QP and QS of Campi Flegrei from the inversion of rayleigh waves recorded during the SERAPIS project

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Summary

Seismic shots recorded during the SERAPIS experiment were used to search a 1D elastic and inelastic model of the Gulf of Pozzuoli, south of the Campi Flegrei caldera. Waveforms were gaussian filtered in the range 5-8 Hz with a frequency step of 0.5 Hz and a half-width of the filter equal to 0.5 Hz. A clear dispersion of the most energetic propagation mode was revealed. This property of the surface wave in the gulf of Pozzuoli was theoretically reproduced using the classical wave-number technique. To infer the best fit propagation model, we developed a semi-automated procedure of fitting of filtered traces with progressive adjustment of the model. The quality of the fitting was estimated using the semblance among each couple of waveform (synthetic and observed). Our formulation allowed us also to estimate the error on model parameter by mapping the noise on seismograms on the semblance. The obtained 1D model confirms that in average intrinsic Qp at the Campi Flegrei caldera is of the order of 300-500 which is a background value higher than that of other volcanic areas.

This report is a summary of a part of the phd thesis in Earth Sciences at University of Bari of Maria Trabace.

Data analysis

During the SERAPIS experiment, seismic signals produced by a battery of 12, 16-liters air-guns mounted on the oceanographic vessel NADIR (IFREMER) were recorded at a dense array of three-component, sea bottom (OBS) and on land seismographs installed in the bays of Naples and Pozzuoli (Zollo et al., 2003). The experiment was originally deployed to obtain 3D Vp and Vs images from the inversion of arrival times. The receiver array consisted of 70 ocean bottom receivers (OBS) and 84 land stations. In our study only OBS recordings were considered.

The OBS were equipped with 4.5 Hz three-component sensors and a continuous recording device (Judhenerc and Zollo, 2004). Before of the analysis all the waveforms were band-pass filtered between 5 and 15 Hz in order to work in the frequency range where the phase and amplitude response of the instruments deployed were the same.

In this report we discuss the results obtained by analyzing one of the seismic shots recorded during the experiment (Figure 1). We analyzed about 200 tra-

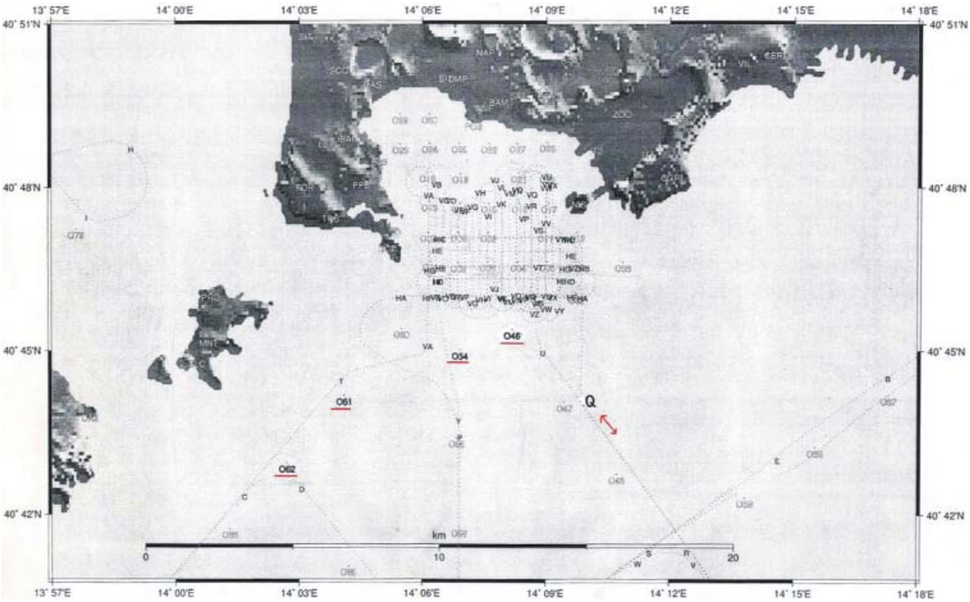

Fig. 1. Shot lines and position of OBs of the seismic experiment SERAPIS. The double arrow denotes the position of the seismic shot used in this study. The location of the four OBS used in this study are marked by a red line.

