Progetto S4

Caratterizzazione sismica dei siti utilizzando tecniche di array

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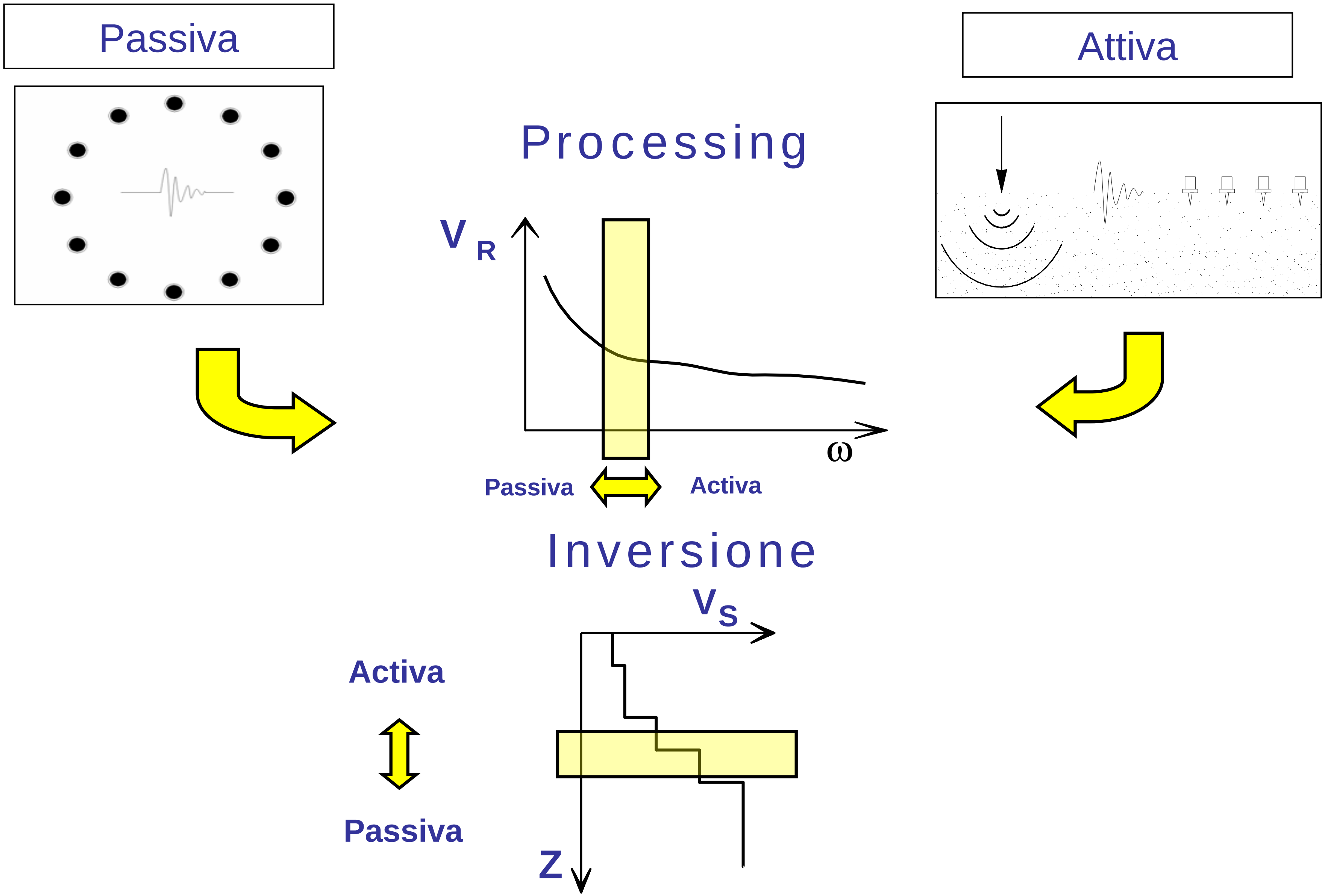
Analisi di onde di superficie

- Tecniche con sorgenti attive

-Tecniche con sorgenti passive

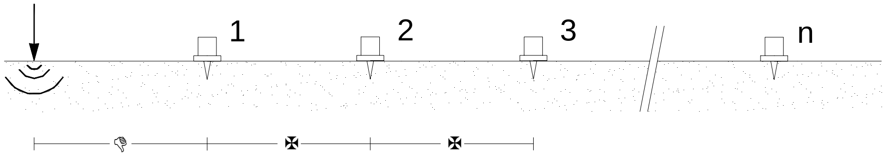
- Interpretazione combinata di tecniche attive e passive

Tecniche basata su sorgenti attive e passive

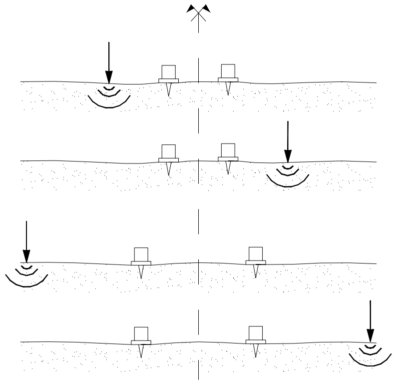


metodi attivi

Multistazione
(f-k, τ-p, ...)



2-stazioni (SASW)



Spectral Analysis of Surface Waves (SASW) requires that:

Two vertical receivers are placed on the ground at equal distance from a fixed centerline

Tests are performed progressively moving the receivers away from the fixed centerline

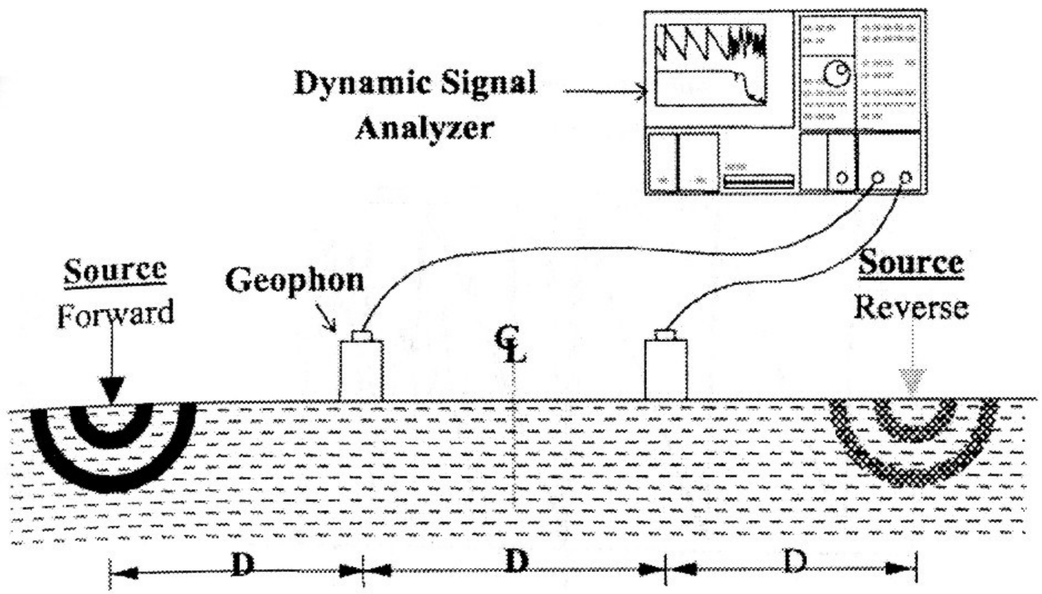


Fig. 1. General configuration of the SASW testing setup.

after Kim and Park (2002)

The phase of the cross-power spectrum, which represents the phase difference between the two receiver signals as a function of frequency, is calculated

