REMOTE SENSING IN SHALLOW-WATER MARINE SEDIMENTS

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1. INTRODUCTION

Underwater acousticians and civil engineers are continuously searching for practical and economical means of determining the physical parameters of the sea floor sediments for applications in environmental and geological research, engineering and underwater acoustics. Over the past two decades an increasing effort has been put into this field both theoretically and experimentally, to determine the geoacoustic properties of the seabed. Recent developments in experimental and forward-inverse modeling techniques indicate that the acoustic wave field in the water column and seismic wave field in the sea floor can be directly measured and by inverse and forward modeling techniques this information can be utilized for remote sensing of the geotechnical characteristics of the sea floor sediments. However in each specific case the experimental and analysis techniques must focus on a specific wave type.

The ability of sea floor sediments to support the propagation of seismic energy depends on the elastic properties of the sediments; namely the bulk modulus (K) and the shear modulus (μ). These and other elastic parameters are related to the compressional (α) and shear (β) velocities by the equations

$$K=\rho(\alpha^2-4\beta^2/3)$$
$$\mu=\rho\beta^2$$

where ρ is the bulk density. As can be seen, if compressional and shear wave velocities of sediments can be measured concurrently with a measurement of either bulk density or acoustic impedance ($\alpha\rho$), the elastic parameters of the sediments can be completely characterized.

The marine sediments also play an important role in problems related to the propagation of acoustic waves, especially in shallow water areas. In these applications, the geoacoustic properties of the sea floor defined by the compressional and shear wave velocities, their attenuation, together with the knowledge of the material bulk density, and their variation as a function of depth, are the main parameters needed to solve the acoustic wave equation. To be able to determine these properties of the seabed, remote sensing techniques have been developed using inversion of both acoustic and seismic data. Here, current methods to obtain compressional and shear wave velocities are briefly described (Ref. 1).

2. REMOTE SENSING TECHNIQUES

These techniques are based on the use of a signal received in a shallow water channel. Fig.1 illustrates a characteristic shallow water signal from an explosion received by a hydrophone close to the sea bottom. Here we briefly describe three different techniques to extract information relative to bottom parameters from these signals.

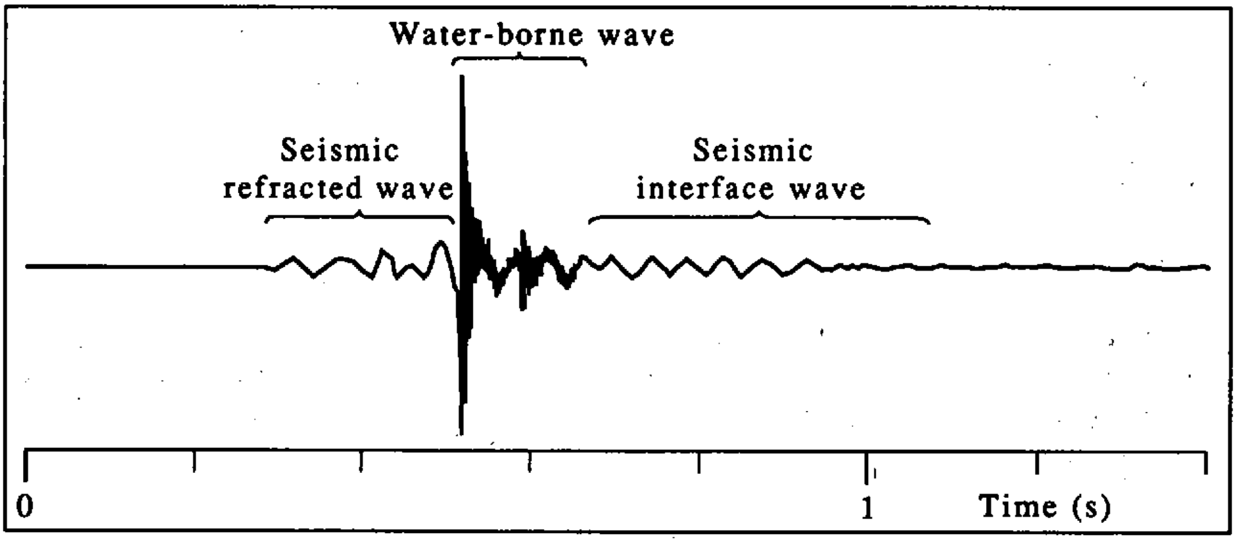

Fig. 1: A characteristic shallow-water signal.

