpretation of the dispersion spectrum, which introduces a high risk of error and increases non-uniqueness (Dal Moro, 2019a).

One of the essential issues is the assumption that the fundamental mode dominates the recorded wavefield. While this is generally valid for normally dispersive media, higher-order modes can become equally or even more energetic in irregularly dispersive environments, leading to ambiguous or misleading dispersion images (Tokimatsu et al., 1992; O'Neill and Matsuoka, 2005; Dal Moro, 2019a, 2023). Specific conditions under which higher modes tend to dominate include, for instance, the presence of a stiff surface layer or abrupt changes in stiffness at depth, which can cause mode jumps or steep velocity trends in the dispersion curve.

(Foti et al., 2018). In such cases, a dispersion curve may appear continuous but actually represent multiple overlapping modes, which complicates its interpretation.

Misidentifying a higher mode as the fundamental mode can lead to significant errors in velocity profiling, as the fundamental mode is defined as the lowest frequency, longest wavelength mode with the simplest displacement pattern and maximum sensitivity to shallow layers (Foti et al., 2018). Higher modes, in contrast, typically involve more complex waveforms, deeper energy penetration, and multiple nodal planes, leading to fundamentally different interpretations of subsurface structure if incorrectly assumed to be the fundamental mode (Dal Moro, 2014).

The separation and identification of multiple modes is a complex task requiring dense spatial sampling, long spreads, and appropriate array geometry (Foti, 2000; Socco and Strobbia, 2004). Even with suitable acquisition, energy transfer between components, limited spectral resolution, and proximity of modes may prevent clear mode discrimination. Consequently, an effective dispersion curve is often preferred, extracted as peak-energy ridges in the velocity spectrum, provides an objective representation of the wavefield but requires fundamentally different interpretation. Misinterpreting these as modal dispersion curves can lead to incompatible datasets and biased inversion results, especially when trying to fit multiple observables (Foti et al., 2018; Dal Moro, 2020a).

A further limitation arises from the subjective nature of modal dispersion curve picking. Since the velocity spectrum is interpreted manually, the extracted curve is inherently subjective and different analysts may derive different results (Zhang and Chan, 2003). Recently, several attempts for automatic algorithms based on deep learning showed promising results (Dai et al., 2021; Wang and Alkhalifah, 2021; Khosro Anjom et al., 2023).

The energy distribution between modes is not constant across frequencies and depends on multiple factors such as source type, offset, attenuation, and lateral heterogeneity (Foti et al., 2018). To address this variability, recommend verifying consistency during inversion using complementary datasets—such as Love and Rayleigh wave dispersion curves - to reduce ambiguity and confirm mode attribution.

In response to these challenges, various multimodal inversion strategies have been proposed. These include mode-selection criteria based on maximum displacement (Lu et al., 2006), full-waveform based approaches incorporating higher and leaky modes (Forbriger, 2003a, 2003b), and specialized misfit functions designed to target modal zeros (Maraschini and Foti, 2010).

Finally, higher modes are not only problematic but can also offer useful sensitivity, particularly for deeper layers. For instance Yin et al. (2014), demonstrated that higher-mode data penetrate deeper than the fundamental mode for both Rayleigh and Love waves, and Liang et al. (2008) showed that inversion stability may decrease in the presence of low-velocity layers, where fundamental mode sensitivity is limited.

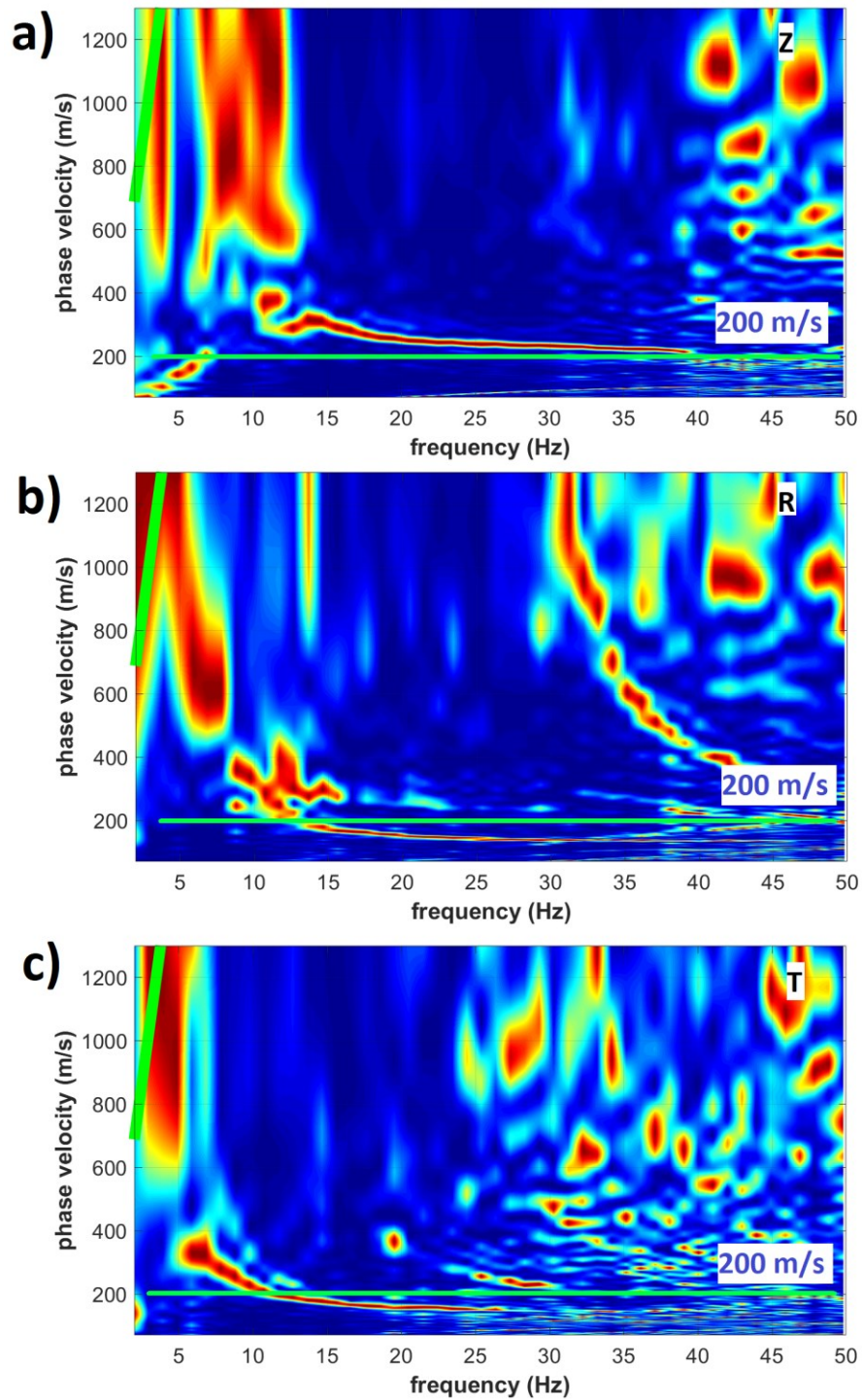


Figure 47: Dispersion spectra from active seismic data collected at the same site, showing a) vertical (Z), b) radial (R), and c) transverse (T) components. The green line at 200 m/s marks a common reference to illustrate the differences in phase velocity across components. The Z component a) shows a higher phase velocity trend in the low-frequency range, suggesting potential influence from higher modes, while the R b) and T c) components more clearly capture the fundamental mode.

Figure 47 presents dispersion spectra for three components of active seismic data acquired using MASW at the same site: a) vertical (Z), b) radial (R), and c) transverse (T). A green line at 200 m/s is included as a reference to highlight the phase velocity differences across components. It is evident that the Z component a) exhibits a significantly higher phase velocity trend in the low-frequency range compared to the R b) and T c) components. This discrepancy suggests that the Z component is likely influenced by higher modes, while the R and T components capture the fundamental modes more accurately, indicated by their lower phase velocities at comparable frequencies. Such mode misidentification can lead to substantial errors in the resulting V_s model if only the Z component is used for dispersion inversion, potentially resulting in an overestimation of near-surface stiffness. This example underscores the importance of joint component analysis in surface-wave studies, as relying on a single component can obscure the true fundamental mode and introduce significant uncertainty into the velocity structure, see Chapter 3.3.2 and Case study 3.

3.2.3 Inversion

The inversion of extracted experimental dispersion curves is a final of fundamental steps in dispersion analysis of surface-waves, aiming to estimate 1D vertical shear-wave velocity (V_s model) profile of the subsurface. The inversion process seeks a subsurface model whose theoretical dispersion response adequately matches the experimentally observed phase velocities. Recently, Gosselin et al. (2022) presented an extensive review of current inversion methodologies in seismic site characterizations, which can serve as a suitable starting point for exploring more detailed sources.

Since the phase velocity of surface waves varies with frequency due to subsurface layering, each frequency component samples a different depth interval. Lower frequencies are sensitive to deeper layers, while higher frequencies investigate shallower structures. This depth–frequency relationship allows reconstruction of the V_s profile, although the dispersion curve itself does not directly represent V_s values but rather an integrated elastic properties of the medium. However, this relationship is non-linear and often non-unique, which is why inversion represents a mathematically ill-posed problem (Luke et al., 2003; Dal Moro et al., 2007). The challenge lies in deriving a physically possible V_s model from the limited and frequency-dependent information encoded in the observed dispersion data. In this context, the accuracy of the inversion strongly depends on the quality of the input data, the choice of inversion strategy, and the underlying assumptions about the wave propagation and subsurface structure (Dal Moro, 2014; Foti et al., 2018). Most of the methods require explicit matching of observed and model-predicted dispersion curves (modal curves or effective curves) in the inversion.

To address the limitations of conventional dispersion-curve-based inversion, alternative approaches such as Full Velocity Spectrum (FVS; Dal Moro, 2014) or full-waveform analysis (FWA; (Dou and Ajo-Franklin, 2014) can be employed.

The inversion process includes the following main steps as highlighted in Figure 48. First, the model parameterisation step involves defining the geometry and physical properties of the subsurface. Typically, the medium is represented as a stack of homogeneous, horizontal layers described by shear-wave velocity (V_s), compressional-wave velocity (V_p), density (ρ), and thickness. Among these, V_s is the primary parameter of interest, while V_p and density are commonly estimated from empirical relationships (Ohta and Goto, 1978; Park et al., 1999). The number of layers and the way their properties are allowed to vary strongly influence both the resolution and the stability of the inversion.

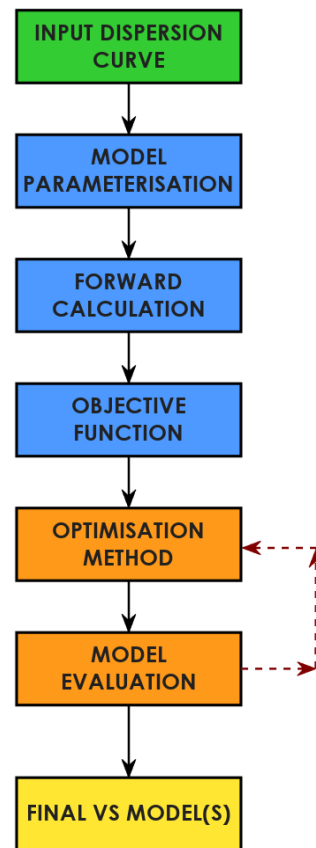


Figure 48: Simplified surface wave dispersion curve inversion workflow.

In the second step, forward modelling is used to compute the theoretical dispersion curve corresponding to a given parameterised model. This curve expresses phase velocity as a function of frequency and is compared against the observed extracted dispersion data. Several numerical approaches are available for this task, and their performance depends on the frequency range, model complexity, and desired accuracy (Foti et al., 2018). The prediction of surface-wave dispersion in laterally homogeneous models is typically carried out using efficient numerical techniques, such as the (Knopoff, 1964; Liu, 2010; Meehan, 2018). A reliable and efficient forward modelling method is critical for the success of the inversion process, as it is called multiple times during optimisation.

The third element of the workflow is the definition of an objective function, which quantifies the misfit between observed and theoretical dispersion data. This function typically takes the form of a weighted least-squares error, where different weights may be applied depending on the uncertainty or reliability of specific dispersion data points. The choice of objective function affects the sensitivity of the inversion and can influence the final solution, especially when dealing with noisy or sparse data.

Next, an optimisation method is applied to search the parameter space and identify models that minimise the objective function. Depending on the chosen algorithm, the optimisation can be local (efficient but sensitive to initial conditions) or global (more robust but computationally expensive). Common approaches include gradient-based methods, Monte Carlo simulations, genetic algorithms, or the Neighborhood Algorithm (Sen et al., 1995; Sambridge, 1999a, 1999b; Wathelet, 2008). This step is inherently iterative, as multiple models are evaluated and refined throughout the process.

Finally, the selection of the final model(s) involves identifying one or more solutions that best fit the observed data while satisfying geological plausibility and smoothness criteria. In many cases, the outcome of the inversion is not a single model but a family of acceptable models, reflecting the non-uniqueness of the inverse problem. Statistical analysis or post-inversion constraints may be applied to select representative models and assess uncertainty (Foti et al., 2018). The quality of the final result depends not only on the data and algorithms used but also on the informed interpretation of the solution ensemble. The following sections provide a brief discussion of the most essential aspects of the inversion process.

3.2.3.1 Input Parameters and Model Parametrization

A fundamental aspect of surface-wave inversion is the definition of the model structure, which involves choosing both the geometry and the physical parameters of the subsurface layers. Given a predefined 1D subsurface model structure, the inversion process aims to identify the set of model parameters that minimizes an objective function quantifying the mismatch between experimental and theoretical dispersion curves. The most critical parameter to be inverted is the shear-wave velocity (V_s), as it governs the dispersion characteristics of Rayleigh waves (Xia et al., 1999). Other parameters such as V_p , q and layer damping are often fixed using empirical relationships or assumed based on typical soil properties, primarily because their influence on the dispersion curve is generally minor compared to V_s (Nazarian, 1984; Xia et al., 1999). V_p is usually expressed in the form of Poisson's ratio (ν), which defines the relationship between compressional (P-wave) and shear (S-wave) velocities, see

Table 8 (Foti and Strobbia, 2002; Kumar Thota et al., 2021).

Table 8: Typical Range of Poisson's ratio for various soil types. After Das (2011).

<i>Type of backfill soil</i>	<i>Poisson's ratio (ν)</i>
Loose sand	0.2-0.35
Dense sand	0.3-0.4
Sandy soil	0.15-0.25
Silt	0.3-0.35
Unsaturated clay	0.35-0.4
Saturated clay	0.5
Clay with sand and silt	0.3-0.42

The objective function is typically defined using a mathematical form - most commonly the root mean square error (RMSE), to represent the difference between corresponding data points. In practical terms, the solution to the inverse problem is taken as the set of parameters that yields a forward-modelled dispersion curve most consistent with the observed site-specific experimental data (Foti et al., 2014).

The layering scheme is another key consideration - models may use a fixed number of layers with predefined thicknesses or adopt flexible parameterizations where layer boundaries and thicknesses are also subject to inversion. Some approaches incorporate smoothing constraints, or monotonic Vs increases with depth to stabilize the solution and avoid overfitting (Foti et al., 2018).

Using too few parameters, such as an insufficient number of layers in 1D models or grid cells in 2D and 3D cases, may lead to underfitting, where essential features of the subsurface structure remain uncertain, and parameter estimates become biased. On the other hand, using an excessive number of parameters increases the risk of overfitting, where the model begins to capture noise rather than meaningful data trends, introducing artificial or poorly constrained features. In addition, a higher number of parameters generally leads to increased computational demands (Gosselin et al., 2022).

3.2.3.2 Classification of Inversion Methods

Inversion methods used in surface-wave analysis can be broadly divided into local and global optimization approaches (also called search methods, Table 9), depending on how they explore the parameter space and address the non-linearity and non-uniqueness of the problem (Foti et al., 2018; Gosselin et al., 2022). However, one of the earliest and most simplified techniques is the so-called $\lambda/3$ or $\lambda/2$ method, which estimates the shear-wave velocity as approximately 110% of the Rayleigh phase velocity at a depth corresponding to one-third or one-half of the wavelength (Socco and Strobbia, 2004) and might be used as well for the initial model. This empirical approach for rapid interpretation was first introduced by (Abbiss, 1981).

Table 9: list of algorithms for inverse problems. Adopted from (Garofalo et al., 2016b).

Label	Algorithm
NA	Neighborhood Algorithm
MC	Monte Carlo method
GA	Genetic Algorithm
SA-GA	Simulated Annealing and genetic algorithm in an hybrid heuristic search method
NLS	Non-linearized least-squares algorithm
LLS_EYE	Linearized Least Square and trial and error procedure
EYE	Trial and error procedure

Local search methods (LSM) aim to find the nearest minimum of the objective function starting from a single initial model (Lai, 1998; Xia et al., 1999; Foti et al., 2009; Socco et al., 2009). These techniques are typically fast and computationally efficient, since they involve only a limited number of forward-model evaluations for Rayleigh wave propagation. They often rely on gradient-based algorithms such as the Levenberg–Marquardt method (Levenberg, 1944; Marquardt, 1963). However, LSM are sensitive to the starting model and prone to getting trapped in local minima, making them less suitable when little a priori information is available or the solution space is complex (Socco et al., 2010).

Global search methods (GSM), on the other hand, are designed to explore a wider region of the parameter space and are thus more robust against local minima. These approaches include Monte Carlo sampling (Socco and Boiero, 2008), Genetic Algorithms (GA, Yamanaka and Ishida, 1996), Simulated Annealing (SA, Yamanaka, 2007), and the Neighborhood Algorithm (NA, (Wathelet, 2008)). Although they are computationally more demanding, they offer a better chance of capturing multiple plausible solutions and characterizing uncertainty, particularly in cases with limited data or complex subsurface conditions (Foti et al., 2011; Gosselin et al., 2022).

In addition to this general classification, methods can also be distinguished as deterministic or stochastic, and as modal inversion (based on dispersion curves) or full-waveform inversion (which uses entire waveforms), as discussed above. While modal inversion is the most widely used in practice due to its lower computational requirements and more straightforward implementation, full-waveform inversion has gained attention in recent years for its potential to extract additional information, though it remains limited by computational cost and sensitivity to noise (Dou and Ajo-Franklin, 2014).

Among the variety of inversion techniques available, Least-Squares Minimization (LSM) and Genetic Algorithms (GA) represent two conceptually distinct and widely used approaches in surface-wave inversion.

3.2.3.2.1 Least-Squares Minimization

Least-Squares Minimization (LSM) is a local, gradient-based method that minimizes the squared differences between observed and theoretical dispersion curves (Marquardt, 1963; Wiggins, 1972; Park et al., 1998; Foti, 2000; Picozzi and Albarello, 2007). The misfit is typically expressed as a weighted least-squares function, where the weights can account for measurement uncertainty or frequency-dependent sensitivity. The algorithm iteratively adjusts the model parameters, typically using derivatives of the misfit function, until a local minimum is reached. LSM methods are efficient and provide rapid convergence when the initial model is close to the true solution (Menke, 2018). It is one of the most widely used approaches to invert surface-wave dispersion data, as implemented in early MASW studies (Xia et al., 1999). However, their main limitation is the risk of convergence to local minima and their inability to explore the broader model space. As such, they are best suited for problems with relatively simple parameter spaces or when reliable a priori models are available (Socco et al., 2010).

3.2.3.2.2 Genetic Algorithm

In contrast, Genetic Algorithms (GA) are global, population-based search methods inspired by the principles of natural evolution (Stoffa and Sen, 1991; Rodríguez-Zúñiga et al., 1997; Louis et al., 1999; Dal Moro, 2014; Zhou et al., 2024). A GA begins with a randomly generated population of models, which evolve over successive generations through processes analogous to genetic crossover, mutation, and selection. At each generation, models are evaluated based on their fitness, typically defined by their ability to reproduce the observed dispersion data. GA can efficiently explore large and complex model spaces and are less sensitive to initial conditions. They are particularly useful for highly non-linear or multimodal problems, where multiple plausible solutions may exist. The main drawbacks of GA include their higher computational cost and the need for careful tuning of algorithm parameters (e.g., population size, mutation rate) to balance exploration and convergence (Cercato, 2009; Foti et al., 2011; Gosselin et al., 2022).

As a conclusion, LSM and GA illustrate the trade-off between computational efficiency and robustness. In practice, hybrid approaches are often employed, where a global method such as GA is used to broadly explore the solution space, followed by a local LSM method to refine the best-fitting models (Picozzi and Albarello, 2007; Foti et al., 2018).

3.2.3.3 Challenges and Practical Aspects

Despite its widespread use, surface-wave inversion remains a complex and inherently uncertain process, with several challenges that must be carefully addressed to obtain reliable and interpretable results. One of the most fundamental issues is the non-uniqueness of the inverse problem (Foti et al., 2009; Dal Moro, 2023): multiple subsurface models may yield very similar dispersion curves, especially when the available data are limited in bandwidth or quality. This ambiguity makes it essential to assess model uncertainty and, where possible, constrain the inversion using independent information such as borehole data or geotechnical profiles (Foti et al., 2011; Gosselin et al., 2022).

Another key challenge is the resolution limit imposed by the data, which is closely linked to the range of wavelengths present in the observed dispersion curve. While high-frequency data improve resolution near the surface, low-frequency content is necessary to image deeper layers. In practice, this often results in good resolution of the upper tens of meters, with increasing uncertainty at greater depths (Socco and Strobbia, 2004; Foti et al., 2011). Moreover, mode misidentification can lead to significant errors in the inversion outcome. Higher modes are especially important in complex stratigraphies or where velocity inversions occur, but they are not always easy to extract or identify unambiguously in the data (Gosselin et al., 2022).

From a practical standpoint, the choice of inversion strategy should reflect the site conditions, data quality, and intended application. For example, local methods may be appropriate for well-constrained problems with high-quality data, whereas global or hybrid approaches are preferable when facing greater uncertainty or complex geology. The parametrization of the model - including the number of layers, fixed vs. free parameters, and depth discretization - also has a major impact on stability and interpretability. Over-parameterization may lead to instability and non-physical results, while overly simplistic models can misrepresent the true subsurface (Socco et al., 2010).

A key consideration in any inverse problem is whether the assumed model structure appropriately reflects the actual physical system. For instance, adopting a laterally homogeneous model in the presence of significant lateral heterogeneity can introduce modelling errors that bias the inversion results or prevent meaningful interpretation entirely. In such cases, more complex two- or three-dimensional representations of the subsurface may be required to capture the relevant features (Dal Moro, 2014; Gosselin et al., 2022).

3.3 Surface Wave Methods

While the theoretical basis of dispersion and the inversion of dispersion curves have been comprehensively introduced earlier, this section aims to provide a practical overview of the most established methodological frameworks. These are categorized primarily based on the type of velocity they retrieve: phase velocity or group velocity, and secondarily based on the nature of the source, i.e., whether the method is active or passive.

Surface-wave methods have evolved significantly over the last decades, with early developments such as the Spectral Analysis of Surface Waves (SASW) technique (Nazarian, 1984; Stokoe, 1994) paving the way for more robust and widely adopted approaches like the Multichannel Analysis of Surface Waves (MASW), introduced by Park et al. (1999) and further refined by Xia et al. (1999). These active-source methods have proven to be effective in retrieving shear-wave velocity profiles at shallow depths and are now considered standard practice in many engineering applications (Socco et al., 2010; Mazanec and Valenta, 2023) but also bring some risks as discussed in Dal Moro (2023).

The rise of passive methods, primarily based on microtremor recordings, has expanded the applicability of surface-wave techniques to deeper targets and urban environments where active sources may be impractical. Among these, the Spatial Autocorrelation (SPAC) method, first proposed by Aki (1957) and further developed by Okada and Suto (2003), has become a foundation of ambient vibration analysis due to its robustness and simplicity. Recent reviews and blind-test comparisons, such as those conducted by Foti et al. (2018) and Asten and Hayashi (2018), confirm that SPAC and its extensions (e.g., ESAC) remain among the most widely used passive array techniques for retrieving dispersion characteristics from ambient vibrations.

3.3.1 Methods Classification

Surface-wave methods can be classified based several criteria's, such as on the type of velocity they analyze - phase velocity (e.g., MASW, ESAC) or group velocity (e.g., FTAN, MAAM). Another distinction is made by source type, where active-source methods (e.g., MASW) use controlled seismic energy, while passive methods (e.g., SPAC, HVSr) rely on ambient vibrations. Finally, surface-wave techniques differ by wave type, targeting either Rayleigh waves, which involve vertical and radial particle motion, or Love waves, which involve transverse horizontal motion.

Phase vs. Group Velocity

The vast majority of studies rely on the estimation of phase velocity, which defines the propagation velocity of individual frequency components. As summarized in Dal Moro and Mazanec (2024), phase velocities from passive data can be derived using various techniques, including correlation and interferometric correlation and interferometric analyses (Snieder, 2008; O'Connell and Turner, 2011; Wapenaar et al., 2011; Van Dalen et al., 2015; Asano et al., 2017), refraction microtremors (ReMi, Louie, 2001), spatial autocorrelation (SPAC, Okada and Suto, 2003; Asten, 2006), extended spatial autocorrelation (ESAC Ohori et al., 2002), miniature array analysis of microtremors (MAAM, Cho et al., 2006, 2013), and f-k analysis (Poggi and Fäh, 2010). Currently, the most widely used approach for estimating shear wave velocity from active surface waves data is the Multichannel Analysis of Surface Waves (MASW), introduced by Park et al. (1998, 1999). Phase velocity methods are generally preferred due to their robustness, the relative simplicity of acquisition and processing, and the direct relationship between phase velocity and shear-wave velocity distribution in layered soils (Ciancimino et al., 2018; Hayashi et al., 2022). It is also the dominant focus of this paper, although the advantages that group velocity analysis brings to site characterization should be emphasized.

By contrast, group velocity-based approaches focus on the velocity of the wave packet envelope (Figure 31), which can also be linked to the subsurface stiffness structure. Group velocity methods are commonly used in seismology and much less widely used in near-surface geophysics. However, they can be applied using direct time-phase approaches, such as Multiple Filter Technique (MFT, Dziewonski et al., 1969), Frequency-Time Analysis (FTAN, Levshin et al., 1972), Wavelet Transform (WT, Kulesh et al., 2005) or the classical frequency-wavenumber ($f - k$, Capon, 1973). Group velocity estimation may be advantageous in certain scenarios, particularly when phase unwrapping is challenging or when signal coherence is limited (Socco and Strobbia, 2004).

Source Type

A second major distinction among surface-wave methods pertains to the type of seismic energy utilized. Active-source methods use controlled excitation (e.g., sledgehammer or weight drop) to generate surface waves in a targeted frequency band. This category includes SASW (Sheu et al., 1988; Stokoe, 1994), MASW (Park et al., 1999), and more recent hybrid methods that jointly analyze Rayleigh and Love waves or integrate refraction data (Socco et al., 2010; Dal Moro, 2020a). Active methods are particularly effective for shallow investigations (<30 m), offering high resolution and signal-to-noise ratios. The main limitations relate to limited low-frequency content, reduced penetration depth, and logistical constraints in urban or inaccessible environments.

