6th International IAEG Congress / 6ème Congrès International de AIGI, © 1990 Balkema, Rotterdam. ISBN 90 6191 130 3

In situ rapid assessment of dynamic soil characteristics for microzonation investigations

Estimation rapide in situ des caractéristiques dynamiques du sol, appliquée dans les investigations microzoniques

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ABSTRACT: In this research we present a geophysical technique for the in-situ assessment of the dynamic properties of soil sites, which can be used very efficiently during microzonation investigations. By using a digital multichannel seismograph and a portable microcomputer we analyze in addition to the first seismic arrivals and the later arriving surface waves (ground role). This analysis consists in evaluating the corresponding surface wave dispersion curves which they are automatically inverted to obtain the shear wave velocity and the modulus of each sublayer. Data collection and analysis using the above technique are discussed in detail and some examples are presented.

RESUME: Dans cette recherche nous presentons une technique geophysique pour l'estimation in-situ des proprietes de sol, utilisees tres efficacement pendant les investigations microzoniques. Cette technique consiste a calculer les courbes correspondantes aux ondes superficiellent dispersees, qui sont automatiquement inversees, pour obtenir la velocite des ondes de cisaillement et le module de chaque couche du sous-sol. La collection des donnees et l'analyse en utilisant la technique mentionnee ci-dessus, sont discussees en detail et plusieurs cas etudiees sont presentees.

1 INTRODUCTION

The knowledge of the dynamic elastic properties of soil consist an essential prerequisite to the design of major structures and are very important in geotechnical engineering. They are useful in designing facilities such as vibration machine foundations and they are the key parameters during the development of microzonation studies and the assessment of the liquefaction potential during earthquake shaking.

Despite the fact that the analytical methods used in earthquake engineering have improved substantially during the last two decades, the capability of determining soil properties in-situ has not followed this pattern. It is obvious that the use of very sophisticated constitutive models incorporated into large finite element programs many times creates doubts about the final results when accurate soil properties are not available.

Since shear wave velocities are directly related to the stiffness of the material through which the wave propagate it is possible to derive material properties such as shear modulus or density in situ from measured wave velocities. This consists the principle behind the use of seismic techniques to assess in-situ soil parameters and there are many investigation techniques such as the surface refraction method, the crossohole method the downhole method etch.

Despite the diversity of applications of seismic velocity studies of soils, problems do arise when dependence is solely placed on velocity, (Tselentis et al 1988), and there is a need to extract as much information as possible from the seismic wave forms encountered. Furthermore, the assessment of shear wave velocity profiles with depth is a difficult task if one employs the surface refraction method and becomes very expensive if we have to use boreholes as it is required for the crosshole and downhole techniques.

An alternative will be to study the dispersion characteristics of surface waves as it was proposed by Nazarian (1984). In layered media the velocity of propagation of surface waves depends on the wavelength of the wave and it is obvious that different wavelengths sample different sections of the layered media.

The principle of the present technique is to generate surface wave signals and invert the corresponding dispersion curves in order to asses the shear wave velocity and modulus profile of the soil.

2 PRINCIPLE OF THE METHOD

Consider the simplified case consisting of only two geophones and a shot point (Fig.1).

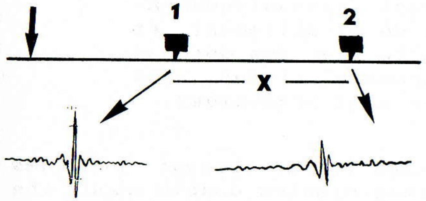

FIG.1:Principle of the method

Dispersion of surface waves between the two sites [1] and [2] can be generally represented by the process of deconvolution of the signal observed at the second geophone with the signal observed at the first geophone. In the frequency domain this process can be written as

$$F21(\omega)=F2(\omega)/F1(\omega)= \tag{1}$$
$$= F2(\omega)F1^{*}(\omega)/|F1(\omega)|^{2}$$

The spectrum $F21(\omega)$ contains both the interstation phase delay and attenuation information. The corresponding time signal is

$$f_{21}(t)=\frac{1}{2\pi}\int_{-\infty}^{\infty}[F_2(\omega)\cdot F_1{}^{*}(\omega)/|F_1(\omega)|^2]\cdot\exp(i\omega t)\,d\omega$$
$$=\frac{1}{2\pi}\int_{-\infty}^{\infty}(A_2(\omega)/A_1(\omega))\cdot\exp[i(\omega t-\phi_2+\phi_1)]\,d\omega$$
$$=\frac{1}{2\pi}\int_{-\infty}^{\infty}\bar{A}(\omega)\cdot e^{-i\phi(\omega)}\cdot e^{i\omega t}\,d\omega$$

where $\bar{A}(\omega)=A2(\omega)/A1(\omega)$ and $\phi(\omega)=\phi2(\omega)-\phi1(\omega)$. The function f21(t) is equivalent to a signal that would be observed at the second station for a zero phase shift, δ-impulse source situated at the first station (Dziewonski and Hales 1972).