ces. At each trace we applied a Gaussian filter having a central frequency fc ranging from 5 to 10 Hz with a step of 0.5 Hz and a narrow bandwidth α=0.5 Hz (Dziewonski and Hales, 1972). The filter function is given by:

$$H(f) = \exp\left[-\pi^2 \frac{(f - f_c)^2}{\alpha^2}\right] \tag{1}$$

The filter function (1) could introduce systematic errors when the medium is strongly dispersive (i.e. group velocity rapidly varies with frequency); in this case the error can be reduced by increasing α, with an increase in the degree of uncertainty on the velocity estimate and the risk of the interference between adjacent modes (Dziewonski and Hales, 1972).An example of the filtering of traces is shown in Figure 2 for the OBS #61. Figure 2 indicates that there is a strong frequency dependence of the arrival time of the maximum energy, which is a clear dispersive effect. Many researchers (for instance Yao

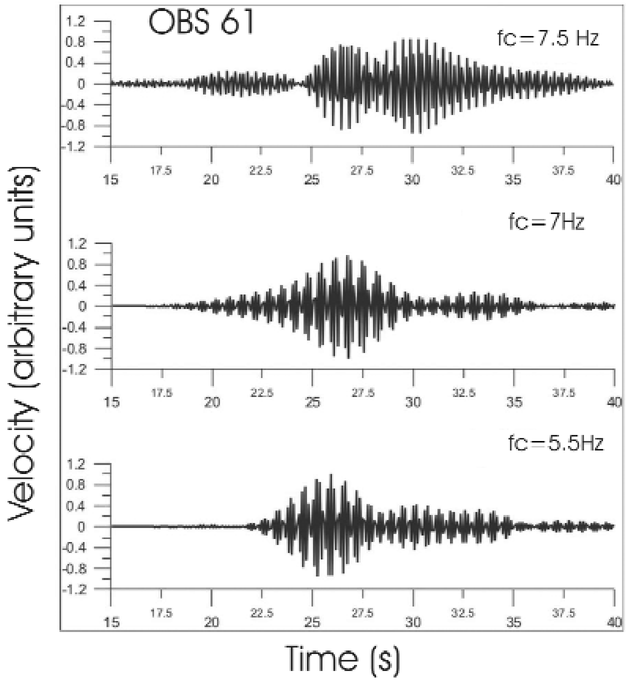

Fig. 2. Gaussian filtered recordings at the OBS 61 (α=0.5 Hz). The central frequency fc of the filter are reported on each seismogram.

and Dorman, 1992) have used the arrival time t* of the maximum of the wave packet to estimate group velocity U at each frequency through the relation:

$$U(f) = \frac{R}{t^*} \tag{2}$$

where R is the source to receiver distance. As discussed in Dziewonsky and Hales (1972) the previous equation furnishes unbiased estimates of group velocity only for an unimodal surface wave.

Since, as an effect of the filtering, several modes of propagation are inferred at the same central frequency, we strongly suspect that there is a bias in the arrival time of each group of each propagation mode and do not use, as preliminary information, the dispersion curve of group velocity. The existence of a clear dispersion of the most energetic propagation mode was also inferred by analyzing the arrival time of the most energetic mode vs. distance (Figure 3). We inferred the presence of a multiplicity of propagation modes at all the OBS. A summarizing plot of data is reported in Figure 4.