On the other hand, passive-source methods rely on the recording of ambient seismic noise (microtremors) generated by natural or anthropogenic sources. These methods are especially advantageous for deep investigations (up to several hundred meters) and can be implemented with minimal equipment and disturbance to the site. The most prominent passive method is the SPAC technique (Aki, 1957; Okada and Suto, 2003), which estimates phase velocity from spatial coherence functions across 2D arrays. Other passive techniques include the f-k transform (Capon, 1969), ESAC (Ling, 1993), miniature array analysis of microtremors (MAAM, Cho et al., 2006, 2013) and the ReMi method (Louie, 2001).

Wave Type

Surface-wave methods can be classified based on the type of wave they analyze: Rayleigh or Love waves. Rayleigh-wave analysis is more commonly used, primarily due to the ease of generating and recording vertical ground motion with standard geophones (Xia et al., 1999; Ólafsdóttir, 2016; Mazanec and Valenta, 2023). However, Love waves, which involve horizontal shear motion, can provide valuable information, particularly for resolving shear-wave velocity structures and assessing subsurface anisotropy. Love-wave phase velocity spectra are generally considered simpler and more straightforward to interpret than those of Rayleigh waves, making them highly valuable for surface-wave data analysis (Takahashi, 1955; Safani et al., 2005, 2006; Kolínský et al., 2011; Dal Moro et al., 2015; Gaždová et al., 2015; Dal Moro and Mazanec, 2024).

As noted by Dal Moro and Mazanec (2024), while the analysis of the transverse (T) component from passive data is not widely adopted in near-surface geophysics, it is relatively common in seismological studies. Numerous authors have successfully extracted dispersion curves for both Rayleigh and Love waves from microtremor correlation (Trampert and Woodhouse, 1995; Bensen et al., 2008; Lin et al., 2008; Snieder, 2008; Behr et al., 2011; Boxberger et al., 2011; Mordret et al., 2013; Poli et al., 2013; Zigone et al., 2015; Asano et al., 2017).

3.3.1.1 Active Methods

3.3.1.1.1 Spectral Analysis of Surface Waves (SASW)

The SASW method, developed in the 1980s (Nazarian, 1984; Sheu et al., 1988), is among the earliest techniques to systematically utilize phase velocity dispersion for geotechnical site characterization. It employs a pair of receivers and a controlled source (typically a sledgehammer) to determine surface-wave phase differences as a function of frequency (Joh, 1996). SASW was the cornerstone for the development of the now much more widely used multichannel methods.

3.3.1.1.2 Multichannel Analysis of Surface Waves (MASW)

Multichannel Analysis of Surface Waves (MASW), introduced by Park et al. (1999) and further developed by Xia et al. (1999), is among the most widely used active-source methods for determining near-surface shear-wave velocity profiles. Unlike earlier approaches such as SASW, MASW employs a linear array of multiple receivers (Figure 49), which allows for better wavefield sampling and more reliable estimation of dispersion characteristics across a broad frequency range (Socco and Strobbia, 2004; Foti et al., 2018).



Figure 49: Field photo of active MASW data acquisition using simultaneous recording with horizontal and vertical geophones and a sledgehammer as the seismic source.

MASW follows the general workflow of surface-wave dispersion analysis as outlined in Chapter 3.2 and Figure 33. The choice of array length and receiver spacing is critical to optimize the frequency bandwidth and depth sensitivity (Dal Moro, 2014). When properly configured, MASW can provide reliable V_s profiles to depths of up to 30–50 meters in standard applications. The multichannel wavefields recorded by linear array is first transformed into a dispersion image in the frequency–phase velocity domain, typically using transformation techniques such as phase shift (Park et al., 1998) or slant stack (McMechan and Yedlin, 1981). In practice, most MASW surveys rely on the analysis of the vertical component of Rayleigh waves, generated by vertical impacts and recorded using vertical-component geophones (Garofalo et al., 2016a). However, Love-wave phase velocity spectra are often less complex and easier to interpret (Dal Moro et al., 2015; Dal Moro and Mazanec, 2024) and, under specific conditions, can exhibit higher modes, which can be valuable for the characterization of deep layers (Dal Moro, 2020a).

The fundamental mode of Rayleigh waves is usually extracted from the dispersion image and used for inversion. While this simplifies the process, it introduces potential biases, as showed in Figure 47, especially in complex subsurface conditions where higher modes may dominate or interfere (Dal Moro, 2023; Dal Moro and Mazanec, 2024). The dispersion curve is typically inverted using linearized least-squares (LSM) or global optimization algorithms such as genetic algorithms (GA), depending on the user's preference, data quality, and required robustness (Foti et al., 2018).

The MASW method has been recognized as an efficient tool for investigating the elastic properties of near-surface materials and has found widespread application, both as a standalone technique and as part of joint analyses, in various fields (Mazanec and Valenta, 2023), including bedrock mapping (Miller et al., 1999; Park, 2016; Rahimi et al., 2020a), soil stiffness estimation (Ólafsdóttir, 2016), marine environment studies (Kaufmann et al., 2005), fault detection (Gaždová et al., 2015), hydrogeological investigations (Suto, 2012; Pasquet et al., 2015), landslide assessment (Suto et al., 2016; Strelec et al., 2017), liquefaction potential analysis (Lin et al., 2004; Rahimi et al., 2020b), karst detection (Debeglia et al., 2006), and general geotechnical site characterization (Penumadu and Park, 2005; Park, 2013; Abudeif et al., 2019; Wood et al., 2019). MASW has also become a common method for estimating the average shear-wave velocity in the upper 30 meters of the subsurface (V_{s30}), as demonstrated by numerous studies (Kanlı et al., 2006; Odum et al., 2013; Hollender, 2019; Rahman et al., 2019; Sairam et al., 2019; Gaytan et al., 2020; Dovbnich et al., 2022; Mazanec and Valenta, 2023), as discussed in Chapter 2.4.

3.3.1.2 Passive Methods

Seismic noise has been documented since the earliest days of instrumental seismology in the nineteenth century. Until the mid-twentieth century, research on seismic noise remained mostly qualitative, as advancements were constrained by the limitations of available instrumentation and data processing methods. One of the earliest comprehensive reviews on the nature and origin of microseisms was conducted by Gutenberg (1911), marking a significant milestone in the study of seismic noise. A review of the historical development of this field is provided in the study by Bonnefoy-Claudet et al. (2006).

3.3.1.2.1 Spatial Autocorrelation Method (SPAC)

The SPAC method, first proposed by Aki (1957) and extensively developed by Okada and Suto (2003), estimates the dispersion characteristics of surface waves by analyzing the spatial coherence of ambient vibrations recorded at multiple locations. The SPAC technique relies on the principle that the azimuthal average of the spatial autocorrelation function follows the form of a Bessel function, with its argument determined by the phase velocity, frequency, and the spacing between receivers. Traditional SPAC acquisition employs a circular or polygonal array (Figure 37) of vertical-component sensors, typically with a central reference station. The method assumes an isotropic and horizontally homogeneous ambient noise field, where

seismic energy arrives equally from all azimuths. This configuration enables the estimation of the SPAC coefficient, defined as the normalized spatial coherence between two sensors. The spatial autocorrelation function is modelled as a zeroth-order Bessel function of the first kind, allowing phase velocities to be estimated from the frequency-dependent correlation values:

$$SPAC(r, \omega) = Re \left[\frac{S_{fg}(\omega)}{S_{ff}(\omega)} \right] \approx J_0 \left(\frac{\omega \cdot r}{c(\omega)} \right) \quad (3.27)$$

where r is the interstation distance; ω is the angular frequency, $S_{fg}(\omega)$ is the cross-power spectrum between stations f and g . $S_{ff}(\omega)$ is the autopower spectrum of the reference station, $c(\omega)$ is the phase velocity of the surface wave at the ω . By inverting this relation across multiple frequencies, a phase velocity dispersion curve can be derived and used for subsurface shear-wave velocity profiling.

Although the above formulation assumes circular geometry, SPAC can also be applied to linear arrays using a simplified approximation. In such cases, the spatial coherence is modeled using a cosine function rather than a Bessel function, reflecting the loss of azimuthal averaging in linear geometries:

$$SPAC_{1D}(r, \omega) \approx \cos \left(\frac{\omega \cdot r}{c(\omega)} \right) \quad (3.28)$$

This 1D cosine approximation is valid when the ambient wavefield is dominated by unidirectional propagation.

While this approach enables analysis without circular deployments, it introduces notable limitations. The most critical is the violation of the isotropy assumption, which can lead to bias and artifacts in the estimated dispersion curve - particularly at low frequencies or in multimodal conditions. Empirical evidence shows that coherence values along a linear array can vary significantly between sensor pairs, indicating strong sensitivity to source directionality. To mitigate these limitations, possible solutions might be: (a) use longer recording times to improve statistical averaging, (b) align the array with known dominant noise directions (if possible), (c) cross-validate results with alternative methods (e.g., FK, ESAC), or (d) perform joint inversion with active-source data for enhanced model reliability (Boxberger et al., 2011; Foti et al., 2018; Hayashi et al., 2022).

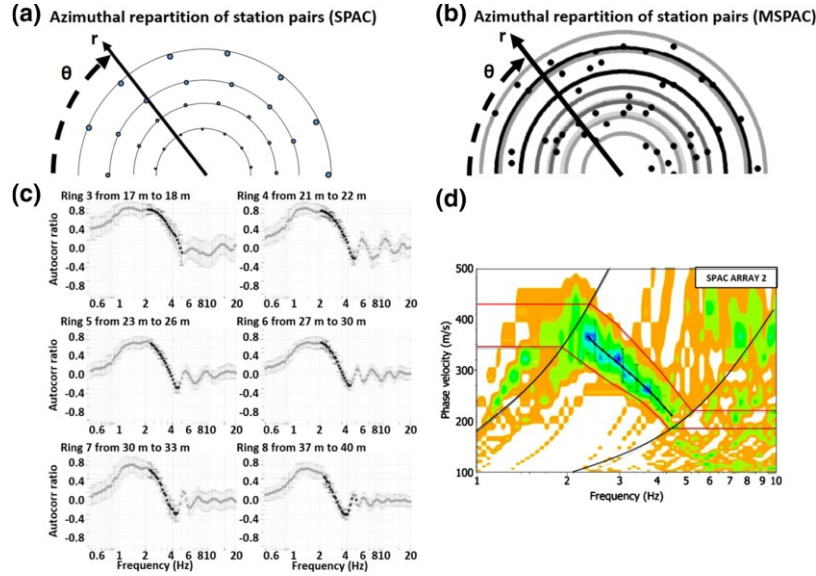


Figure 50: Example of SPAC techniques: a) azimuthal distribution of station pair vectors for the standard SPAC method, where r denotes the constant interstation distance for each ring and the azimuth of the station pair; b) the same for the MSPAC method, where each ring is defined by a range of r values (shown in gray); c) spatial autocorrelation functions for six different rings shown in b); (d) corresponding dispersion image, with manually picked dispersion curve shown in black with error bars. Autocorrelation estimates in c) used to derive phase velocities within the area bounded by the black and red curves in d) are indicated in black. Adapted from Foti et al. (2018).

Traditional 2D geometry SPAC is particularly suited for low-frequency analysis, enabling the investigation of deeper velocity structures when large array apertures and low-frequency sensors are employed (Tada et al., 2007; Asten et al., 2014; Hayashi and Craig, 2017; Asten and Hayashi, 2018). The method has been applied in several large-scale studies, such as the InterPACIFIC project, see Figure 50, where it showed high consistency in the derivation of dispersion curves (Garofalo et al., 2016b, 2016a; Foti et al., 2018).

Several extensions of the SPAC method have been proposed to accommodate more flexible array geometries or improve resolution, including: ESAC (Extended SPAC) Okada et al. (1997), MMSPAC – for large, nested, non-circular arrays, see Figure 50 (Bettig et al., 2001; Köhler et al., 2007; Foti et al., 2018), krSPAC – handling irregular arrays (Asten et al., 2019) or 3C-SPAC – incorporating three-component data (Hayashi et al., 2022). These variants keep the assumption of isotropic wavefield propagation and thus benefit from circular or symmetrical array geometries.

3.3.1.2.2 Extended Spatial Autocorrelation (ESAC)

The Extended SPAC (ESAC) method, introduced by Ling (1993) and Okada and Suto (2003), is conceptually related to the classical SPAC approach, designed to enhance flexibility in array design. Rather than expressing the spatial autocorrelation as a function of frequency (as in conventional SPAC), ESAC plots the autocorrelation against interstation distance for fixed frequencies. This formulation allows for the use of linear, irregular or nested arrays and facilitates the analysis of a broader range of sensor separations.

As shown in Galiana-Merino et al. (2016) assuming an array of N stations, the initial step of the ESAC method involves calculating the normalized cross-spectra $\rho_{jn}(f)$ for each station pair, using the following expression (Ohori et al., 2002; Okada and Suto, 2003):

$$\rho_{jn}(f) = \frac{\frac{1}{W} \sum_{w=1}^W \text{Re} \{S_{jn}^w(f)\}}{\sqrt{\frac{1}{W^2} \sum_{w=1}^W S_{jj}^w(f) \sum_{w=1}^W S_{nn}^w(f)}} \quad (3.29)$$

where f is the frequency; $S_{jn}^w(f)$ denotes the cross-spectrum between the j -th and n -th stations for the w -th time segment; $S_{jj}^w(f)$ and $S_{nn}^w(f)$ are the corresponding power spectra at stations j and n ; and W is the total number of segments into which the recorded signal is divided. The set of normalized cross-spectra computed for all station pairs yields the experimental azimuthally averaged spatial correlation:

$$\rho(f, r) = \{\rho_{jn}(f)\} \quad (3.30)$$

where j and n denote all possible pairs of stations within the array, and r represents their corresponding interstation distances. In the case where each frequency corresponds to a single phase velocity, Aki (1957) originally showed that the azimuthally averaged spatial correlation takes the form of a Bessel function of the first kind, J_0 , whose argument depends on the frequency, interstation distance, and the Rayleigh-wave phase velocity $c(f)$:

$$\rho(f, r) = J_0 \frac{2\pi f r}{c(f)} \quad (3.31)$$

Therefore, the phase velocity $c(f)$ at each frequency can be estimated through an iterative grid-search procedure that fits the experimental azimuthally averaged spatial correlation to the theoretical values of the Bessel function (Galiana-Merino et al., 2016).

Although ESAC often is applied to the vertical component of Rayleigh wave, it can be applied to any component (Boxberger et al., 2011). The ability of ESAC to operate on non-circular, even linear arrays is a major advantage, particularly in urban or industrial environments where full azimuthal coverage is difficult or impossible to achieve, see Chapter 3.4.3, Dal Moro and Mazanec (2024) and Dal Moro (2020). In such cases, the re-use of existing linear MASW array layouts greatly simplifies data acquisition logistics and enhances the feasibility of joint active-passive analysis).

3.3.1.2.3 Miniaturized Ambient Array Measurements (MAAM)

Miniature Array Analysis of Microtremors (MAAM) is a passive seismic technique designed to capture wide frequency range of surface wave dispersion using small-scale arrays. Originally developed by (Cho et al., 2006, 2013) for shallow site characterization, MAAM allows efficient deployment in constrained or noisy environments. Its compact, usually triangular, geometry requiring just a small number of geophones (typically 4 or 6) minimizes the influence of lateral heterogeneities and facilitates straightforward logistics (Dal Moro, 2020a).

From a methodological standpoint, MAAM shares mathematical foundations with SPAC, requiring a circular array configuration with a central reference sensor used for noise-to-signal ratio (NSR) correction (Cho et al., 2013; Cho and Iwata, 2021). This compensation significantly enhances the reliability of the dispersion curve, especially at low frequencies where noise contamination is more likely. The NSR correction is performed using a parameter $\varepsilon(f)$, defined as the ratio of incoherent noise to coherent microtremor signal. The processing yields an effective dispersion curve derived from the azimuthally averaged cross-spectral phase differences, which needs to be modelled using the effective wavefield approach rather than a modal dispersion framework (Tokimatsu et al., 1992; Lunedei and Albarello, 2009; Dal Moro, 2020a).

As shown in Figure 51, typical MAAM setup uses four vertical component geophones, three positioned at the vertices of an equilateral triangle and one at the centre, recording ambient vibrations. For near-surface applications, setup using four geophones arranged within a 2–4 meters aperture and a 15–20 minutes recording duration generally provides reliable ambient vibration data within the 3 to 30 Hz frequency range. The technique assumes spatial homogeneity and isotropy of the microtremor field, but its small aperture ensures that even in the presence of significant microtremor energy from nearby sources, usable high-frequency content can be extracted (Dal Moro, 2020a; Dal Moro and Mazanec, 2024).

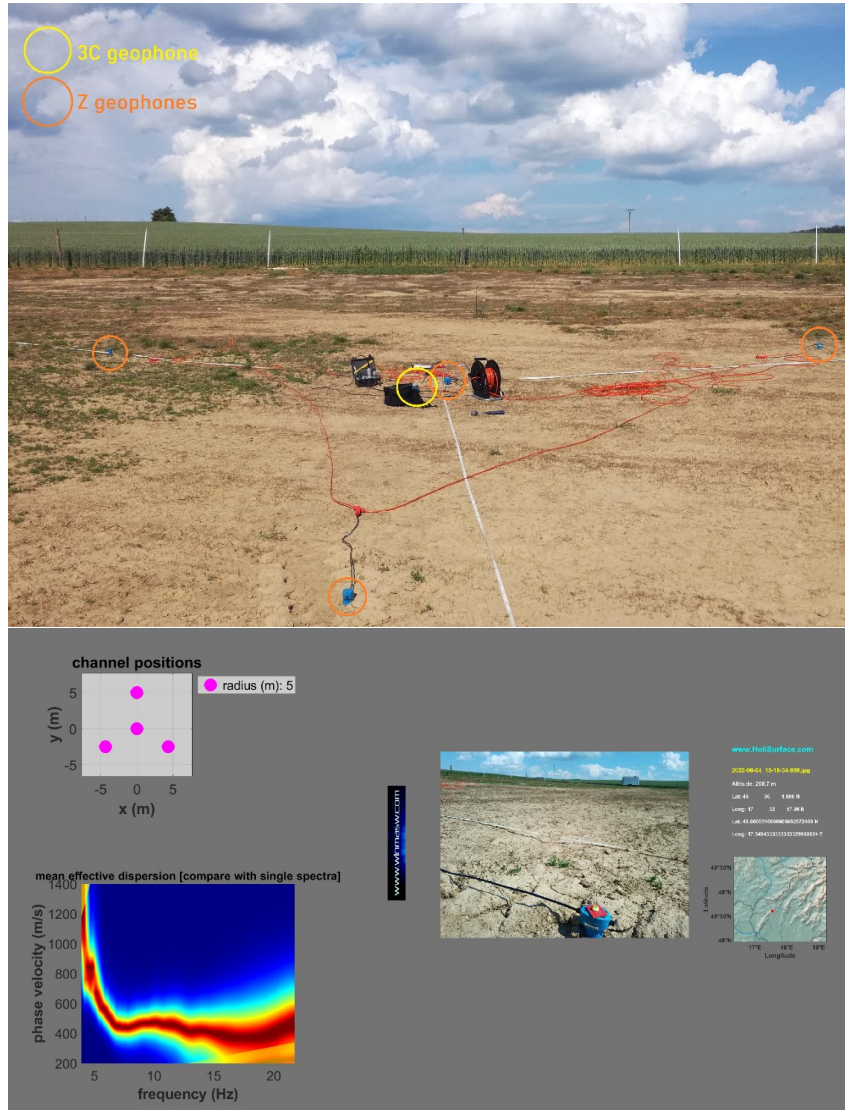


Figure 51: MAAM acquisition layout (upper panel) and related channel position, field notes and pictures and resulting effective dispersion curve (lower panel) in the WinMasw software. Example from the data acquisition in Dobrá Voda, Slovakia.

Despite its advantages, MAAM presents certain limitations. Its' very small array aperture restricts the maximum resolvable wavelength and consequently the investigation depth, generally limited to the top 10–20 meters. Furthermore, the method is sensitive to instrumental inconsistencies - careful matching of sensor responses and quality ADCs are required. As emphasized in Dal Moro (2020a), discrepancies in geophone characteristics or misalignment can introduce significant spectral artefacts. This necessitates strict quality control protocols during both acquisition (e.g., the use of huddle tests; (Cho et al., 2006) and processing (e.g., selective windowing and sensitivity analyses of spectral stability).

While MAAM excels in retrieving the dispersion of Rayleigh waves (Z-component), it is not suitable for direct retrieval of Love-wave dispersion. Attempts to extract Love-wave characteristics from MAAM configurations (Tada et al., 2010) have shown limited success. For this reason, complementary measurements (e.g., ESAC using the T-component in linear arrays) are often employed in joint inversion frameworks (Dal Moro and Mazanec, 2024).

Importantly, MAAM is highly compatible with joint analysis approaches. When combined with HVSR and Love-wave dispersion data, it contributes valuable constraints to the inversion, particularly in the shallow range (Dal Moro, 2023). In noisy environments, such as heavy industrial zones, MAAM has proven capable of retrieving robust and reliable Rayleigh-wave dispersion curves that correlate well with other observables (Dal Moro and Mazanec, 2024).

3.3.1.2.4 Frequency–Wavenumber (f – k) Analysis

The frequency–wavenumber (f – k) analysis is a classical array-based technique originally developed by Capon (1969) and later adapted for geophysical and microtremor applications (Asten and Henstridge, 1984; Socco and Strobbia, 2004). As discussed in 3.2.2.1.1, the f – k transform is a mathematical tool that maps wavefield data from the time–space domain into the frequency–wavenumber domain, allowing for the identification of propagating energy modes.

When used as a passive surface wave method, f – k analysis applies this same transformation to ambient vibration data recorded over an array. For each frequency f , the spatial Fourier transform of the wavefield reveals peaks in wavenumber k , corresponding to coherent plane waves traveling across the array. These energy peaks form dispersion ridges in the f – k domain, which can be inverted to obtain phase velocity as:

$$c(f) = \frac{f}{k_{max}(f)} \quad (3.32)$$

where $k_{max}(f)$ is the wavenumber at which maximum energy is observed, see Figure 52.

Unlike SPAC-based methods, this approach does not assume isotropy of the ambient wavefield. Instead, f – k analysis works effectively even when ambient noise arrives predominantly from a single direction. A key advantage of f – k analysis is its ability to resolve multiple propagation modes (e.g., fundamental and higher Rayleigh modes), which appear as separate energy ridges in the dispersion image. On the other hand, the method is sensitive to array geometry and requires sufficient aperture and receiver density to prevent spatial aliasing and ensure spectral resolution. Therefore, the separation of different modes is highly challenging (Dal Moro, 2020a).

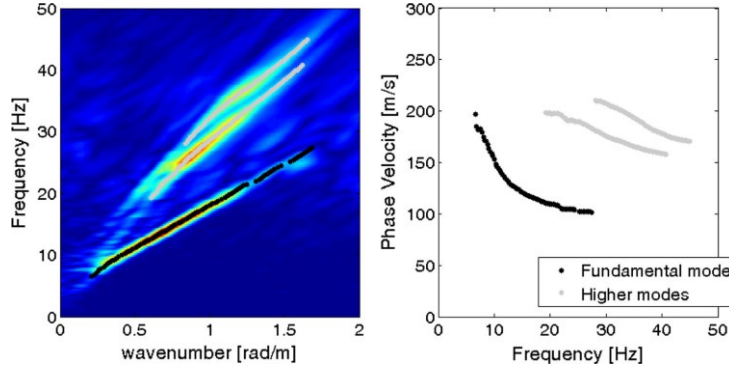


Figure 52: Example of $f-k$ analysis on real data (Mirandola site, InterPACIFIC project). Left: maxima of the $f-k$ spectrum corresponding to the fundamental mode (black) and higher modes (gray). Right: the associated dispersion curves plotted in the phase velocity–frequency domain. Adapted from Foti et al. (2018).

Modern implementations often include high-resolution techniques, such as beamforming or Capon’s adaptive processing, which improve the clarity of dispersion images and enhance performance in noisy settings (Socco and Strobbia, 2004; Yoon et al., 2005; Garofalo et al., 2016b; Cheng et al., 2018; Foti et al., 2018).

3.3.1.2.5 Refraction Microtremor (ReMi)

The Refraction Microtremor (ReMi) method, proposed by Louie (2001), is a passive surface wave technique that estimates Rayleigh-wave phase velocities from ambient vibrations using linear arrays of vertical-component geophones. Its main appeal lies in its logistical simplicity, as it can be applied using the same acquisition setup as standard seismic refraction surveys.

Fundamentally, ReMi is based on the slant stack ($\tau-p$) transform, previously described in Chapter 3.2.2.1.2, which converts seismic data from the time–space ($t-x$) domain into the intercept time–slowness ($\tau-p$) domain. In ReMi, this principle is applied in the frequency domain, where the wavefield is analyzed using a frequency–slowness beamforming approach to identify the dominant surface wave components.

For each frequency f , the method scans over a range of slowness values p and computes the stack energy using the relation:

$$S(f, p) = \left| \sum_{n=1}^N U_n(f) \cdot e^{-i2\pi f p x_n} \right|^2 \quad (3.33)$$

where: $S(f, p)$ is the energy in the frequency–slowness domain, $(U_n f)$ is the complex Fourier amplitude of the signal at sensor n , x_n is the position of sensor n , p is the trial slowness, N is the number of sensors.

The maximum energy at each frequency corresponds to the most coherent slowness $p(f)$, from which the phase velocity is obtained as $c(f) = 1/p(f)$. In this way, ReMi applies a spectral form of the slant stack to ambient noise to extract dispersion information.

Despite its efficiency and former popularity (Kaiser and Smith, 2005; Akkaya and Özvan, 2019; Kurtuluş et al., 2020), ReMi has been subject to methodological criticism. It implicitly assumes that ambient Rayleigh waves arrive uniformly from all azimuths, yet real-world noise fields are often highly directional, especially in urban or industrial areas. This can result in misinterpretation or suppression of modal energy, leading to unreliable dispersion curves (Cox and Beekman, 2011; Strobbia et al., 2011; Dal Moro, 2014, 2020). Additionally, ReMi does not explicitly separate surface wave modes and may mix body and surface waves when applied in noisy or complex settings. Studies by Garofalo et al. (2016a), Foti et al. (2018), and Hayashi et al. (2022) have shown that ReMi can produce significant errors when the ambient field deviates from its theoretical assumptions. As such, ReMi should only be used when the geometry and ambient field characteristics are well understood, or ideally in combination with other passive or active methods for cross-validation (Dal Moro, 2014).

3.3.1.2.6 Horizontal-to-Vertical Spectral Ratio (HVSr)

Although not a surface-wave method in the strictest sense, the Horizontal-to-Vertical Spectral Ratio (HVSr) technique is one of the most widely used tools for passive seismic site characterization. It is conceptually linked to surface-wave phenomena through its sensitivity to Rayleigh wave ellipticity and is therefore commonly incorporated into surface-wave analysis workflows (Fäh et al., 2001; Lunedei and Albarello, 2009).

