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IMAGING OF THE WEATHERED ZONE AND ESTIMATION OF Q IN SEDIMENTS

Abstract. Phase-consistent filtering together with frequency time analysis (multiple filtering) is used to obtain the group velocity dispersion curve of the fundamental mode and first few higher modes of Rayleigh waves from seismic exploration data. The group velocities are then inverted to obtain the shear wave velocity distribution to a depth of about 30 meters. Comparison of the experimental traces with the complete synthetic seismograms, computed using structural parameters of the inversion results as input, allow us to estimate che anelastic properties of the medium: Q (quality factor) values are observed to vary from 5 to 20.

INTRODUCTION

In seismic exploration for hydrocarbons, the source and the geophones are deployed on the surface of the earth in such a way as to attenuate the undesired signals. One of these unwanted signals is given by the fundamental and higher modes of Rayleigh waves and is termed ground roll in seismic literature. Ground roll not attenuated during the acquisition stage obscures the reflected signals and makes them difficult to interpret. The seismic analyst consequently eliminates it by suitable processing techniques. However, ground roll does in reality contain information on the medium through which it propagates. In this paper we show how this information can be retrieved. The method commonly known as frequency - time analysis (FTAN) followed by phase - consistent filtering is employed to obtain the group velocity dispersion curve of the ground roll. Since the dispersion is primarily controlled by the shear wave velocity of the medium, the information retrieved from the data is inverted to obtain the S-wave velocity distribution versus depth.

Finally, by comparing experimental and synthetic seismic sections, constructed by inversion of the dispersion data, it is possible to estimate che anelasticity of the medium.

In seismic exploration, knowledge of the variation with depth of the S-wave velocity can be employed in what is termed static correction (Mari, 1984), to enhance the quality of seismic data by improving the continuity of the reflected signals. Furthermore, when the density is known, the shear modulus can be determined, which, in combination with the quality factor Q, is important in studies of foundation vibrations, effects of earthquakes and slope stability (Gabriels, et al., 1987). Knowledge of the attenuation is important in the acquisition, processing and interpretation of high-resolution seismic data, vertical seismic profile data and borehole sonic log data, as well as in the deduction of other physical properties such as permeability (Klimentos and McCann, 1990).

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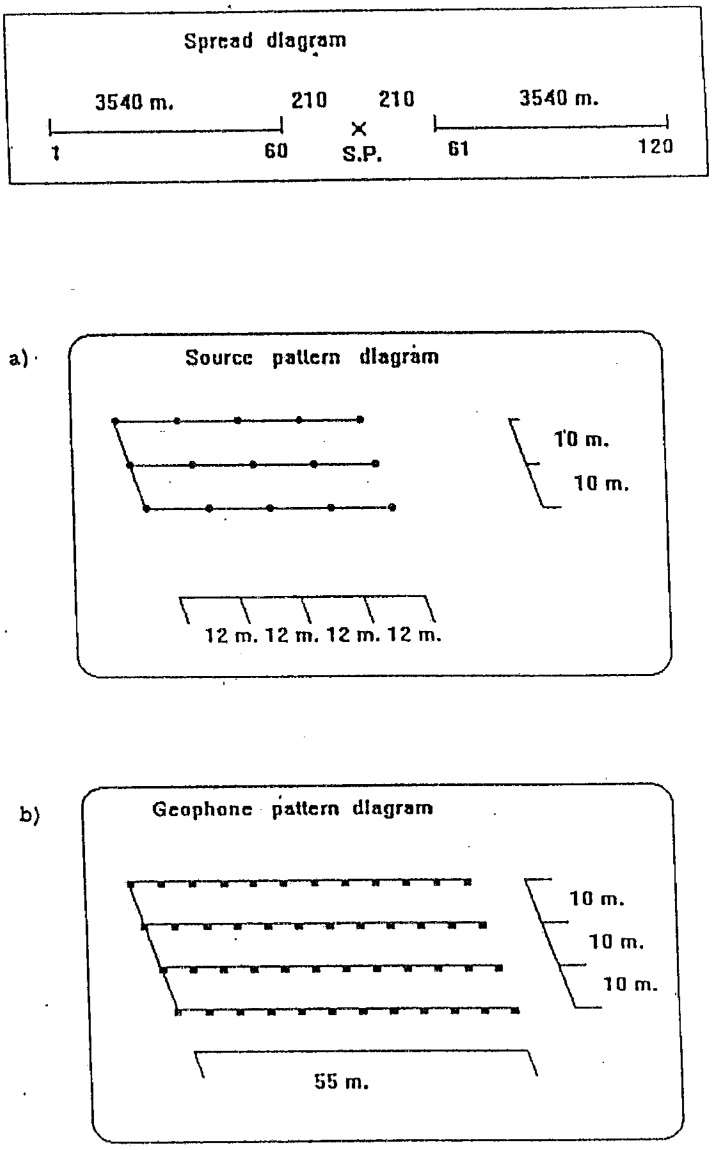


Fig. 1 - Spread configuration, a) source and b) geophone patterns.