We may note from the above equation that the entire phase information of the signal f21(t) is contained in the numerator

$$F2(\omega)F1^{*}(\omega)$$

which represents the crosspower spectrum and corresponds to the cross correlation between the two signals. Thus to evaluate the required relative phase difference between the two geophones, one has to determine the imaginary part of the product of the linear spectrum at geophone [2] (output) with the complex conjugate of the linear spectrum at geophone [1] (input).

This process is very unstable in the presence of noise. A good tool for assessing the quality of the observed signals and detect the presence of any noise, is to evaluate the coherence function $\gamma(f)$ which is defined by

$$\gamma^2(f)=\frac{F21(f).F21^{*}(f)}{F11(f).F22(f)} \tag{2}$$

It is obvious from [2] that the coherence function will be equal to unity for all frequences, except for the case that some noise is present.

From each measurement we select the range o frequencies which correspond to a coherence function close to one. In the selected frequency range, the phase of the cross power spectra can be used to calculate phase velocities and associated wavelengths for each frequency. If the distance between the two receivers is x, the phase velocity for a given frequency f can be evaluated from

$$Vph=x/t \tag{3}$$

where

$$t=\Phi/(360.f) \tag{4}$$

is the travel time and Φ is the corresponding phase difference. The wavelength can be determined directly from the following equation

$$Lph=Vph/f \tag{5}$$

The plot of the wavelength versus the phase velocity is the required dispersion curve.

After constructing the dispersion curve, the next step is to invert it and evaluate the shear wave velocity and modulus profile. This is achieved by constructing dispersion curves for specified underground properties and compare them with the experimental dispersion curves through a semiautomatic process. Techniques for evaluating dispersive curves for a given layering are well known in the literature (Harkrider 1964). In the present work we solved the Rayleigh wave period equations in order to find the dispersion curves for phase velocity. The classical Haskell Thomson approach was reformulated using Dunkin's (1965) methodology in order to eliminate numerical instabilities at higher frequencies and the phase velocities were determined to an accuracy better of 0.00000001 by using the algorithm proposed by Herrman (1971).

3 FIELD PROCEDURE

The general configuration of a typical field layout is shown in Fig.2. The seismic line consists of 12 or 24 vertical component geophones with natural frequency of 1 Hz spaced at 1m-2m. The signal generator is at a distance from the first geophone which is equal to the geophone spacing. From each (24 geophone), layout we can select 12 pairs of geophones in such a way that the distance between these geophones is equal to the distance between the shot point and the first geophone of the pair.

Furthermore, we obtain measurements by reversing the location of the source. We carry also a few measurements with the shot point set at 10m from the first geophone, selecting the corresponding geophone pairs with same methodology.

At close shot distances we use small hand held hummers and as the shot is set at further distances we employ a large sledge-hammer.

The recording device is a 24-channel ABEM Terraloc signal enhancement seismograph with digital storage facilities. The whole sequence of measurements and data collection is automatically controlled by a portable microcomputer connected to the seismograph. The data processing scheme can be done either directly on the field or later in the office according to the scheme presented at Fig.3.

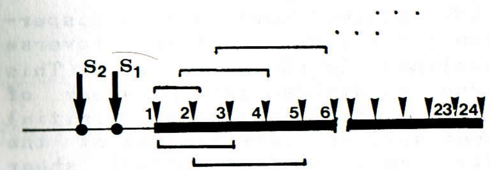

Fig.2:Geophone layout

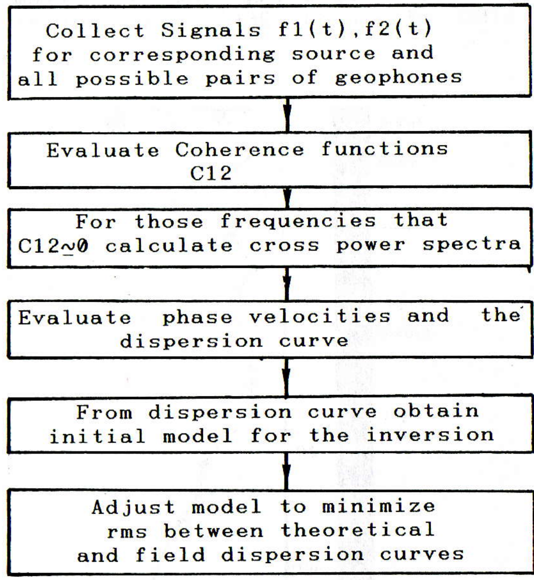

Fig.3:Data processing scheme

4. APPLICATION

A borehole was drilled down to a depth of 50m. The geological profile based upon the borehole data and the N-values have been shown in Fig.4. From this profile it is obvious that the geological formations are alternatively clay and fine sands in varying thicknesses. the sand is occasionally mixed with gravels.

