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INVERSION FOR SHEAR WAVE VELOCITY IN SEDIMENTS AND ESTIMATION OF NEAR-SURFACE ANELASTIC PARAMETERS

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ABSTRACT: Standard seismic refraction data recorded in North-East Italy, were used to define the shear wave velocities in sediments by means of the inversion of the Rayleigh wave dispersion relation. The group velocity for the fundamental mode and the phase velocity for the fundamental and first few higher modes were measured in the frequency range from 12 to 28 Hz. The inversion of these data gave a well defined S-wave velocity structure of the weathered zone to a depth of 38 meters. Using this information in the computation of complete synthetic seismograms for anelastic media, allowed a qualitative estimation, in the same depth range, of the distribution with depth of the quality factor, Q.

Key words/phrases: Anelastic parameters, dispersion, inversion, Rayleigh waves, shear wave velocity

INTRODUCTION

The knowledge of the distribution of the shear wave velocity (Vs) and of the anelastic parameters versus depth can be utilised to enhance the quality of seismic data which usually suffer significant degradation by highly inhomogeneous and poorly compacted near surface weathered layer. This information, together with the density, can be used to determine the shear modulus, a constant that is strictly correlated to lithology and useful in earthquake engineering to define the effects of earthquakes, foundation vibrations and slope stabilities (Gabriels *et al.*, 1987; Jongmans and Demanet,

1993). Moreover Poisson's ratio is obtainable when compressional wave velocity (Vp) variation with depth is known from independent data.

Rayleigh waves, which are often generated during seismic reflection and refraction surveys, contain important information about the physical properties of the medium through which they propagate, mainly the shear wave velocity. In seismic prospecting the application of the techniques that permit the detection and analysis of surface waves and their inversion are relatively few. The source-generated noise has been studied in several areas of Texas having distinctly different near-surface layering and velocities by Jolly and Mifsud (1971). Ten years later Essen *et al.* (1981) examined the propagation of surface waves in marine sediments paying attention to the Vp/Vs ratio and Al-Hussein *et al.* (1981) analyzed dispersion patterns of ground roll in Saudi Arabia to determine the best field acquisition system in reflection surveys. In the same year McMechan and Yedlin (1981) used a method developed earlier by Nolet and Panza (1976) in earthquake seismology for surface wave analysis. Mari (1984), Szelwis and Behle (1986) estimated the static correction for S-wave seismic profiling from the dispersion properties of Love and Rayleigh waves, respectively. *In situ* measurements of shear wave velocity to a depth of 50 m by the recognition of six higher-modes of Rayleigh waves in the frequency-wave number domain were made by Gabriels *et al.* (1987). Al-Eqabi and Herrmann (1993) demonstrated that a laterally varying shallow S-wave structure, derived from the dispersion of the ground roll, can explain observed lateral variations in the direct S-wave arrivals, while Jongmans and Demanet (1993) used the information derived from seismic refraction data to define the dynamic characteristics of soils, by the employment of the same technique. In addition they later found Qs by determining the frequency dependent attenuation factor of the surface waves. Tilahun Mammo *et al.* (1995) used the residuals of ground roll present in reflection seismic data recorded in a desert area to obtain an estimation of the quality factor using complete synthetic seismograms.

In this paper it is shown how well defined shallow shear wave velocity structure can be obtained using the inversion of dispersion relations and how the depth variation of the quality factor can be inferred from numerical modelling of the whole signal.

DATA ACQUISITION AND PROCESSING

The data were recorded in North-East Italy along two lines perpendicular to each other having 200 m and 150 m of length, respectively. The source used was cannon mini-bang buried at a depth of 0.50 m. A 24 channel seismograph (Bison mod. 7024) was used with inter-receiver spacing of 1 m. Data of 4 s length were obtained having a sampling interval of .004 s. Figures 1a and b show the raw data for two perpendicular lines at the same site. These wave-fields were then respectively transformed into frequency-wave number domain via 2-D Fourier transformation (Figs 2a and b). The Nyquist frequency (125 Hz) was greater than the maximum frequency found in the data and therefore no aliasing occurred. The Nyquist wave-number was 0.5 1/m and thus it was possible to have a sufficiently narrow spatial sampling for the wave--lengths under consideration. In Fig. 2a the fundamental (0) and few higher modes (1=first and 2=second) can be seen whereas in Fig. 2b only the fundamental mode (0) is clearly visible. This is due to the fact that the number of channels for the seismic refraction survey of Fig. lb is small (144). This fact hampered a sufficient sampling in the wave number domain with the consequent problem of a poor resolution for the higher modes.