ANALYSIS OF RAYLEIGH WAVES

Diverse methods exist for the determination and analysis of Rayleigh modes. These methods utilize group velocities or phase velocities of the fundamental mode and/or higher modes depending on the scope of the study. Gabriels et al. (1987) used the frequency-wavenumber domain to measure the dispersion properties of Rayleigh modes and obtained the phase velocity for the fundamental mode and the first 5 higher modes. Their method has an advantage in that it shows good resolution for higher modes. But the weakness of this method is that it requires the acquisition of a large amount of data and subsequently a large amount of computer time is needed in the processing stage (Jongmans, 1991). Moreover it suffers from aliasing whenever both the spatial and temporal samplings are not sufficiently small.

Another method used to calculate the phase velocity dispersion curve takes into consideration the entire wavefield (McMechan and Yedlin, 1981). The method involves two linear transformations: slant stack and 1-D Fourier transform. Slant stack produces a wavefield in the phase slowness-time intercept (p-t) plane in which phase velocities are separated ($p=1/c$,

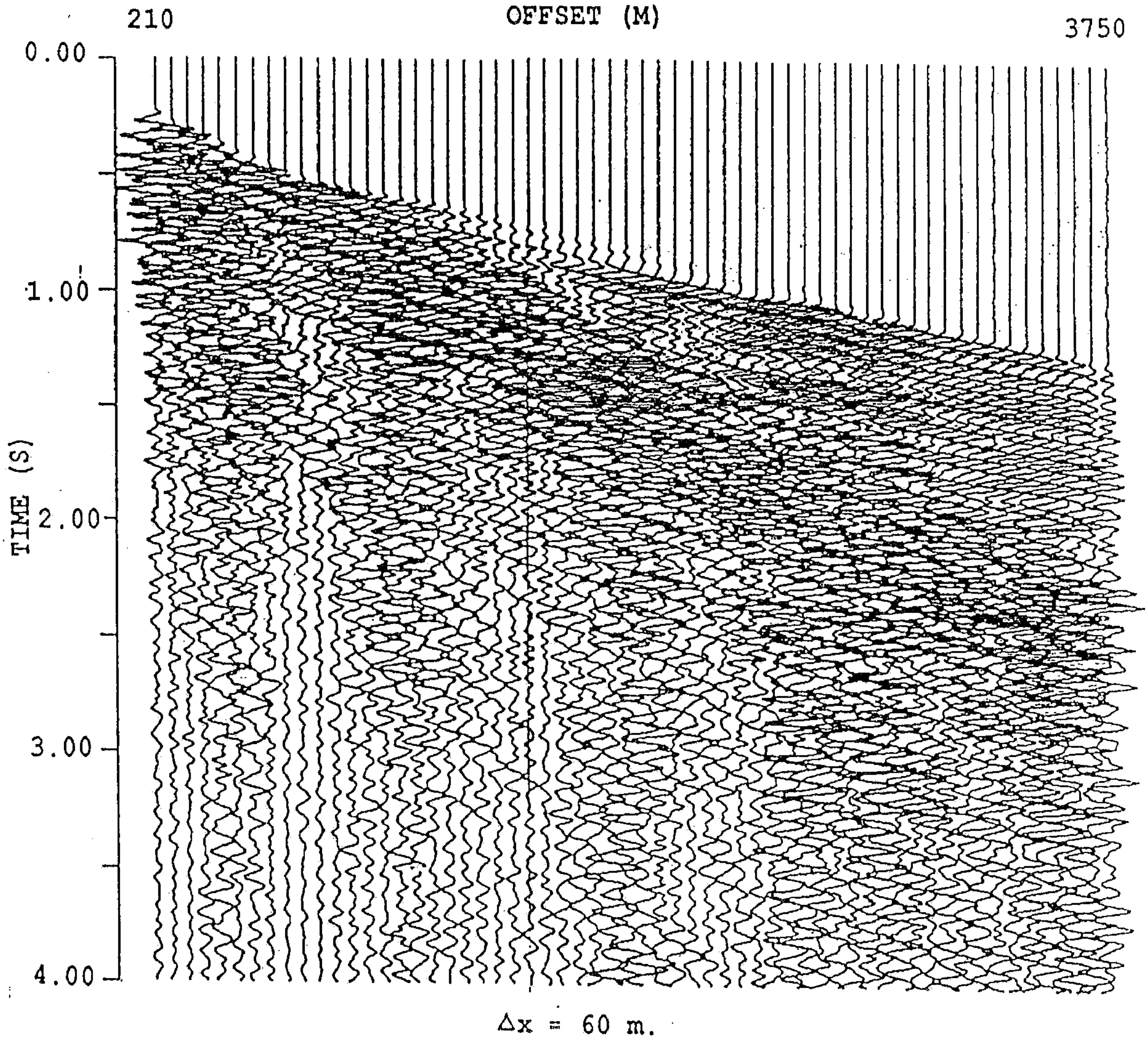


Fig. 2 - Example of available data sets (channels 61 to 120 from shot n. 1).

c=phase velocity). The frequency associated with each phase velocity is then obtained by a 1-D Fourier transform over t. Thus the data wavefield is linearly transformed from the time-distance domain into the slowness-frequency (p-ω) domain where dispersion curves are imaged. This method again suffers aliasing whenever sampling rates in both time and offset are not sufficiently high.