The time delay *t*(*f*) between the receiver fro each frequency is calculated by:

$$t(f) = \frac{\theta_{xy}(f)}{2\pi f} \qquad (1)$$

θ_{xy} is the phase shift of the cross-power spectrum in radian

f is the frequency

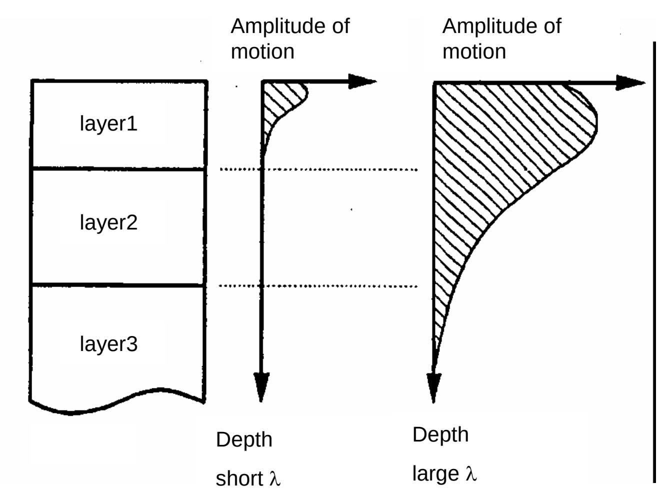
The phase velocity V_R is then calculated by:

$$V_R = D/t(f) \qquad (2)$$

D is the distance between the two receivers

The corresponding wavelength λ_R of the surface wave that determined the depth of investigation (of about $\lambda_R/3$, $\lambda_R/2$) is:

$$\lambda_R = V_R/f \qquad (3)$$



The procedure described before is based on Fourier Transform

At each frequency the phase difference between the two sinusoidal component with constant frequency and amplitude over the whole record length must be in the range +/- 180°

Therefore in the conventional SASW the unwrapping process of relative phase angle is required (adding or subtracting the correct number of 360° cicles) before evaluating the phase velocity

$$\theta_{unwrap}(f) = 2\pi n + \theta_{unwrap}(f) \qquad (4)$$

At each frequency the phase difference between the two sinusoidal component with constant frequency and amplitude over the whole record length must be in the range +/- 180°

In order to perform the reliable unwrapping process is necessary to obtain the correct phase information at frequencies lower than the frequencies of interest

That is the phase velocity at a certain frequency cannot be obtained independently from the phase information at frequencies lower than it.

1) the energy generated by an impact source is band limited, with low signal to noise ratio at very low and very high frequency

2) Geophones act as high-pass filters

Therefore the S/N ratio might be low at low frequencies

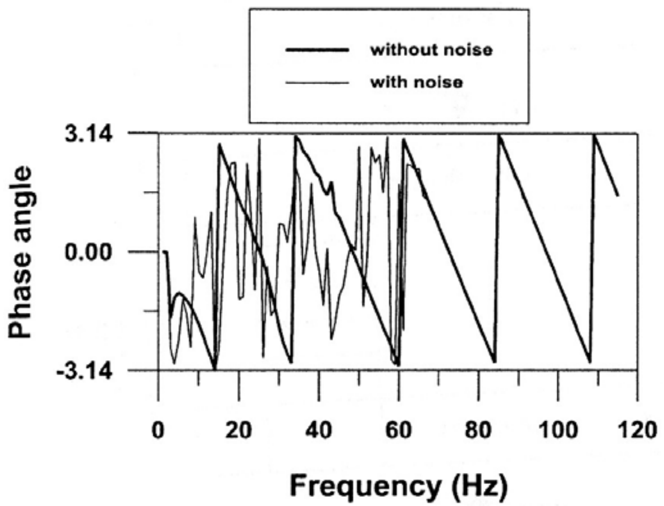
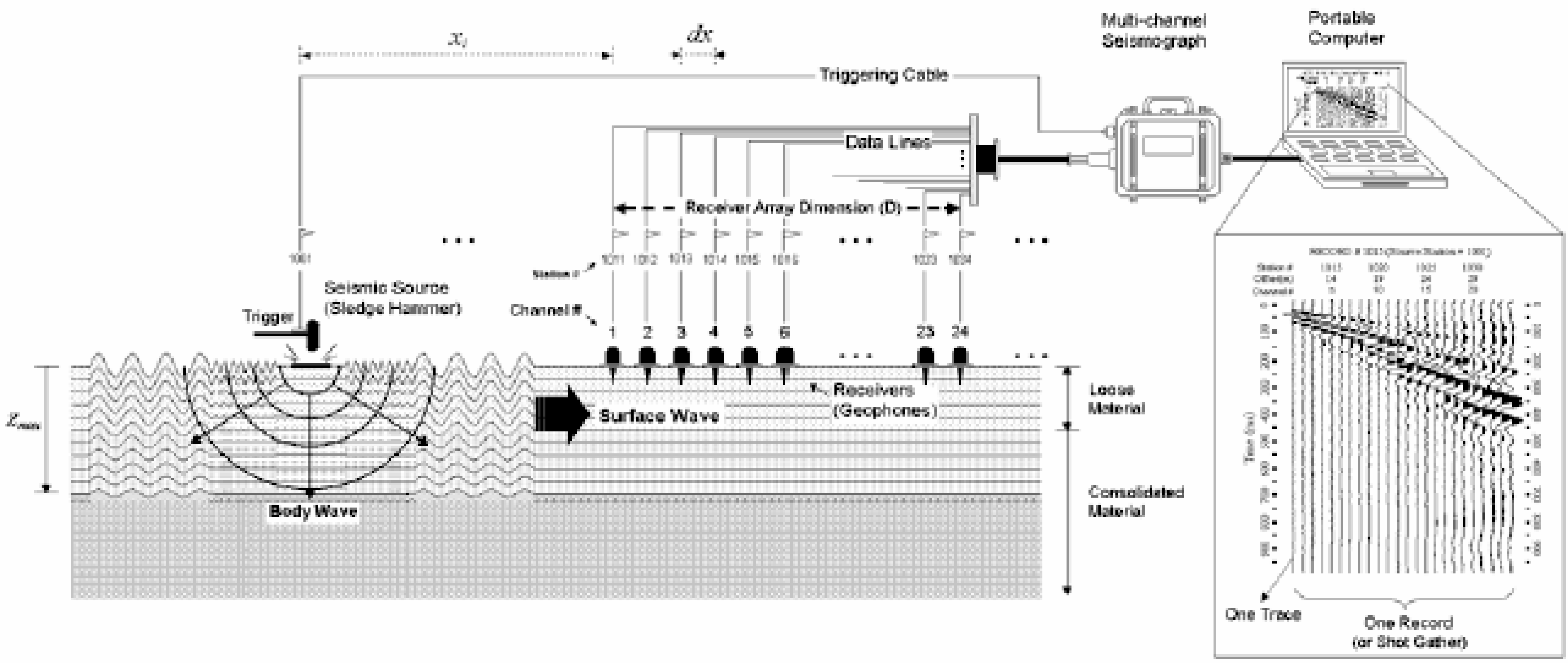


Fig. 2. Phase spectrum of signals with noise and without noise.

after Kim and Park (2002)

Multichannel Analysis of Surface Wave was found to be more efficient for unraveling the dispersive properties (Park et al., 1996).



The dispersion spectrum S(ω,*v*)from multichannel surface wave data is determined by the equation

$$S(\omega,v)=\int e^{-i(k-\omega/v)x}A(x,\omega)\,dx=\int e^{-i(\omega/V-\omega/v)x}A(x,\omega)\,dx \qquad (5)$$

A(ω) is the normalized energy spectrum fro each receiver, *k* the wavenumber, ω the circular frequency, *v* the assumed phase velocity and *V* the phase velocity for a given frequency