We discarded all the traces for which the filtered traces show unwanted arrivals preceding the first arrival on unfiltered traces and experienced a great difficult in separating the surface wave contribution. The main reason for this difficult is that the 4.5 Hz cutoff of the instruments does not allow us to study

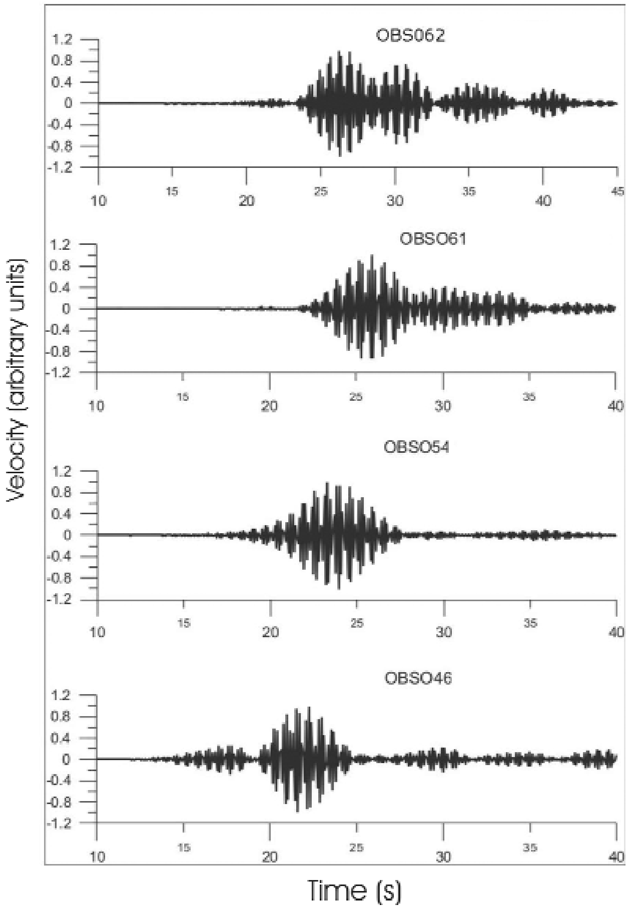

Fig. 3. Plot of Gaussian filtered traces (fc=5.5 Hz, α=0.5 HZ) at four OBS. This figure clearly shows the dispersion vs. distance of the most energetic propagation mode.

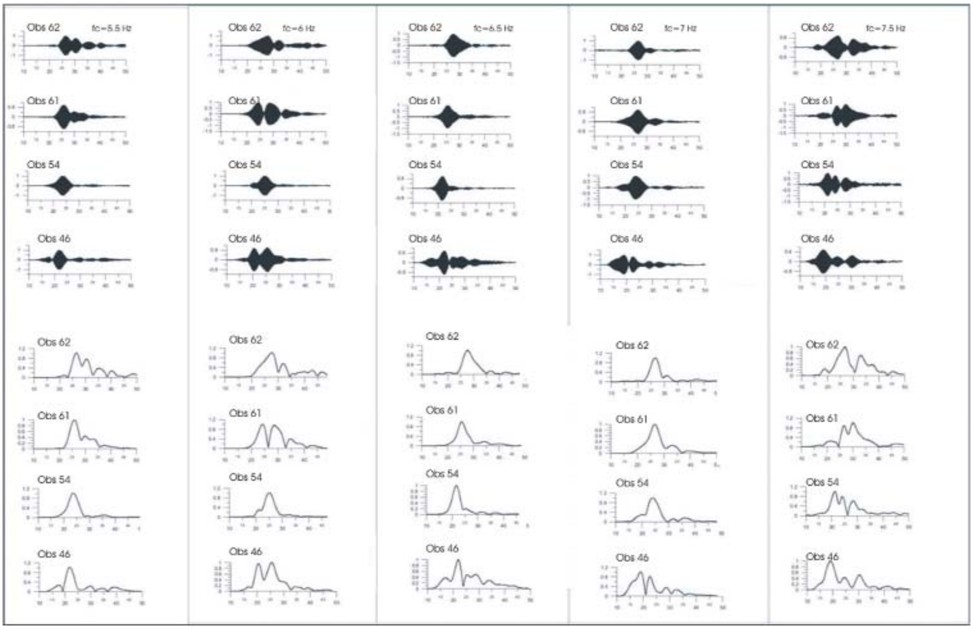

Fig. 4. Summary of filtered Gaussian traces and their absolute values.