Techniques developed for remote sensing of the uppermost sediments (25 to 50 m below the sea floor) utilize broad-band sources (small explosives) and an "L" shaped array of sensors with geophones deployed on the seafloor and hydrophones vertically within the water column. Recently we have implemented an additional procedure by attaching a small explosive source array (Fig. 2) to the geophone array. This approach improves the vertical resolution for refracted waves. To obtain estimates of the bottom properties as a function of depth, both acoustic (water-borne) and seismic (refracted compressional, shear and interface) waves are analyzed. For acoustic-data forward modelling (exact solution to the wave equation) and for seismic data inverse modelling based on modified versions of the inversion techniques developed by earthquake seismologists and geophysicists to study dispersed Rayleigh-waves and refracted waves have been carried out. Fig. 2 illustrates the basic experimental setup where generally a two-ship configuration is used for acoustic measurements and single-ship configuration for seismic measurements. Fig. 3 shows the range stack of signals received by an array of hydrophones and geophones, permitting studies of both acoustic and seismic waves. These data are analyzed and via forward and inverse modelling techniques compressional and shear wave velocities are obtained as a function of depth in the seafloor. Moreover, studies of attenuation and lateral variability are also possible using the same data set (Ref. 2).

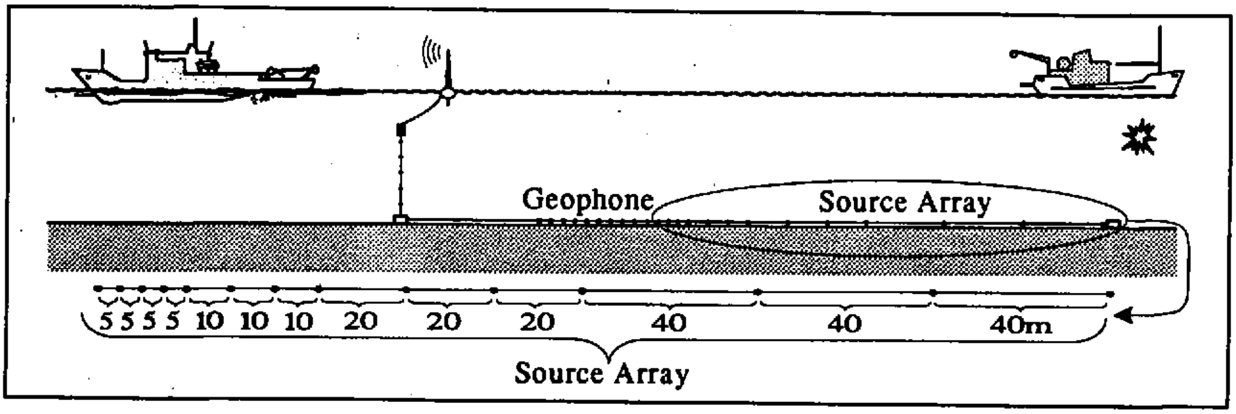

Fig. 2: Experimental set-up.