Frequency-time analysis (Levshin *et al.*, 1972) was applied to single channel (e.g. Fig. 3a) to measure group velocity. The algorithm we used, derived from multiple filter analysis (Dziewonski *et al.*, 1969) employs a system of narrow-band Gaussian filters through which is passed the spectrum of the signal. Figure 3b shows the consequent bi-dimensional representation of the signal envelope as a function of the group velocity and of the frequency. On the diagram, the dispersion curve is defined by the ridge crest. Because of the time overlapping of the higher modes, this technique is in general suitable only for the fundamental mode. In our case the fundamental mode is visible in the left lower part of Fig. 3b for group velocities around 0.200 kms^{-1} and frequencies not exceeding 30 Hz. A band-pass filtering with windowing (Ratnikova, 1990; Shapiro, 1992) was utilised to separate the fundamental mode (Figs 3c and d) from the available signal.

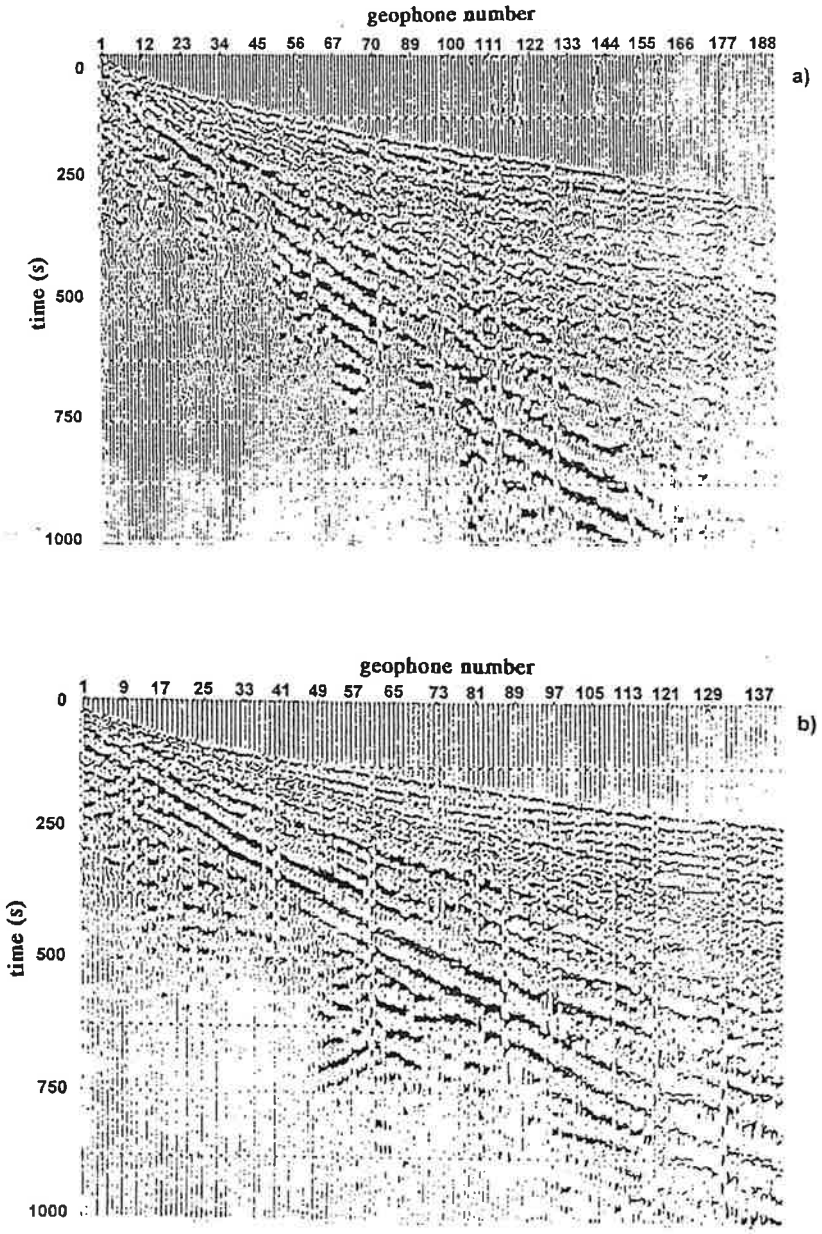

Fig. 1. **a) Seismic refraction data with 200 m and b) 150 m long array of receivers.** The inter-geophone spacing for both of them is 1m.