the contribution of the Rayleigh waves below this frequency where it may be dominant. Moreover, we were unable to obtain filtered traces showing a clear contribution of the fundamental mode of propagation for frequencies of the gaussian filter greater than 7.5-8 Hz, owing to the superimposition of different modes of propagation. Another problem we experienced was the effect of high frequency noise (> 8Hz) on the filtered traces which tends to produce fictitious arrivals. The lower limit (5Hz) imposed by the instrument will limit our analysis only to a very thin layer below the sea bottom.

After the analysis of the waveforms we selected four stations which show a clearly readable and significant contribution of the Rayleigh waves and, from each trace, we extracted 5 gaussian filtered traces with fc= 5.5, 6, 6.5, 7 and 7.5 Hz and α=0.5 Hz. The source to receiver distance is reported in Table 1.

Tab. 1. Identification number of the Obs used in this study and their distance from the shot.

Obs #	Source to receiver distance (km)
46	9.5
54	10.3
61	12.8
62	14

Data inversion

The inversion technique used to model Vp, Vs, Qp and Qs is a refinement of the de Lorenzo et al. (2003) technique. It is based on the waveform fitting. The philosophy is to find the 1D elastic and anelastic model which best reproduces the entire waveform content in each frequency range where the Gaussian filter has been applied. This allows us to avoid the introduction of a subjective criterion of classification of the several propagation modes of surface waves which can be, at least partially, superimposed and then produce a bias in group velocity measurement. We use as data five Gaussian filtered waveforms (fc=5.5; 6; 6.5; 7; 7.5 Hz) with a bandpass width α=0.5 Hz at the four OBS. The time window considered is T=25 s for each seismogram. Since the time sampling of the four OBS is dt=4 ms, the number of data available for each filtered trace is N=6250. The total number of data is then Ntot=6250 x 20= 125000. Since we consider five layer with four unknown parameters the problem is clearly overdetermined.
The theoretical seismograms are computed using the discrete wave-number technique (Bouchon, 1981) considering a 1D layered anelastic medium and a point-like source whose frequency content is about equal to the inverse of the rise time τ of a ramp-like displacement source. In our calculation τ was fixed to 5 ms to simulate the sources of the SERAPIS experiment. Many studies on the inversion of Rayleigh waves (e.g. Nolet, 1990; and references therein) have pointed out the non-uniqueness of the solution. In the attempt to avoid the inference of a velocity and inelastic model which corresponds to a local minimum, we propose a forward modeling which uses as a priori information the 1D velocity model previously computed for the gulf of Pozzuoli (Judhenerc and Zollo, 2004). Considering the average frequency content of our data (f=6Hz) and a maximum value of phase velocity of 2 km/s (which is a typical maximum value for Vs of oceanic sediments) we can obtain a first order estimate of depth penetration of our signals of the order of 1 km. For this reason we have considered an initial model of two layers (a layer over an halfspace). The elastic and anelastic properties of the layers are reported in Table 2.
In the modeling we have subdivided the first layer in 4 layers of equal thickness (0.25 km). The procedure consists of evaluating the matching between theoretical and observed seismograms varying one of each model parameter (Vp, Vs, Qp and Qs) of each layer at a time.

Tab. 2. The initial model.