When *v* is equal to *V* , S(ω,*v*) is maximized

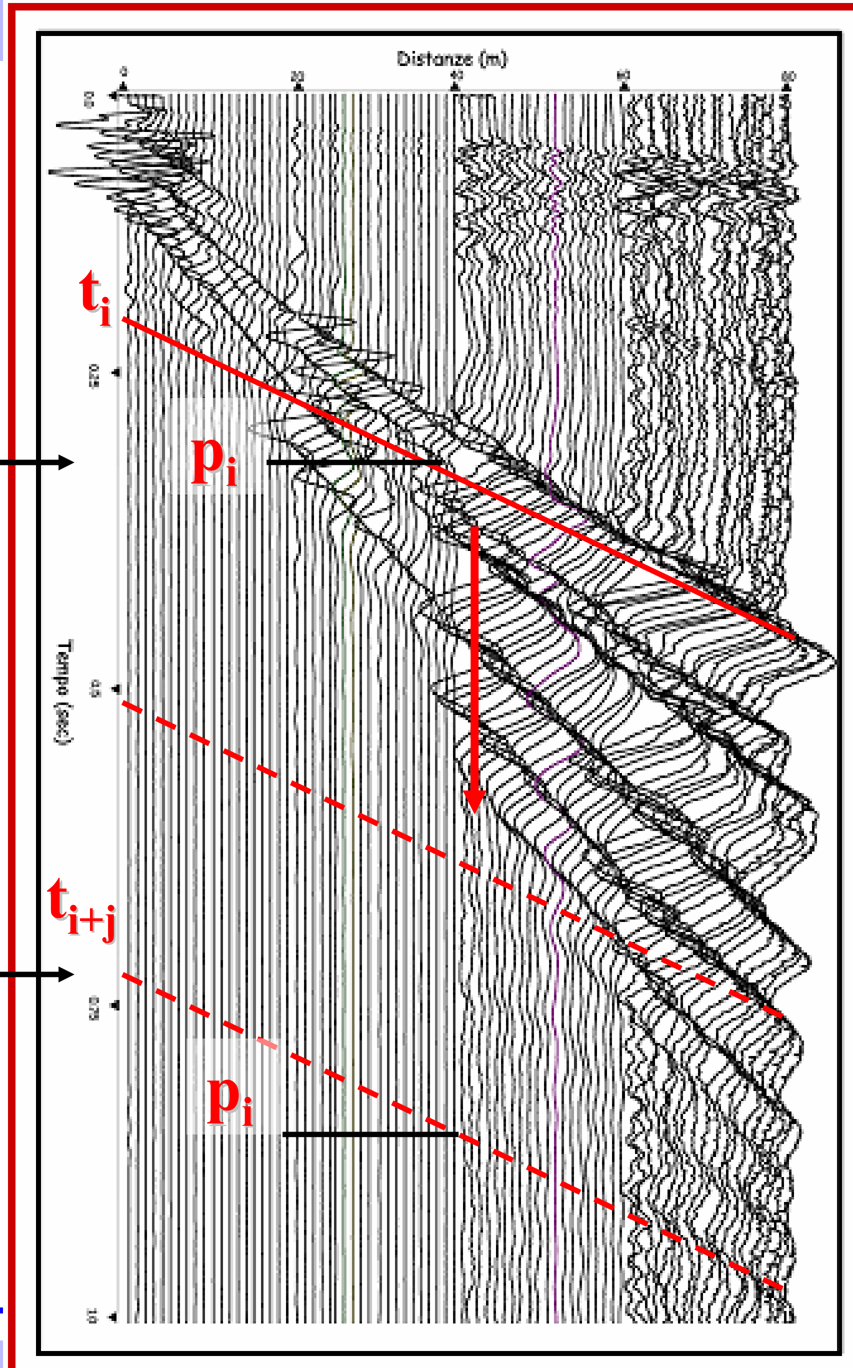
Slant-Stack (p,τ transformation)

$$A(p,\tau) = \sum_{j=1}^{nx} A(x = jdx, t = \tau + px)$$

$\Sigma A_i(p_i, t_i)$

For each fixed value of P, the operation is repeated for each t

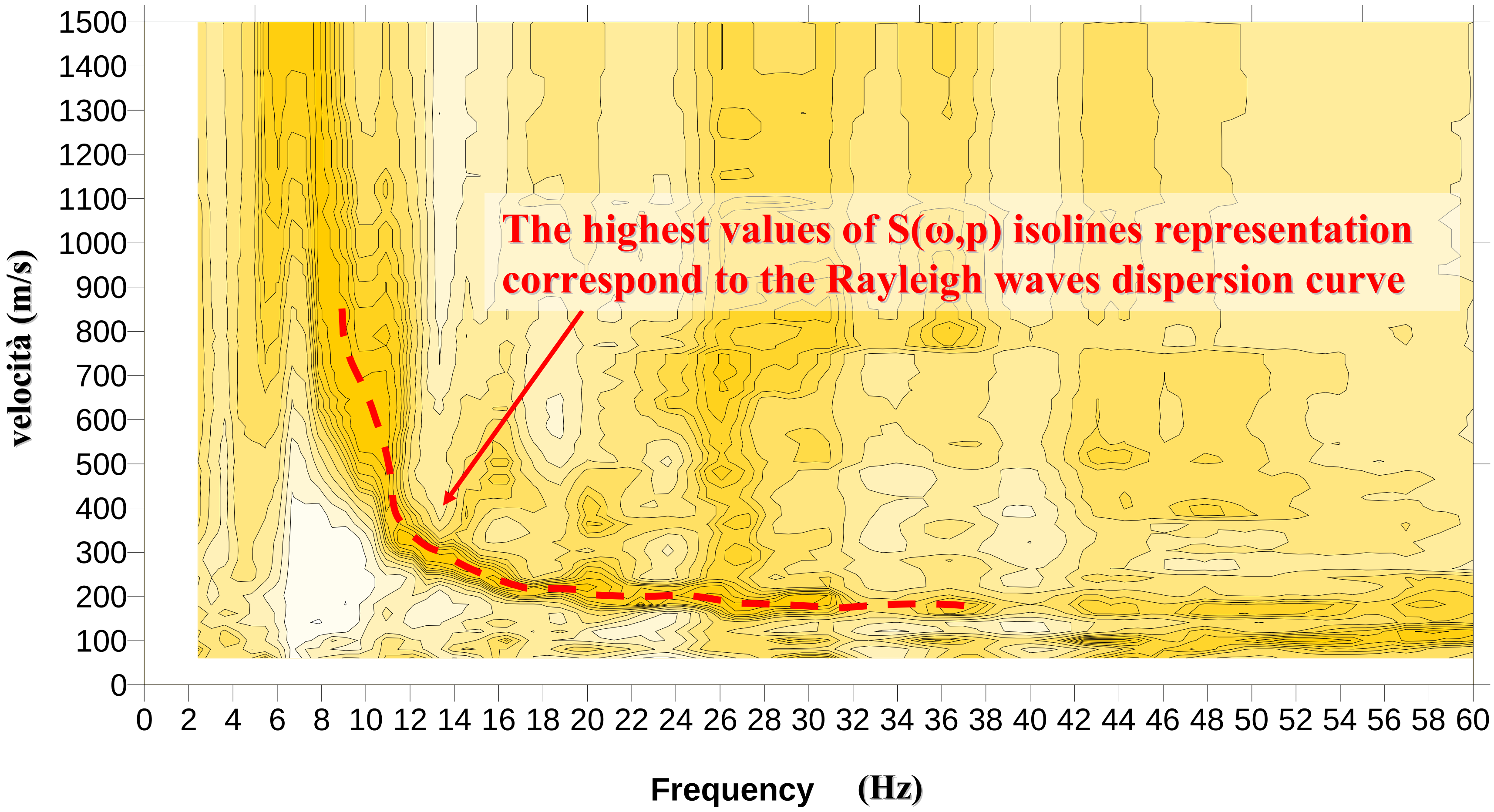
$\Sigma A_{i+j}(p_i, t_{i+j})$



Fourier transformation

$$F_A(p,f) = \int_\tau A(p,\tau) \cdot e^{-i2\pi f\tau} dt$$

after Picozzi (2005)



after Picozzi (2005)