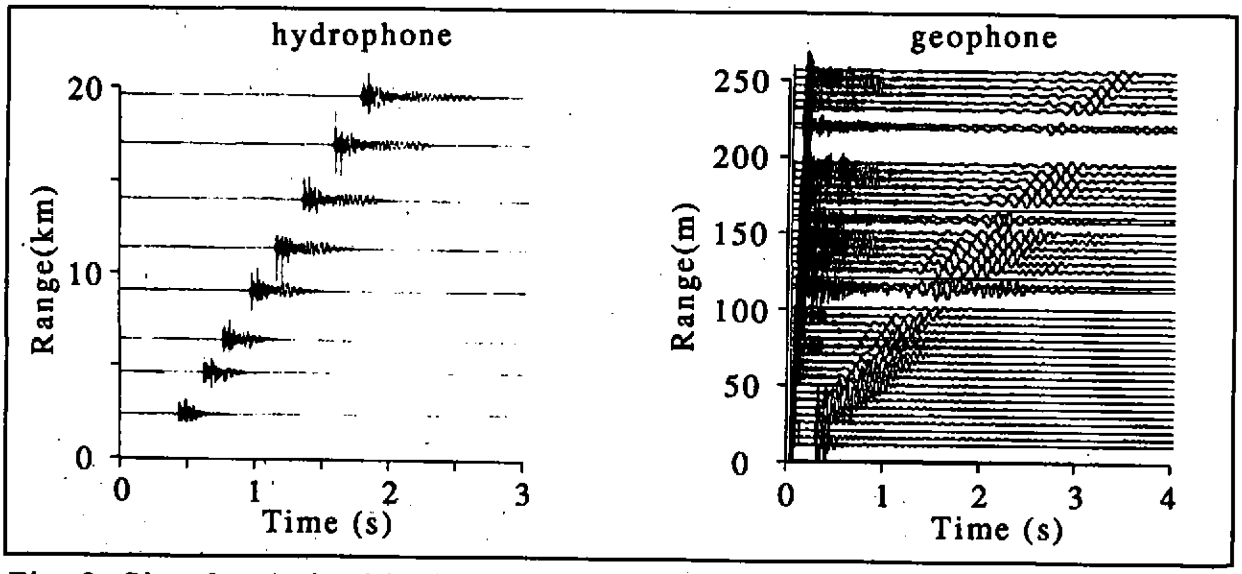

Fig. 3: Signals received by hydrophone and geophone arrays.

2.1. Transmission Loss Techniques

The seafloor is known to be the controlling factor in low-frequency shallow water acoustic propagation (Refs. 3,4). Forward modelling is performed with models giving exact solutions to the wave equation, i.e. SAFARI (Ref. 5) where, p- and s-wave velocities, the attenuation factors associated with these waves, and the sediment density are the main input parameters. Acoustic energy propagating through a shallow water channel interacts with the sea floor causing partitioning of waterborne energy into different types of seismic and acoustic waves. The propagation and attenuation of these waves observed in such an environment are strongly dependent on the physical characteristics of the sea bottom. *Transmission Loss* (TL), representing the amount of energy lost along an acoustic propagation path, carries the information relative to the environment through which the wave is propagating. Fig. 4 shows a comparison of TL data and model predictions together with the input parameters used at 400 Hz. The effects of changes in bottom parameters are also shown in the figure. This technique becomes extremely useful when seafloor information is sparse.

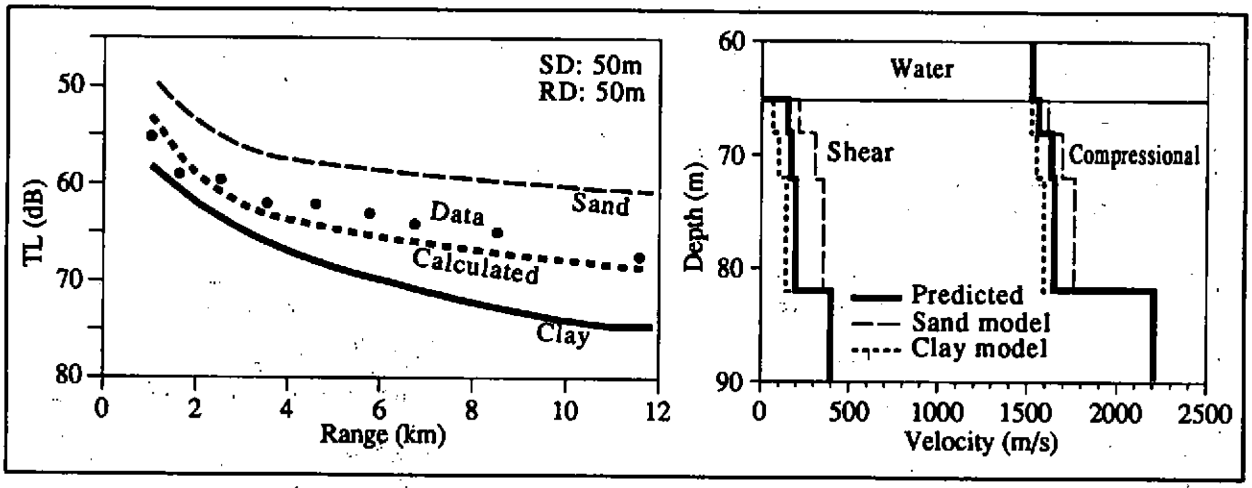

Fig. 4: Comparison of TL data and SAFARI prediction. Effects of changing bottom parameters from sand to clay is also shown together with the input profiles utilized for SAFARI predictions.