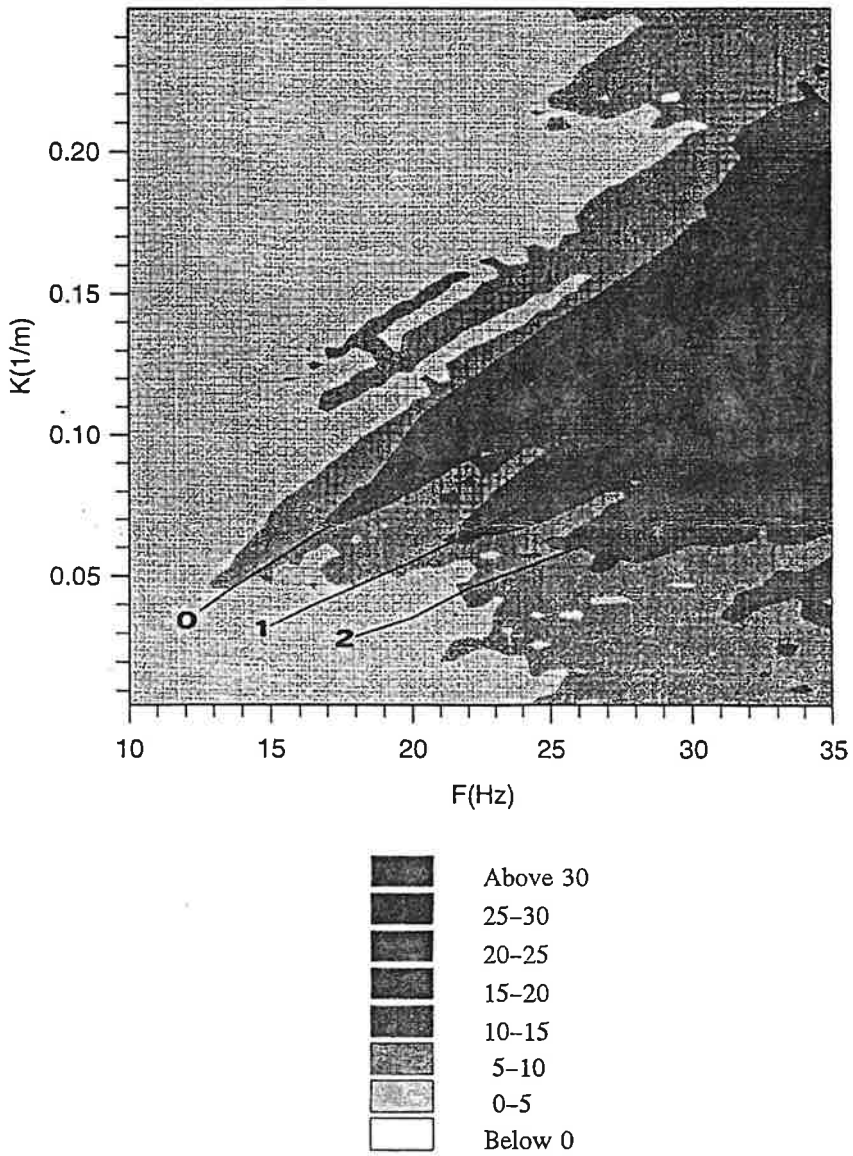

Fig. 2a. 2-D Amplitude spectrum of the data of Fig. 1a. (0, fundamental mode.)