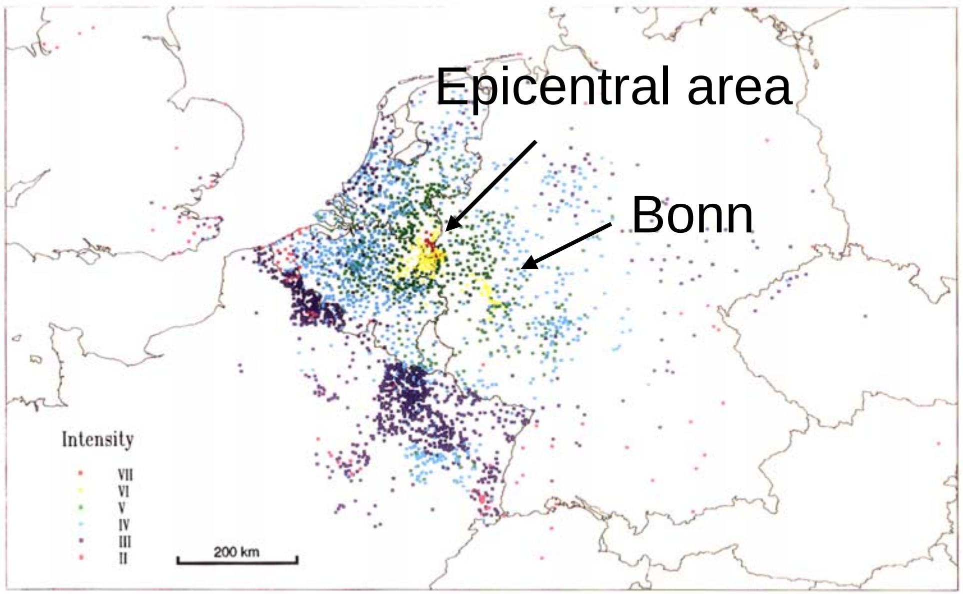
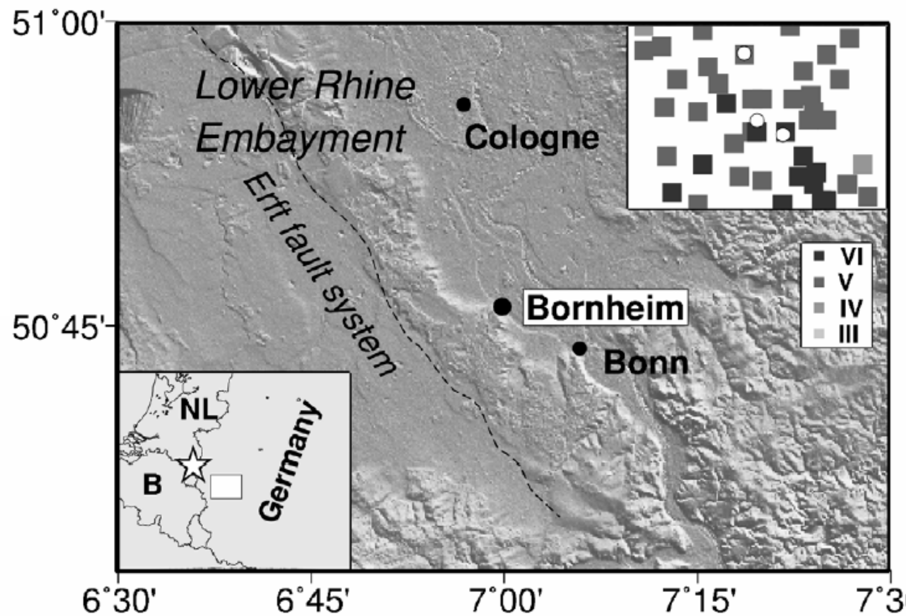


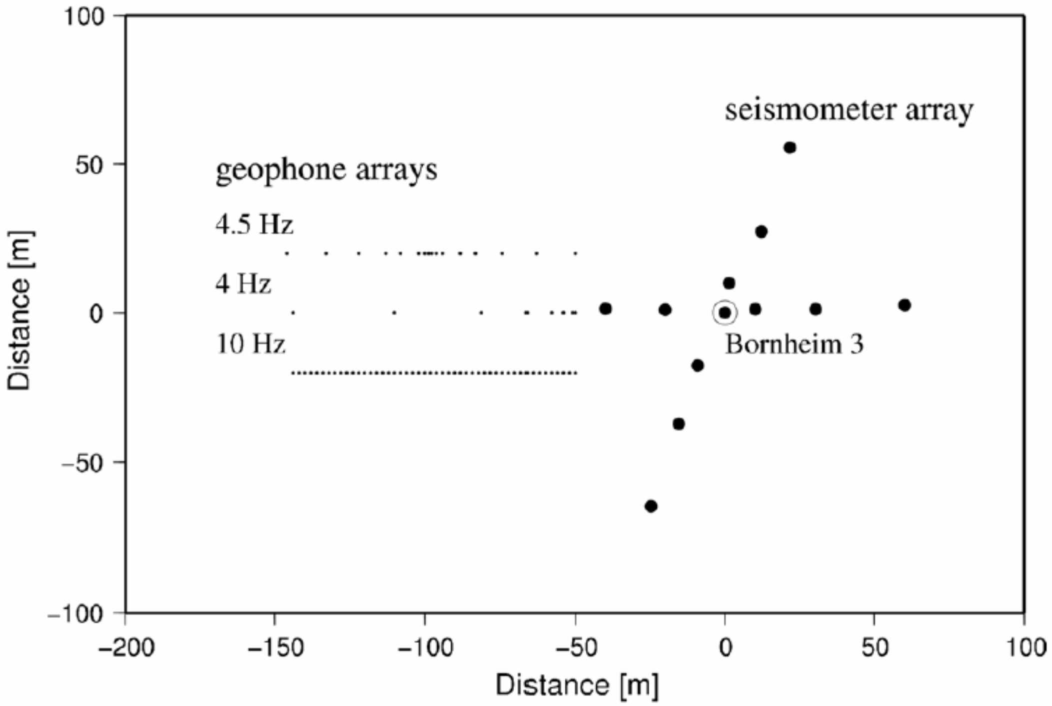
Fig. 1. Intensity map for the 1992 Roermond earthquake. Intensities are assigned according to the MSK-scale.

Courtesy of G. Grünthal

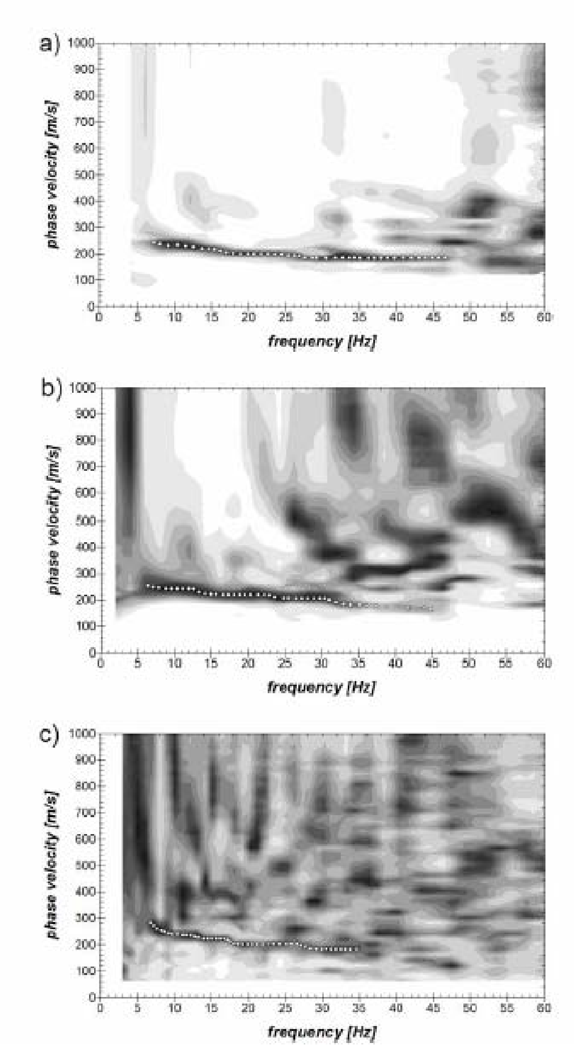
The intensities locally increased in the Bonn area



After Richwalski et al. (2007)



After Richwalski et al. (2007)



4 Hz geophones

4.5 Hz geophones

10 Hz geophones

Power spectrum functions for linear geophone arrays from MASW analysis (scale is arbitrary, white representing low values and black high values). The picked phase velocity values are shown as white dots. (a) 4 Hz, (b) 4.5 Hz, (c) 10 Hz

After Richwalski et al. (2007)

1) ESAC, F-K based methods

2) Refraction Microtremors (ReMi)

Aki assumed that noise represents the sum of waves propagating in a horizontal plane in different directions with different powers, but with the same phase velocity for a given frequency. He also assumed that waves with different propagation directions and different frequencies are statistically independent. A spatial correlation function can therefore be defined as:

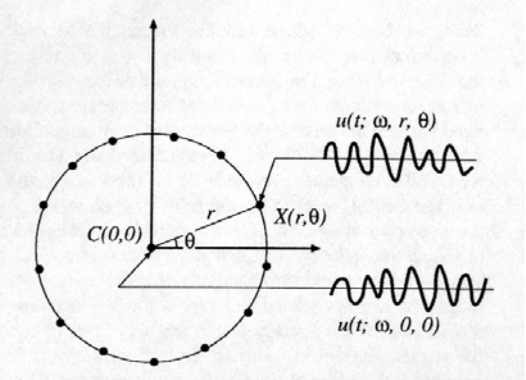


Figure 1. A schematic description of the symbols for the observation sites and the vertical component of the microseisms. $u(t;\ \omega,\ 0,\ 0)$ and $u(t;\ \omega,\ r,\ \theta)$ denote the harmonic waves at frequency ω, which are obtained at the center $C(0, 0)$ of the array and site $X(r, \theta)$ on the circle shown by "●".

after Morikawa et al. (2004)

$$\phi(r,\lambda) = < u(x,y,t)(x + r\cos(\lambda), y + r\sin(\lambda), t) > \qquad (1)$$

$u(x, y, t)$ is the velocity observed at point (x,y) at time t; r is the inter-station distance; λ *is* the azimuth and < > denotes the ensemble average

The space-correlation function for one angular frequency ω_0, normalized to the power spectrum, will be of the form

$$\phi(r,\omega_o) = J_0\left(\frac{\omega_0}{c(\omega_0)}r\right)$$

J_0 is the zero order Bessel function.
$c(\omega)$ is the frequency-dependent phase velocity

For every couple of stations (fixed the distance r) the function $\phi(\omega)$ can be calculated in the frequency domain by means of (Malagnini et al., 1993; Ohori et al., 2002; Okada, 2003):

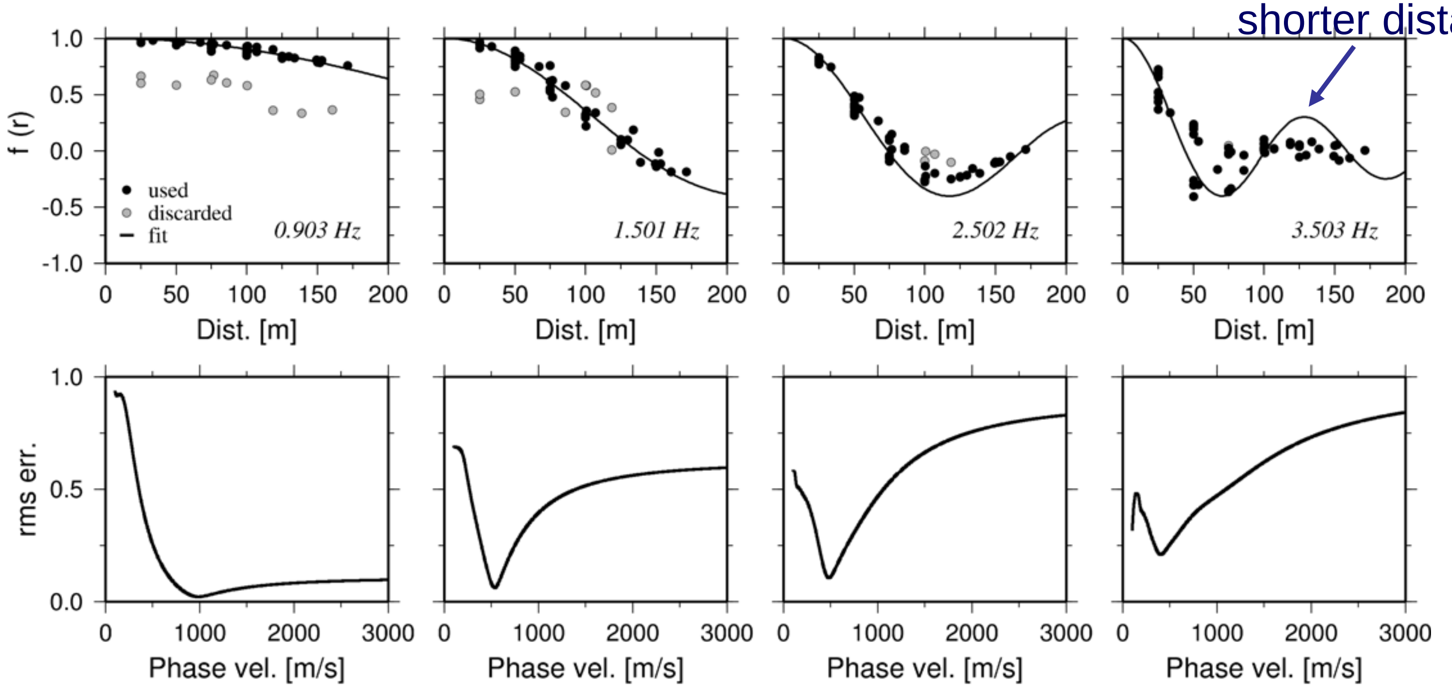
$$\phi(\omega) = \frac{\frac{1}{M}\sum_{m=1}^{M} \mathrm{Re}\left({}_mS_{jn}(\omega)\right)}{\sqrt{\frac{1}{M}\sum_{m=1}^{M} {}_mS_{jj}(\omega) \sum_{m=1}^{M} {}_mS_{nn}(\omega)}} \qquad (5)$$

where ${}_mS_{jn}$ is the cross-spectrum for the *m*th segment of data, between the *j*th and the *n*th station, and *M* is the total number of used segments. The power spectra of the *m*th segment at station *j* and station *n* are ${}_mS_{jj}$ and ${}_mS_{nn}$, respectively.

Spatial correlation values $\phi(\omega)$ are plotted as function of distance. A grid search procedure is applied to find the $c(\omega)$ that gives the best fit to the data

$$RMS = \sqrt{\frac{\sum_{n=1}^{N}\left[\bar{\rho}(r_n, f_o) - J_o\left(\frac{2\pi f_o r_n}{c(f_o)}\right)\right]^2}{N}}$$

High frequencies lose coherency at shorter distances



The estimate of the frequency-wavenumber (F-K) spectra *Pb(f,k)* by the **Beam Forming Method** is given by:

$$P_b(f,k) = \sum_{l,m=1}^{n} \phi_{lm} \exp\{ik(X_l - X_m\}$$

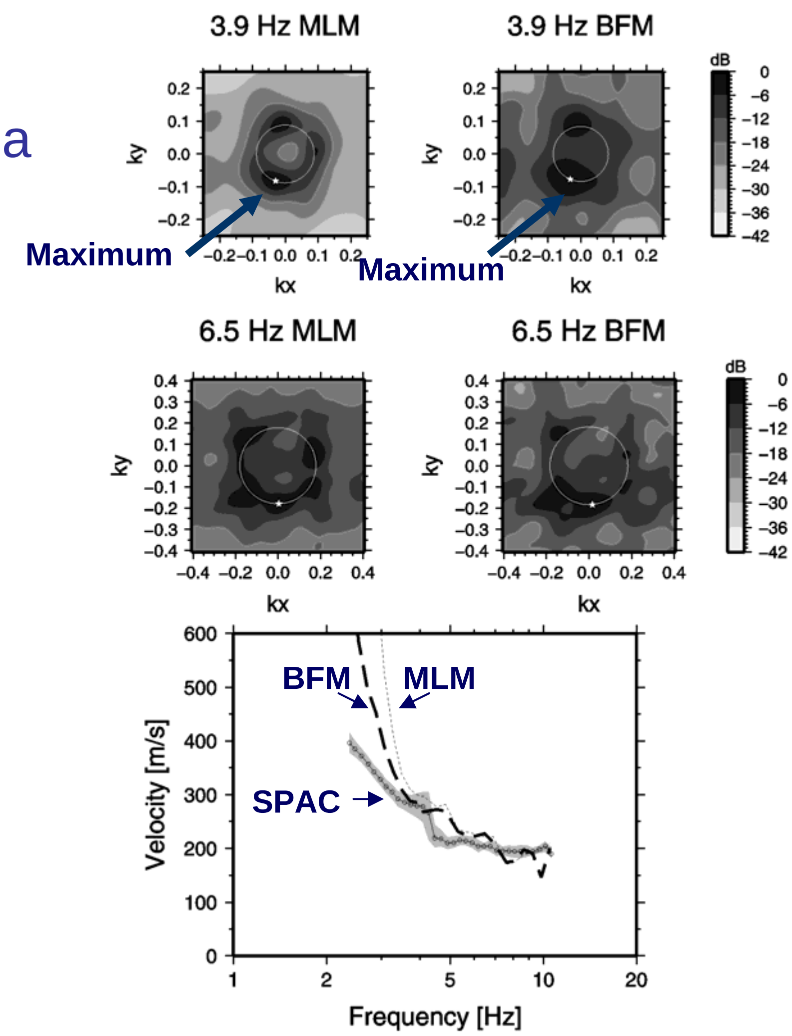
The **Maximum Likelihood Method** gives the estimate of the F-K spectra *Pm*(*f*,*k*) as:

$$P_m(f,k) = \left(\sum_{l,m=1}^{n} \phi_{lm}^{-1} \exp\{ik(X_l - X_m\} \right)^{-1}$$

Where *f* is the frequency, *k* the two-dimensional horizontal wavenumber vector, *n* the number of sensors. Φ_{lm}^{-1} is the element of the corresponding inverse of the matrix Φ_{lm} , X_i and X_m, are the coordinated of the *l*-th and the *m*-th sensors, respectively.