2.2. Refraction Techniques

Fig. 5 shows the expanded early portion of the geophone data shown in Fig. 3, where the first arriving energy out to a range of about 250 m has been fitted with a curve showing that p-waves refracted through the sediments just beneath the seafloor travel faster as they penetrate more deeply into the sediment. A p-wave velocity-depth curve for the upper part of the seafloor can be derived from the first arrivals (Fig. 5) using the classical Herglotz-Bateman-Wiechert integration (Ref. 6). In this method the slope of the travel-time curve (fitted parabola) gives the rate of change of the range with respect to time which is also the velocity of propagation of the diving p-wave at the level of its deepest penetration into the sediment. At each range Δ, the depth corresponding to the deepest penetration is then calculated using the following integral

$$z(V) = 1/\pi \int_0^{\Delta} \cosh^{-1}\left(V(dt/dx) \right) dx,$$

where

$$1/V = (dt/dx)_{x=\Delta}$$

The result is the solid velocity-depth curve shown in Fig. 5.

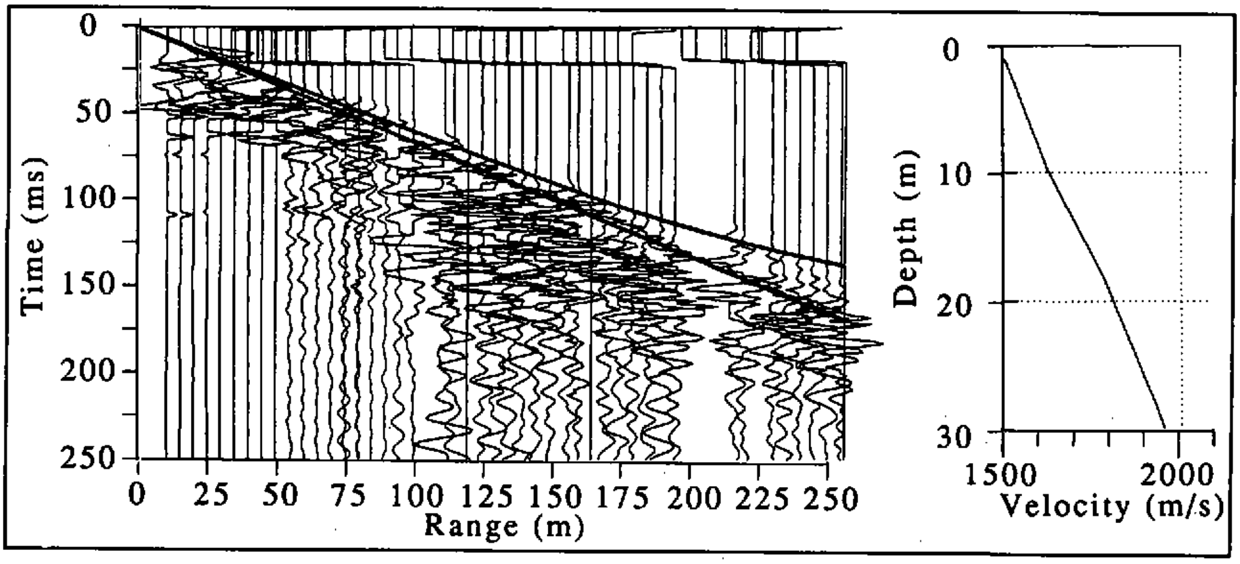

Fig. 5: Expanded early portion of the data shown in Fig.3, and velocity-depth curve predicted by the model.

2.3. Interface Wave Techniques

In order to obtain a shear wave velocity-depth profile from the data, later arrivals corresponding to interface waves traveling in a dispersed fashion may be utilized. Figure 6 shows the expanded late arriving portion (interface wave) of the geophone data shown in Fig. 3. The portion of each individual signal corresponding to the interface-wave arrival can be processed using multiple filter analysis (Ref. 7) to create a group velocity dispersion diagram (Gabor diagram). The result of applying this technique to a dispersed signal is a