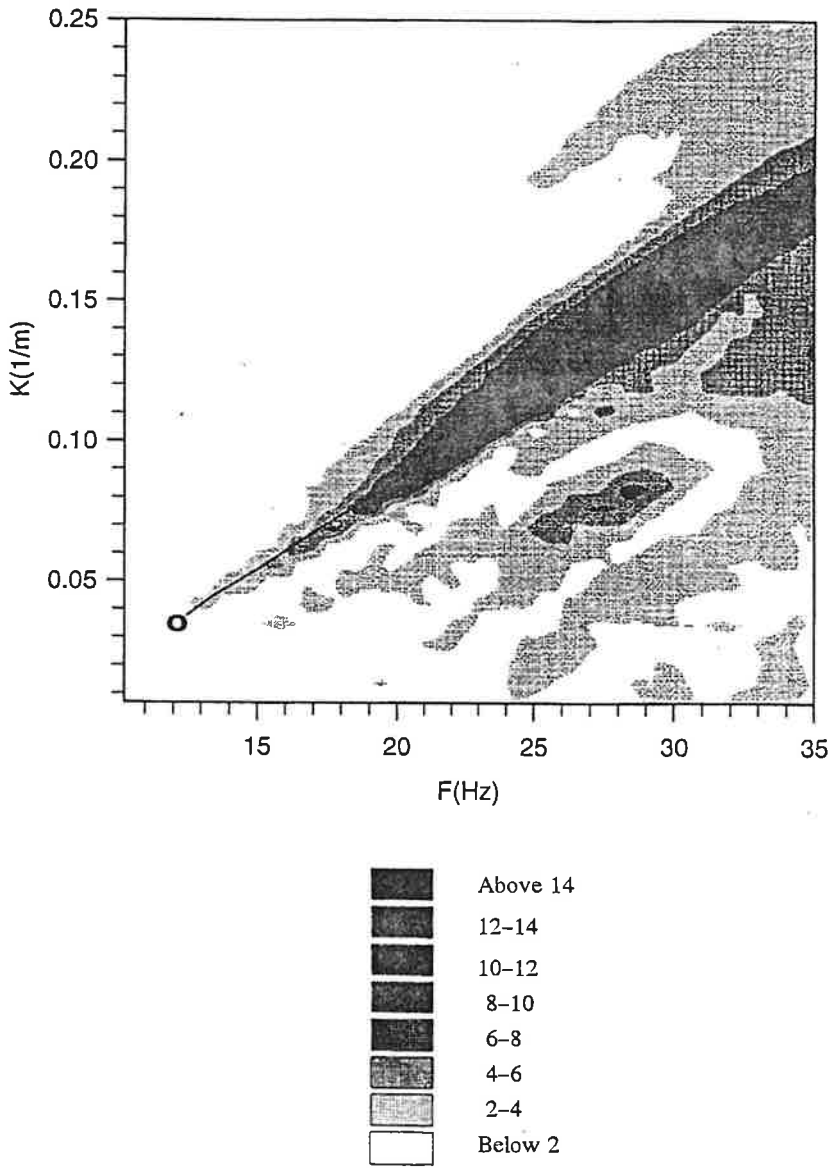

Fig. 2b. 2-D Amplitude spectrum of the data of Fig. 1b. (0, fundamental mode; 1, first higher mode; 2, second higher mode.)