The method we employed to analyze the Rayleigh waves is FTAN (Levshin et al., 1972), which is an improved version of the multiple filter technique first introduced in digital form by Dziewonski et al. (1969). In this work we used FTAN followed by phase-consistent filtering (also called phase-matched filtering or phase equalization) to facilitate signal selection (Herrmann and Russell, 1990). Bandpass filtering and windowing are also used where necessary. This method, which measures group velocity dispersion curve of a single seismic trace, doesn't employ much computer time and is free from the aliasing problem.

DATA ACQUISITION AND PROCESSING

The data were collected in North Africa in September 1987. The spread configurations for both the sources and the geophones are shown in Fig. 1. For each shot, 120 receiver group arrays were employed. Sixty geophones were used in each group array.

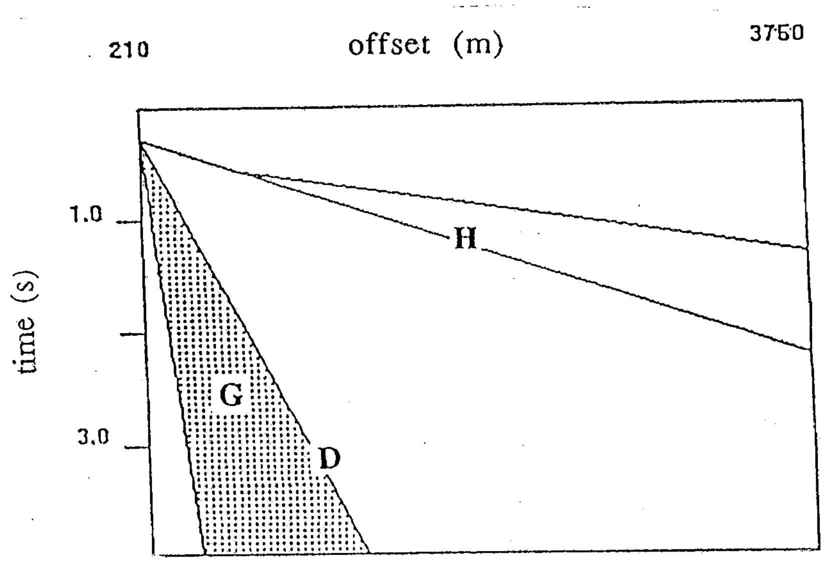


Fig. 3 - Schematic representation of main phases visible in the section: H) head wave, D) direct wave, G) ground roll.

Fifteen explosions, with a total weight of 18.75 kg, were fired simultaneously at a depth of 3 metres. Four-second-long time series having a sampling interval of 2 ms were collected. The frequency content of the data ranges from 4 Hz to 50 Hz. Fig. 2 shows an example of 60 traces acquired in the experiment. In this section the head waves, direct waves and ground roll are identified (Fig. 3).

In Fig. 4 three seismic traces extracted from the data are shown. In all these traces the ground roll can be observed. Denoting these Rayleigh modes with X(t) we can compute the spectrum S(ω):

$$S(\omega)=\int_{-\infty}^{+\infty} X(t)\exp(-i\omega t)\,dt.$$

We pass the spectrum through a system of narrow-band Gaussian filters H (ω) with varying central frequency ω_i:

$$H(\omega)=\exp\left(-\alpha\left(\frac{\omega-\omega_i}{\omega_i}\right)^2\right),$$

where α is the parameter controlling the resolving power of the filters.

The frequency-time representation of the output function is

$$Y(\omega_i, t)=\int_{-\infty}^{+\infty} S(\omega)\exp\left(-\alpha\left(\frac{\omega-\omega_i}{\omega_i}\right)^2\right)\exp(i\omega t)\,d\omega.$$

The diagram $|Y(\omega_i, t)|$ is the signal envelope (ridge crest) and is useful for the determination of group velocities. On the diagram log $|Y(\omega_i, t)|$, the dispersion curve $\tau(\omega)$ follows the ridge crest. $\tau(\omega)$ is given by (see Ratnikova, 1990)