HVSr was first formalized by Nakamura (1989) and is based on the ratio of horizontal-to-vertical spectral amplitudes of ambient vibrations recorded at a single station. Unlike array-based dispersion methods, HVSr does not require wavefield transformations or knowledge of wave propagation direction. Instead, it is used to identify the fundamental resonance frequency of a site (Figure 53), typically associated with significant impedance contrasts, such as the sediment-bedrock interface, which manifests as a distinct peak in the HVSr curve (Borges et al., 2016; Anbazhagan et al., 2019; Thabet, 2019; Stanko and Markušić, 2020; Rigo et al., 2021; Farazi et al., 2023). This makes HVSr particularly relevant for assessing site effects and potential amplification of seismic ground motion (Chávez-García et al., 1996; Gosar, 2007; Poggi et al., 2012; Burjánek et al., 2014; Michel et al., 2014; Stanko et al., 2017; Putti and Satyam, 2020; Chaaraoui et al., 2021).

The HVSr function is defined as:

$$HVSr(f) = \frac{\sqrt{H_T^2(f) + H_R^2(f)}}{V(f)} \quad (3.34)$$

where $H_T(f)$ and $H_R(f)$ are the Fourier amplitude spectra of the two orthogonal horizontal components (T-transverse and R-radial), and $V(f)$ is the spectrum of the vertical component.

A key advantage of HVSR is that it is not an array-based method. It requires only a single recording point and therefore offers exceptional deployment flexibility — particularly in urban, indoor, or logistically constrained environments (see chapter Case study 3 from Dal Moro and Mazanec, 2024). In practical applications, HVSR is often used as a consistency check in dispersion inversion (Garofalo et al., 2016a)

Nevertheless, its use as a standalone method for determining the V_s profile is strongly discouraged, as emphasized by Dal Moro (2019a). This caution is based on several factors:

1. **Non-uniqueness:** Different subsurface models can yield identical HVSR curves, making it impossible to derive a unique shear-wave velocity profile without additional constraints. However, this non-uniqueness is a general issue in geophysical methods.
2. **Source sensitivity:** HVSR results depend on the ambient wavefield characteristics, which may vary with season, weather, or time of day.
3. **Complex wavefield contributions:** While often linked to Rayleigh wave ellipticity, the HVSR response may also be influenced by P-wave velocities, attenuation properties (Q), Love wave s , and body waves (Bonnefoy-Claudet et al., 2008; Lunedei and Albarello, 2009; Dal Moro, 2014, 2015).

In modern practice, HVSR is most powerful when integrated into joint inversion schemes, where it complements the dispersion information by adding sensitivity to vertical velocity contrasts and resonance phenomena (Dal Moro et al., 2007; Castellaro, 2016; Dal Moro and Mazanec, 2024; Olafsdottir et al., 2024).

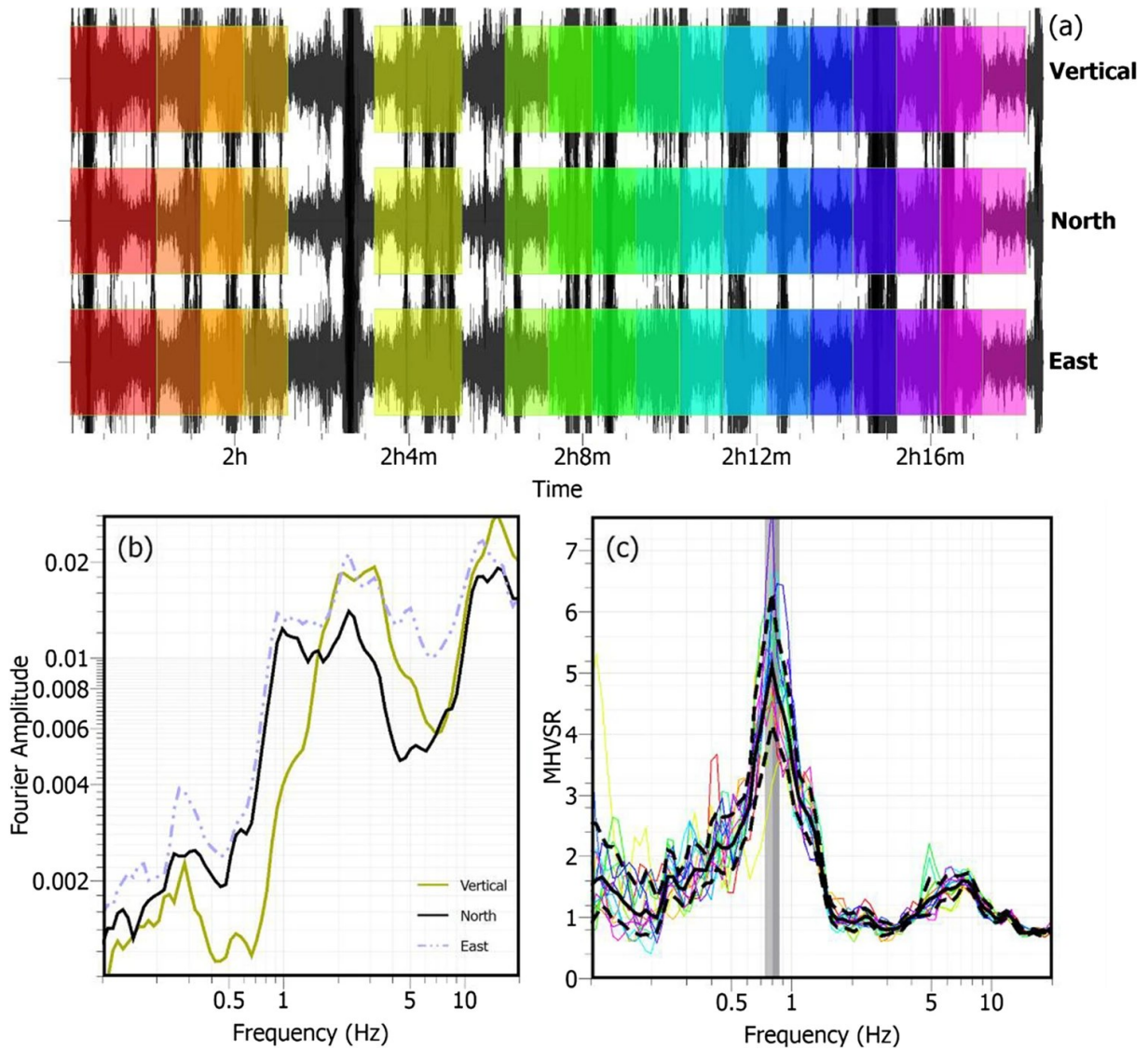


Figure 53: Illustrative example of the HVSR method: a) Segmentation of the microtremor time series into multiple time windows. b) Time window-averaged Fourier amplitude spectra (FAS) for each of the three recorded components. c) Resulting average microtremor horizontal-to-vertical spectral ratio (MHVSR) curve, represented by a solid black line, with the standard deviation indicated by dashed lines. The peak frequency of the average HVSR curve is highlighted with vertical gray shading, reflecting the variability in peak frequency values across individual time windows. This example, calculated using the Geopsy software package (Wathelet et al., 2020), is from a site in Metro Vancouver, British Columbia, Canada. Adapted from Molnar et al. (2022).

3.3.1.3 Group Velocity-Based Methods

While the majority of surface-wave methods rely on the analysis of phase velocity, a smaller set of techniques utilize group velocity - the propagation velocity of energy envelop, to infer subsurface shear-wave velocity structures. These methods focus on the travel-time behavior of wavefield envelopes and can be implemented either in the time domain or via frequency-time transformations. Although less commonly used in engineering-scale site characterization, group velocity methods offer certain advantages in specific acquisition and processing contexts. As summarized in Dal Moro and Mazanec (2024) group velocities derived from passive data are commonly employed in crustal-scale investigations (e.g., (Ritzwoller and Levshin, 1998; Levshin et al., 1999; Cotte and Laske, 2002; Pontevivo and Panza, 2002; Kolínský, 2004; Fang et al., 2010; Asano et al., 2017; Lu et al., 2018)), while in near-surface applications, they are typically extracted from active-source measurements (Nunziata et al., 1999, 2009; Luo et al., 2011; Gaždová et al., 2015; Kolínský et al., 2016; Dal Moro, 2023).

Group velocity analysis of surface waves can be conducted using various techniques, which can be broadly categorized into:

- (a) direct time-phase approaches,
- (b) indirect Fourier-based methods.

Direct approaches track the propagation of wave energy over time, allowing for the extraction of group velocity dispersion curves. Among these, the Multiple Filter Technique (MFT) applies a series of narrow-band filters to isolate dispersive wave components (Dziewonski et al., 1969), while Frequency-Time Analysis (FTAN) extends this concept by mapping the wave energy in the time-frequency domain, improving the identification of dispersive trends (Levshin et al., 1972; Natale et al., 2004). Additionally, Wavelet Transform (WT) methods provide an alternative means of decomposing wave signals to extract group velocity dispersion (Kulesh et al., 2005). Indirect approaches often involve spectral and wavefield transformation techniques. The frequency-wavenumber ($f - k$) method analyzes wave energy distribution in the frequency and wavenumber domain, allowing for the identification of both phase and group velocities (Lacoss et al., 1969; Capon, 1973). Beamforming, a variant of $f - k$ analysis, applies adaptive filtering to enhance resolution (Rost and Thomas, 2002). Another widely used technique is cross-correlation and empirical Green's function (EGF) analysis, which extracts group velocity information from ambient vibrations by computing cross-correlations between station pairs (Campillo and Paul, 2008). The selection of a suitable method depends on data availability, station configuration, and the frequency range of interest.

Group velocity approaches are typically more intuitive in the time domain, as they reflect the actual movement of surface-wave energy across the array. For example, time-domain MASW variants interpret the arrival times of energy maxima recorded at different sensors to estimate group velocity dispersion (Socco and Strobbia, 2004). This can be particularly useful when dealing with highly attenuative soils or when the phase information is obscured by noise or

complex waveforms. However, group velocity estimation is generally more sensitive to signal distortion and less stable than phase velocity analysis, especially in noisy or multimodal conditions (Socco and Strobbia, 2004; Dal Moro, 2014).

In practice, group velocity dispersion is often used as a complementary dataset rather than a stand-alone basis for inversion. It can help constrain the identification of modal branches in complex dispersion images or assist in quality control when comparing results from time-domain and frequency-domain approaches. Furthermore, some inversion schemes integrate both group and phase velocity information to improve model convergence and reduce non-uniqueness (Foti et al., 2018; Dal Moro, 2020).

The definition of group velocity becomes ambiguous in multimodal wavefields, especially when higher modes are present or energy partitioning is not well understood (Hayashi et al., 2022). Therefore, careful processing and interpretation are required when group velocity methods are applied in layered or strongly heterogeneous media.

3.3.2 Joint Analysis of Surface Waves

The choice of surface-wave method must reflect specific survey goals, site conditions, and technical constraints. Each method provides a different trade-off between resolution, penetration depth, sensitivity to wavefield properties, and interpretational robustness. However, due to the inherent non-uniqueness of inversion in surface-wave methods (and e.g., in geophysics in general), and the modal ambiguities frequently contend with in complex subsurface environments, the joint analysis of multiple datasets has become an increasingly standard practice (Dal Moro, 2020). As illustrated in Figure 54, joint analysis (or joint inversion) refers to the simultaneous or integrated use of two or more complementary geophysical observables.

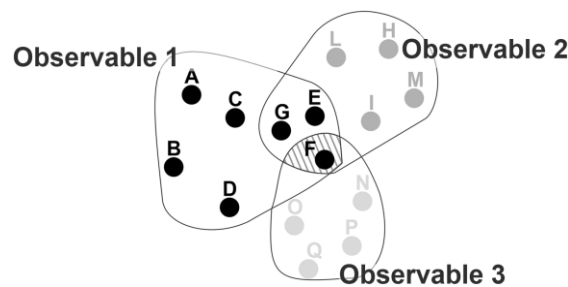


Figure 54: Schematic illustration of general ambiguity in applied geophysics and the role of joint analysis of multiple independent observables in reducing it. While observable 1 may be explained by subsurface models A–G and observable 2 by models E–M, etc., combining multiple observables allows the solution space to be constrained more tightly, in this case converging in region F. Adapted from Dal Moro (2019a).

Foti (2012) described that the combined use of geophysical methods can serve multiple purposes: (1) comparison for validation or calibration; (2) data integration and fusion to combine complementary information about the same subsurface medium; (3) the use of one method as a priori information to support another; and (4) joint inversion, allowing for the simultaneous interpretation of multiple datasets. Such a combinations of geophysical

approaches is very common (Vozoff and Jupp, 1975; Dobróka et al., 1991; Hayashi et al., 2005; Ivanov et al., 2006; Hu et al., 2009; Garofalo et al., 2015, 2022; Špaček et al., 2017; Blecha et al., 2018; Tábořík, 2022; Klanica et al., 2024).

In context of surface waves analysis typically combine dispersion data from Rayleigh and Love waves, group and phase velocities, and Horizontal-to-Vertical Spectral Ratios (HVSr). This approach aims to reduce the non-uniqueness of inversion results and improve the resolution and reliability of the resulting shear-wave velocity models (Dal Moro, 2014, 2015, 2020, 2023). Joint analysis that combines both active and passive data have proven to be highly effective, as they purchase the high-frequency resolution of active methods with the low-frequency depth penetration of passive methods. Joint inversion strategies have been shown to improve model reliability and reduce the ambiguity associated with mode misidentification and non-uniqueness (Dal Moro, 2014, 2015, 2019b, 2020; Hayashi et al., 2022; Mazanec and Valenta, 2023; Dal Moro and Mazanec, 2024; Olafsdottir et al., 2024; Abdelrahman, 2025). Joint inversion of HVSr and dispersion curves is now standard practice, improving model stability and depth resolution (Foti et al., 2018; Dal Moro, 2020; Hayashi et al., 2022).

Joint analysis has been extensively explored by Giancarlo Dal Moro, and numerous studies are available that demonstrate specific aspects and various methodological approaches to its application. Especially the book Dal Moro (2020) is devoted to this topic and shows case studies using various joint analysis. An example of this joint analysis (R + T components) with the Rayleigh-wave Particle Motion (RPM, (Dal Moro et al., 2015), where the inclusion of Love waves helped resolve a deep stiff layer masked in Rayleigh-wave-only inversion. This methodology has also been applied in the context of multi-offset MASW data and is compatible with Pareto-based optimization to handle multiple objectives simultaneously. Joint use of Rayleigh and Love waves is especially useful in stratified media where mode conversion or energy partitioning complicates interpretation, and it is particularly beneficial when higher modes are observable Dal Moro (2020).

An example of a basic joint analysis combining active and passive data can be found in Chapter 3.4.2, while an advanced approach incorporating three complementary methodologies is presented in Chapter 3.4.3.

3.4 Case studies

This chapter presents a series of case studies that illustrate selected aspects and practical capabilities of surface-wave analysis based on real-world data. The first example is intentionally simple and demonstrates how surface-wave dispersion can be effectively integrated into conventional refraction seismics. The second case introduces a basic joint analysis that, with minimal effort, increases the number of observables and reduces the non-uniqueness of the inversion. The final and most advanced example focuses on a complex industrial environment, where three distinct joint inversion strategies are applied to showcase state-of-the-art surface-wave approaches. Together, these case studies highlight three levels of methodological complexity and are designed to demonstrate the versatility of surface waves to both less experienced users and advanced practitioners.

3.4.1 Case Study 1 – The Receivers

Motivation: showing that even with standard 10 Hz vertical geophones, refraction data might be easily enriched using surface wave analysis.

This Case study 1 is part of the article Mazanec and Valenta (2023), which focused on the simplest approach suitable as an add-on to a standard refraction survey – using low-frequency vertical component geophones. Case study 1 refers to a study from landslide area in north Czechia, comparing two seismic datasets acquired along the same seismic profile with linear geometry. Active MASW data were acquired simultaneously using two types of vertical geophones with different eigenfrequencies: 4.5 Hz (RTClark geophones RTC-4.5Hz-395) and 10 Hz (Geospace, GS-20DX). As shown in Figure 56, both geophone types were placed at identical positions along the seismic profile, in close proximity to each other, ensuring comparability of their recordings.

The seismic profile extended 92 meters with geophones spaced at 4-meter intervals, using 24 sensors of each type. Recording time was 1s, see the shot gathers in (Figure 56). Refraction data were collected using 10 kg sledgehammer impacts on a metal plate placed every 8 meters along the profile. To ensure a high signal-to-noise ratio for surface-wave analysis, eight vertically stacked shots were performed at both ends of the line.

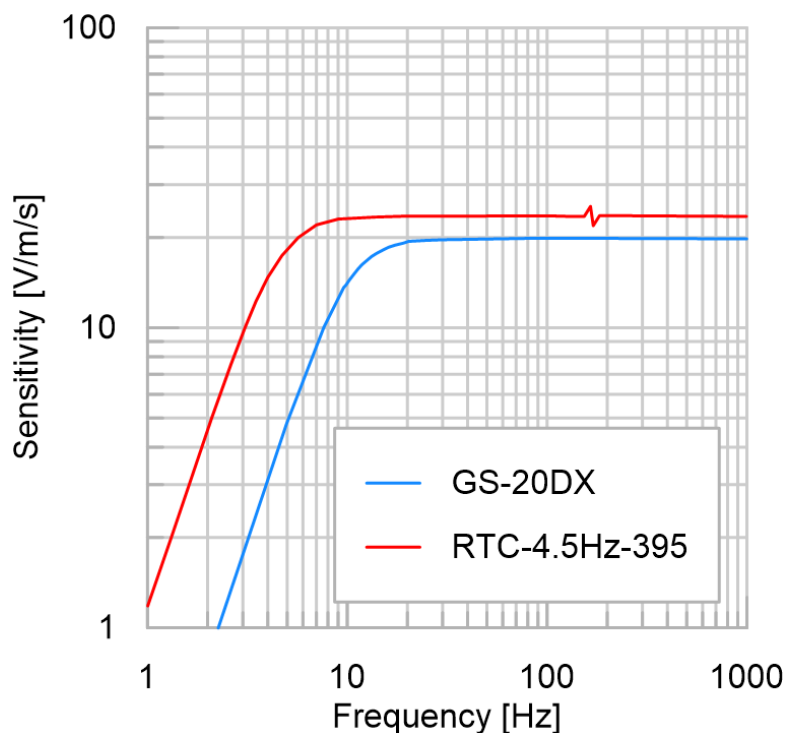


Figure 55: response curves for 4.5 Hz (RTClark RTC-4.5Hz-395) and 10 Hz (Geospace, GS-20DX) geophones used in the test. Adapted from Mazanec and Valenta (2023).

Figure 55 illustrates the response curves of the geophones used in this study. As noted by Dean and Shem (2018), the primary difference between the 4.5 Hz and 10 Hz geophones lies in their natural frequency, while their sensitivity remains comparable. The response curves indicate a rapid decrease in sensitivity below the natural frequency. For instance, at 5 Hz, the signal recorded by the 4.5 Hz geophone is approximately three times stronger than that from the 10 Hz geophone. However, despite the reduced signal amplitude, data from the higher-frequency geophone can still be recovered, albeit with a lower signal-to-noise ratio.

To compare the two datasets (4.5 Hz and 10 Hz vertical geophones) and demonstrate the applicability of surface wave analysis to refraction data, Figure 56 presents records from the 94 m shot point. The middle panels (b, c) show nearly identical shot gathers in terms of arrival times, amplitudes, and noise levels. The lower panels (d, e) display dispersion images computed using the Phase-Shift method (Park et al., 1998), revealing highly similar spectral patterns. The 4.5 Hz geophones exhibit stronger amplitudes around 5 Hz and better visibility of higher modes, while the 10 Hz geophones might provide slightly more clear definition of the fundamental mode at higher frequencies (20–30 Hz), aiding shallow Vs resolution. Overall, both datasets yield comparable trace and dispersion characteristics.

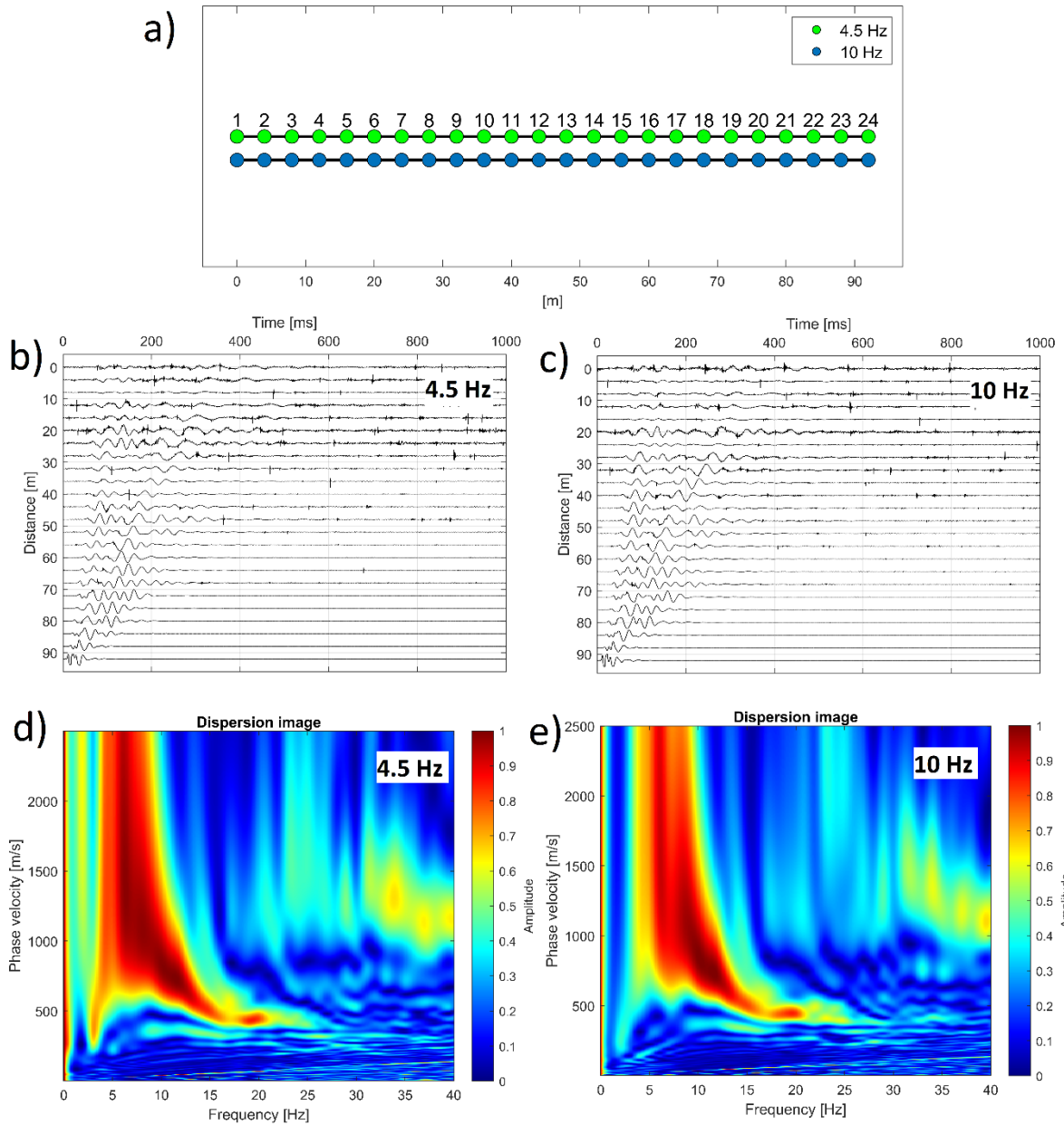


Figure 56: Comparison of two datasets: a) geophone array geometry; b) and c) shot gathers from the 94 m shot position for 4.5 Hz (b) and 10 Hz (c), respectively; d) and e) surface waves dispersion images for 4.5 Hz (d) and 10 Hz (e) geophones. Adapted from Mazanec and Valenta (2023).

The fundamental mode dispersion curves from both datasets (Figure 57) were inverted using a non-linear least squares method, and the resulting Vs profiles are plotted together for comparison (10 Hz: red line; 4.5 Hz: blue dashed). In the uppermost 5 m, both models exhibit similar trends with differences below 20 m/s. Between 5–11 m, the 4.5 Hz model shows a slight velocity inversion, unlike the steadily increasing trend in the 10 Hz model. From 11–24 m, the models remain largely consistent, except for a velocity inversion near 20.8 m, where variations reach up to 40 m/s. In the final section (24–30 m), the 4.5 Hz model indicates a stronger increase, but differences do not exceed 25 m/s.

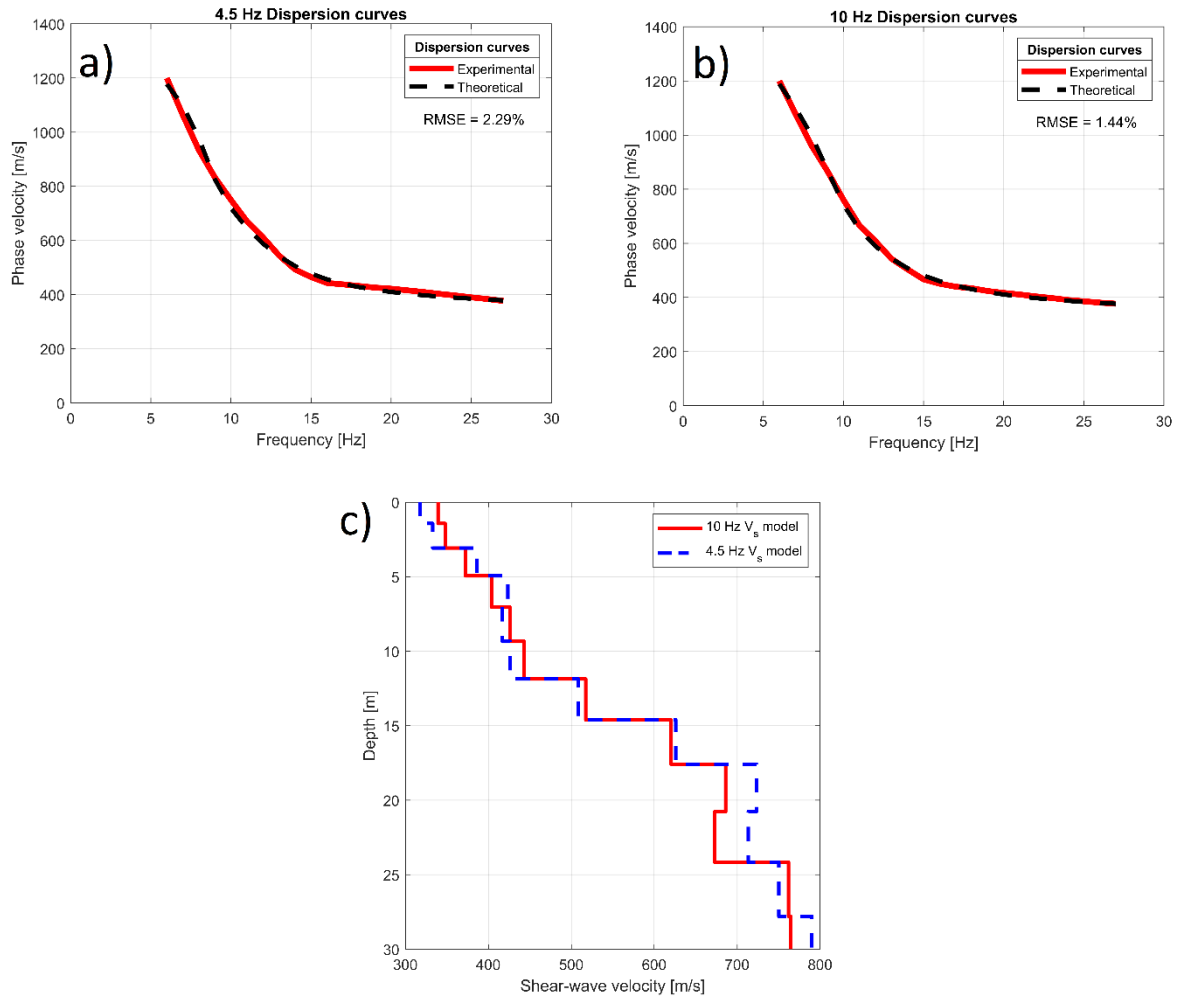


Figure 57: a) 4.5 Hz and (b) 10 Hz theoretical (black dashed line) and experimental (red solid line) Vs models' associated dispersion curves and RMSE for both datasets; c) final Vs models for both 4.5 Hz (blue dashed line) and 10 Hz (red solid line). Adapted from Mazanec and Valenta (2023).