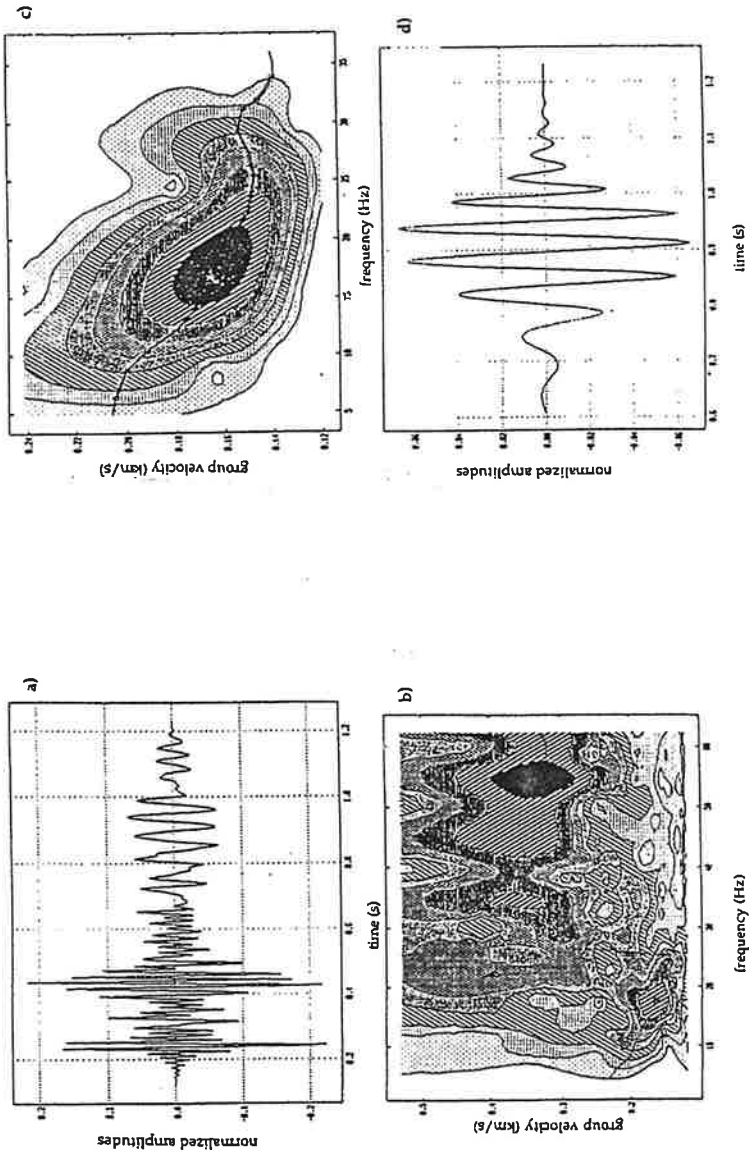

Fig. 3. a) Signal recorded at the offset of 144 m. Low frequency amplitudes belong to the fundamental mode.
b) bi-dimensional frequency-time representation for the signal of Fig. 3a. Instantaneous amplitudes of the fundamental mode are clearly circumscribed as function of frequency and group velocity.
c) group velocity dispersion curve for the fundamental mode after floating point filtering.
d) the uncontaminated signal of the fundamental mode retrieved after processing.

INVERSION FOR SHEAR WAVE VELOCITY AND ATTENUATION

Since the dispersion curves of the modes are sensitive to the shear wave velocity these can be inverted to determine the S-wave velocity distribution with depth. The non-linear inversion was carried out using a modified version of the Hedgehog method developed by Keilis-Borok and Yanovskaja (1967) and discussed in detail by Panza (1981). Only a brief discussion will be given here.

An initial model represented by a set of parameters (shear and compressional wave velocities and densities in a layered earth) whose magnitudes vary with depth was considered. We constructed the compressional wave velocity model from the analysis of the first arrivals (Fig. 4), using the generalised reciprocal method (GRM) (Palmer, 1981) while densities were inferred from the literature. Perturbing some of the model parameters at a time in a systematic manner varying each parameter by multiples of a basic increment, phase and/or group velocities corresponding to all frequencies under consideration were computed. These theoretical velocities were then compared with experimental velocities of equal frequencies. If the root mean square error of the entire data set was less than a critical value (in our case 14 ms^{-1}) and if no individual calculated velocity differed from its experimental counterpart of the same frequency by more than a certain value (see Table 1) the model was accepted as a solution. The inversion procedure was divided into two steps in order to reduce the number of variables investigated in the multidimensional parameters space. First we had inverted group and phase velocity for the fundamental mode to fix approximately the shear wave velocity in the near-surface 4 m. Then phase velocities for fundamental, first and second higher modes were inverted varying the S-wave velocity from 4 to 38 m. The different solutions of the inverse problem were contained within the interval defined by the dashed lines (Fig. 5). At the bottom of the model, where the Vs velocity was equal to 1190 ms^{-1}, the uncertainty in the shear wave velocity is around 35 ms^{-1}. In the next step we compute complete synthetic seismogram (Panza, 1985; Panza and Suhadolc, 1987) for the structural model shown in Fig. 5 (solid line), using the same acquisition array employed for the field data. These synthetic seismograms were used to define the anelastic properties of the medium (Craglietto *et al.*, 1989) which otherwise, because of local site effects, scattering and lateral lithological changes (Mokhtar *et al.*, 1988), are difficult to obtain by direct measurement.