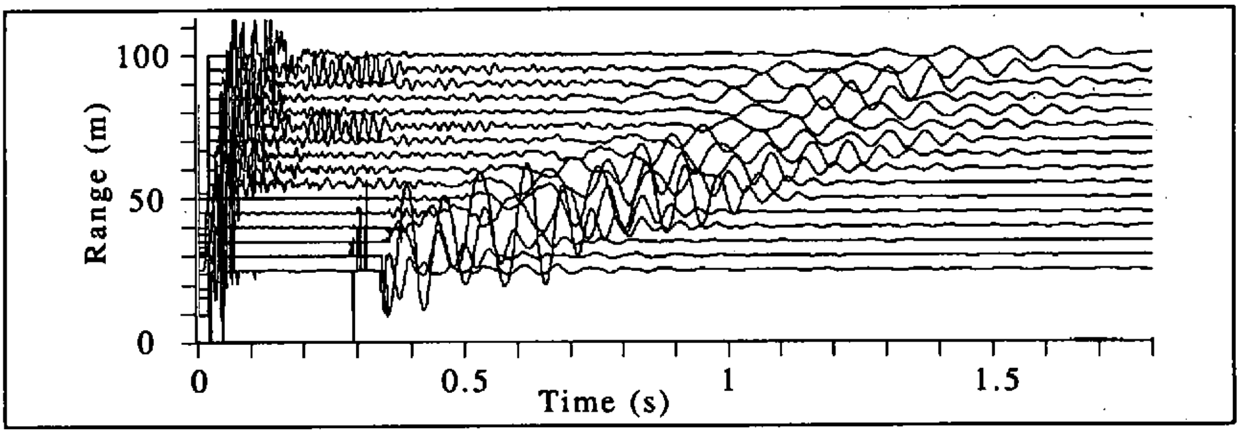

Fig. 6: Expanded Interface Wave portion of the data shown in Fig. 3.

filtered time signal whose envelope reaches a maximum at the group velocity arrival time for a selected frequency. The envelope is computed by taking the quadrature components of the inverse Fourier transform of the filtered signal. Filtering is carried out at many descrete frequencies over a selected frequency

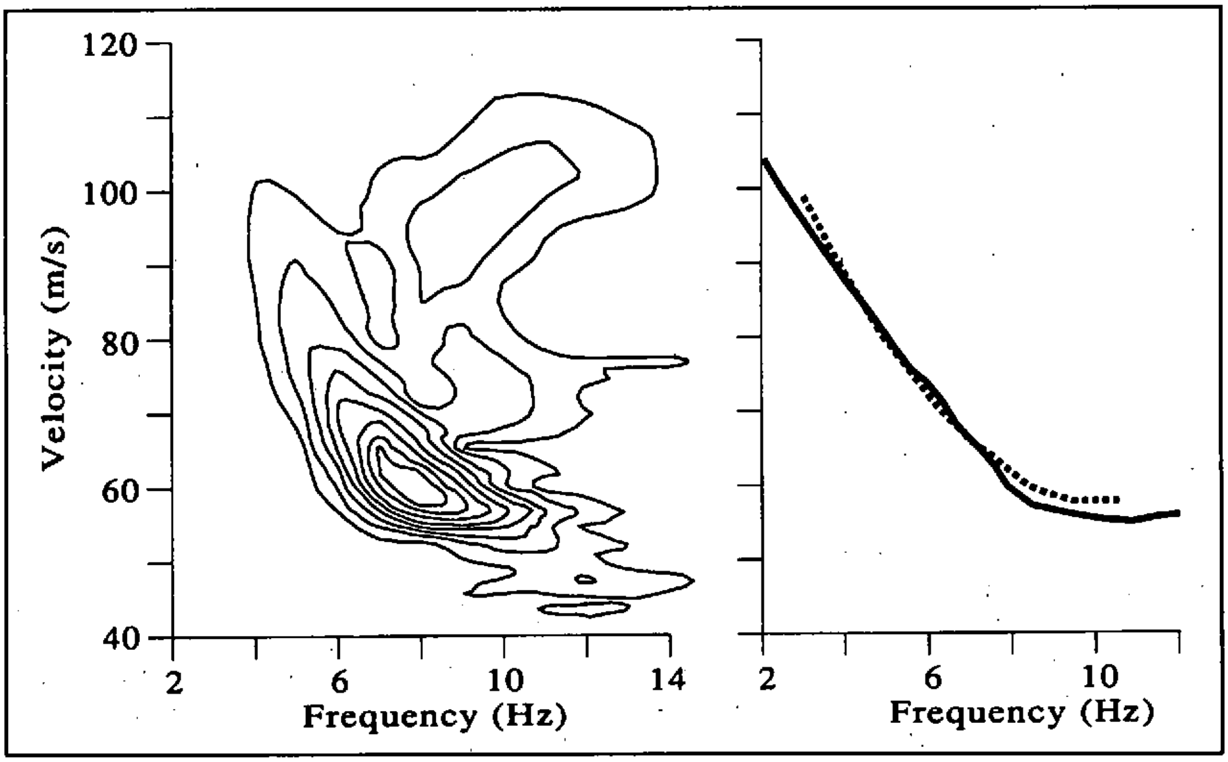

Fig. 7: Dispersion diagram, measured and predicted dispersion curves.