Both Vs models show very similar trends with only minor differences, including matching velocity drops at 11.8, 14.6, and 17.6 m, and a gentle inversion at 20.8 m. Figures 6b and 6c compare the experimental and theoretical dispersion curves, with RMSE values of 2.3% (4.5 Hz) and 1.4% (10 Hz), confirming their similarity. These results demonstrate that standard 10 Hz geophones, commonly used in refraction surveys, can reliably support basic surface-wave analysis alongside the typical 4.5 Hz sensors. Furthermore, dispersion analysis can be applied retrospectively to suitable refraction datasets, helping reduce solution ambiguity.

The results demonstrate that standard 10 Hz geophones, commonly used in shallow refraction surveys, are suitable for basic surface-wave dispersion analysis alongside the recommended 4.5 Hz sensors. As shown in Figure 56 and Figure 57, differences in recordings, dispersion spectra, and the resulting 30 m Vs models are minimal. Although discrepancies may increase with depth due to geophone sensitivity limits, surface-wave analysis can also be effectively applied retrospectively to suitable refraction datasets, helping reduce model ambiguity. To achieve an investigation depth of up to 30–50 m, geophones with a natural frequency of 4.5 Hz

or less are commonly used (Xia et al., 1999). However, based on a study carried out by Park et al. (2002), 10 Hz and 40 Hz geophones can be used to record surface waves of frequencies as low as 5 Hz and 10 Hz, corresponding to a maximum investigation depth of about 30 m and 15 m depth, for most of the common soil sites. These findings are consistent with our observation in the Case study 1 (Figure 57).

These observations are not intended to be universally applicable but rather to demonstrate that basic forms of surface-wave dispersion analysis are accessible and can serve as a valuable complement to refraction seismics, as illustrated, for example, in Case Studies 2 and 3 of Mazanec and Valenta (2023).

3.4.2 Case Study 2 – Basic Joint Analysis of Rayleigh-Wave Passive and Active Data

This case study, adapted from Mazanec and Valenta (2023), demonstrates the most basic form of joint surface-wave analysis, using only a single linear geophone array to acquire three complementary datasets: active-source Rayleigh-wave records, passive ambient noise for SPAC analysis, and P-wave travel-time data for refraction tomography (Redpath, 1973; Sheehan et al., 2005; Azwin et al., 2013). All datasets were acquired using with 24 vertical-component 4.5 Hz geophones spaced at 4 m, forming a 92-meter-long array.

The site is located in northwestern Czechia at the SKAC seismic station (Figure 58), which is part of the RINGEN monitoring network (Fischer et al., 2023; Káldy and Fischer, 2023). Geologically, it lies within the Merboltice Formation of the Bohemian Cretaceous Basin (Šafanda et al., 2020).

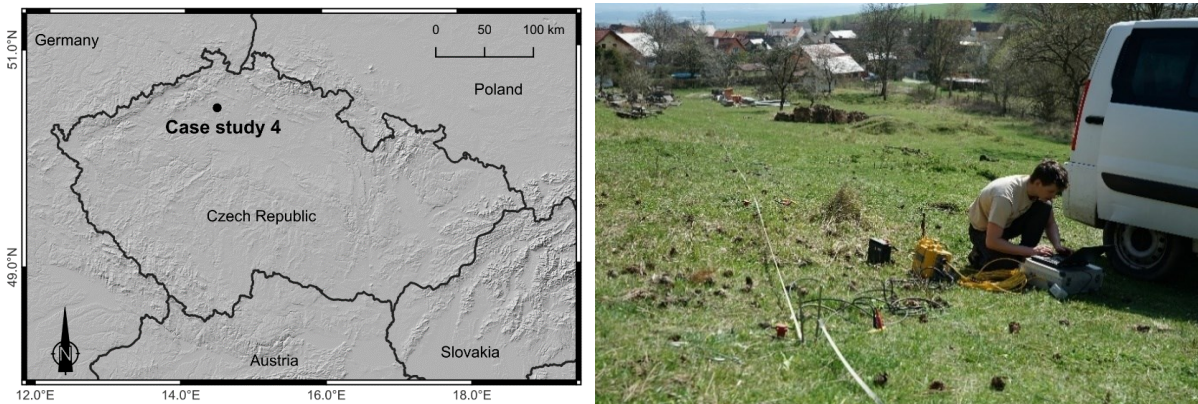


Figure 58. Left panel: map showing the location of the SKAC seismic station (Case study 4 adapted from Mazanec and Valenta 2023); right panel: field photo taken during data acquisition.

The active-source acquisition was conducted using a 10 kg sledgehammer and metallic plate as a trigger, with shot points located at -10, -2, 22, 46, 70, 94, and 102 meters, each repeated eight times to enable vertical stacking. To minimize near- and far-field effects and reduce the ambiguity associated with vertical-only geophones, two source offsets were placed on each side of the array. This configuration yielded four independent dispersion images (two from each side), allowing for an assessment of lateral heterogeneity and evaluating the validity of a 1D approximation. Ideally, the resulting 1D Vs model should represent the average shear-wave velocity distribution beneath the site. To further verify the assumption of lateral homogeneity, a shallow 2D velocity section was obtained from P-wave first-arrival tomography. Following the active acquisition, a 20-minute ambient vibration record was collected using the same array without reconfiguration. Passive data were recorded with a 4 ms sampling interval, while active data used a 0.063 ms sampling rate and 1 s time window. This setup illustrates a low-cost and effective joint workflow that integrates basic MASW, SPAC, and tomography within a standard seismic refraction survey design.

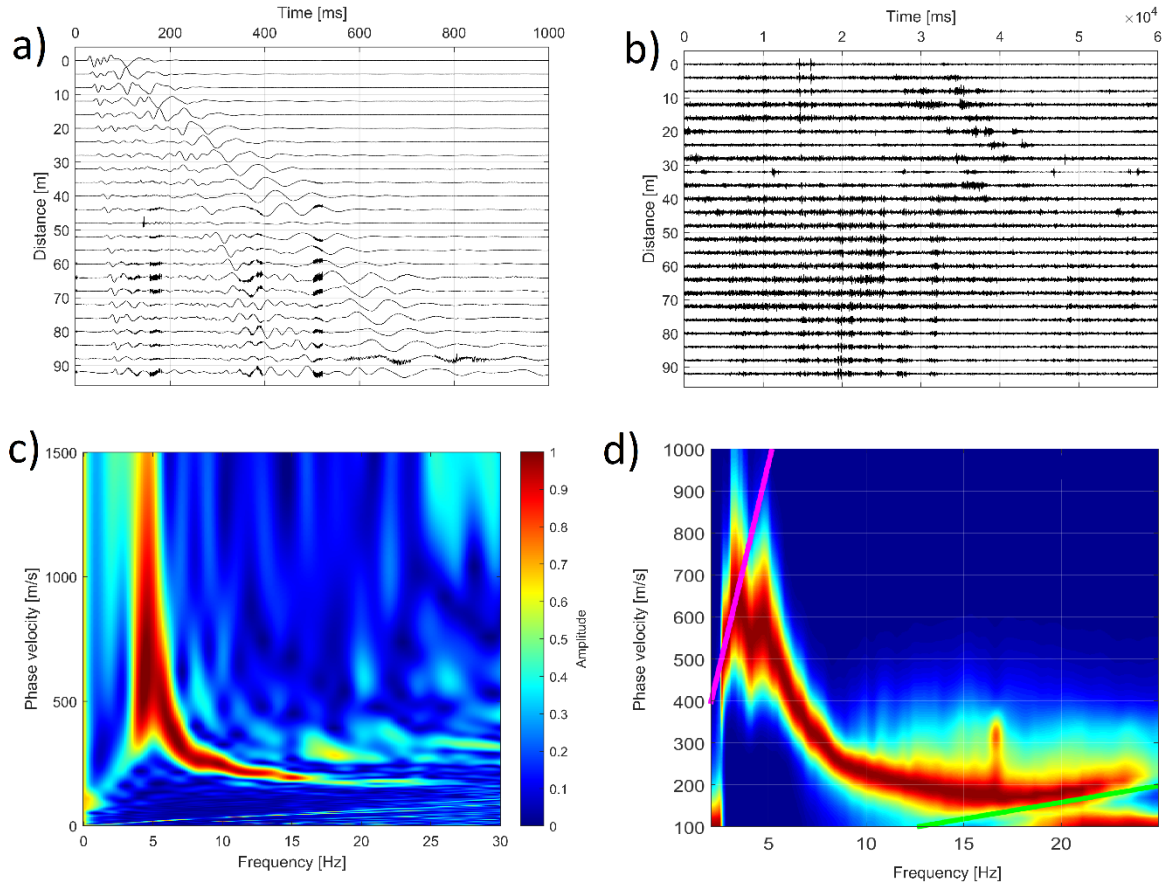


Figure 59: SKAC site: a) shot gather from active data acquired at the -10 m source position; b) 60-second ambient vibration traces; c) Rayleigh-wave dispersion spectra; d) passive dispersion spectra derived using the SPAC method. Adapted from Mazanec and Valenta (2023).

The Rayleigh-wave dispersion image obtained from the active data was calculated using the phase shift method, as shown in Figure 59. passive data were processed using the SPAC method, yielding a dispersion spectrum shown in Figure 59. The two partial dispersion curves: one covering the high-frequency range from MASW and the other the low-frequency range from SPAC, were subsequently merged into a composite experimental dispersion curve (Figure 61). This joint dispersion curve was then inverted using a least-squares method to obtain the V_s profile.

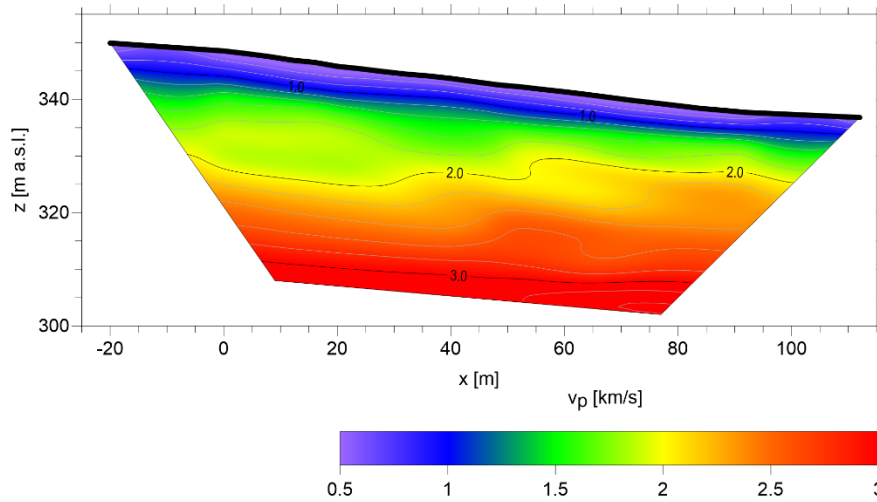


Figure 60: P-waves travel time tomography model of the SKAC site. Adapted from Mazanec and Valenta (2023).

The resulting shear-wave velocity model (Figure 61) reveals three prominent velocity contrasts at approximate depths of 18 m, 30 m, and 44 m, likely reflecting lithological transitions within the sedimentary sequence. The upper two interfaces show a strong correlation with the P-wave velocity model derived from travel-time tomography (Figure 60), while the deepest interface remains invisible to refraction methods, likely due to their limited sensitivity at depth.

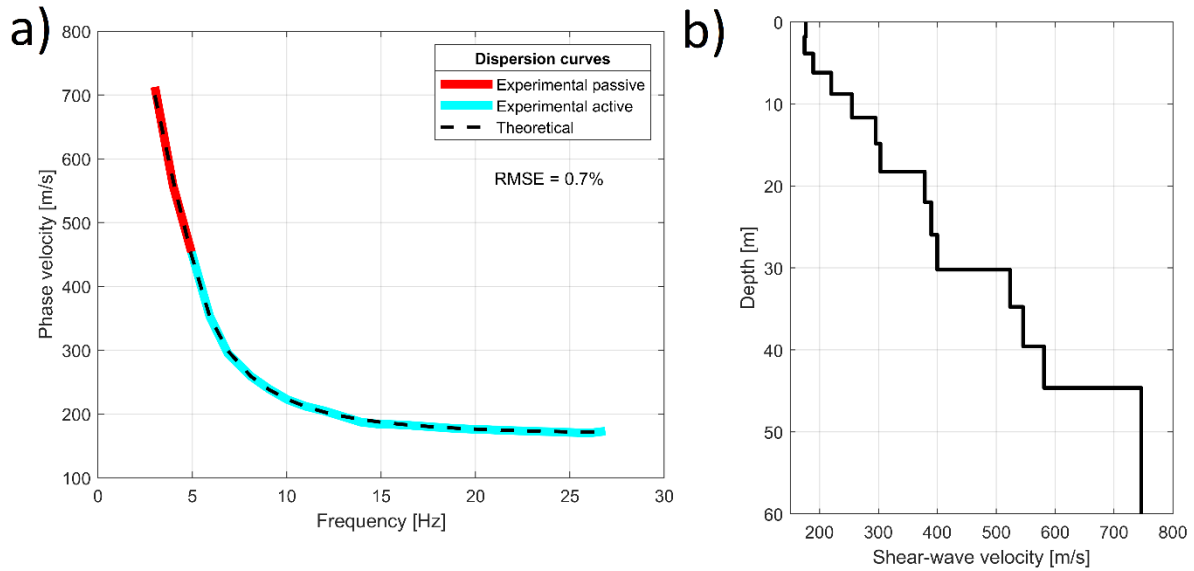


Figure 61: Results of the joint inversion of passive and active MASW Rayleigh-wave phase velocities: (a) inverted V_s model; (b) corresponding experimental dispersion curve (red bold line indicates the passive-data-dominated segment, blue line represents active MASW data), theoretical dispersion curve (dashed), and inversion RMSE. Adapted from Mazanec and Valenta (2023).

This case study effectively illustrates how a basic joint analysis, based on a linear receiver array, can significantly improve subsurface imaging and depth penetration, particularly when active and passive methods are integrated. It further emphasizes that surface-wave methods can complement conventional refraction data, especially in resolving deeper or subtle velocity transitions that would otherwise remain undetected.

While the linear array configuration is highly practical, especially for combining MASW, SPAC, and tomography from a shared geometry, it is important to note that SPAC analysis based on linear arrays carries inherent limitations, particularly regarding azimuthal bias and reduced ability to average over source directionality. Therefore, results from SPAC should be interpreted with caution when applied in non-isotropic noise fields or ideally validated through complementary methods or stability testing (Dal Moro, 2020).

3.4.3 Case Study 3 – Advanced Multicomponent Analysis in Industrial Site

This case study, based on Feruglio (1936; Cavallin et al., 1987; Dal Moro and Mazanec, 2024), demonstrates an advanced implementation of multicomponent surface-wave analysis conducted at a challenging industrial site characterized by strong anthropogenic noise and complex subsurface conditions. It illustrates a comprehensive integration of multiple techniques, wave types, and components, specifically designed to overcome the limitations of individual methods when used in isolation under adverse environmental circumstances. The objective was to derive a robust shear-wave velocity model by combining active and passive data from both vertical (Z) and transverse (T) components, incorporating both group and phase velocity information. The complex dataset was analyzed through three different joint inversion scenarios, enabling comparative evaluation to identify the most suitable approach for such conditions. This methodology proves particularly effective for seismic site characterization in urban environments, which are often difficult to assess using conventional techniques.

3.4.3.1 Site and Acquisition Setup

The site is situated in the eastern part of the Lower Friulian Plain (north Italy), where the shallow subsurface is composed of thick Quaternary alluvial deposits overlying Tertiary marine units, as typically observed in this sector of the Venetian-Friulian basin (Feruglio, 1936; Cavallin et al., 1987). The lithostratigraphy includes anthropogenic fill, silty sands, and fine-grained clayey layers, interbedded with gravel and sand lenses—reflecting an alternation of fluvial and marine sedimentation regimes, as visible in the borehole (Regione Autonoma Friuli Venezia Giulia, IT; borehole#1610060) approximately 200 m east from the investigated site (Figure 62). The water table lies at shallow depth (approximately between 1 m and 3 m), influencing wave propagation and amplification (Martelli and Granati, 2010).

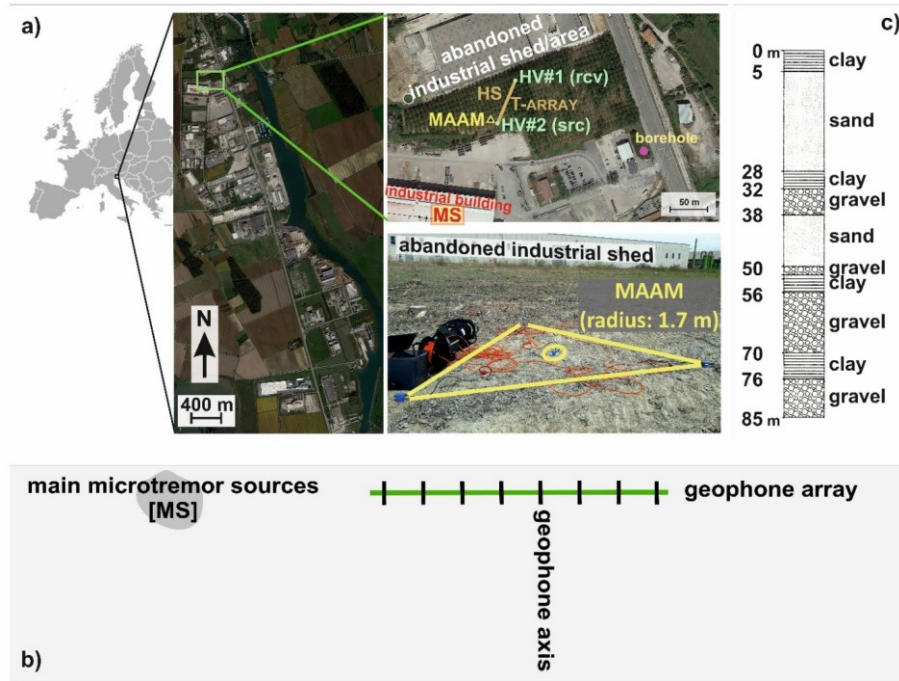


Figure 62: Site overview and acquisition layout. a) Map showing the survey location with deployed methods (HS, MAAM, T-component ESAC, and two HVSR points); MS marks the main source area of local microtremors. Inset: photo of the MAAM triangular array. b) Linear array used for T-component ESAC, aligned with dominant microtremor source direction (c) Stratigraphy from the closest borehole (#1610060). Adapted from Dal Moro and Mazanec (2024).

A set of active and passive datasets was acquired to compare the outcomes of various approaches based on the joint analysis of multiple observables (

Table 10). The underlying assumption is that, in general, increasing the number of independent observables leads to reduced uncertainties and ambiguities in the resulting solution. The acquisition setup included 4.5 Hz horizontal and vertical geophones and a set of 5 Hz 3C nodal sensors.

Passive data were acquired to define three observables: the Rayleigh-wave dispersion curve from miniature array analysis of microtremors (MAAM), the Love-wave dispersion curve from extended spatial autocorrelation (ESAC), and the horizontal-to-vertical spectral ratio (HVSR). Active data for the holistic surface-wave analysis (HS) were extracted from a hybrid acquisition setup using only two 3C geophones, which also enabled HVSR computation at two locations. The resulting observables were integrated using three different approaches:

- (#1) joint inversion of Rayleigh-wave dispersion and HVSR,
- (#2) joint inversion of Rayleigh- and Love-wave dispersion together with HVSR, and
- (#3) joint inversion of multicomponent group velocities, HVSR, and Rayleigh-wave particle motion (RPM) curves.

Table 10: Summary of the techniques used to extract information about surface-wave propagation. Modified from Dal Moro and Mazanec (2024).

<i>technique</i>	<i>notes</i>
MAAM (Z component)	equipment: four vertical-component geophones; triangular array (radius 1.7 m); record time 20 min; investigated frequency range: 2.6-18 Hz
ESAC (T component)	equipment: six horizontal-component geophones; linear array (44 m length); record time 11 min; investigated frequency range: 3-20 Hz
HVSR (3C)	equipment: two 3C geophones; record time 20 min considered frequency range: 0.6–19 Hz
HS (3C)	equipment: 3C geophone; offset: 44 m investigated frequency range: 5.5-41 Hz

Active data were acquired using a single 3C sensor for the application of the Holistic Analysis of Surface Waves (HS) approach (Dal Moro et al., 2018; Dal Moro, 2019a, 2020a), i.e., the joint inversion of multicomponent group velocities together with the RPM and HVSR curves. Passive data were collected to derive three key observables: the HVSR curve, the Rayleigh-wave dispersion curve from the vertical (Z) component using the Miniaturized Ambient Array Method (MAAM) and the Love-wave dispersion curve from the transverse component through the ESAC technique. Three complementary acquisition setups were used:

1. MAAM (Miniaturized Ambient Array Measurements):

The MAAM data were acquired using a triangular array of four vertical component geophones (radius 1.7 m), with three receivers at the triangle's vertices and one at the centre for estimating and compensating the noise (Figure 62 and Figure 63). Ambient vibrations were recorded for 20 minutes and frequency range of 2.6–18 Hz was investigated.

2. Linear Array for Love-wave analysis:

To capture transverse-component microtremors, a linear array of six horizontal geophones was deployed over a total length of 44 m. The array, strategically aligned with the direction of dominant local microtremor sources (MS; see Figure 62), recorded ambient vibrations for 11 minutes. This configuration aimed to ESAC analyses, specifically targeting Love-wave dispersion under conditions of strong directional noise (Figure 62). The investigated frequency range was from 3 to 20 Hz.

3. Hybrid acquisition for HVSR and HS:

For high data acquisition efficiency unique *hybrid* acquisition setup designed to optimize field procedures was conducted. This combines active group data acquisition for HS as well as HVSR ambient vibrations recording. Practically, two 3C sensors were deployed at the ends of the linear array to record 20 minutes of ambient vibrations for HVSR analysis (Figure 62). Part of these 20 minutes was also active data acquisition for the HS. The investigated frequency range for HVSR spanned from 0.6 to 19 Hz and for the HS was 5.5-41 Hz.

3.4.3.2 Observables

Various observables were derived from the complex acquired dataset, as presented in the following section. It should be noted that the data used for the HS approach are discussed in greater detail later in Section "Approach #3".

MAAM

Figure 63 summarize the dispersion retrieved from MAAM (Z component). In addition to the Rayleigh-wave dispersion curve obtained using the noise compensation routine described by Cho et al. (2013).

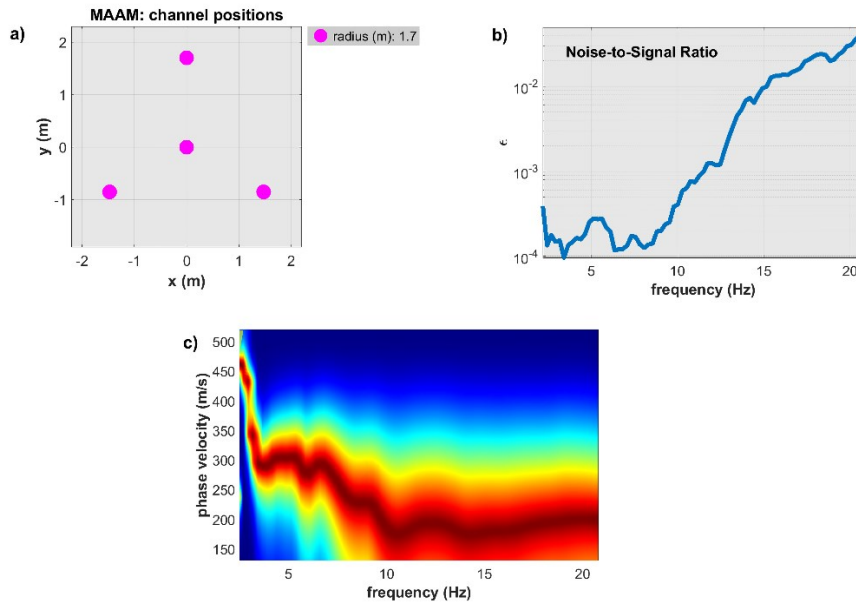


Figure 63: Rayleigh-wave dispersion (vertical component) obtained via MAAM: a) geophone array; b) Noise-to-Signal (ϵ) curve (Cho et al., 2013); c) effective Rayleigh-wave dispersion. Adapted from Dal Moro and Mazanec (2024).

ESAC

The dispersion analysis of the transverse (T) component obtained using the ESAC method is presented in Figure 64. It demonstrates that even with relatively short arrays, a limited number of channels is sufficient to retrieve reliable information within the frequency range relevant for near-surface investigations (see

Table 10). This is supported by the inter-receiver spacing shown in Figure 64, which confirms adequate wavelength sampling. The effectiveness is partly attributed to the simpler propagation behavior of Love waves (Safari et al., 2005, 2006; Dal Moro, 2014).

Aliasing, however, may cause the appearance of spurious straight lines with slopes opposite to those of the actual dispersion curve when using a small number of channels (Dal Moro, 2020a). These artefacts are easily recognizable and can be removed using a basic frequency–wavenumber (f – k) filter (Zhou, 2014), which targets and eliminates components with physically implausible slopes (Figure 64).

The dashed line in Figure 64e and Figure 64f, computed following the formulation by Ohori et al. (2002), estimates the lowest reliable frequency under the assumption that the maximum resolvable wavelength is approximately twice the largest sensor spacing. In this case, the threshold frequency is around 3 Hz, below which the phase velocities exhibit an abrupt and unreliable drop (Figure 64e).