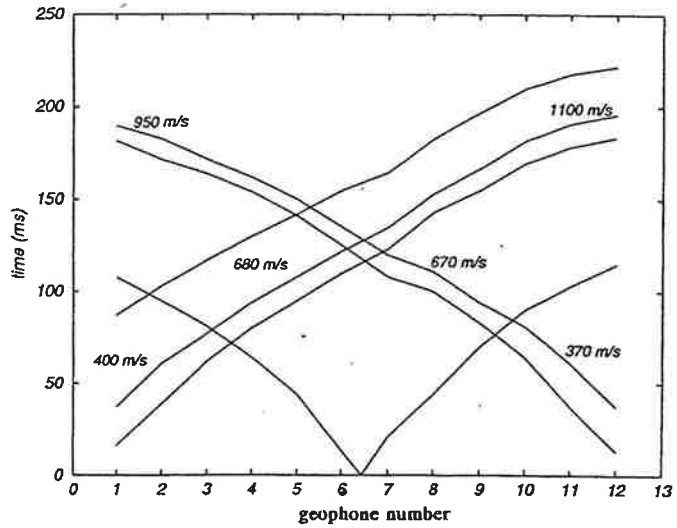

Fig. 4. Travel time curves derived from refraction seismic to obtain a near–surface compressional wave velocity model.

Table 1. Dispersion function and observational errors used in the inversion. First five points belong to the second higher mode, the following six to the first higher mode and the last eight to the fundamental mode.

	Frequency (Hz)	Phase vel. (ms^{-1})	Error (ms^{-1})
Second higher mode	18.0	620	15
	20.0	563	12
	22.0	488	11
	24.0	455	11
	26.0	433	10
First higher mode	15.0	460	40
	16.5	412	20
	17.5	397	15
	20.0	363	12
	22.5	333	11
	25.0	320	11

Table 1. (Contd.)

	Frequency (Hz)	Phase vel. (ms^{-1})	Error (ms^{-1})
Fundamental mode	12.5	329	25
	13.5	300	15
	15.0	277	10
	16.5	257	9
	17.5	248	7
	20.0	232	5
	22.5	219	4
	25.0	213	4
	σ=14 ms^{-1}	ave. r.m.s. error	

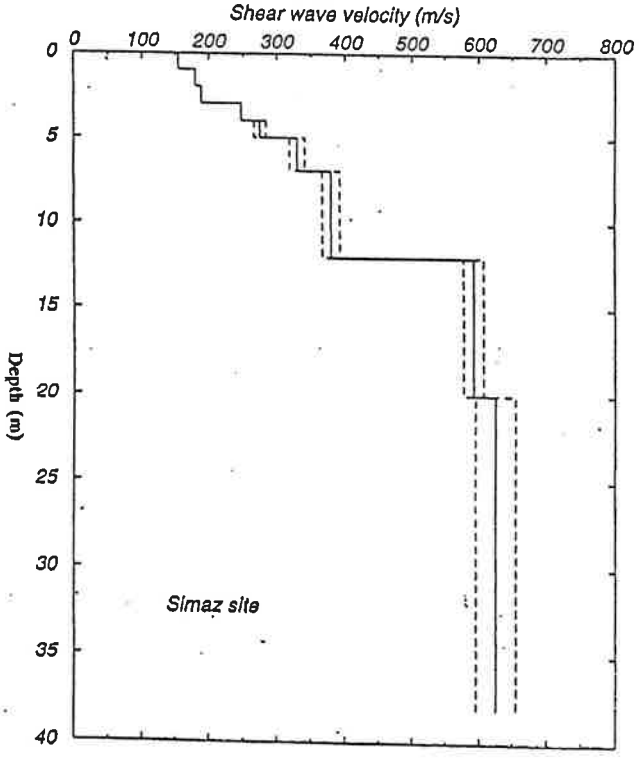

Fig. 5. Models for shear wave velocity. Dashed lines include the interval of possible solutions, solid line represents the model used in computations.

Very low (Fig. 6a) and very high (Fig. 6b) values of Q, respectively 5 and 50 for the shallow part of the model were used initially to generate the synthetic seismograms. Subsequently the distance between these extreme values had been reduced until a qualitative satisfactory match is reached between the synthetic and field data (Fig. 6c). On the basis of the similarity of signal envelopes, Q values ranging from 5, near to the surface, to 25 at the bottom of the model were finally selected (Table 2).