$$\tau(\omega)=\frac{r}{U(\omega)}-\phi_s'(\omega),$$

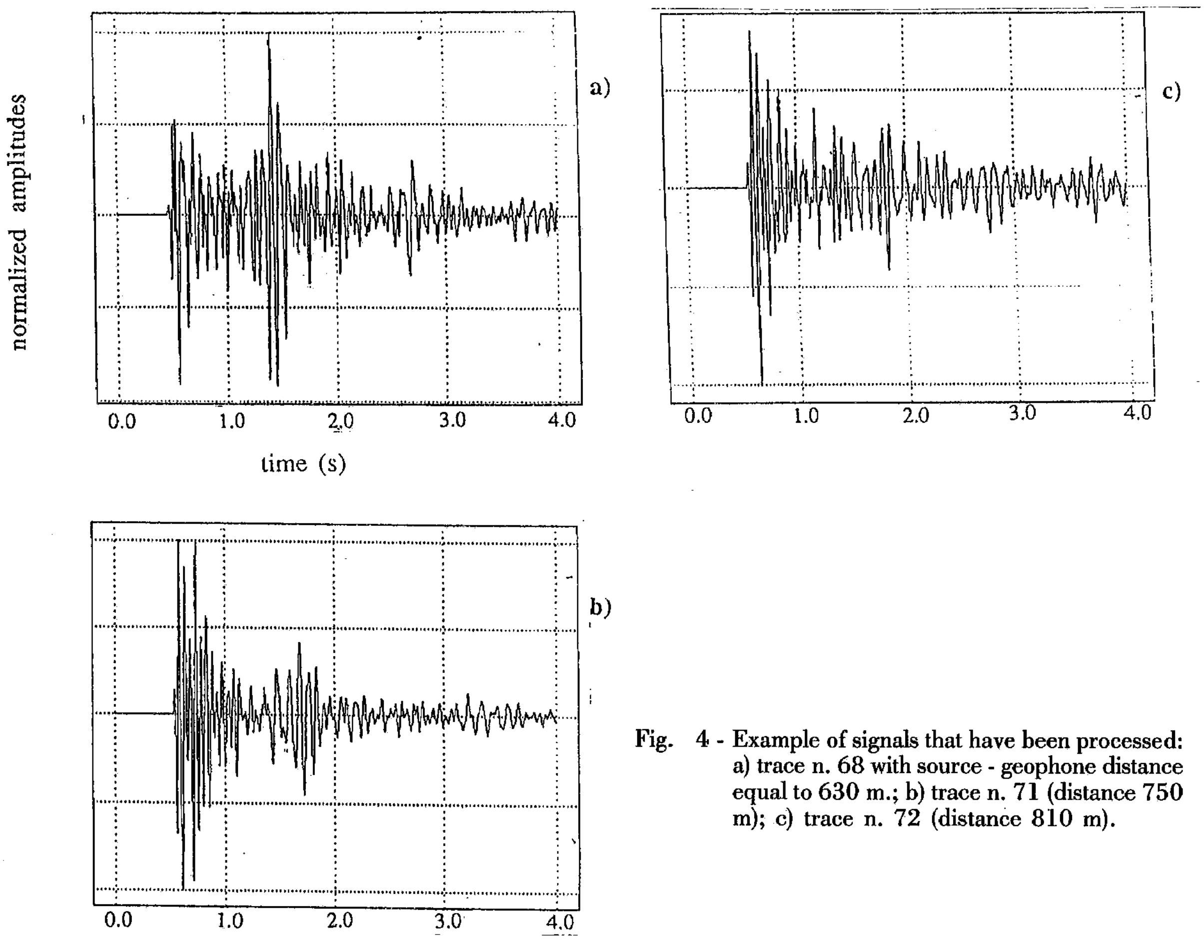


Fig. 4 - Example of signals that have been processed: a) trace n. 68 with source - geophone distance equal to 630 m.; b) trace n. 71 (distance 750 m); c) trace n. 72 (distance 810 m).

where r is the distance the wave has travelled, U is the group velocity and ϕ_s' is the derivative of the spectral phase at the source.

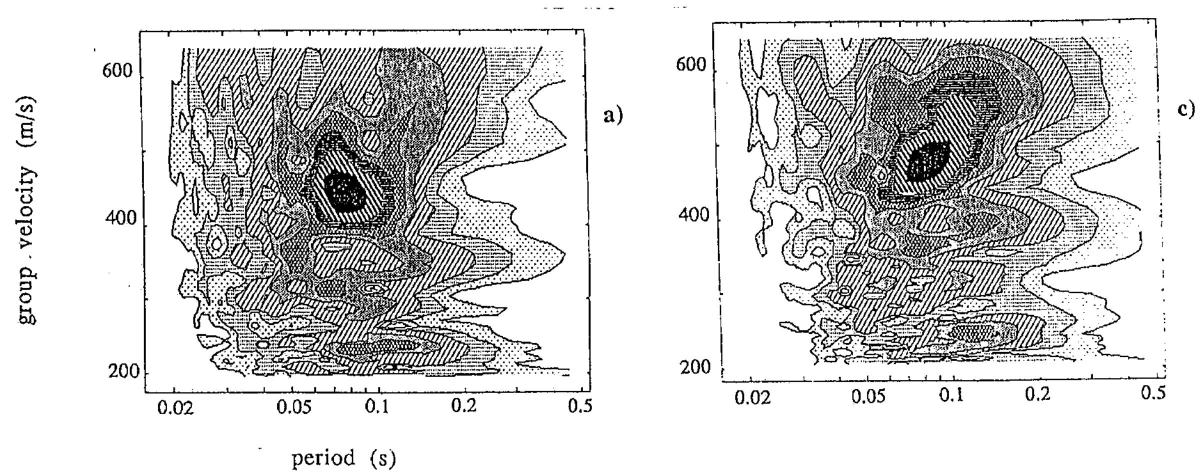
The diagrams for the three traces after the application of deconvolution to remove the instrument responses, are shown in Fig. 5. In all these diagrams one can see three zones of maximum amplitude concentrated around the velocity values of 230 m/s, 300-320 m/s and 450-470 m/s corresponding, respectively, to the fundamental mode and to the first few higher modes of Rayleigh waves. The similarity in the three diagrams indicates the absence of significant lateral heterogeneity along the sampled path. Once the dispersion curve of each mode is approximately determined, we can transform the signal into the undispersed one using phase-consistent filtering. The scope of this processing step is to remove arrivals not pertinent to the mode under consideration, which we call here noise. The procedure is as follows: on the diagram log $|Y(\omega_i, t)|$, approximate group time $\tilde{\tau}(\omega)$ values are identified at some discrete frequencies and subsequent cubic spline interpolation is performed. The approximate phase delay ($\tilde{\Phi}(\omega)$) of the signal is then calcultated from the group time (Ratnikova, 1990):

$$\tilde{\Phi}(\omega) = -\int_{-\infty}^{+\infty} \tilde{\tau}(\eta)\, d\eta + C_1\omega + C_2,$$

where C_1 and C_2 are constants that control the time shift (Shapiro, 1992).