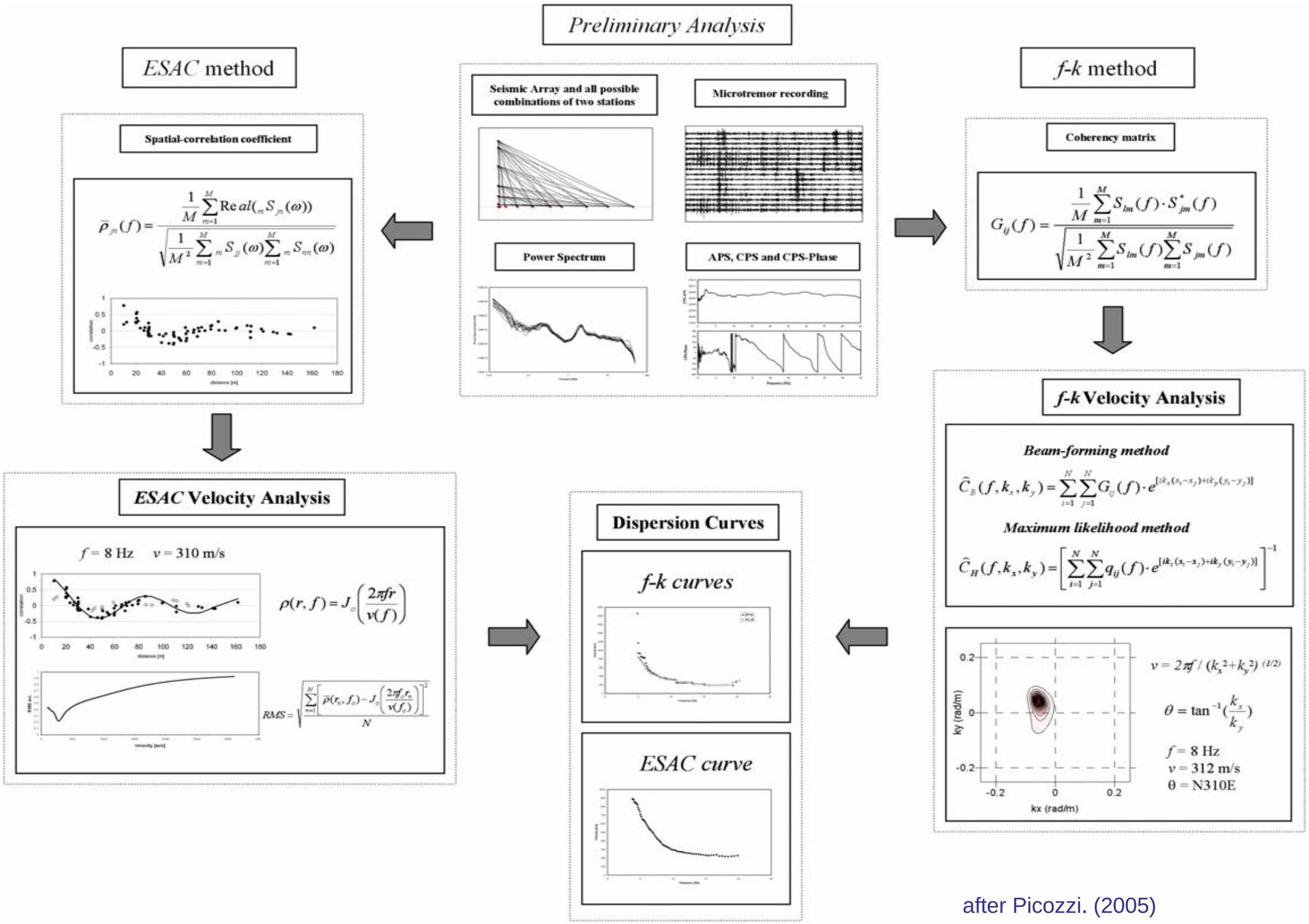
From the peak in the F-K spectrum occurring a coordinates k_{xo} and k_{yo} for a certain frequency f_0 the phase velocity c_0 can be calculated by:

$$c_0 = \frac{2\pi f_0}{\sqrt{k_{xo}^{\ 2} + k_{y0}^{\ 2}}} \qquad (9)$$



The F-K spectra *Pm*(*f*,*k*) methods allow to check the azimuthal distribution of the noise sources at each frequency.

The ESAC method (under the condition that noise is stationary in space and time) appears to be more suitable than F-K analysis for providing reliable dispersion curves over a larger –extended toward lower frequency- frequency range (i.e. larger depth of investigation up to more than 2 times the array dimension)



after Picozzi. (2005)

The inversion can be carry out linearizing the problem, that is calculating the Jacobian matrix that links the model parameters to the phase velocity.

$$C=J\Delta V$$

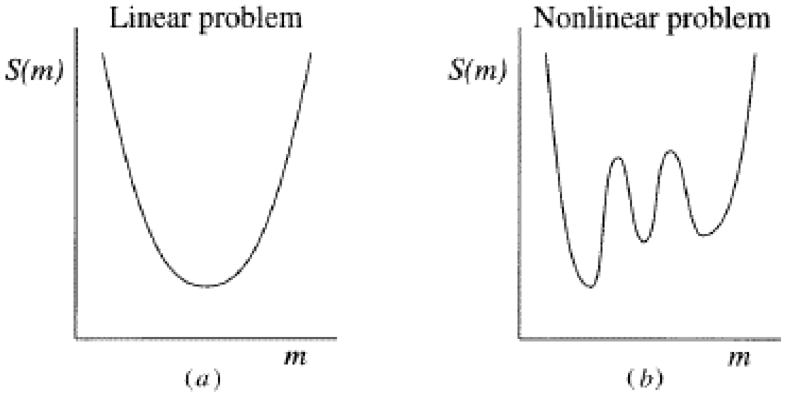
Where C= vector whose i elements are the $c_{obsi}(\omega)$-$c_{calci}(\omega)$

J=matrix with i rows and j (number of unknown, e.g. S-wave velocity in each layer) whose elements are $\delta\, c_{calci}(\omega)/\, \delta Vs_j$
ΔV is an array whose j elements are the correction values of the starting j layer velocities

The inverse problem can be solved using Singular Value Decomposition (SDV, Press et al., 1986) and the RMS of differences between observed and theoretical phase velocities is generally minimized.
However the final results strongly depends on the starting model!

Other methods can be used to solve the non-linear problem.

Parolai et al. (2005), following Yamanaka and Ishida (1996) adopted the non-linear optimization method that uses a genetic algorithm (e.g. Goldberg, 1989). In contrast to linearized inversion schemes, this method requires only an evaluation of the functions, not their derivatives.