Layer id	Thickness (km)	Depth of the top (km)	Density (g/cm^3)	Vp (km/s)	Vs (km/s)	Qp	Qs
1	1	0	1.9	2.86	1	300	100
2	∞	1	2.3	2.86	1.65	300	100

The quality of the fit is quantified by computing the value assumed by the semblance operator s (Telford et al.,1990). For a couple of signals s is defined as:

$$s_j = \frac{\sum_{i=1}^{npt}\left[U_j^{est}(t_i)+U_j^{obs}(t_i)\right]^2}{2\sum_{i=1}^{npt}\left(U_j^{est}(t_i)\right)^2+\left(U_j^{obs}(t_i)\right)^2} \tag{3}$$

where U_j^{est} and U_j^{obs} are respectively the theoretical and the observed seismogram at time ti; by considering N traces the semblance will be:

$$s = \sum_{j=1}^{N} s_j \tag{4}$$

To quantify the error on model parameter we map the error on data in the model parameter space. The error on semblance due to the noise on data is:

$$\Delta s = \left|\frac{\partial s}{\partial U^{obs}}\right|\Delta U^{obs} \tag{5}$$

$$s_j = \frac{\sum_{i=1}^{npt}\left[U_j^{est}(t_i)+U_j^{obs}(t_i)\right]^2}{2\sum_{i=1}^{npt}\left(U_j^{est}(t_i)\right)^2+\left(U_j^{obs}(t_i)\right)^2} \tag{6}$$

$$\frac{\partial s}{\partial U^{obs}} = \frac{2s\left[U_j^{est}(t_i)+U_j^{obs}(t_i)\right]}{\sum_{i=1}^{npt}\left[U_j^{est}(t_i)+U_j^{obs}(t_i)\right]^2}+\frac{2sU^{obs}(t_i)}{\sum_{i=1}^{npt}\left[U_j^{est}(t_i)+U_j^{obs}(t_i)\right]^2} \tag{7}$$

The error on the observed seismogram is estimated in L1 norm by using the relationship:

$$\Delta U^{obs} = \frac{1}{T_1}\int_0^{T_1}\left|U^{obs}\right|dt$$

where T1=5 sec is the time duration of the signal, considered as the noise, which precedes the first arrival on each seismogram. The error on semblance

can then be mapped in the model parameter space to determine all the parameters which lie in the range [s-Δs;s]. This will allow us to estimate the average value of each parameter and the uncertainty affecting it.
Figure 5 shows the matching between observed and theoretical waveforms for the initial velocity model for a central frequency fc=7 Hz. Figure 6 shows the variations of the semblance with varying the model parameters during the inversion procedure.