Now we add

$$\int_{-\infty}^{+\infty} \tilde{\tau}(\eta)\, d\eta - C_1\omega - C_2$$



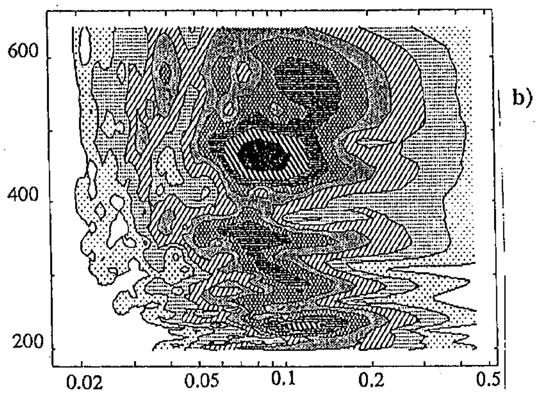


Fig. 5 - Frequency time analysis diagram for the traces shown in Fig. 4. Darker areas correspond to larger amplitudes.

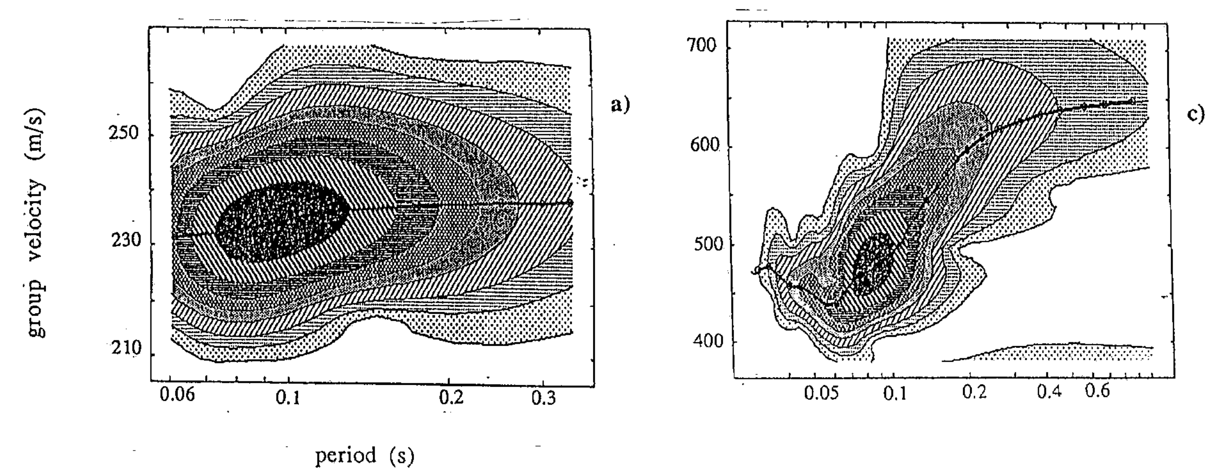
to the true phase spectrum $\phi(\omega)$ of the signal. If the phase spectrum ($\tilde{\Phi}(\omega)$) is a reasonable approximation to the true phase spectrum $\phi(\omega)$, then $|\phi(\omega)-\tilde{\Phi}(\omega)|$ is sufficiently close to zero and the inverse transformation will give a compressed or undispersed signal shifted to a certain instant of time, which is controlled by the constants C_1 and C_2. The noise will be either spread out or, if compressed, is compressed only slightly and located away from the signal. The noise is then removed by application of an amplitude time window to obtain a filtered signal which is subsequently transformed into the frequency domain (Herrmann et al., 1990). The final step is to restore the original phase by adding

$$-\int_0^{+\infty} \tilde{\tau}(\eta)\, d\eta + C_1\omega + C_2,$$

to the phase spectrum $\phi(\omega)$. The output of this final step is seen in Fig. 6.