Additional support for the spatial localization of the main microseismic source, positioned in the MS area identified in Figure 62a, is provided by comparing the mean phase velocity spectra computed via the phase-shift method for forward and reverse propagation directions in Figure 64g. The spectrum for the forward direction is well defined and aligns closely with the ESAC-derived dispersion curve shown in Figure 64f, further confirming the validity of the adopted assumptions.

Figure 64c show comparison between experimental spatial autocorrelation functions (blue dotted curves) and theoretical Bessel functions of the first kind (red solid curves) for three selected frequencies. The best-fitting Bessel function defines the Rayleigh-wave phase velocity at each frequency, as used in the ESAC method.

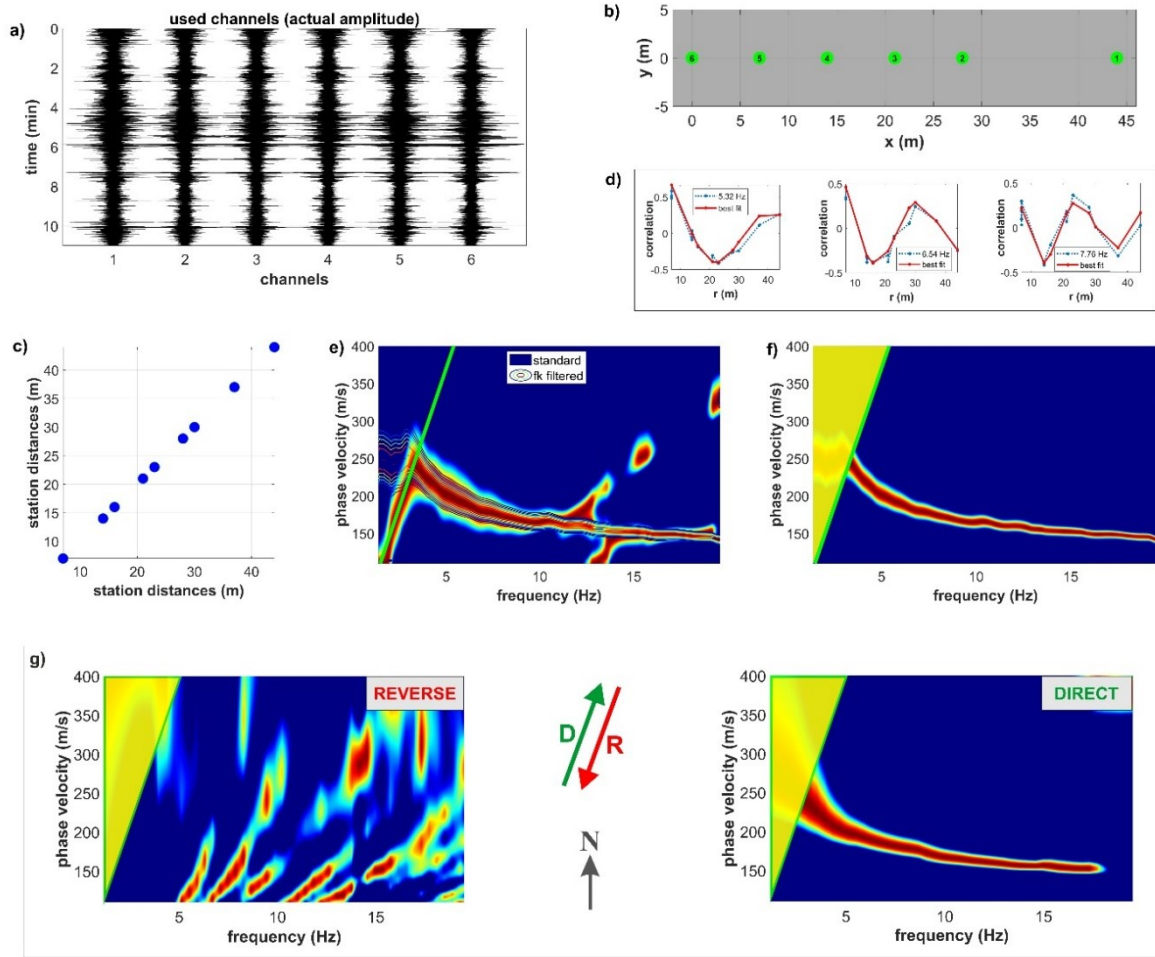


Figure 64: T-component (Love wave passive data): a) field recordings; b) sensor layout map; c) inter-receiver distances; d) Bessel function fits from ESAC for three selected frequencies, showing experimental curves (blue dashed) and the theoretical curves corresponding to the identified phase velocities (red solid); e) phase-velocity spectrum with background colors representing the standard ESAC output and contour lines showing the result after applying an f - k filter; f) f - k -filtered ESAC spectrum (same as the contours in panel e); g) phase-velocity spectra computed using the phase-shift method for both forward and reverse propagation directions showing the effect of MS. In all spectra, the lowest reliable frequencies are also indicated based on the criterion proposed by Ohori et al. (2002). Modified from Dal Moro and Mazanec (2024).

HVSR

To assess the potential presence of significant lateral variations, the acquisition setup included HVSR measurements at both ends of the array (Figure 62a). Although the resulting curves exhibit some expected differences (Figure 65a), the variations are minor enough to justify the assumption of a 1D subsurface structure. For the joint analyses presented in the following sections, the average of the two HVSR curves is used as a representative input (Figure 65b).

The presence of significant monochromatic industrial noise within the frequency range relevant to this study (approximately 0.6–25 Hz) was assessed by analyzing the mean coherence functions for all three sensor combinations (UD–NS, UD–EW, and EW–NS) (Figure

65c), along with the quasi-unsmoothed amplitude spectra (Figure 65d) (Dal Moro, 2020b). The comparison of these datasets revealed several narrow-band industrial components (20 Hz, 25 Hz, 33.5 Hz, 41 Hz, 43.6 Hz, and 44.4 Hz), demonstrating the usefulness of coherence functions in detecting such sources, as they can exhibit distinct peaks even when the corresponding amplitude signatures are weak. For more information about the industrial signals, see Dal Moro and Mazanec (2024) page B215.

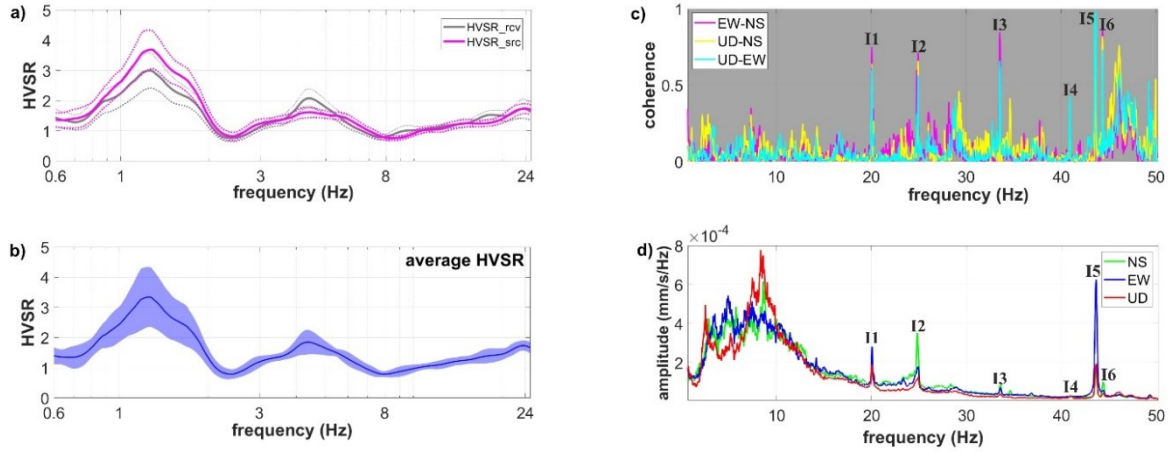


Figure 65: HVSr analysis: a) HVSr curves recorded at the two endpoints of the HS array; b) average HVSr curve with associated standard deviation; c) mean coherence functions computed for the indicated sensor combinations at HV#2 (UD refers to the vertical Up–Down component); d) mean amplitude spectra of the three components at HV#2, smoothed by 0.5%. Dominant monochromatic industrial peaks are marked with "I" labels. Adapted from Dal Moro and Mazanec (2024).

3.4.3.3 Inversion of the considered observables – three different approaches

To identify the most effective procedures for reliably constraining the Vs profile, three alternative approaches are considered for the data inversion. The observables were derived from passive data and their integration through two alternative joint inversion strategies: the first (1) combines Rayleigh- and Love-wave phase velocities with HVSr (three observables), while the second (2) includes only Rayleigh-wave phase velocities and HVSr (two observables). A third (3), fundamentally different approach is also explored, which relies on passive and active data recorded by a single 3C geophone. This method enables the extraction of group velocity spectra for multicomponent data (considering only the vertical and radial components of Rayleigh waves), along with the Rayleigh-wave particle motion (RPM) and HVSr curves, resulting in a total of four observables.

For all combinations of observables, joint inversion is performed using a multi-objective evolutionary algorithm (MOEA). This algorithm is an optimization technique designed to simultaneously minimize multiple, potentially conflicting objective functions. Instead of seeking a single best solution, it generates a set of Pareto-optimal models that represent trade-

offs between the objectives, more can be found in literature (Fonseca and Fleming, 1993; Yapo et al., 1998; Zitzler and Thiele, 1999; Dal Moro et al., 2007; Huang et al., 2015) and from Dal Moro and Mazanec (2024). page B215.

Dispersion data obtained from passive data (approach #1 and #2) are modeled considering the effective dispersion curve, whereas dispersion obtained from multicomponent active data (approach #3) is analyzed according to the FVS technique, which also allows the RPM curve to be modeled (Dal Moro, 2019b). HVSR is computed according to Arai and Tokimatsu (2004), i.e., considering the contribution of Rayleigh and Love waves.

3.4.3.3.1 Approach #1: Joint analysis of Rayleigh- and Love-wave dispersion with the HVSR

The first approach in Figure 66 relies on the joint inversion of three observables derived exclusively from passive data: Rayleigh-wave dispersion obtained through MAAM, Love-wave dispersion from ESAC, and the HVSR curve. During the inversion, shear-wave velocity anisotropy was allowed by distinguishing between VSV and VSH, accounting for potential directional differences in subsurface properties (Babuska and Cara, 1991; Pugin et al., 2013; Russell et al., 2019; Becker, 2020). This flexibility enabled a consistent fit across all observables, including Rayleigh and Love wave dispersion as well as the HVSR curve.

The V_s profile shown in Figure 66d indicates that the shallow layers down to approximately 30 m are constrained by all three observables, while the deeper section is informed solely by the HVSR curve, as the dispersion data extend only to around 3 Hz (see

Table 10: Summary of the techniques used to extract information about surface-wave propagation. Modified from Dal Moro and Mazanec (2024).). The velocity increase at roughly 50 m corresponds to the HVSR peak near 1.25 Hz, whereas the subtle low-velocity layer at approximately 40 m is linked to the HVSR trend around 2.3 Hz.

It is worth noting that the dispersion curve derived from MAAM shows strong agreement with both the Love-wave dispersion and the HVSR curve, providing mutual validation and confirming the overall consistency of the dataset.

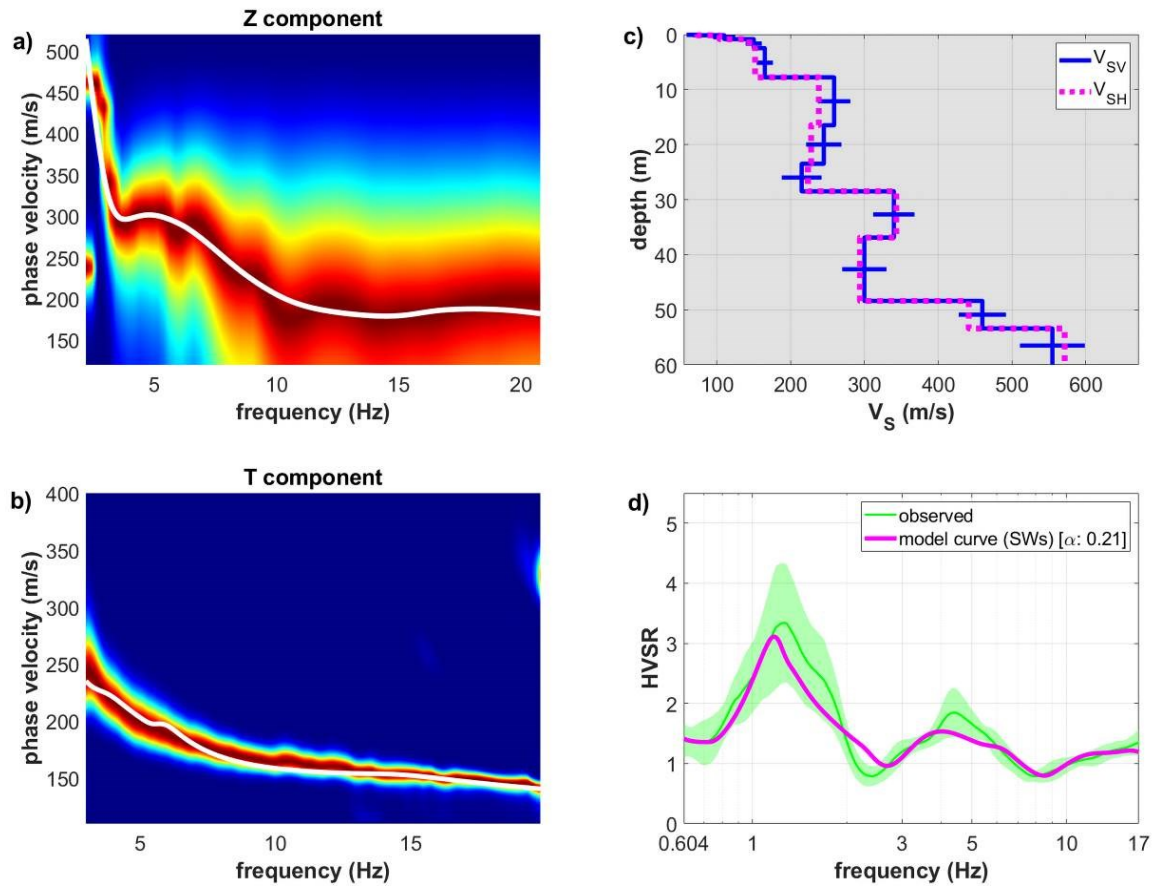


Figure 66: Approach #1 (passive data only): Joint inversion of Rayleigh- and Love-wave dispersion curves combined with HVSR. a) Effective phase velocity dispersion curve for the Z component (Rayleigh waves); b) Effective phase velocity dispersion curve for the T component (Love waves); c) Retrieved shear-wave velocity profile (V_{SV} and V_{SH}), including standard deviations derived from the final Pareto-optimal model set; d) HVSR curves—measured and modelle. In panels a and b, the background colours represent observed field data, while the overlaid curves correspond to the effective dispersion curves of the inverted model. In panel d, the α parameter (0.21, i.e., 21%) denotes the modeled proportion of Love waves within the ambient vibration wavefield, treated as an inversion variable following Arai and Tokimatsu (2004) and Dal Moro (2019a). Modified based on Dal Moro and Mazanec (2024).

3.4.3.3.2 Approach #2: Joint analysis of Rayleigh-wave dispersion and HVSR

This approach represents the common joint analysis of Rayleigh-wave dispersion and HVSR (Arai and Tokimatsu, 2005; Picozzi and Albarello, 2007; Asten et al., 2014). The obtained results

are summarized in Figure 67. Although the misfit values appear satisfactory and may suggest a reliable result, comparison with the solution incorporating both Rayleigh and Love waves in approach #1 (see Figure 66 and related discussion) reveals that relying solely on Rayleigh waves can lead to inaccurate models, typically characterized by overestimated shear-wave velocities. This topic is covered in more detail in Dal Moro and Mazanec (2024) page B217.

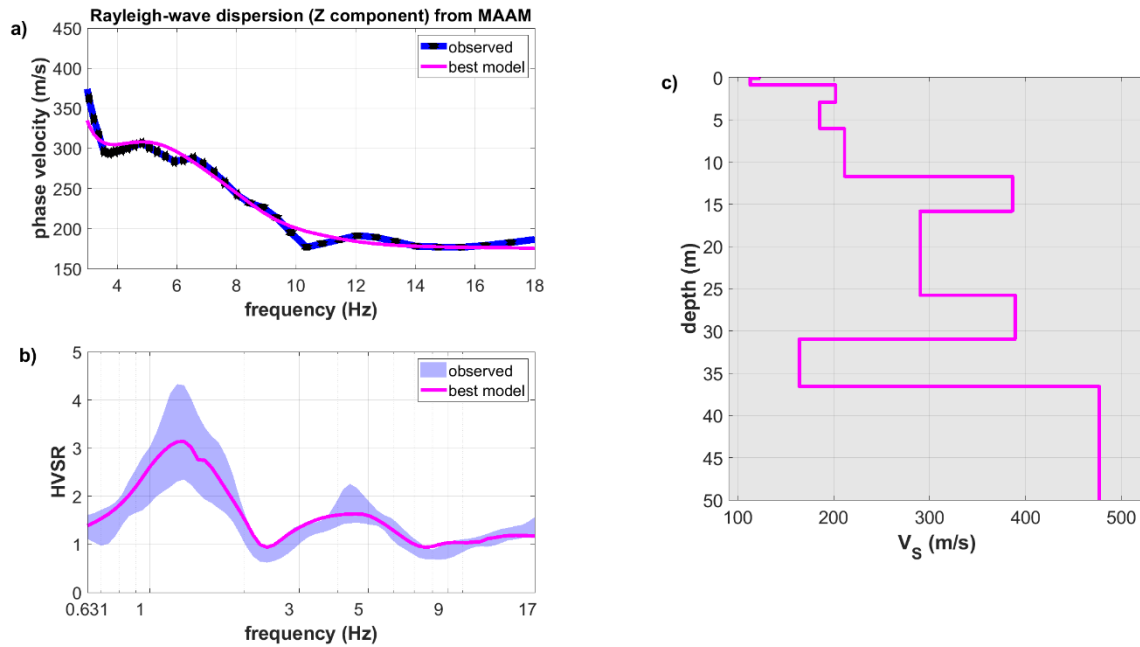


Figure 67: Approach #2: Results of the joint inversion of the effective phase velocity dispersion curve for the Z component (a) and the HVSR (b). Due to the ambiguities inherent in both the Rayleigh-wave dispersion and the HVSR curve, the resulting shear-wave velocity profile (c) tends to overestimate the actual velocities. For comparison, refer to the results of approach #1 presented in Figure 66. Adapted from Dal Moro and Mazanec (2024).

3.4.3.3 Approach 3: Holistic analysis of Rayleigh waves and HVSR

An additional advanced surface-wave analysis approach involves computing group velocities from multicomponent (single-offset) data and jointly interpreting them with the Rayleigh-wave Particle Motion (RPM) curve, which characterizes the true particle motion associated with Rayleigh-wave propagation (Dal Moro et al., 2017). This combined analysis provides valuable constraints for refining the subsurface model (Dal Moro et al., 2018). It is relatively uncommon and complex within near-surface geophysics, but it deserves attention. Here, it is covered only briefly, with reference to a more detailed description in Dal Moro and Mazanec (2024), page B218.

To optimize field time and efficiently acquire data for computing HVSR curves at the two endpoints (Figure 68), an unconventional hybrid acquisition procedure was employed, combining both passive and active data collection. Two 3C geophones were placed 44 meters apart and recorded ambient vibrations for 20 minutes. During the first 2 minutes, active source signals were generated by repeatedly striking near one of the geophones using a 10 kg sledgehammer (see layout in Figure 68a). These impulses were later extracted and stacked to

form a conventional active dataset, while the remaining portion of the recording was used to calculate the HVSR curves at the two endpoints (S and R in Figure 68a).

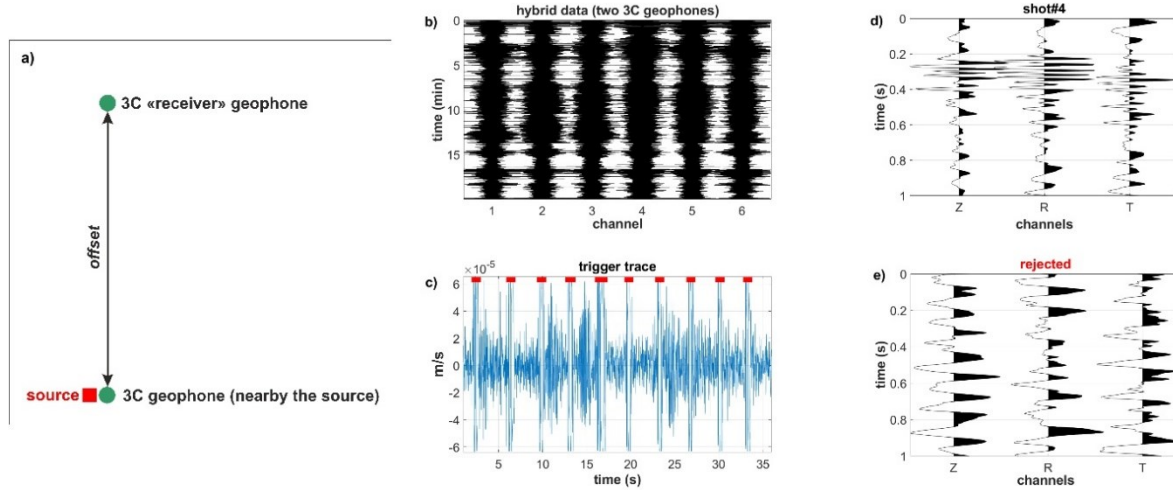


Figure 68: Extraction of active data from hybrid recordings using two 3C geophones: a) Acquisition geometry; b) All six recorded channels (the first three correspond to the receiver geophone, the last three to the 3C geophone near the source); c) Section of the fifth channel (radial component of the source-side geophone) used as a trigger to identify the origin time of each active shot; d) Example of an extracted shot at the receiver geophone showing the Z, R, and T components, with time zero aligned to the trigger trace; e) Example of a low-quality shot excluded during preprocessing and not used in the stacking of active data. Adapted from Dal Moro and Mazanec (2024).

In the preprocessing good shots (e.g., Figure 68d) were selected and the bad ones (e.g., Figure 68e) rejected. The good ones were then stacked to increase the SNR and use the obtained traces for the computation of the group velocity spectra with MFA.

The Rayleigh-wave particle motion (RPM) curve characterizes the elliptical movement of particles in the vertical–radial plane caused by Rayleigh wave propagation. By analyzing the frequency-dependent ratio and orientation of vertical and radial components, RPM provides additional constraints on subsurface shear-wave velocity structure. This observable is particularly useful for identifying velocity contrasts and improving inversion robustness when combined with other surface-wave observables (Dal Moro et al., 2017; Dal Moro, 2019a).

In the HS approach, the dispersion information contained in the group velocity spectra is not interpreted in terms of individual modal curves but is instead analyzed using the multicomponent Full Velocity Spectrum (FVS) method (Dal Moro, 2014, 2019a, 2019b).

Figure 69 summarizes the extracted observables—group velocity spectra for the Z and R components, along with the RPM curve—and presents the HS solution, which is obtained through joint inversion of the multicomponent group velocities, RPM, and HVSR curves. The inclusion of the HVSR helps to extend the depth of investigation within the resulting Vs profile.

The resulting Vs profile (Figure 69) appears very similar to that obtained through the joint inversion of phase velocity data from the Z and T components combined with the HVSR (approach #1; see Figure 66). In both cases, the HVSR peaks at approximately 1.2 Hz and 4.4 Hz can be linked to shear-wave velocity increases at depths of around 48 m and 8 m, respectively.

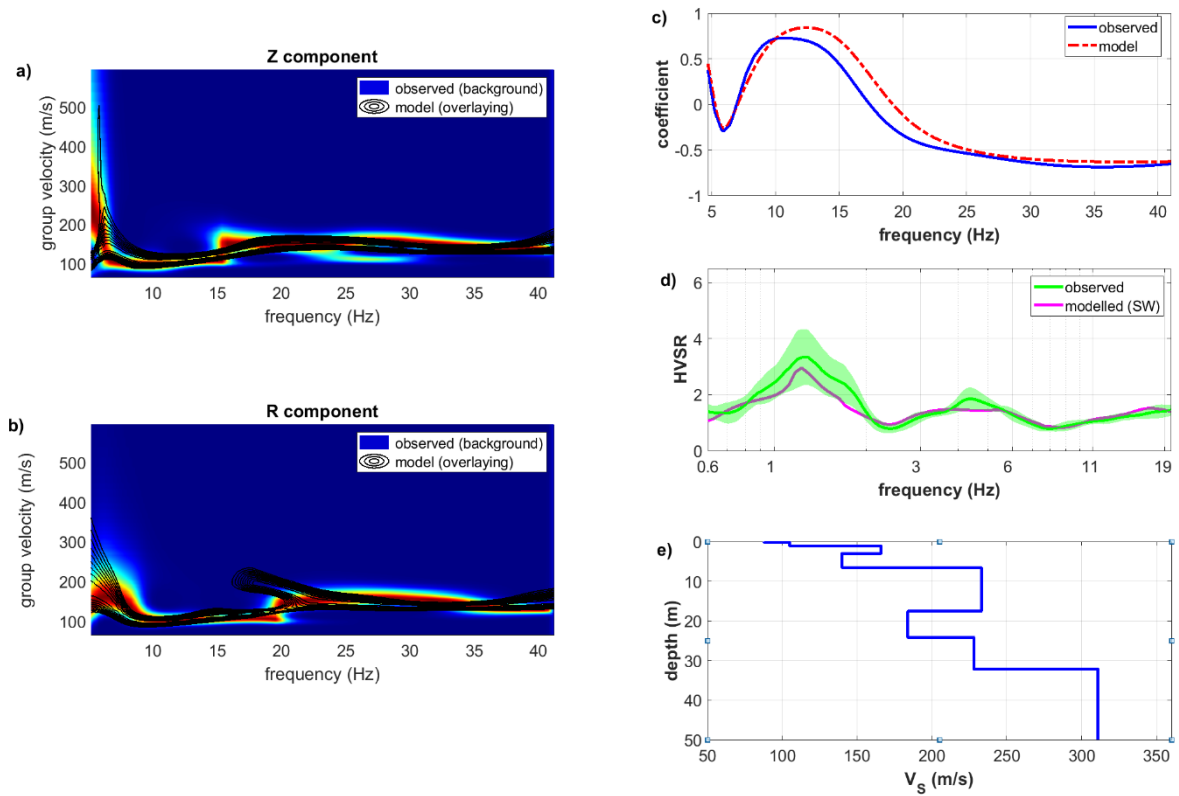


Figure 69: Holistic analysis of the group velocities of the vertical (Z) (a) and radial (R) (b) components (FVS approach – background colors represent the field data while the overlying black contour lines the velocity spectra of the identified model) jointly with the RPM (c) and HVSR (d) curves. The obtained V_s profile (e) is entirely similar to the one obtained while considering the joint analysis of the phase velocities of the Z and T components. Adapted from Dal Moro and Mazanec (2024).