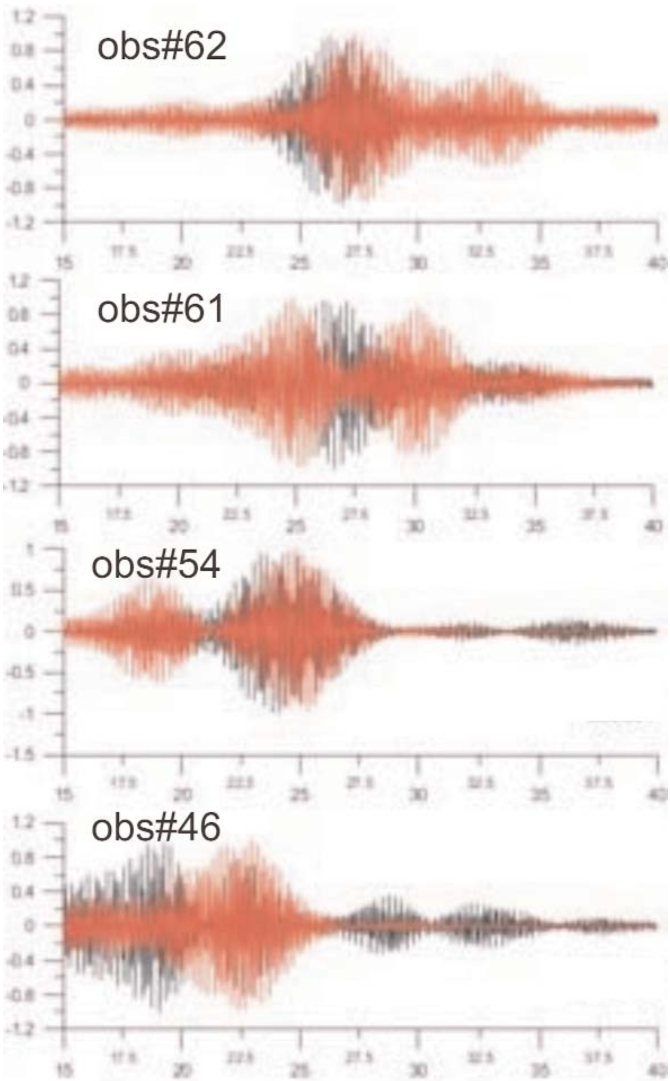

Fig. 5. Waveform fitting at fc= 6 Hz using the initial elastic and inelastic velocity model. The observed waveforms are plotted with a black line; the synthetic waveforms with a red line.