INVERSION FOR SHEAR-WAVE VELOCITY STRUCTURE AND Q

From Fig. 6 one can see that FTAN analysis allows separation of the various modes. The fundamental mode is shown in (a). The higher modes in (b) and (c). The higher modes are interpreted as first and third respectively, by comparing their dispersion curves with the synthetic ones, computed for a structural model consistent with the fundamental mode dispersion. Due to the structural properties, in the frequency band considered, the second higher mode is not excited, and the modes of higher number cannot be well separated due to their very similar group velocities. Therefore we decided to invert the dispersion curves of the fundamental and the first higher mode, keeping the dispersion curve of the third higher mode only for a qualitative control of the inversion results. Once the modes are identified, the next step is to perform their



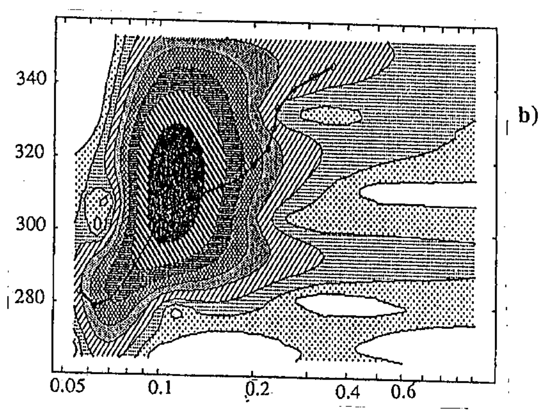


Fig. 6 - Example of the dispersion curves obtained; a) fundamental mode, b) first higher mode and c) third higher mode.

Table 1 - Frequency values with observation error used in the inversion. The first five values belong to the fundamental mode, while the next six to the first higher mode.

Frequency (Hz)	Error (m/s)
7.09	10
8.92	8
11.11	7
13.69	6
16.66	5
8.13	25
8.47	20
9.34	18
11.11	15
16.66	12
19.23	9

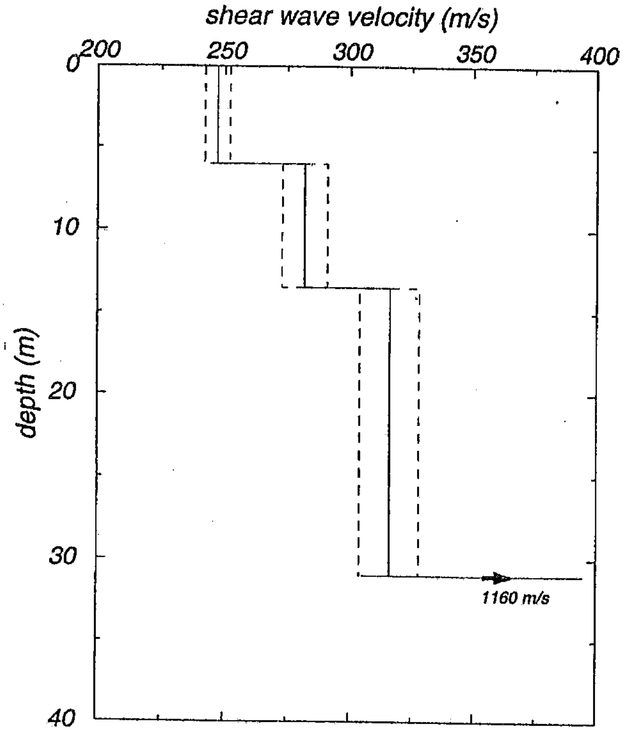


Fig. 7 - Schematic representation of the solution to the inverse problem. The solid line represents the model used in calculations and dashed lines contain the possible solutions. 1160 m/s represents S-wave velocity in the bedrock.

inversion. The dispersion curves of Rayleigh modes are primarily sensitive to the shear wave velocity and therefore can be inverted to find the S-wave velocity distribution with depth. The inversion is undertaken by the hedgehog method, developed by Keilis-Borok and Yanovskaja (1967), and discussed in detail by Panza (1981).

In this procedure an initial structural model is considered, which is represented by a set of parameters whose values vary with depth. We constructed the model from an analysis of the first arrivals, sonic and density logs. Perturbing the part of the initial model that is most sensitive to the available data, group velocities corresponding to all periods determined in the data processing are computed. These velocities are then compared with the observed velocities of the same periods. If the difference between the computed and the observed values for any of the periods exceeds the pre-set threshold value, the model is rejected and another model in the neighbourhood of the first is considered, and the procedure repeated. If, on the other hand, the difference is below the pre-set limit for the longest period, the procedure extends to the next shorter period. If the difference is below the threshold limit for all periods the model is accepted, provided that the r.m.s., σ, of the entire data set is less than the pre-set threshold. In our case all models with 8 m/s are accepted. The error at each frequency point was estimated on the basis of repetition of dispersion measurements for more traces, and is given in Table 1.

The different solutions of the inverse problem are contained within the stripe defined by the dashed lines shown in Fig. 7; in the same figure the solid line represents the starting model used in computations. Values of the V_p/V_s ratio versus V_s obtained from the inversion are given as small solid circles in Fig. 8; as can be seen from the figure these values are in quite good agreement with the measurements, made by Stumpel et al. (1984), for dry and partially saturated sands.