Finally, Figure 70 compares the V_s profiles resulting from the three approaches. The solutions from approaches #1 and #3 show a high degree of similarity, while the profile derived from the approach #2—joint inversion of the Rayleigh-wave dispersion curve and HVSR alone (using only two observables) exhibits a notable deviation from the other two.

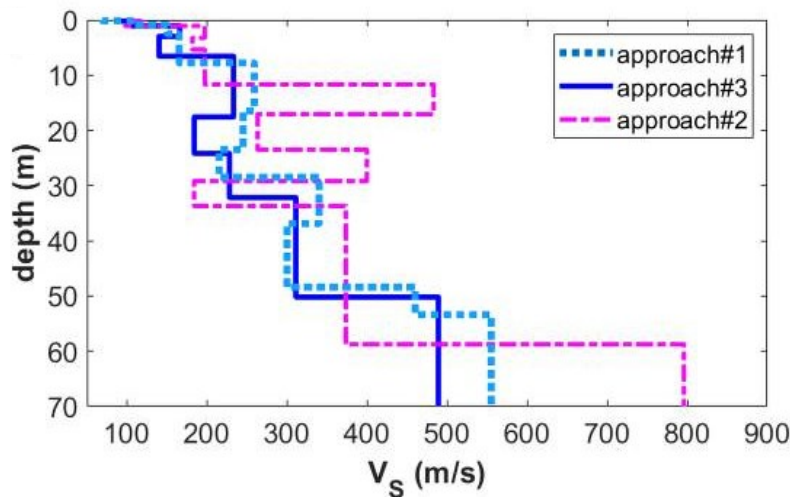


Figure 70: Comparison of the V_s profiles obtained through the three considered approaches (#1-#3). Modified from Dal Moro and Mazanec (2024).

3.4.3.4 Discussion and Conclusion

The data and analyses presented in the preceding sections lead to the following key observations.

- 1) Despite the high-amplitude microtremor noise from ongoing industrial activity, surface-wave analysis yielded high-quality data suitable for constraining the inversion—something that would be difficult to achieve using lower-amplitude body-wave methods such as refraction or reflection. Passive techniques such as MAAM, which require careful attention during both data acquisition and processing, proved to perform remarkably well even under conditions of intense industrial activity.
- 2) When dealing with passive data, surface-wave modeling cannot rely on a modal dispersion curve-based approach; instead, it is essential to adopt a framework based on the mathematics of the effective dispersion curve.
- 3) The intrinsic ambiguity of Rayleigh-wave dispersion—particularly in its effective form, which is also influenced by parameters such as VP and QS (see Dal Moro and Mazanec (2024), Appendix A)—often necessitates the inclusion of additional observables to resolve otherwise unavoidable uncertainties. In this context, Love waves can play a crucial role in achieving more robust and unambiguous subsurface models, especially when analyzing phase velocities. Importantly, Love-wave dispersion can be reliably extracted from passive data, even when using linear array configurations.
- 4) Even when employing a joint approach designed to reduce model ambiguity, a certain level of uncertainty in the resulting Vs profile remains unavoidable.
- 5) Despite the use of simple field procedures and a single 3C geophone, the FVS-based analysis of multicomponent group velocities and RPM curves enables the retrieval of robust S-wave velocity profiles. This intrinsically multimodal and objective approach effectively addresses spectral distortions, and its combination with HVSR further constrains the inversion and extends depth sensitivity.

These results confirm that incorporating multiple observables—especially from different wave types and components—significantly improves model reliability in noisy or structurally complex environments. Presented study showed possible ways of obtaining robust Vs model in challenging environment, such as noisy urban area and highlighted related issues and recommendations. The acquired insights and methodologies can be effectively applied to site response mapping in urban areas. For more details, see Dal Moro and Mazanec (2024).

4 SITE CONDITION CHARACTERIZATION USING SURFACE WAVES

This chapter demonstrates how surface-wave methods can be effectively applied for the characterization of local site conditions, bridging the theoretical and methodological background introduced in Chapters 2 and 3. While Chapter 2 reviewed the concept and importance of site effects in seismic hazard assessment, and Chapter 3 detailed the principles and processing of surface waves, this section integrates these topics through three applied case studies. Each example illustrates how surface-wave dispersion analysis can contribute to evaluating subsurface stiffness, quantifying proxies such as V_{s30} , and ultimately supporting seismic hazard assessments or engineering site classifications. The studies vary in complexity, from basic surface-wave surveys at seismic stations (Section 4.1), through the integration of V_{s30} into a national seismic hazard map and its correlation with topographic proxies (Section 4.2), to the development of a rapid and inversion-free method for V_{s30} estimation (Section 4.3). Together, they emphasize the versatility and relevance of surface-wave analysis in modern site response characterization.

4.1 Surface Waves Site Characterization of WEBNET Seismic Stations

This chapter presents the site characterization of seismic stations within the West Bohemia Seismic Network (WEBNET) network, based on surface wave analysis. The work builds upon the study by Mazanec et al. (2024) and serves as the foundation for the analysis in Chapter 2.5, which uses the derived parameters, particularly V_{s30} and H_{800} , to investigate the relationship between shear-wave velocity and seismic amplification. In addition to supporting amplification analysis, the aim of this study was to characterize the station sites in detail and derive realistic 1D shear-wave velocity profiles, which may also prove useful for determining station-specific constants.

4.1.1 Site description

The West Bohemia/Vogtland region (Figure 71), located along the Czech German border, forms the western margin of the Bohemian Massif and lies within a transition zone between three major Variscan units: the Teplá-Barrandian, Moldanubian, and Saxothuringian (Babuška et al., 2007; Kolínský et al., 2011; Fischer et al., 2014). The area is intersected by two dominant neotectonic structures: the ENE–WSW trending Eger Rift and the NNW–SSE striking Mariánské Lázně Fault (Prodehl et al., 1995; Dèzes et al., 2004; Blecha et al., 2018). The bedrock consists predominantly of Proterozoic to early Paleozoic metasedimentary rocks of low to medium metamorphic grade (Czech Geological Survey, 2004).

This region is covered by high-quality WEBNET, which offers a dense, instrumentally homogeneous array designed to monitor weak local seismicity, covering over 900 km² with 25 broadband, three-component stations (Horálek and Vavryčuk, 1985; Institute of Geophysics, 1991; Fischer et al., 2010, 2014). Seventeen accessible stations were selected for this study (as marked in Figure 71).

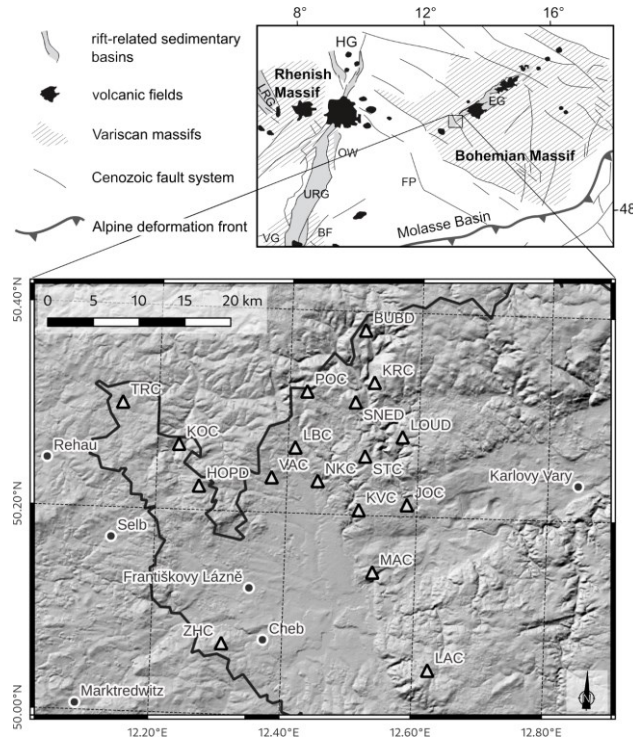


Figure 71: Simplified tectonic map of Central Europe based on Dèzes et al. (2004), with an inset showing a digital elevation model of the West Bohemia/Vogtland region. WEBNET seismic stations are indicated by triangles with corresponding station codes. Major cities are shown as black circles. Abbreviations: BF – Black Forest, EG – Eger (Ohře) Graben, FP – Franconian Platform, HG – Hessian Grabens, LRG – Lower Rhine (Roer Valley) Graben, OW – Odenwald, URG – Upper Rhine Graben, VG – Vosges. Adapted from Mazanec et al. (2024).

4.1.2 Surface Waves Analysis for Vs30 Mapping

To derive shear-wave velocity (V_s) models and estimate V_{s30} for 17 WEBNET stations, the Multichannel Analysis of Surface Waves method was applied. To ensure that the 1D V_s model accurately represents the station location, the seismic profile was centred on the station position. Data acquisition was performed using a 24-channel linear array with 4.5 Hz vertical geophones. Receiver spacing ranged from 3 to 5 meters, depending on local terrain and site conditions. To ensure a high NSR, eight vertically stacked 10 kg sledgehammer impacts on a metal plate were conducted at two offset distances (typically 2.5 and 10 m) at both ends of each profile. Additionally, shot gathers for seismic refraction tomography were recorded from five in-profile shot points. The sampling time was set to 0.063 ms and acquisition time ranged from 1 to 2 seconds to capture all wave types.

Shot gathers from both ends of each profile were transformed into the phase velocity–frequency domain using the Phase-Shift method to generate Rayleigh wave dispersion spectra. Fundamental mode dispersion curves were picked in the ~5–45 Hz range (Figure 72), typically averaged from four source positions.

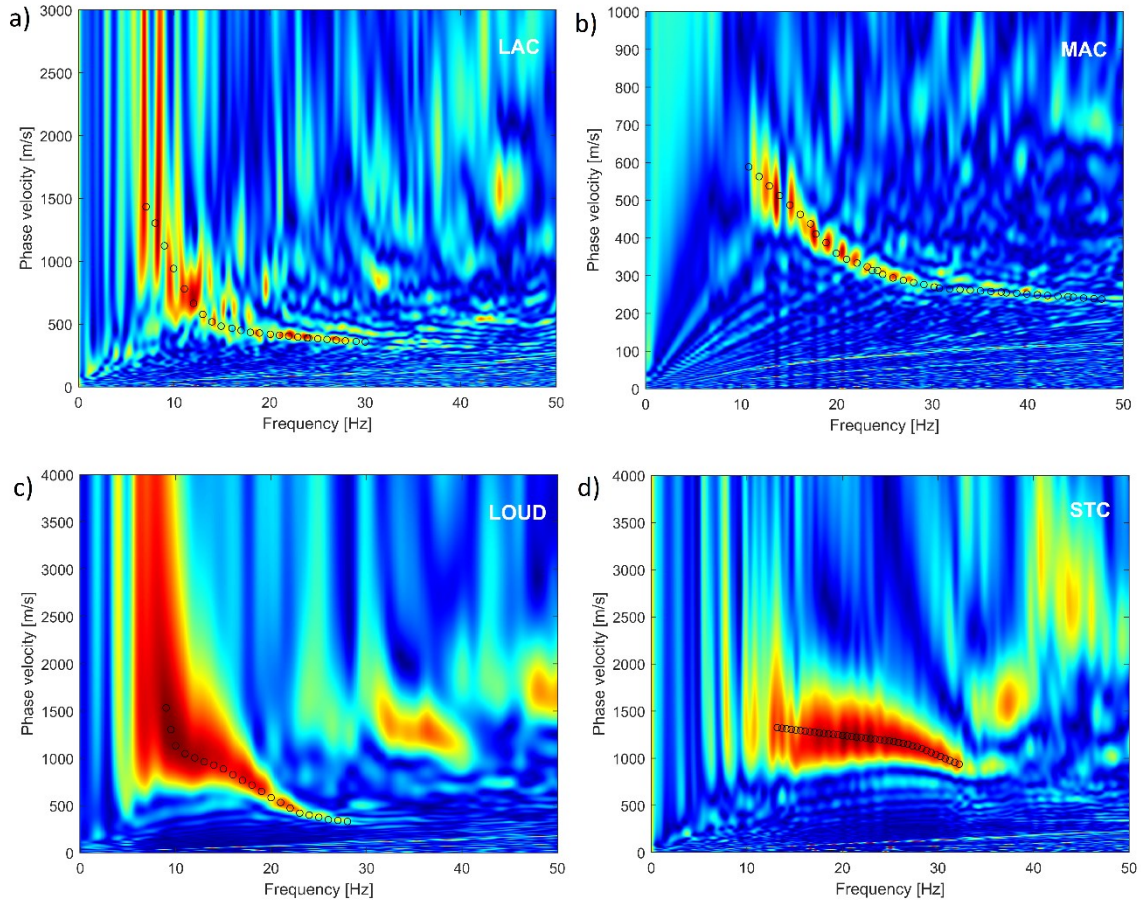


Figure 72: Rayleigh-wave dispersion spectra with picked fundamental mode dispersion curves (black circles) for four WEBNET sites: a) LAC; b) MAC; c) LOUD and d) STC. Adapted from Mazanec et al. (2024).

In the inversion step, first an initial V_s model was constructed and its dispersion curve calculated. Using a non-linear least squares method (LSM), the model was iteratively adjusted to minimize the misfit with the observed curve. Regularization was applied to control model smoothness and prevent over-parametrization. The root means square error (RMSE) between observed and theoretical dispersion curves quantified the fit (Table 11). Finally, the V_{s30} was estimated and $H800$ values were retrieved from the V_s models (Table 11).

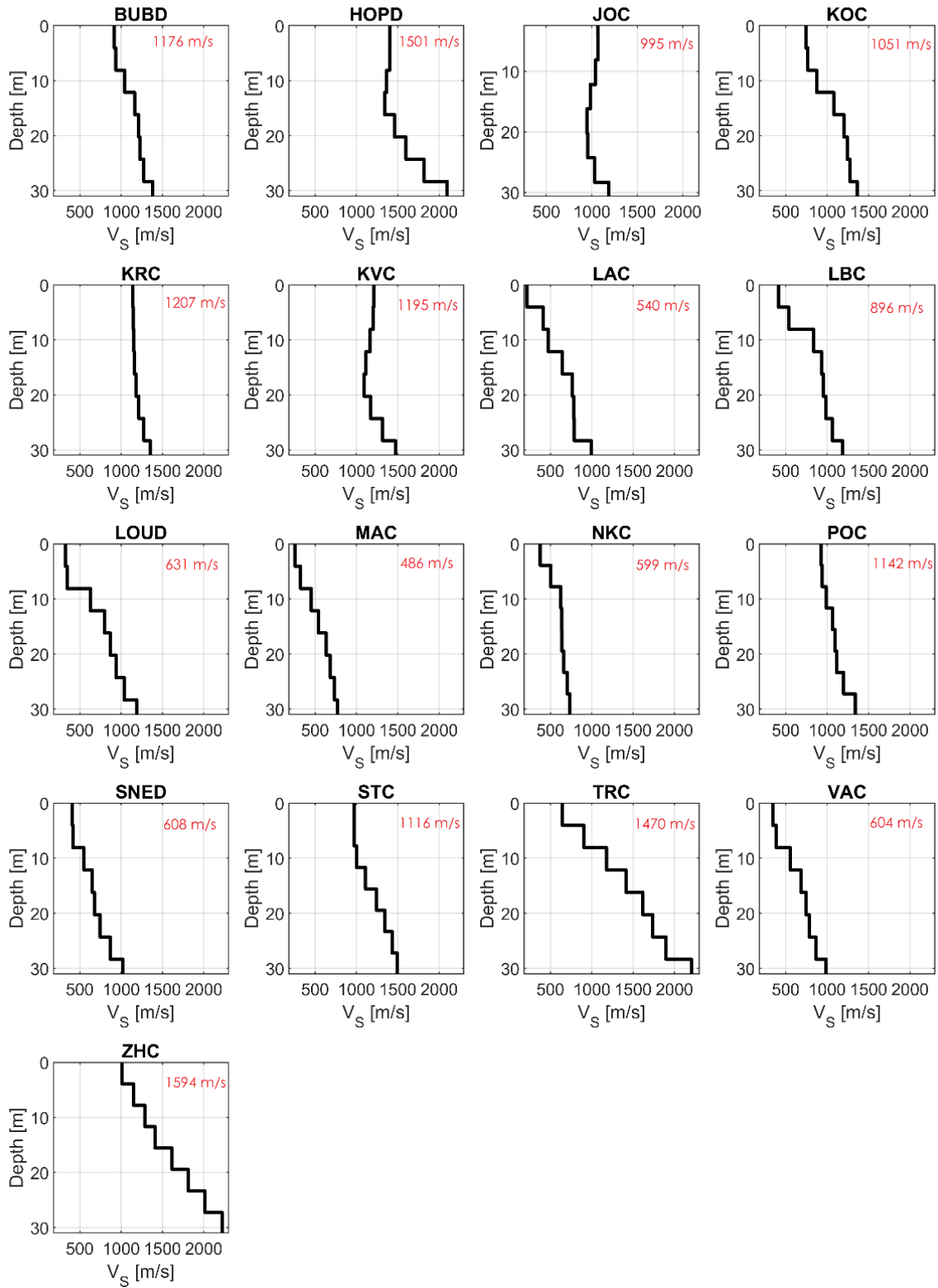


Figure 73: Shear-wave velocity profiles for 17 WEBNET sites obtained using the MASW method. Vs30 values are highlighted in red. For additional details, see Table 11. Modified from Mazanec et al. (2024).

To further constrain the inversion, the resulting shear-wave velocity models were cross-validated using P-wave velocity refraction tomography based on the same dataset, supplemented by additional shot points within the geophone array.

Table 11: Site characterization of 17 WEBNET stations, including location name, station code, geographic coordinates, actual seismic station installation depth, RMSE of the surface wave inversion, Vs30 estimated from MASW, site classification according to EC8-1 (CEN 2004), H800 parameter, bedrock geology based on the 1:50 000 geological map (Czech Geological Survey, 2004), and bulk density derived from the 1:200 000 density mapping (Blizkovsky and Novotny, 1981). Adapted from Mazanec et al. (2024).

Name	Station code	Coordinates		depth [m]	RMSE [%]	Vs30 [m/s]	EC8 class	H800 [m]	Bedrock geology	Density [g/cm ³]
		[°N]	[°E]							
Bublava	BUBD	50.38	12.51	1.5	0.6	1176	A	0	quartzite	2.7
Horní Paseky	HOPD	50.22	12.27	1.6	1.3	1501	A	0	paragneiss	2.67
Josefov	JOC	50.21	12.58	2	0.7	995	A	0	schist	2.65
Kopaniny	KOC	50.26	12.23	3	0.7	1051	A	7	quartzite	2.68
Kraslice	KRC	50.33	12.53	2.5	0.3	1207	A	0	quartzite	2.7
Květná	KVC	50.20	12.51	3	0.5	1195	A	0	mica schist	2.65
Lazy	LAC	50.05	12.62	4	1.7	540	B	26.8	metamorphosed granite	2.53
Luby	LBC	50.26	12.41	4.5	0.4	896	A	8.9	phyllite	2.77
Loučná	LOUD	50.28	12.58	1.6	1.6	631	B	13.2	mica schist	2.61
Chlum sv. Máří	MAC	50.14	12.54	4	0.6	486	B	>30	mica schist	2.6
Nový Kostel	NKC	50.23	12.45	5	0.3	599	B	>30	mica schist	2.7
Počátky	POC	50.32	12.43	2	1.1	1142	A	0	phyllite	2.77
Sněžná	SNED	50.31	12.50	1.5	0.6	608	B	23.7	phyllite	2.7
Studenec	STC	50.26	12.52	3	0.5	1116	A	0	phyllite	2.68
Trojmezí	TRC	50.30	12.14	2	1.3	1470	A	3.7	phyllite	2.7
Vackov	VAC	50.23	12.38	5	0.7	604	B	23.7	mica schist	2.61
Zelená hora	ZHC	50.07	12.31	1.5	0.9	1594	A	0	phyllite	2.77

4.1.3 Discussion and conclusion

The surface wave analysis conducted at 17 seismic stations of the WEBNET network yielded detailed 1D shear-wave velocity models (Figure 73) and key site parameters, including Vs30 (Figure 74), H800, density, and interpreted lithology (Table 11). These results provided a consistent framework for site classification and form the basis for the amplification analysis presented in Chapter 2.5. In addition to enabling comparisons with recorded ground motion data, the obtained profiles offer valuable input for estimating station-specific parameters (e.g., site transfer functions or station constants) that are critical for advanced seismological processing and modelling.

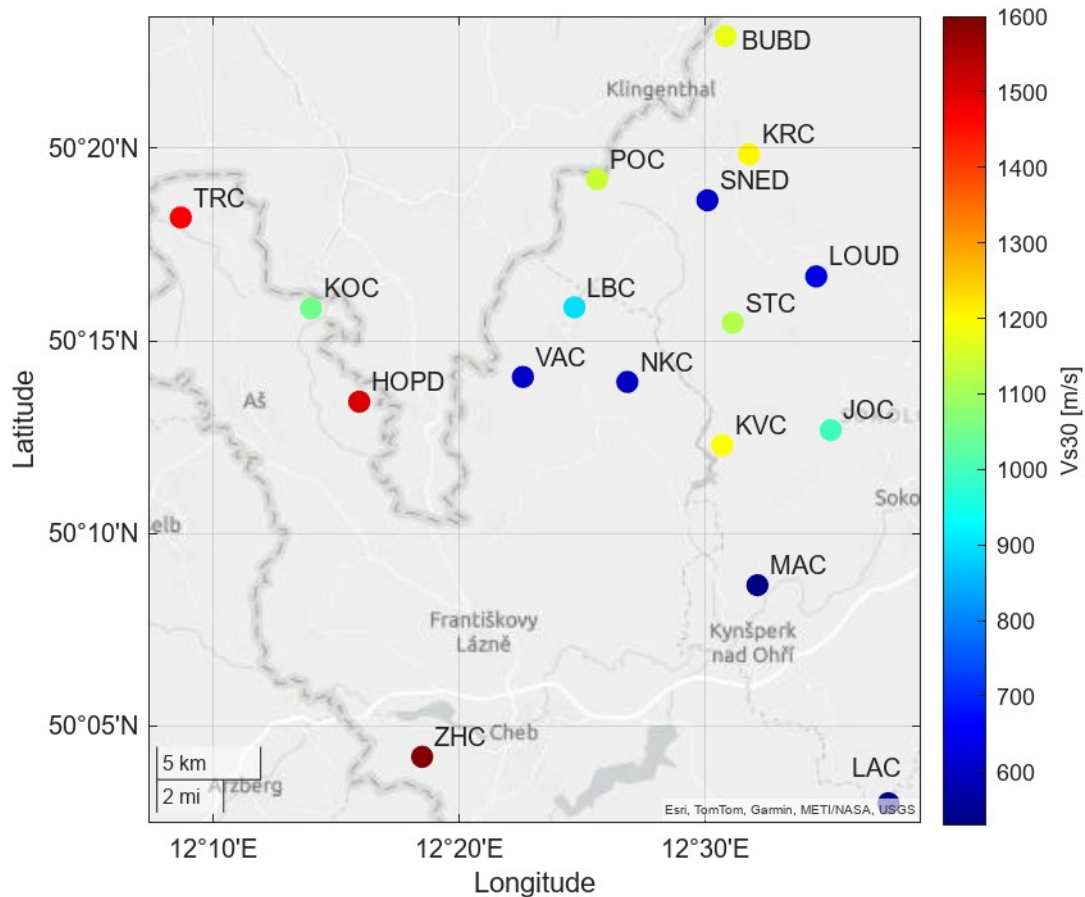


Figure 74: Spatial distribution of Vs30 values derived from surface wave analysis at 17 stations of the West Bohemia Seismic Network (WEBNET). Each station is color-coded according to its estimated Vs30 value, illustrating the variability of near-surface shear-wave velocities across the network. The color gradient reflects increasing stiffness of the subsurface, with lowest values (dark blue) corresponding to softer sediments and highest values (dark red) indicating hard rock conditions. This overview highlights the heterogeneity of local site conditions in the West Bohemia/Vogtland region.

The study demonstrated that surface waves analysis can be employed to classify soil conditions and infer site classes. One intuitive approach is to relate Vs30 to density and lithological type. While a trend is observed, particularly among acidic rocks and their derivatives such as quartz-rich sediments and low-grade metamorphics, this correlation becomes unreliable when mafic lithologies are included. These high-density mafic rocks,

typically of volcanic origin, tend to weather more intensely due to their lower resistance to chemical breakdown, as described by Bowen's reaction series (Bowen, 1922). The subsequent reduction in mechanical stiffness, especially near the surface, leads to lower V_{s30} values and, in turn, higher amplification potential (Barton, 2006).

A remarkable observation is that all EC8 Class B stations in this dataset exhibit shallow low-velocity zones and steep near-surface velocity gradients within the upper 10 meters. This pattern suggests that near-surface weathering is a key factor influencing local shear-wave velocities and thus site amplification. Despite relatively small variations in bulk density across the stations and comparable lithological settings, the differences in V_{s30} and inferred amplification imply that mechanical degradation of the uppermost layers - rather than rock type alone - is likely the dominant control.

Overall, this analysis provided a robust and internally consistent dataset for site classification within the WEBNET network. The V_{s30} and $H800$ values derived from this work were directly used in the subsequent amplification study (Chapter 2.5), enabling an integrated evaluation of how local conditions influence seismic ground motions. Moreover, the detailed site profiles contribute to improving the seismic characterization of this dense monitoring network and can be directly applied in seismic hazard modeling, station corrections, and ground-motion simulations.

4.2 Seismic Hazard Map of the Czech Republic

A new seismic hazard map for the Czech Republic has been recently developed using probabilistic (PSHA) approach, Vackář et al. (2024), currently under review. As a part of this study one key step was to characterize the local site conditions in the national scale.

The V_{s30} parameter is typically determined in the field through dispersion analysis of Rayleigh and Love waves (as previously discussed). However, due to the high spatial variability of local near-surface conditions, obtaining densely spaced V_{s30} measurements over large regions is impractical. As a result, indirect proxies are often employed and later converted to V_{s30} , such as terrain slope. This proxy assumes that low V_{s30} values are associated with unconsolidated, erosion-prone sediments that amplify seismic waves and tend to occur in areas with gentle slopes (Wald and Allen, 2007). With the increasing availability of high-resolution digital elevation models (DEMs) globally, this method has become an important tool for large-scale V_{s30} mapping.