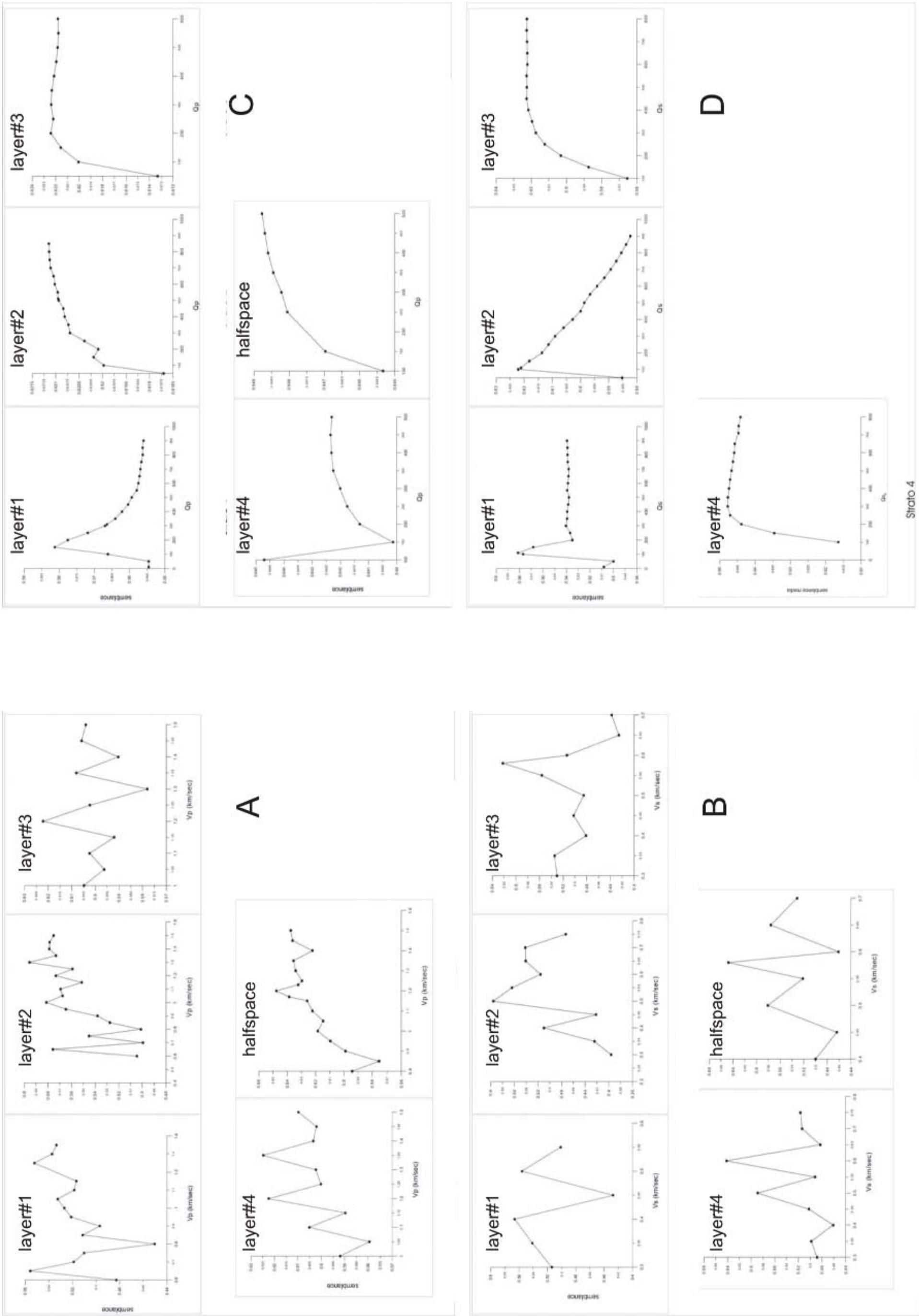

Fig. 6. Variation of the semblance with varying **(a)** Vp of each layer; **(b)** Vs of each layer; **(c)** Qp of each layer; **(d)** Qs of each layer.

The final velocity and inelastic model is reported in Table 3, together with the error on each parameter. The matching between observed and theoretical Gaussian filtered waveforms at fc=7Hz is shown in Figure 7. To better evaluate the quality of fitting we use the absolute value of the waveforms which allows us to enhance the energy content of the signals.

Tab. 3. Final velocity and inelasticity model.

Thickness (km)	Vp (km/s)	Vs (km/s)	Density (g/cm^3)	Qp	Qs
0.25	1.3± .05	0.4± 0.1	1.5± 0.1	150± 20	110± 20
0.25	1.3± .05	0.5± 0.1	1.2± 0.2	900±200	100± 10
0.25	1.2±0.05	0.58± .1	1.7± 0.2	200± 10	450±150
0.25	1.2± 0.1	0.58± .1	1.7± 0.1	200± 80	450±200

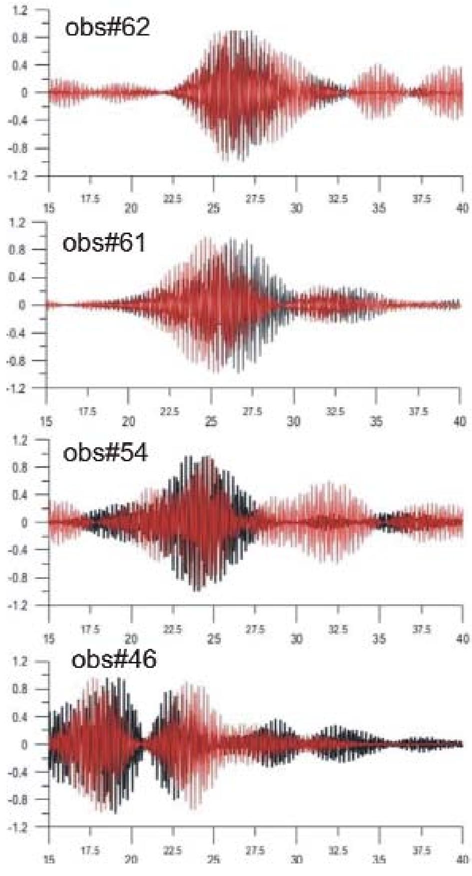

Fig. 7. Waveform fitting at fc=7 Hz using the final elastic and inelastic velocity model. The observed waveforms are plotted with a black line; the synthetic waveforms with a red line.

Figure 8 shows the comparison between the absolute values of filtered theoretical and observed waveforms for fc= 7 Hz. Since we did not perform the deconvolution for the instrumental response we compared the unit normalized traces.