The availability of the distribution versus depth of the elastic parameters, and of the density allows us to compute complete synthetic seismograms (Panza, 1985) and therefore to determine the anelastic properties of the medium (Craglietto et al., 1989). Since local site effects, scattering and lateral lithological changes prevent meaningful direct measurements of amplitude decay

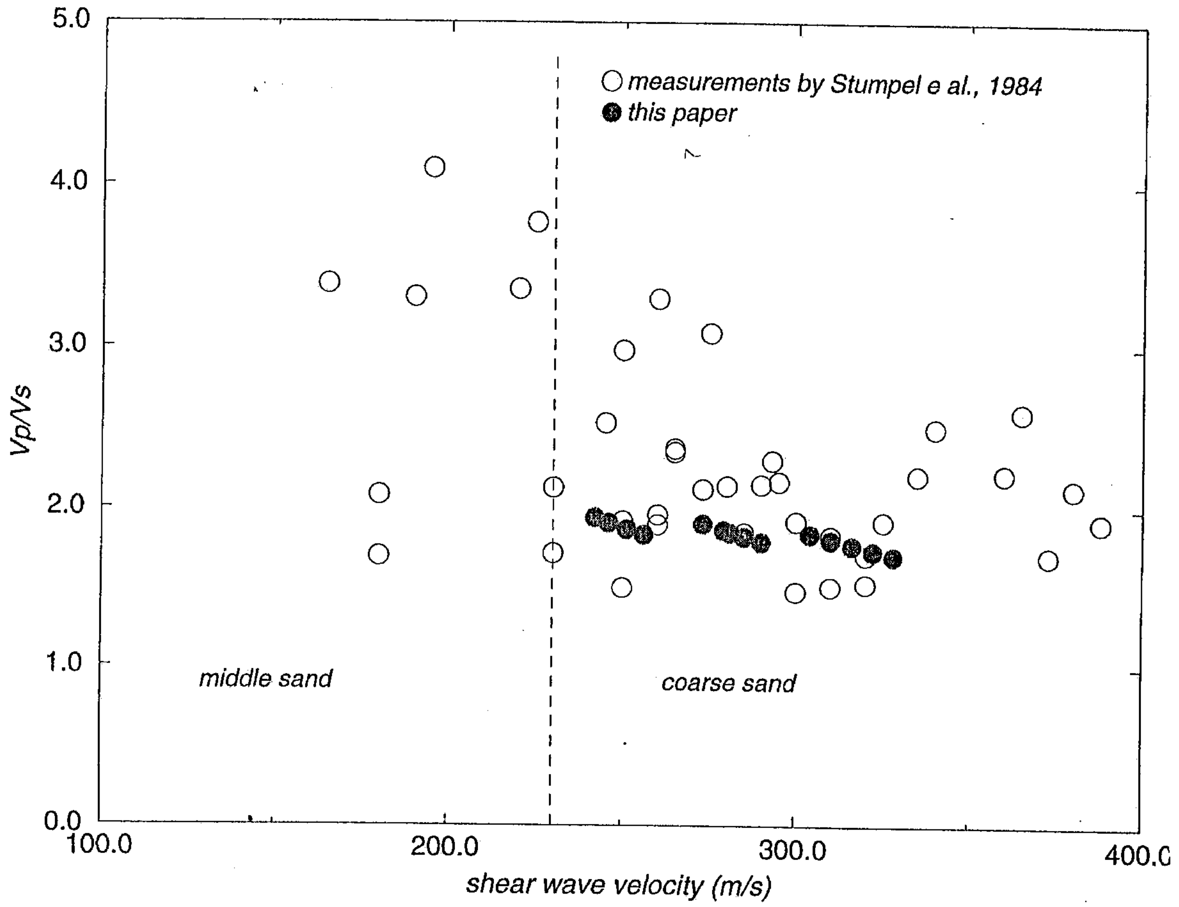


Fig. 8 - Vp/Vs ratio versus Vs for sands (dry and partially saturated) (from Stumpel et al., 1984) in relation to our results.