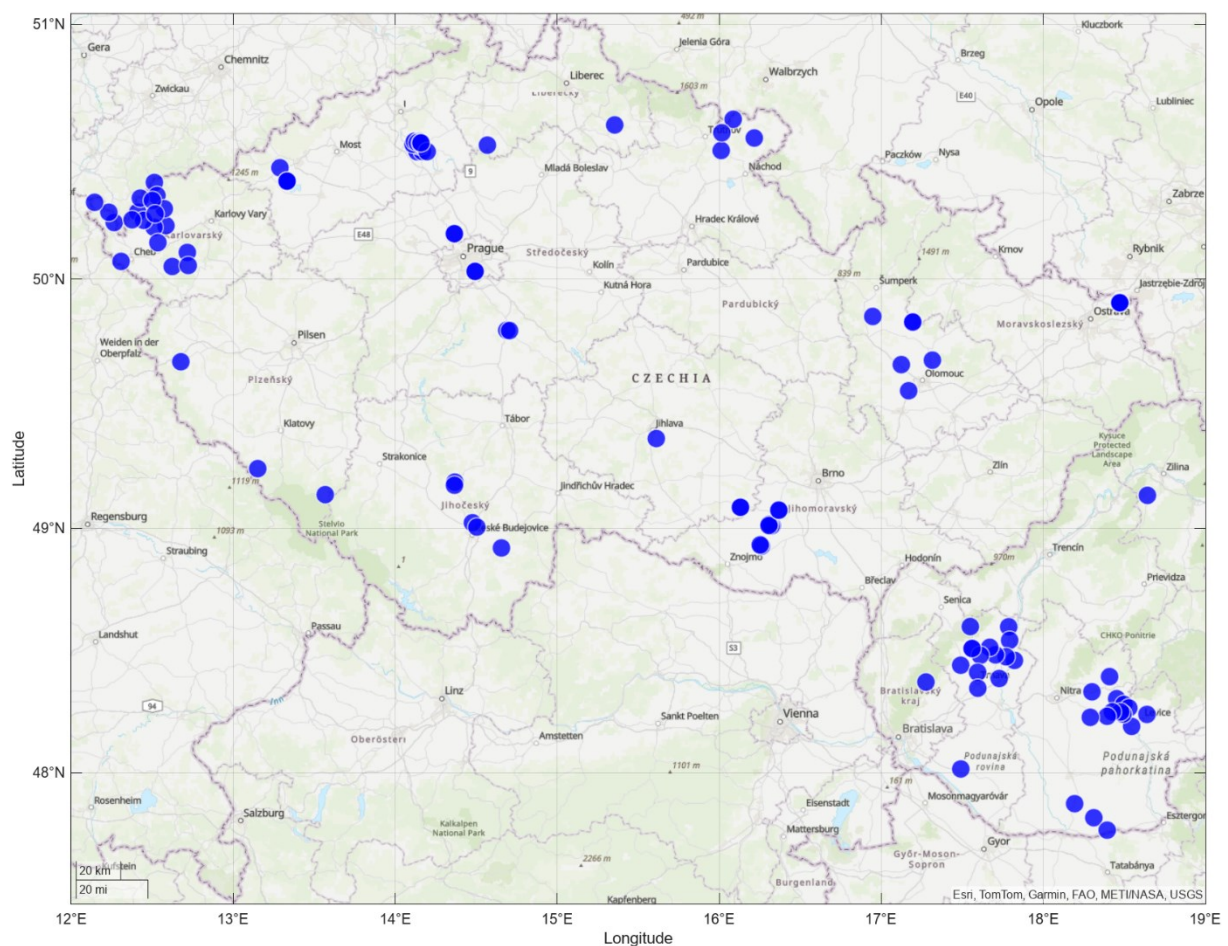


Figure 75: Map of locations with in-situ V_{s30} measurements conducted using surface wave methods. Measurement sites are marked with blue points.

4.2.1 Topography-based Vs30 estimations

Topography-based Vs30 estimations, while useful for large-scale mapping, are subject to considerable uncertainty due to their indirect nature and oversimplified assumptions. These estimates often suffer from systematic errors related to regional geological variability, such as the presence of soft sediments in steep terrain or stiff materials in flat areas, which violate the basic premise of the slope–Vs30 correlation. To improve their reliability, locally in-situ estimated Vs30 data can be used to calibrate the slope–Vs30 relationship, enabling the development of region-specific regression models that correct for these discrepancies and enhance the accuracy of the derived Vs30 values.

Local site conditions in the new seismic hazard map are based on Vs30 values estimated using the slope-based methodology proposed by Wald and Allen (2007), which relies on digital elevation data. Terrain slope was computed from high-resolution digital elevation models provided by the State Administration of Land Surveying and Cadastre of the Czech Republic and the Geodesy, Cartography and Cadastre Authority of the Slovak Republic. To minimize local biases, the original irregular-resolution grids were first resampled to a uniform 50 × 50 m grid prior to slope calculation. (In contrast, the previous version of the hazard map used the SRTM model, which lacked sufficient accuracy, particularly in forested and urbanized areas).

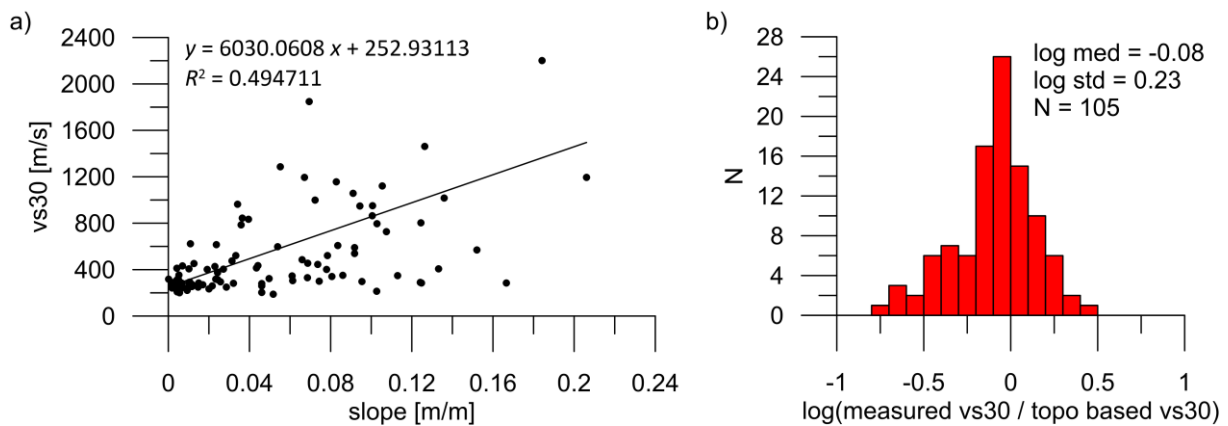


Figure 76: a) Correlation between terrain slope and measured Vs30 values based on linear regression; b) histogram showing the distribution of residuals - the differences between measured and slope-predicted Vs30 values. These results illustrate the variability and limitations of the slope-based estimation method. Adapted from Vackář et al. (2024).

Vs30 values were measured at over 100 locations, see Figure 75, using surface waves analysis (mainly MASW and SPAC). The sites were selected to capture a representative range of geological, lithological, and topographical conditions, from crystalline bedrock to platform sediments, and from lowland to mountainous terrain. A linear regression was then applied to relate terrain slope to the measured Vs30 values, and the coefficient of determination (R^2) was computed (Figure 76a) to enable Vs30 estimation across the entire DEM. The measured Vs30 values were subsequently compared with the regression-based predictions to assess the residuals and overall fit (Figure 76b). While the discrepancies between measured and

predicted values may seem substantial, they are consistent with those reported by Wald and Allen (2007).

To demonstrate the spatial validity and geological consistency of the regionally calibrated topography-based V_{s30} model, Figure 77 presents its comparison with various geological and geomorphological datasets in the example region of Týnec nad Labem (Czechia). A clear agreement is observed between zones of lower V_{s30} values and areas of low elevation and Quaternary sedimentary deposits (panel d), while higher V_{s30} values align with elevated terrains and more consolidated bedrock units. The main fault structures correspond well with strong lateral V_{s30} gradients (panel e), indicating that the V_{s30} model effectively captures major subsurface contrasts relevant for site classification and seismic response assessment.

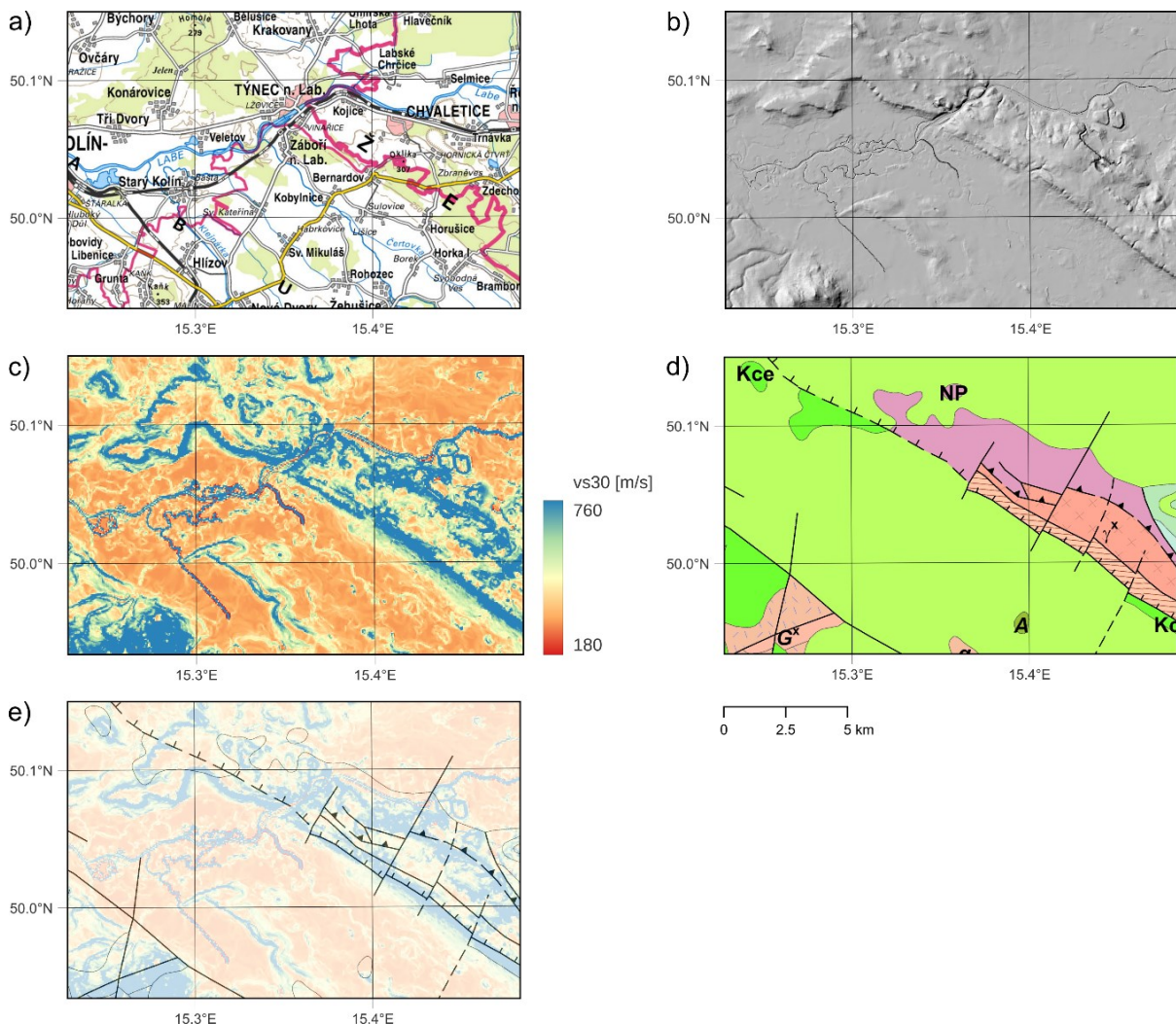


Figure 77: Multilayer spatial characterization of the studied example area near Týnec nad Labem (Czechia): a) Overview map with administrative and transport infrastructure based on Czech base map data; b) Digital elevation model (DEM) showing topographic features and relief variations across the region; c) Estimated V_{s30} distribution map derived from a proxy model based on topographic slope analysis (regionally calibrated). The values range from 180 m/s (soft sediments) to 760 m/s (rocky terrain); d) Geological map displaying major lithological units and tectonic structures, with the trace of the main fault zone emphasized. e) Composite visualization of the V_{s30} map with superimposed tectonic lineaments and selected cross-sections (dashed lines), highlighting the correlation between surface conditions, geology, and seismic site response (author: Jan Valenta).

4.2.2 Discussion and conclusion

The corrected Vs30 values were employed as a proxy for local site conditions in the development of a nationwide seismic hazard model for Czechia. These values were integrated into a high-resolution 50×50 m terrain-slope-based grid and used to assign site classes necessary for selecting appropriate PGA and SA response spectra in seismic hazard calculations. The resulting hazard outputs were implemented into an interactive web platform seismic hazard map of Czechia (www.seismickamapa.cz, Figure 79), which enables users to generate location-specific seismic hazard reports that incorporate site amplification effects based on Vs30.

The topography-based approach to regional Vs30 estimation provides an efficient first-order screening tool for identifying potentially vulnerable sites where local geological conditions may lead to significant ground motion amplification. However, as discussed in previous chapters, such proxy-based methods are subject to limitations, including uncertainty in terrain-slope based Vs30 correlations, local geological heterogeneities, and the smoothing effects of empirical calibration. To mitigate these uncertainties, the approach was calibrated using more than 100 in-situ measurements of Vs30 across various geological environments in Czechia. This regional calibration notably improved the reliability of the model, as shown in Figure 77, which illustrates a reasonable spatial correlation between topography-derived Vs30 and surface geological, tectonic, and geomorphological features in the Týnec nad Labem region.

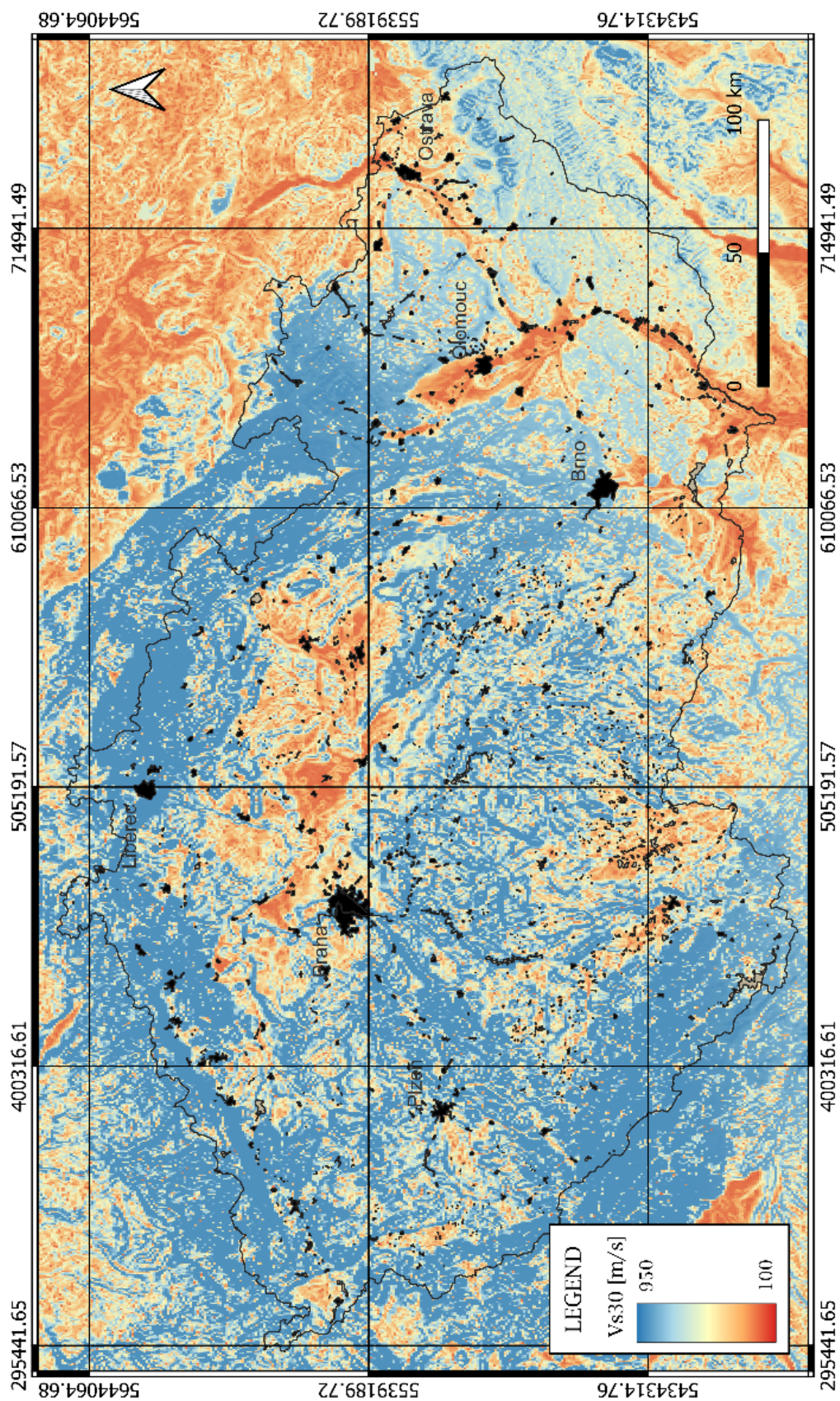


Figure 78: Map of estimated Vs30 values across the Czech Republic, derived using the topography-based method of Wald and Allen (2007) calibrated with locally measured Vs30 data.

The integration of the calibrated Vs30 model ultimately resulted in the creation of a national Vs30 map (Figure 78), which serves as a foundational layer in the interactive seismic hazard application. This tool enables site-specific hazard assessment by combining regional seismicity with local amplification factors and supports the development of location-adapted earthquake-resistant design across Czechia.

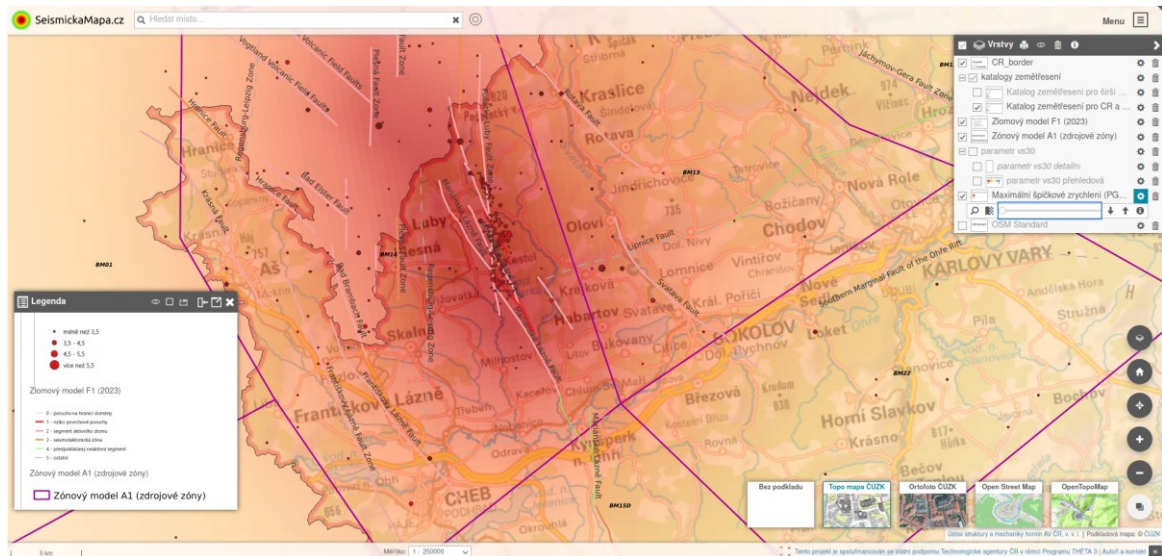


Figure 79: Screenshot of the interactive seismic hazard map interface. Users can adjust the visibility and transparency of individual layers (top right), select different topographic backgrounds (bottom right), and access a comprehensive legend for all layers (left panel). Adapted from Vackář et al., 2024).

4.3 RAPid Vs30 Estimation (RAVE) from surface-wave dispersion

This chapter is based on the study by Mazanec and Dal Moro (2025) submitted for publication and introduces an initial step toward simplified Vs30 estimation, offering a practical alternative to conventional inversion-based methods. Traditional Vs30 estimation methods rely on inverting surface-wave dispersion curves - a process that is computationally demanding, non-unique, and often ambiguous, as discussed in Chapter 3.2.3.3. Comina et al. (2022) highlighted that the primary source of uncertainty in deriving layered Vs models from surface wave testing stems from the inherently ill-posed nature of the surface wave inversion problem. However, RAPid Vs30 Estimation (RAVE), a novel approach that derives Vs30 directly from observed dispersion data without inversion. Two complementary methods are introduced: RAVE-A, which relates Vs30 to the area under the dispersion curve, and RAVE-SF, which uses the phase velocity at a single frequency for estimation.

4.3.1 The RAVE Concept

Two rapid methods for estimating the Vs30 parameter, collectively referred to as Rapid Vs30 Estimation (RAVE) are presented here:

- (1) RAVE-A, which uses the area under the observed dispersion curve,
- (2) RAVE-SF, which relies on the phase velocity at a single selected frequency.

The concepts of both approaches are outlined in Figure 80. When comparing two theoretical Vs models and their corresponding dispersion curves, it becomes evident that the area subtended by the dispersion curves is proportional to the shear-wave velocity model. Specifically, for a higher-velocity models, the dispersion curve exhibits a larger area compared to a lower-velocity models.

As illustrated in Figure 80, the RAVE-A approach enables a quick estimation of Vs30 by calculating the area under the dispersion curve within a defined frequency range (f_{min} to f_{max}). In contrast, the RAVE-SF approach focuses on identifying specific frequencies at which the correlation between observed phase velocity and Vs30 is most pronounced.

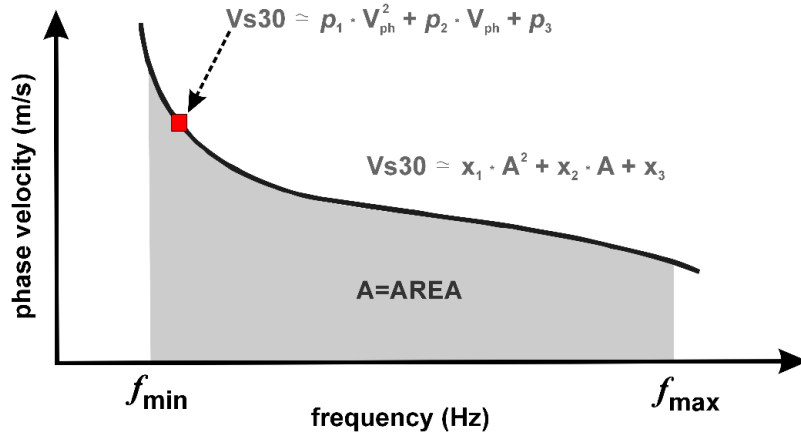


Figure 80: Concept of the RAVE approaches. In RAVE-A, Vs30 is estimated from the area under the dispersion curve between two fixed frequencies (grey shading)—larger areas correspond to higher Vs30 values. RAVE-SF, on the other hand, links Vs30 to the phase velocity at specific, optimally selected frequencies using second-order regression. Adapted from Mazanec and Dal Moro (2025).

4.3.2 Statistical Evaluation of RAVE approaches

To evaluate the relationship between Vs30 and the dispersion-curve characteristics used in the RAVE methods, 1600 randomly generated models, including those with low-velocity layers, were analyzed. For each model, fundamental-mode (FM) and effective dispersion curves (EC) were computed for both Rayleigh and Love waves. Vs30 values were then statistically correlated with the subtended area (RAVE-A, Figure 81 and Figure 82) and phase velocities at selected frequencies (RAVE-SF, Figure 83-Figure 86).

For both approaches, a quadratic regression model was used to describe the correlation, evaluated using the coefficient of determination (R^2) and the standard error (Δ), expressed as the 68% confidence interval. In the figures, model categories are color-coded: light blue for bedrock deeper than 30 m, dark blue for depths between 5 and 30 m, and red for shallow bedrock (< 5 m).

4.3.2.1 RAVE-A

For the statistical correlation between Vs30 and the area subtended by the dispersion curve (A) in the RAVE-A approach, the following quadratic model was applied:

$$Vs30 = x_1 \cdot A^2(\Delta f) + x_2 \cdot A(\Delta f) + x_3 \quad (4.1)$$

where $A(\Delta f)$ is the subtended area in the considered frequency range (Δf , see Figure 80) and x_1 and x_2 and x_3 are parameters obtained from the regression analysis.

Since the analyzed frequency range (Δf) in surface-wave analysis depends on various factors, including array length, receiver configuration, and the method applied, a wide range of frequency bounds was tested (see Mazanec and Dal Moro (2025) in Tables 1–2). Here, we present the relationships only for selected frequency intervals. Specifically, Figure 81 and

Figure 82 show the results of the statistical correlation for the RAVE-A approach within the 1–50 Hz frequency range, for Rayleigh and Love waves, respectively. The statistical analysis shows a distinct trend that, while best fit by a second-order polynomial, is nearly linear. When using the effective dispersion curve instead of the fundamental mode, a greater degree of scatter is observed, although the overall correlation remains relatively strong.

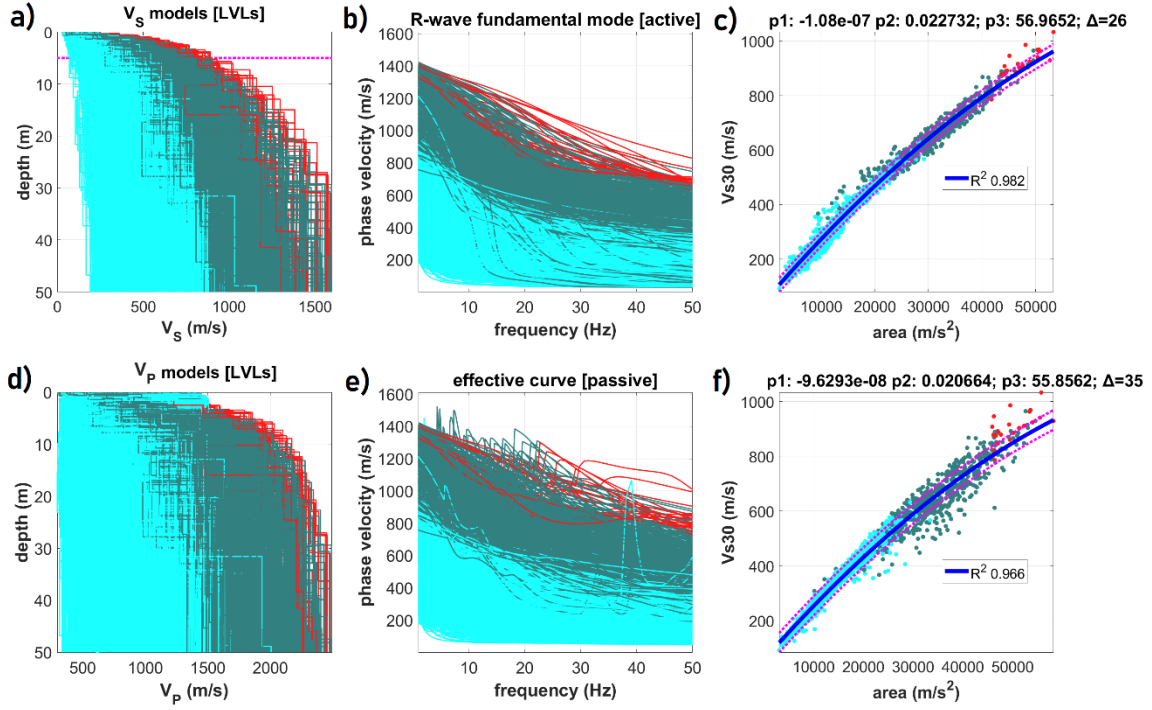


Figure 81: RAVE-A statistical analysis for Rayleigh waves (color-coded based on bedrock depth – see text): a) V_s models; b) FM dispersion curves; c) correlation between the V_{s30} parameter and the area subtended by the FM dispersion curve; d) V_p profiles; e) effective dispersion curves; f) correlation between the V_{s30} parameter and the area subtended by EC dispersion curve. In c) and f) the title provides the regression coefficients x_1 - x_3 (Equation 2) and the average standard error Δ (magenta dotted lines). Adapted from Mazanec and Dal Moro (2025).