band. Once the arrival times are converted to velocity, the envelopes are arranged in a matrix and contoured and dispersion curves are obtained by connecting the maximum values of the contour diagram (Fig. 7).

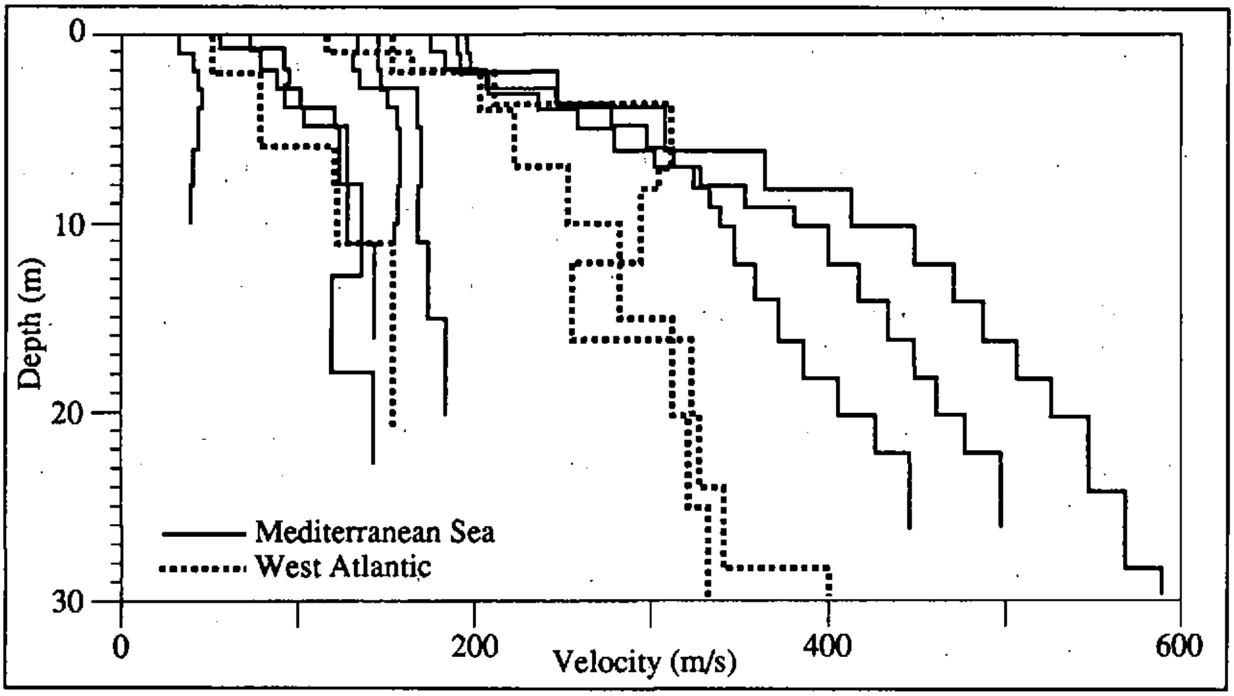

Fig. 8: Summary of shear velocity profiles from the Mediterranean and west Atlantic continental shelves.

Having obtained the dispersion characteristics of the interface waves, we determine the geoacoustic model, as a stack of homogeneous layers with different s- and p-wave velocities for each layer, that allows us to theoretically match the measured dispersion curve (Refs. 8,9). Fig. 8 illustrates examples from the Mediterranean and West Atlantic continental shelf, covering data from soft clays to hard sands.

3. CONCLUSIONS

Experimental and theoretical work has shown that it is possible to determine the geoacoustic properties of sediments by remote sensing techniques. Different techniques briefly described in this paper indicate that seismoacoustic waves observed over specific propagation paths may contain sufficient information to deduce the fundamental characteristics of the sediments. It is evident that still more research needs to be done to develop these techniques for fast and reliable results.

4. REFERENCES

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