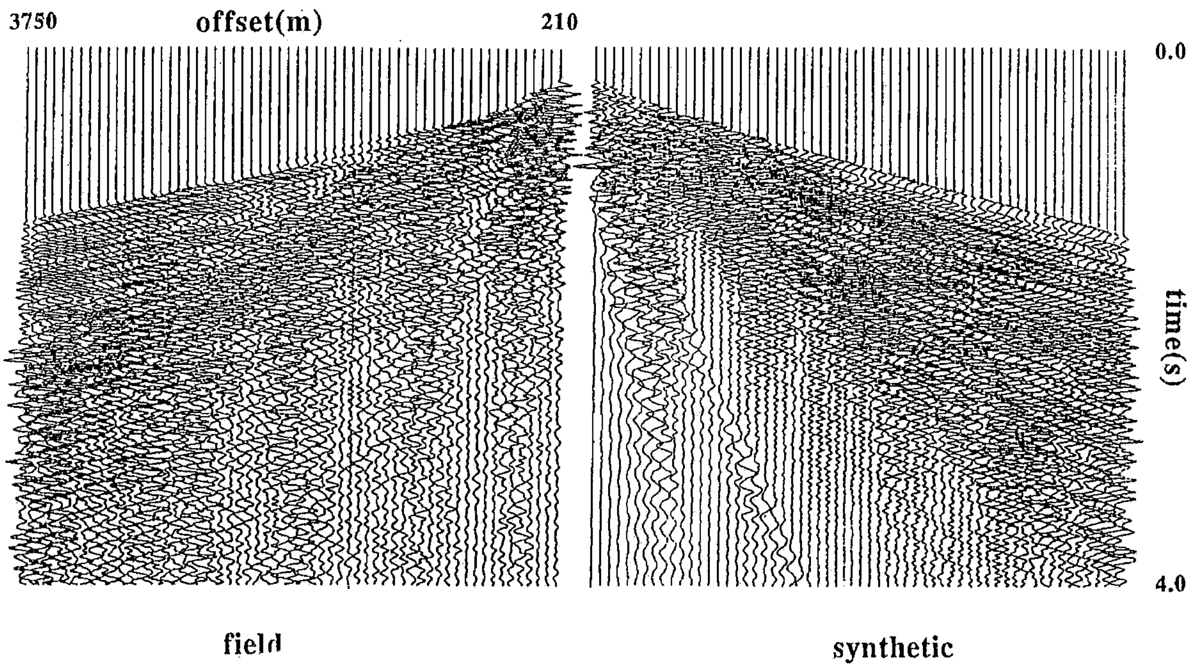


Fig. 9 - Field data and their corresponding synthetic section computed using a single point instantaneous explosive source and a single geophone per channel.

Table 2 - Derived distribution of Q values versus depth.

Depth (m)	Quality factor
0-31	5-20
31-490	25-60
490-2200	25-120

(Mokhtar et al., 1988), quality factor Q values can be estimated by direct comparison of experimental and synthetic data. Initially very high and very low values of Q are selected. Subsequently the range between these extreme values is narrowed until we get a satisfactory match between the synthetic and experimental data, special attention being paid to the similarity of signal envelopes. An example of synthetic data, to be compared with the experimental ones shown in Fig. 2, is given in Fig. 9. Similar signals are obtained using, in the computations, Q values that are variable over the ranges given in Table 2.

CONCLUSION

From seismic reflection data it is possible to retrieve the S-wave velocity distribution versus depth from the dispersion curve of Rayleigh modes, determined using frequency-time analysis and phase-consistent filtering. The method used doesn't suffer from aliasing and doesn't require large amounts of computer time, since data from single stations are used. The result obtained for the elastic properties is comparable to that described by Gabriels et al. (1987), who used the stacking of 256 stations. Here, in addition, with the use of complete synthetic seismograms, we have given an estimate of the quality factor Q.

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REFERENCES

Craglietto A., Panza G.F., Mitchell B.J. and Costa G.; 1989: *Anelastic properties of the crust in the Mediterranean area.* In: American Geophysical Union, Geophysical Monograph 51, IUGG, **6**, pp. 179-196.

Dziewonski A., Bloch S. and Landisman M.; 1969: *A technique for the analysis of transient seismic signals.* Bull. Seism. Soc. Am., **59**, 427-444.

Gabriels P., Snieder R. and Nolet G.; 1987: *In situ measurements of shear-wave velocity in sediments with higher-mode Rayleigh waves.* Geophysical Prospecting, **35**, 187-196.

Herrmann R. and Russell D.; 1990: *Ground roll. Rejection using adaptive phase-matched filters.* Geophysics, **55**, 776-781.

Jongmans D.; 1991: *L'influence des structures géologiques sur l'amplification des ondes sismiques.* Thèse de Docteur en Sciences Appliquées, Universitè de Liège.

Keilis-Borok V. and Yanovskaja T.; 1967: *Inverse problems of seismology (structural review).* Geophys. J.R. Astr. Soc., **13**, 223-234.

Klimentos T. and McCann C.; 1990: *Relationships among compressional wave attenuation, porosity, clay content, and permeability in sandstones.* Geophysics, **55**, 998-1014.

Levshin A., Pisarenko V. and Pogrebinsky G.; 1972: *On a frequency-time analysis of oscillations.* Annales Geophys., **28**, 211-218.

Mari J.L.; 1984: *Estimation of static corrections for shear wave profiling using the dispersion properties of Love waves.* Geophysics, **49**, 1169-1179.

McMechan G. and Yedlin M.; 1981: *Analysis of dispersive waves by wavefield transformation.* Geophysics, **46**, 869-874.

Mokthar T., Hermann R. and Russel D.; 1988: *Seismic velocity and Q model for shallow structure of the Arabian shield from short period Rayleigh waves.* Geophysics, **53**, 1379-1387.

Panza G.; 1981: *The resolving power of seismic surface wave with respect to crust and upper mantle structural models.* In: R. Cassinis (ed), The solution of the inverse problem in Geophysical Interpretation. Plenum press, pp. 39-77.

Panza G.; 1985: *Synthetic seismograms: the Rayleigh waves modal summation.* J. Geophys., **58**, 125-145.

Ratnikova L.; 1990: *Frequency-time analysis of surface waves.* In: Workshop on "Earthquake Sources and Regional Lithospheric Structures from Seismic Wave Data", Nov. 19-30, 1990, I.C.T.P., Trieste.

Shapiro N.; 1992: *Frequency-time analysis of surface waves.* In: Workshop on Three-Dimensional Modelling of Seismic Waves Generation, Propagation and their Inversion, Nov. 30-Dec. 11, 1992, I.C.T.P., Trieste.

Stümpel H., Khaler S., Meissner R. and Milkereit B.; 1984: *The use of seismic shear waves and compressional waves for lithological problems of shallow sediments.* Geophysical Prospecting, **32**, 662-675.