Best model

average
misfit

Minimum
misfit

All tested
models

H/V spectral ratio

Observed data

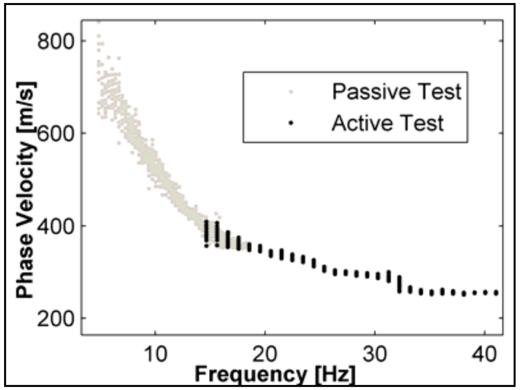
Calculated data

Global search inversion method

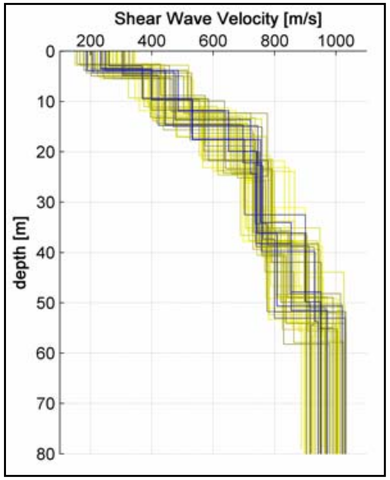
- Randomly generated profiles (V_S vs depth).
- Use of the scaling properties of Rayleigh wave dispersion to optimize the exploration of Model Parameter Space.
- A statistical test is used to select equivalent profiles accounting for data uncertainties.

By the use of MC-algorithm it is possible to study the uncertainty (equivalence) associated with SW-inversion

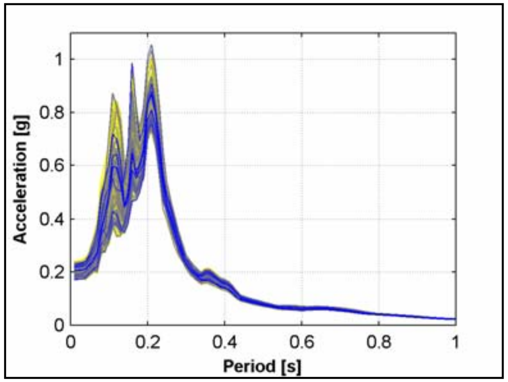
Data



Soil Profile



Local Site Response



(Foti et al., 2008)

Active + passive methods

Methods 1D:

For Vs profile of sediments and bedrock depth:

A B C D E - Active and Passive SW tests.

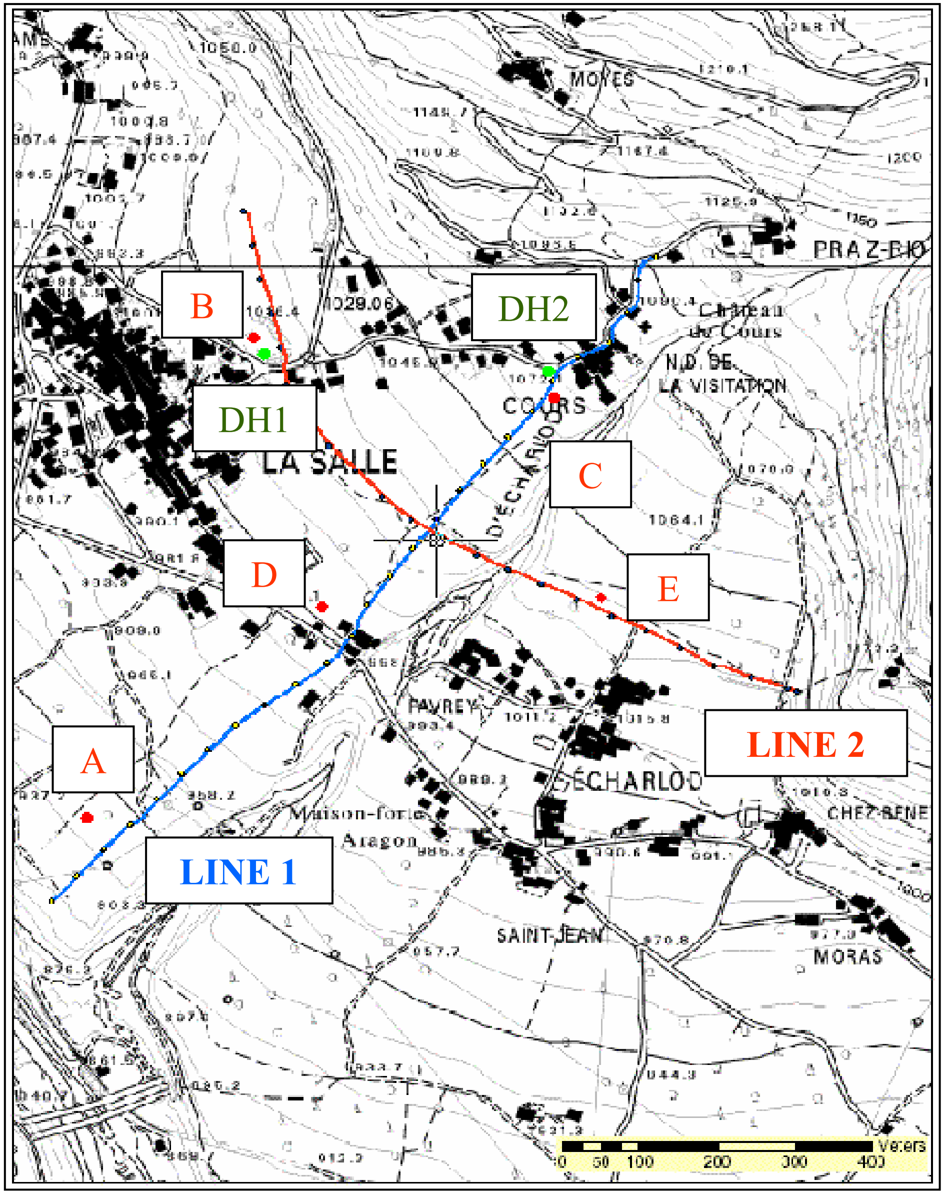
DH1 DH2 –Down Hole Tests (P and Sh waves).

Methods 2D:

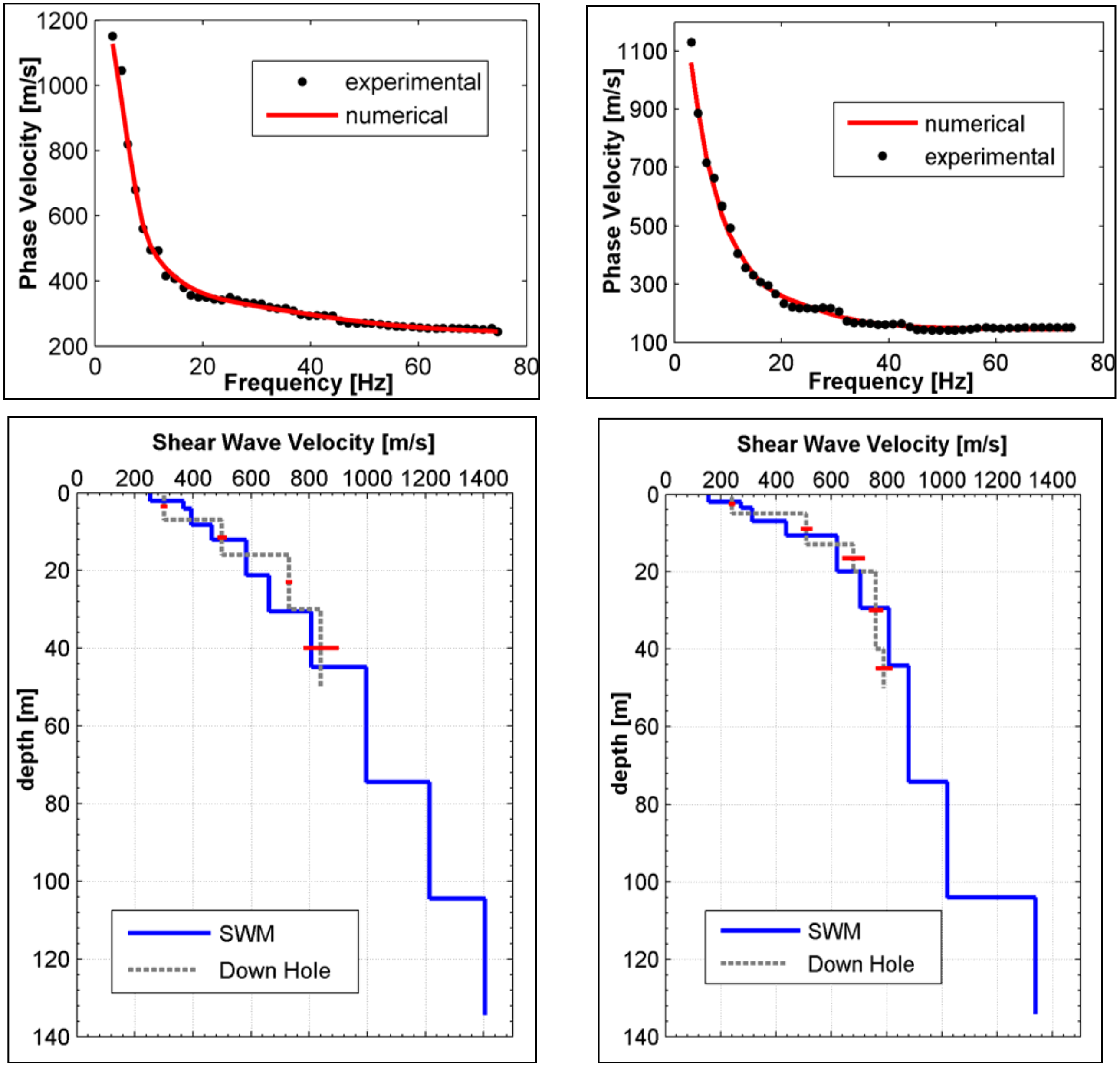
For bedrock depth and morphology:

LINE 1 e LINE 2 seismic reflection tests.

LCI on Ground-Roll



Site B



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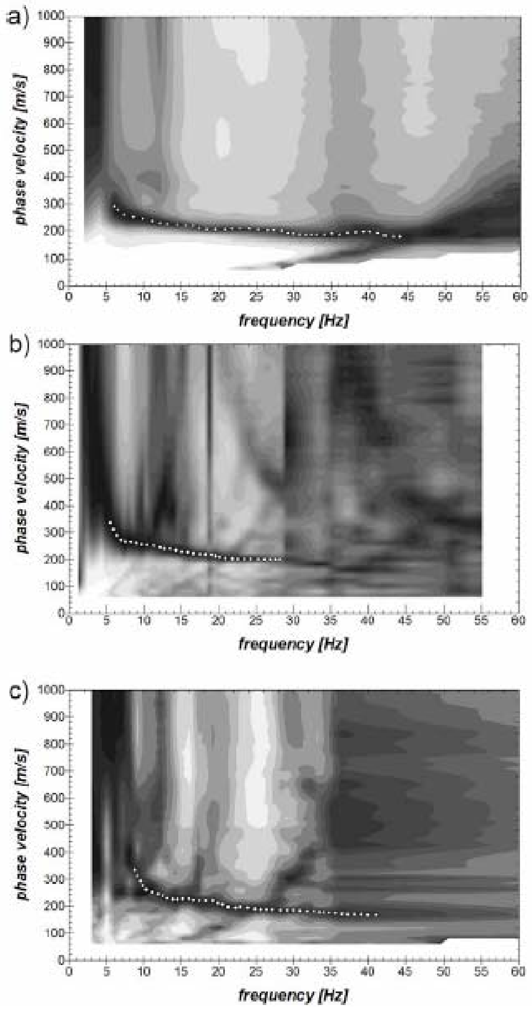

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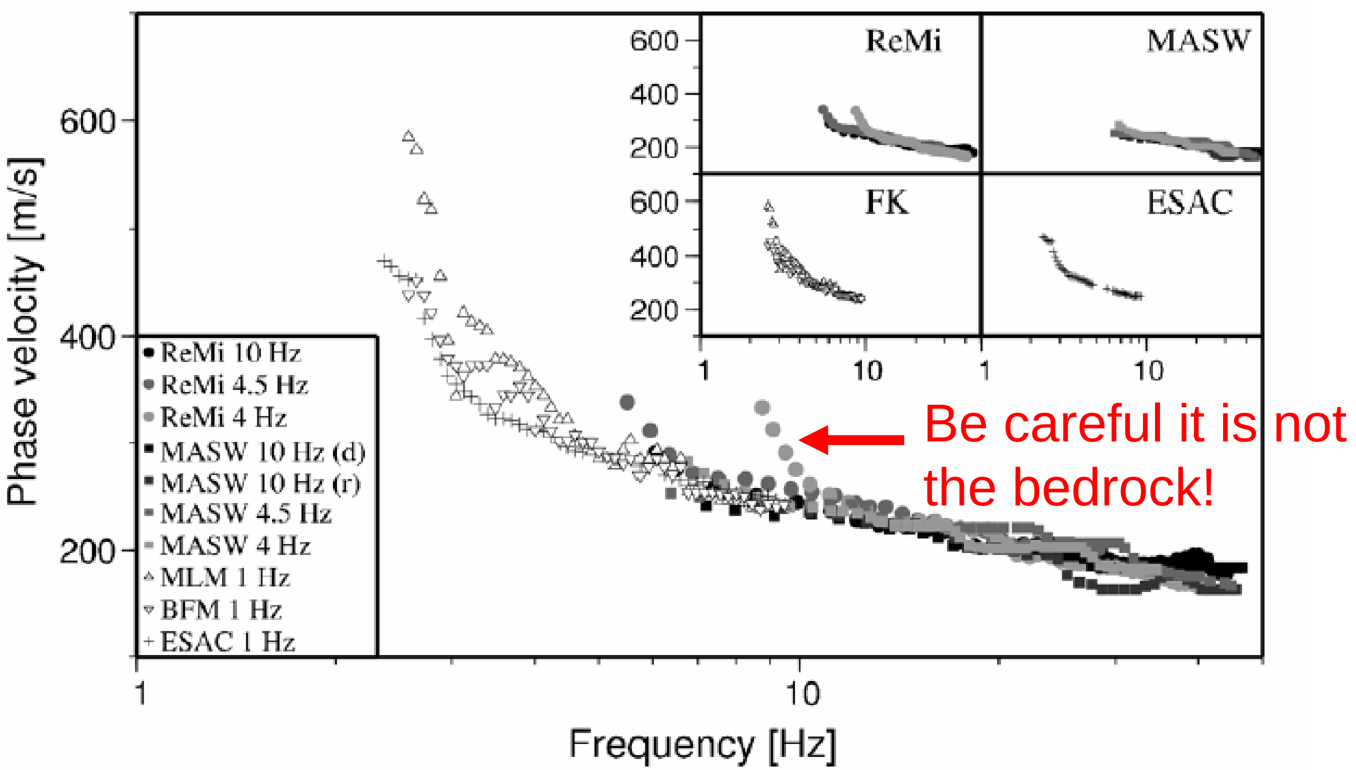
Power spectrum functions for linear geophone arrays from ReMi analysis (scale is arbitrary, white representing low values and black high values). The picked phase velocity values are shown as white dots. (a) 4 Hz, (b) 4.5 Hz, (c) 10 Hz

After Richwalski et al. (2007)

Comparison of all dispersion curves. (d) and (r) refer to direct and reverse shots of the active experiment, respectively.



	SWM	Refraction (S-h waves)
Advantages	Works also where soft and stiff layers alternate Reduced testing time on site Average properties (dynamic behaviour of the whole soil deposit)	Lateral variations (2D maps) Better positioning of interfaces with high stiffness contrast
Disadvantages	Resolution shallow layers Inverse problem 1D Model with plane and parallel layers Resolution decreases with depth	Errors (sometime large errors) due to slow or hidden layers Time consuming acquisition Limited number of layers (2 or 3)

- Any technique that provides wide frequency range
- Avoid ReMi (?)
- Inversion with higher modes if necessary
- Provide basic rules for check results (available wavelength vs depth&resolution)
- Lessons learned from Blind Test in Grenoble (?)
- Lessons learned from NERIES Project (?)

- Le tecniche passive possono essere utilizzate anche per investigare siti con bedrock profondi (>30 m)
- Dato il numero limitato di strumentazione in tal caso si perde in risoluzione superficiale
- In alcuni casi, comunque, l’influenza sulla risposta di sito osservata era non significativa.
- I siti devono essere accessibili
- La banda di frequenza analizzata puo’ essere di molto estesa combinando tecniche attive e passive

Cosa intendiamo per classificazione di sito?

(Vs sino bedrock? Vs30, Vs5 etc)

Tempistica?

Accessibilita'?

Riteniamo opportuno una calibrazione delle tecniche

per ogni tipologia di sito identificata (bacino, bordo bacino, roccia)?