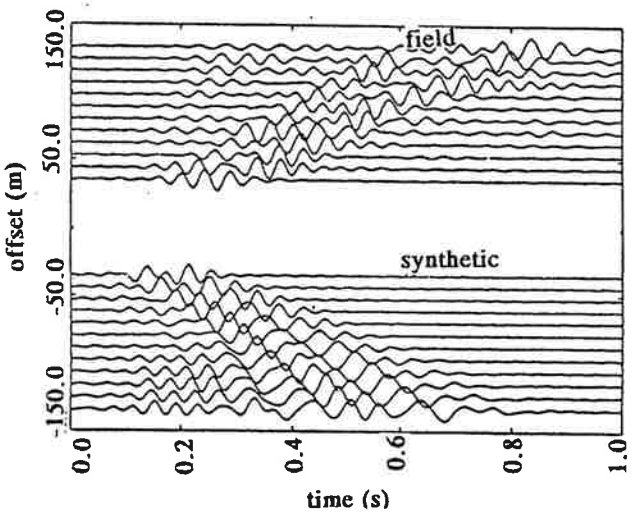

Fig. 6a

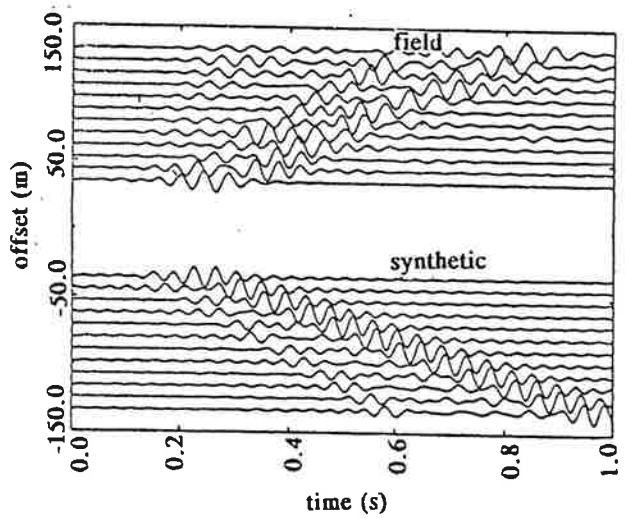

Fig. 6b

Fig. 6c

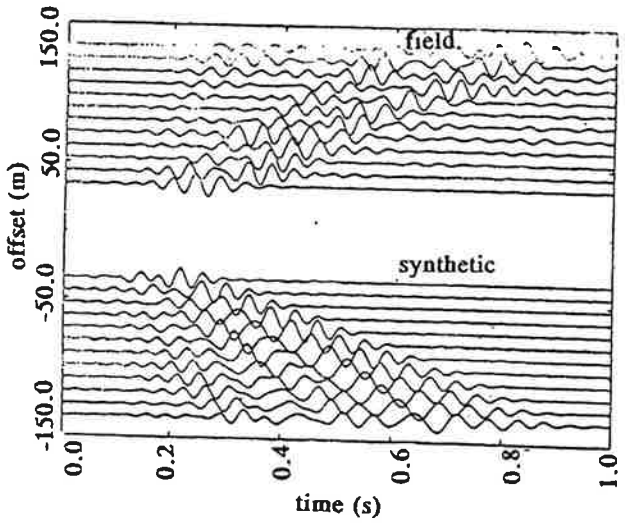

Fig. 6. Examples of synthetic and field data. Synthetic seismograms were computed using different Q values for near surface weathered structure: a) Q=5, b) Q=50, c) Q variable with depth between 5 at the free surface and 25 at 38 m of depth. Field are high cut-off filtered at the frequency of 30 Hz.

Table 2. Estimated Q value with depth.

Interval depth (m)	Q range
0–10	5–10
10–20	10–15
20–38	10–25
>38	>100

CONCLUSION

Shallow seismic velocity structure can be effectively studied using Rayleigh modes obtained from a refraction seismic survey. The result shows the efficiency of analysing dispersion of Rayleigh waves using bi-dimensional Fourier transform and frequency-time analysis.

The inversion result gave a velocity model characterised by Vp\Vs ratios (from 1.5 to 2.25) typical of dry or only partially saturated materials as fine gravel and coarse sand (Stumpel *et al.*, 1984). The satisfactory agreement between the experimental and the synthetic data permitted defining the Q values up to a depth of 38 m. The very low values obtained agree with those determined for the similar soil conditions by Jongmans (1990) using other techniques.

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