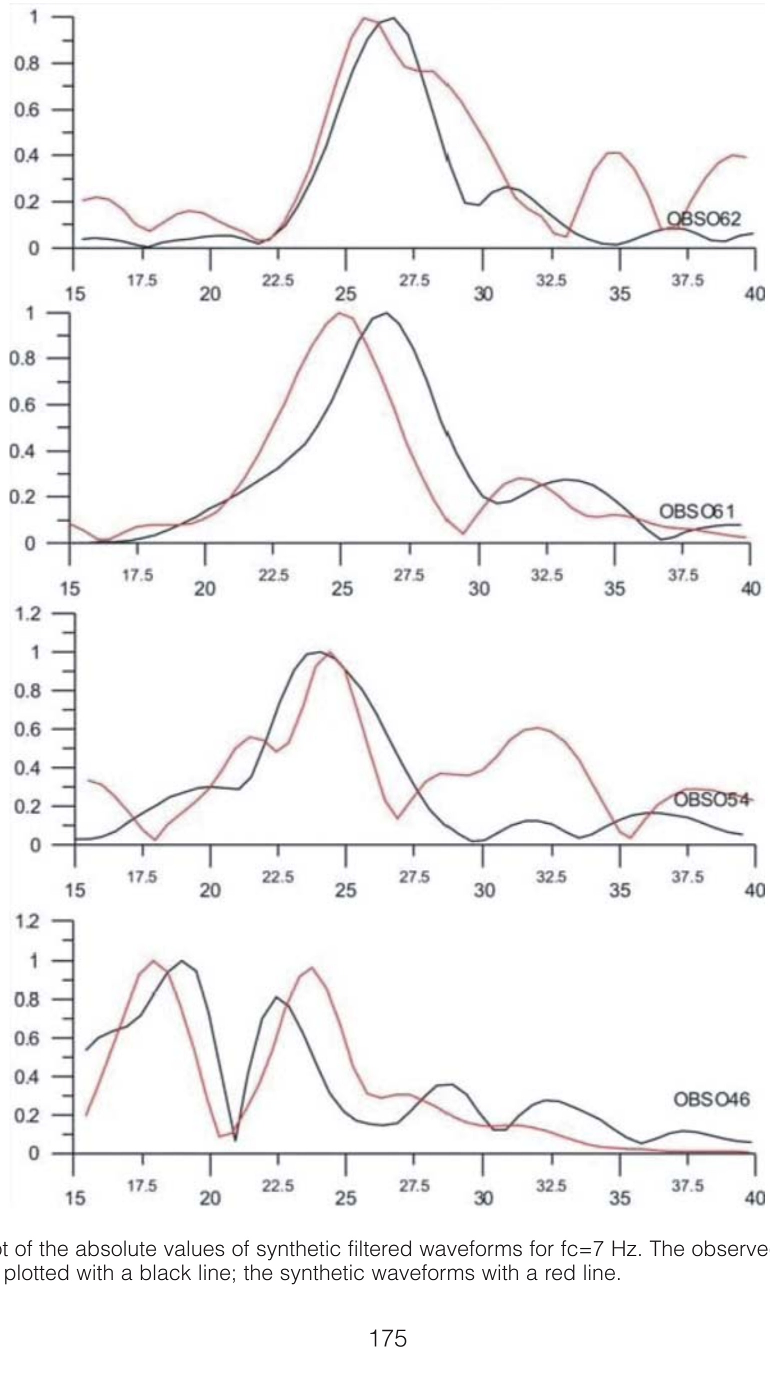

Fig. 8. Plot of the absolute values of synthetic filtered waveforms for fc=7 Hz. The observed waveforms are plotted with a black line; the synthetic waveforms with a red line.

Very interestingly the main energy content is very well reproduced by the model and, more interestingly, the repartition of the energy in at least two propagation modes, is also well reproduced, in particular at the OBS 46. The value of semblance increases of 28% from the initial model to the final one.

Conclusion

Qp values in the four thin layers is generally in the order of 100-200 which is roughly the same magnitude order of the average values previously inferred from a tomographic study of the Campi Flegrei caldera (de Lorenzo et al.,2001), whereas only a very high Qp (Qp=900) is inferred between 0.25 and 0.5 km. Vp values inferred for the first layer are slightly lower than those obtained by Judhenerc and Zollo (2004).

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