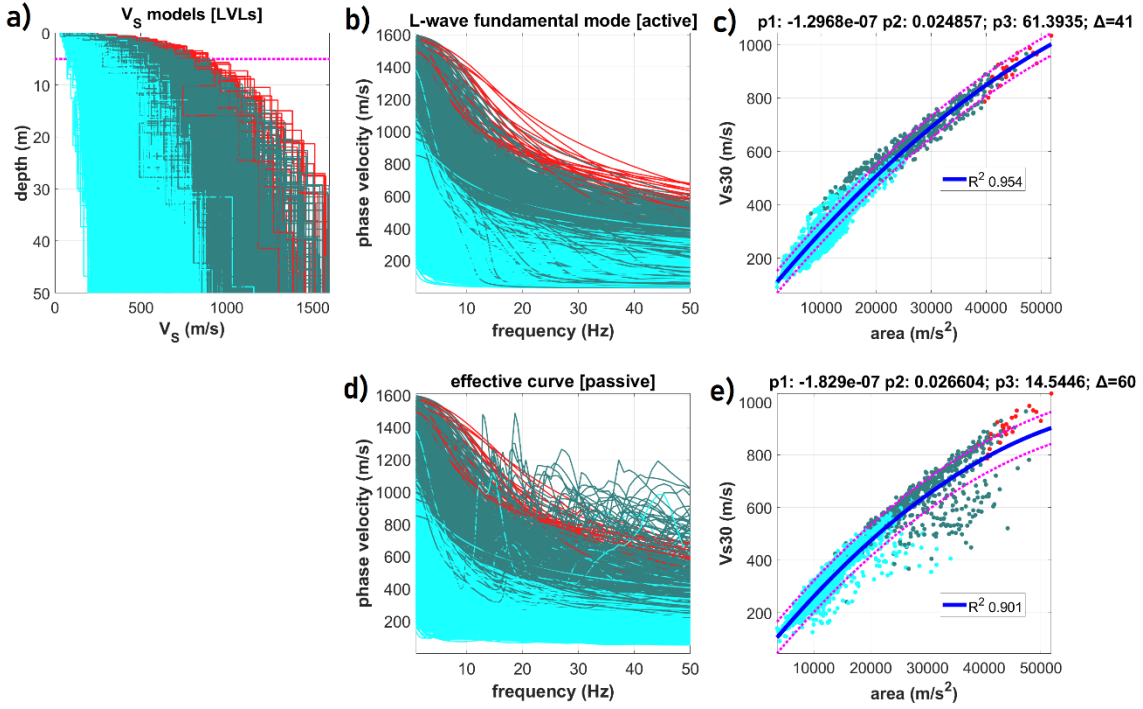


Figure 82: RAVE-A statistical analysis for Love waves: a) V_s models; b) FM dispersion curves; c) correlation between the V_{s30} and the area subtended by the FM dispersion curve; d) effective dispersion curves; e) correlation between the V_{s30} and the area subtended by effective dispersion curve. In c) and e) the title provides the regression coefficients x_1 - x_3 (Equation 2) and the average standard error Δ (magenta dotted lines). Adapted from Mazanec and Dal Moro (2025).

4.3.2.2 RAVE-SF

For the statistical correlation between V_{s30} and the phase velocity value of the dispersion curve at specific frequency values in the RAVE-SF approach, the following quadratic model was applied:

$$Vs30 = p_1 \cdot V_{ph}^2(f) + p_2 \cdot V_{ph}(f) + p_3 \quad (4.2)$$

where V_{ph} represents the phase velocity at a specific frequency (see Figure 80) and p_i are the coefficients of the regression model. The analysis was conducted for both fundamental mode (active data) and effective curve (passive data) for Love and Rayleigh waves within the frequency range of 1–50 Hz. Figure 83-Figure 86 show the four best-fitting models for each wave and source type. Similar to RAVE-A, the RAVE-SF regressions show generally strong correlations between V_{s30} and phase velocity. However, they exhibit more scatter and greater non-linearity, making the application of a quadratic regression model (Eq. 4.2) necessary.

The V_{s30} value obtained through the RAVE-SF approach represents the median of values computed across multiple frequencies from the observed dispersion curve, rather than being based on a single frequency. A set of frequencies closely corresponding to those used in the simulations is selected from the dispersion curve, and the relevant RAVE-SF relationships are applied.

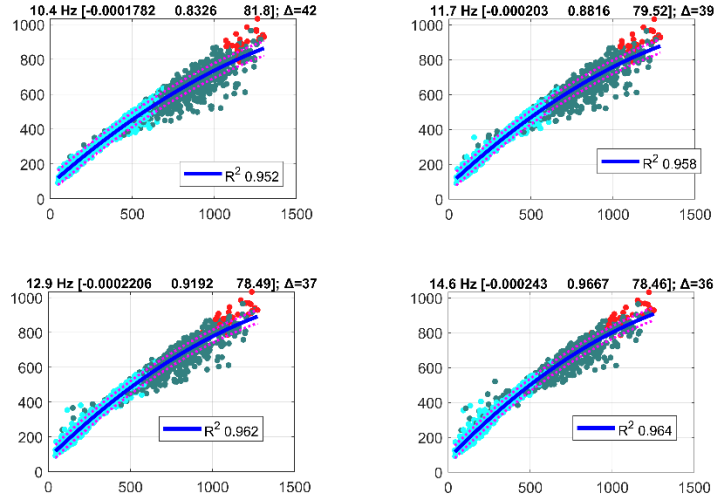


Figure 83: RAVE-SF analysis for Rayleigh-wave fundamental mode (Z component) showing the correlation between Vs30 and phase velocity (VR) at selected frequencies. Data points are color-coded by bedrock depth. Graph titles indicate the frequency, regression coefficients (p1–p3; Eq. 4.2), and average standard error Δ (magenta dashed lines). Modified from Mazanec and Dal Moro (2025).

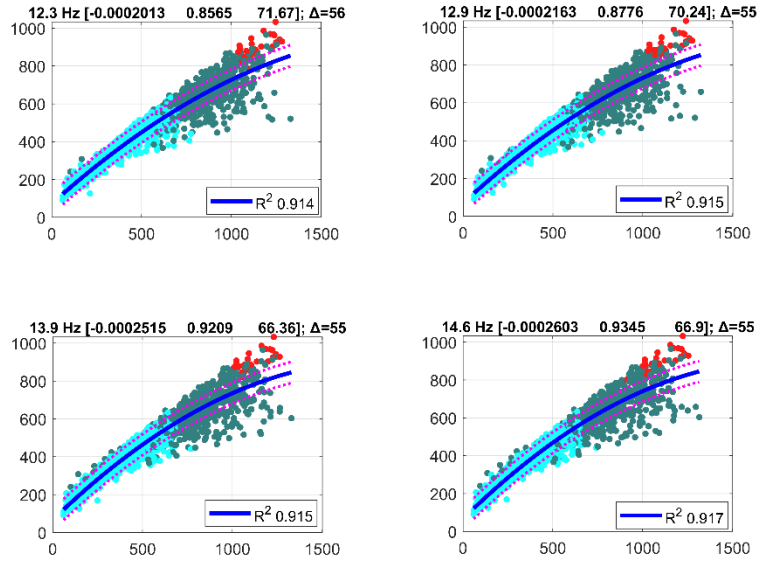


Figure 84: RAVE-SF analysis for Rayleigh-wave effective curve (Z component) showing the same correlation between Vs30 and phase velocity (VR) at selected frequencies as Figure 83. Modified from Mazanec and Dal Moro (2025).

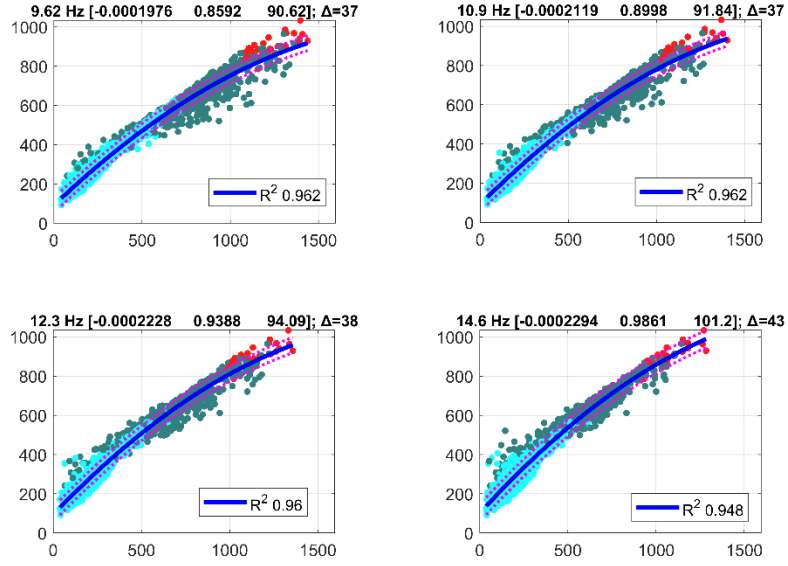


Figure 85: RAVE-SF for Love-wave fundamental mode showing the same correlation between V_{s30} and phase velocity (V_R) at selected frequencies as Figure 83. Modified from Mazanec and Dal Moro (2025).

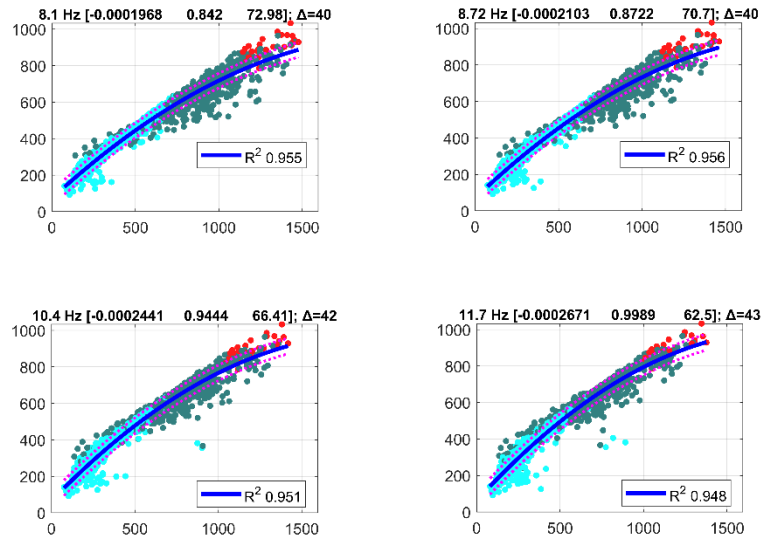


Figure 86: RAVE-SF for Love-wave effective curve showing the same correlation between V_{s30} and phase velocity (V_R) at selected frequencies as Figure 83. Modified from Mazanec and Dal Moro (2025).

4.3.3 Empirical Validation of RAVE on Surface-Wave Field Data

To evaluate the practical applicability of the RAVE approaches a real surface-wave data is presented. The case study site - Purgessimo site, situated approximately 4 kilometres east of Cividale del Friuli in NE Italy and is described in detail in Dal Moro (2014), Appendix, case study #2. The location is a reclaimed basin, where soft sediments overlies bedrock at a depth of roughly 30 meters, with the water table located around 1.5 meters below the surface.

In this case study, the subsurface V_s profile was derived from the joint analysis of active data using transversal and radial components (Love wave and Rayleigh wave radial component). Active seismic data were acquired for all three components (Z, R, and T) using a 23-trace linear array with 3-meter geophone spacing (see Dal Moro, 2014 for details). In this study, joint analysis was carried out based on the phase-velocity spectra of Love waves and the radial component of Rayleigh waves (because it better highlights higher modes presence than vertical component, which allow to further improve the dispersion analysis). These obtained phase-velocity spectra were inverted according to the Full Velocity Spectrum (FVS) technique (Dal Moro, 2019a). The results of the joint analysis are shown in Figure 87, where are also summarized the RAVE results.

The Purgessimo case study showed a strong consistency between RAVE-A and RAVE-SF estimates, with discrepancies remaining within 10%. Comparable agreement was also observed when comparing the average V_{s30} values derived from RAVE methods to those obtained through conventional inversion of the dispersion data.

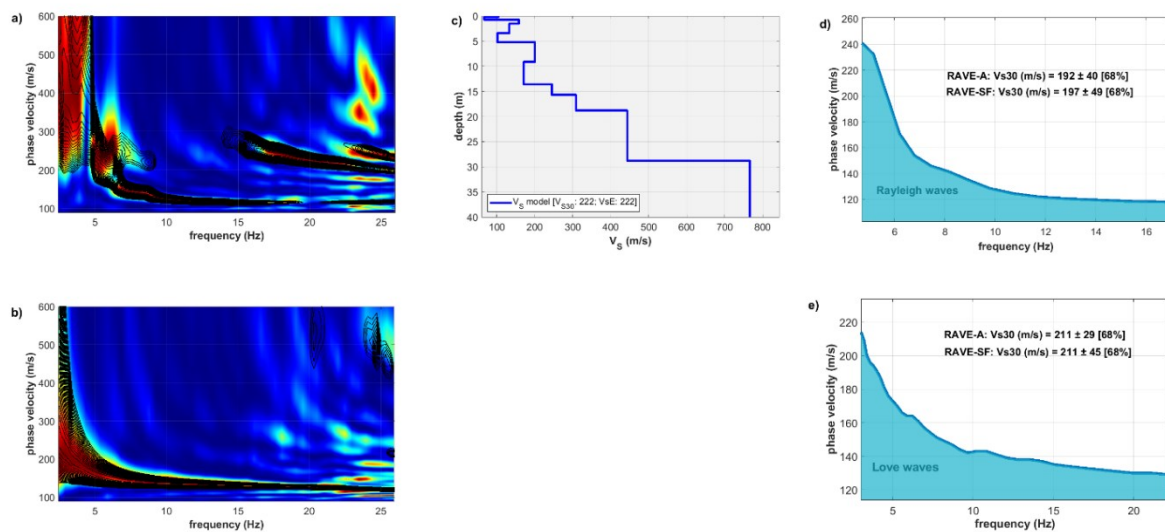


Figure 87: Case Purgessimo. Joint inversion of Love-wave and Rayleigh-wave (radial component) dispersion data. a) Rayleigh-wave radial component phase-velocity spectrum (black contours represent synthetic spectra from the FVS-based model); b) Love-wave phase-velocity spectrum; c) inverted shear-wave velocity (V_s) model; d) RAVE estimation based on Rayleigh fundamental mode; e) RAVE estimation based on Love fundamental mode. Modified from Mazanec and Dal Moro (2025).

4.3.4 Discussion and conclusion

The two introduced RAVE methods both demonstrate strong potential for rapid Vs30 estimation. The relationship between Vs30 and the RAVE-A area or RAVE-SF phase velocity follows a near-linear trend, well described by the quadratic regression models (Eqs. 4.1 and 4.2), with coefficients of determination often exceeding 0.9. The best fit was achieved for the RAVE-A approach using active data (fundamental mode) in the 1–50 Hz range, with an R^2 of 0.98 (Figure 81). Most of the observed scatter and nonlinearity occurs at higher Vs30 values (EC8 class A), which are less critical in seismic hazard assessments.

These findings indicate that combining both RAVE approaches may help detect inconsistencies caused by complex dispersion behavior. A practical solution could be to average the RAVE-A and RAVE-SF estimates; however, when discrepancies occur, RAVE-A is generally more reliable due to its integration over a frequency range rather than relying on individual frequencies. Its performance improves further when applied to both Rayleigh and Love waves, which likely explains the higher deviation in the second case study that used only Rayleigh waves. While RAVE provides a fast estimate of Vs30 based on regression models, it is important to note that even standard inversion methods face similar challenges related to uncertainty and non-uniqueness (Luke et al., 2003; O'Neill et al., 2004; Safani et al., 2005; Dal Moro, 2020a, 2023).

Results indicate that the proposed RAPid Vs30 Estimation (RAVE) method enables efficient large-scale, unsupervised Vs30-based soil classification and serves as a practical tool for standard Vs30 mapping, helping to mitigate several challenges typically associated with the inversion of surface-wave data. The very first tested real world dataset - Purgessimo showed reasonable consistency between RAVE-A and RAVE-SF and conventional Vs30 values. The current implementation of RAVE represents a preliminary framework intended for further refinement and expanded validation through additional case studies and methodological developments.

5 SUMMARY AND CONCLUSIONS

This dissertation presents a comprehensive investigation into the characterization of local site conditions with a particular emphasis on surface-wave methods, which are analyzed both conceptually and practically. The work combines two major thematic areas: the quantification of site effects and the application of surface-wave analysis for determining near-surface shear-wave velocity structures. These themes intersect throughout the thesis, culminating in an applied synthesis that demonstrates how surface-wave methods can directly support and enhance the estimation of seismic site effects.

The thesis opens with a theoretical and conceptual framework that defines the nature of site effects, emphasizing their complexity and dependence on local geological and geotechnical conditions. A thorough review of the mechanisms influencing site amplification is presented, including the role of seismic impedance contrasts, resonance, and wavefield trapping. It is established that shear-wave velocity (V_s) is one of the most influential parameters in controlling seismic response. The early chapters also review traditional and initial methods for estimating site effects, including empirical, geotechnical, geophysical, and analytical approaches. From these methods, a set of site proxies emerges, of which particularly the average velocity in the uppermost 30 meters (V_{s30}) is most commonly adopted in engineering and seismic hazard practice.

A major contribution of this dissertation lies in the critical evaluation of V_{s30} , both as a standalone metric and in combination with other parameters. Chapter 2 includes a comprehensive discussion of the history, utility, and limitations of V_{s30} , contextualized within international guidelines such as Eurocode 8. The thesis also contributes to the ongoing international debate about the validity of V_{s30} as a site-effect proxy by presenting new empirical results from a unique dataset: a homogenous seismic network in West Bohemia (WEBNET). Based on the study Mazanec et al. (2024), V_{s30} values measured through dispersion analysis were directly compared to recorded earthquake amplification, allowing for an independent evaluation of their relationship. Additionally, this study defines a new empirical relationship between V_{s30} and amplification, contributing to the ongoing international debate regarding the adequacy of V_{s30} as a standalone site-effect parameter.

The results demonstrate a clear trend between V_{s30} and relative amplification, though with notable scatter, particularly within EC8 class B sites. This variability was attributed primarily to the presence of shallow low-velocity layers that are not adequately captured by V_{s30} alone. The analysis also confirmed the strong correlation between V_{s30} and the H800 parameter. Analysis resulted in the definition of a new empirical relationship between V_{s30} and amplification, further advancing our understanding of the factors controlling site response.

Building on the need for robust and accessible velocity profiling, the thesis proceeds to extensively review and illustrate surface-wave methods for estimating near-surface shear-wave velocities. The physical basis of Rayleigh and Love wave propagation in vertically heterogeneous media is thoroughly examined, including their dispersive behavior. The methodological section details the complete workflow of dispersion analysis, from acquisition and preprocessing to the extraction of dispersion curves and inversion into Vs models. Emphasis is placed on practical aspects, such as equipment requirements, transformation techniques, and the mitigation of uncertainty and non-uniqueness in inversion.

To ground the theoretical discussion in practical examples, Chapter 3 presents a series of case studies that illustrate various aspects of surface-wave applications. The first study based on Mazanec and Valenta, (2023) demonstrates that even with standard refraction seismic equipment, meaningful surface-wave analysis is possible, allowing for the extraction of Vs30 from conventional P-wave data. This highlights the low barrier to entry for employing surface-wave methods in routine fieldwork. The second case study (Mazanec and Valenta, 2023) introduces a basic joint analysis combining active MASW, passive SPAC, and travel-time tomography, all performed with a single linear geophone array. This approach illustrates how minimal additional effort can lead to a substantial reduction in ambiguity and improved model robustness. The first two case studies are based solely on the analysis of the fundamental mode of the vertical component of Rayleigh waves. Although this approach was discussed in the thesis and its potential risks were highlighted, the analysis was conducted at a time when other types of geophones were not available. To reduce ambiguity, dispersion curves from multiple source offsets were averaged, and the results were compared with P-wave tomography to assess the lateral variability of the site. The ideal case, incorporating more advanced acquisition, is demonstrated in Case Study 3. The third case study (Dal Moro and Mazanec, 2024) represents a state-of-the-art application of joint inversion techniques in a challenging industrial setting, comparing three different approaches to integrate multicomponent dispersion data and HVSR observations. The results exemplify how modern surface-wave analysis can adapt to complex environments and deliver high-resolution subsurface models.

The final chapter brings together the two central themes of the dissertation, site effects and surface-wave analysis, demonstrating their integration in practical seismic hazard applications. Three applied examples are presented. In the first chapter based on Mazanec et al., (2024), surface-wave characterization was conducted at 17 stations of the WEBNET network to derive Vs30 and H800 values, which were subsequently used in the amplification analysis of Chapter 2.5. The results not only supported the correlation between Vs30 and amplification but also contributed to a detailed seismological characterization of this high-quality monitoring network.

The second chapter presents a partial contribution to the development of a seismic hazard map of the Czech Republic (Vackář et al., 2024), with Vs30 estimated from topographic slope as an initial proxy. This estimate was then corrected using over 100 in-situ Vs30 measurements obtained from surface-wave dispersion analysis. The correction significantly improved the reliability of the Vs30 map (as shown in Figure 78), and the final values were used in a PSHA framework to calculate ground motion parameters. The resulting data are accessible via an interactive online platform, which also provides Vs30-based site-specific hazard reports.

The third and final contribution is the introduction of a new methodological framework called RAVE – RAPid Vs30 Estimation (Mazanec and Dal Moro, 2025), which addresses the issue of inversion ambiguity by providing a fast and practical alternative to conventional approaches. RAVE includes two implementations (RAVE-A and RAVE-SF) that relate Vs30 directly to dispersion curve properties and demonstrate high correlation ($R^2 > 0.9$) with standard Vs30 estimates in early case studies. This novel framework has strong potential for large-scale mapping projects and regions where full inversion is impractical or ambiguous.

In summary, this thesis makes several original contributions to the field of site response characterization:

- It critically evaluates the applicability and limitations of Vs30 and related proxies (e.g., H800), using unique seismological datasets and in situ estimated Vs30 values.
- It proposes and demonstrates the use of surface-wave analysis as an efficient and robust tool for deriving site parameters.
- It delivers new data and methodological advancements (e.g., RAVE), which can improve large-scale seismic hazard modeling and site classification practices.

By bridging detailed theoretical exposition with practical field demonstrations, the dissertation illustrates the strong potential of surface-wave analysis in modern seismic hazard workflows, both in research and engineering practice. The integration of multiple datasets and analytical strategies points toward future opportunities in combining ambient noise seismology, dispersion analysis, and site response modeling for high-resolution, cost-effective ground characterization.

6 PERSPECTIVES FOR FURTHER RESEARCH

Several promising directions for future research emerge from the results and observations presented in this thesis. First, while the analysis confirmed that V_{s30} is a valuable proxy for site amplification, the dataset used for correlation-derived from earthquake recordings at WEBNET stations was limited to high V_{s30} values due to the location of stations in competent bedrock. This limitation restricts the generalizability of the proposed V_{s30} amplification relationship, particularly for sites in soft sediments. With the increasing availability of autonomous 3C seismic nodes, future work should focus on expanding this approach to a broader range of site conditions. A strategically designed deployment of such stations in areas with low V_{s30} and soft soils will enable a more comprehensive analysis of amplification behavior, particularly for low-frequency ground motions and in 2D/3D geological settings.

Another critical avenue is the continued development and validation of the RAVE (Rapid V_{s30} Estimation) framework. While initial tests have demonstrated its reliability and simplicity, the method remains in its early stages and requires further validation across diverse geological environments. Future research should focus on systematically analyse RAVE against full inversions, refining its parameterizations, and testing its performance under varying acquisition conditions. In particular, applying RAVE to regional or continental-scale seismic hazard mapping projects would serve as a critical test of its scalability and robustness.

The analysis of surface waves represents a central methodological focus of this dissertation. Most of the presented case studies were constrained by basic single-component seismic equipment, which limited the range of wave types and polarization that could be investigated. With recent access to multicomponent (3C) equipment and more advanced field acquisition capabilities, the author intends to deepen expertise in the broad spectrum of dispersion analysis methods and their joint inversion strategies. This includes not only enhanced Rayleigh-wave modeling but also the systematic use of Love waves, higher modes, and azimuthal variability, which may help characterize velocity anisotropy or 2D site effects in urban environments.

A related research interest is the application of ambient vibration recordings for the dynamic characterization of critical infrastructure (e.g., buildings, bridges, embankments). This approach offers a non-invasive, cost-effective way to assess structural resonance and vulnerability using naturally occurring seismic noise. Combined with surface-wave dispersion and HVSr techniques, this opens the possibility of integrated ground–structure system assessment. In this context, the potential of Distributed Acoustic Sensing (DAS) should be

further explored, especially in borehole and near-surface installations, where its high spatial resolution could improve the imaging of shallow site features that affect amplification but are typically poorly constrained.

Finally, the interplay between surface-wave methods and site proxies provides rich ground for continued research into the multi-parameter characterization of site effects. This includes expanding the analysis to include parameters like f_0 , H800, sediment thickness, and impedance contrasts, potentially leading to multi-variate classification schemes that better reflect local amplification behavior than any single proxy.

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LIST OF PUBLICATIONS

- I. **Surface waves as a cost-effective tool for enhancing the interpretation of shallow refraction seismic data**
Mazanec M., Valenta J.
Acta Geodynamica Et Geomaterialia
(contribution 60%)

- II. **Does VS30 reflect seismic amplification? Observations from the West Bohemia Seismic Network**
Mazanec, M., Valenta, J., Málek, J.
Natural Hazards
(contribution 60%)

- III. **Determination of the profile at a Vs noisy industrial site via active and passive data the critical role of Love waves and the opportunities of multi-component group velocity analysis**
Dal Moro, G., Mazanec, M.
Geophysics
(contribution 30%)

- IV. **Rapid Vs30 Estimation (Rave) from Surface-Wave Dispersion Without Inversion: Two Direct Approaches**
Mazanec, M., Dal Moro, G.
Submitted for publication
(contribution 40%)

- V. **Interactive Seismic Hazard Map of the Czech Republic**
Vackář Jiří, Valenta Jan, Mazanec Martin Málek, Jiří and Prachař, Ivan and Lukešová, Renata and Mrkvová, Barbora
Under review
(contribution 15%)