In order to test the reliability of the present method, the underground distribution of P- and S- waves was obtained by refraction and well shooting techniques respectively. For well shooting observations, a 25Hz three components geophone was used. SH waves were generated by hitting, with a hammer, the end of a slender weighted wooden plate laid down on the ground surface (Fig.5).

The computed velocity distribution of P- and S- waves is shown in Fig.6. Judging from this figure we see that the S- wave velocity profile is more comparable to the type of geological formations than the P- wave velocity profile.

In Fig.7 we show the obtained phase of the cross power spectrum (unwinded by using the algorithm WIND of Sterns 1988). The corre-

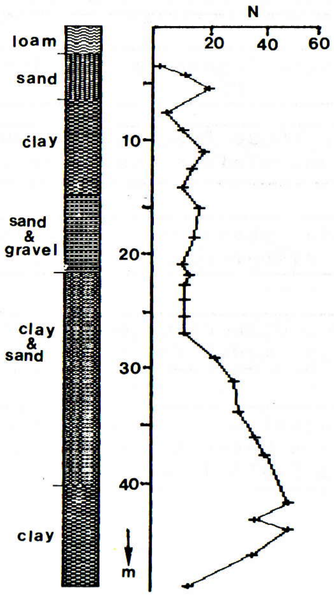

Fig.4:Geological and N profile

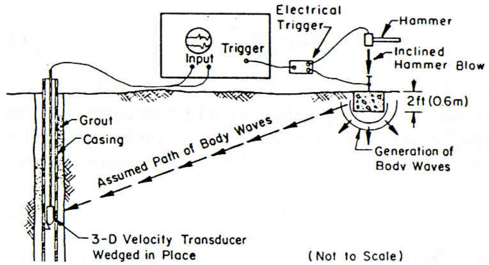

Fig.5: Well shooting technique used

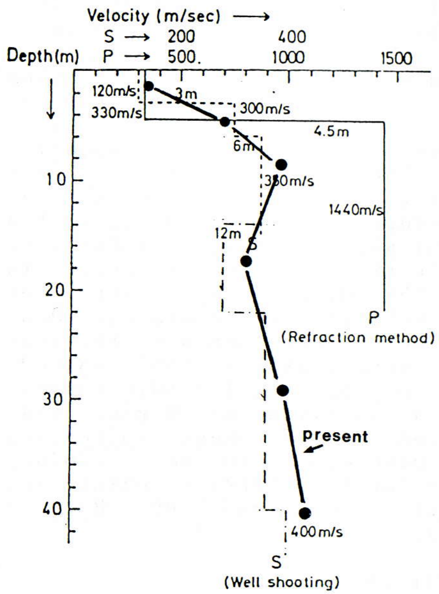

Fig.6:Velocity profile

sponding coherence function which determines which frequencies were used for the analysis is shown in Fig.8.

The obtained surface wave dispersion curve (for direct and reverse shooting) is shown in Fig.9. This curve is divided into a number of layers for estimating the initial shear wave velocity profile at the site. Entering this initial shear wave velocity profile at the numerical algorithm which determines the dispersion curve for a specified earth model it was possible to improve the obtained initial model by minimizing the error between the theoretical and field dispersion curve. The Final result is depicted

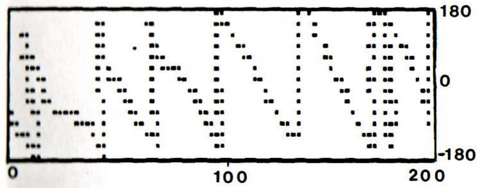

Fig.7:Phase of crosspower spectrum

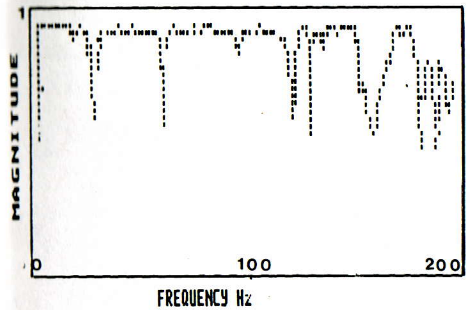

Fig.8:Magnitude of coherence function

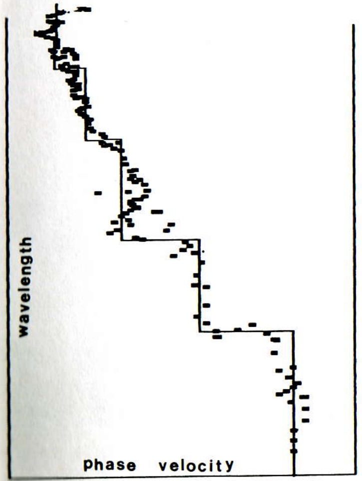

Fig.9:Dispersion curve

in Fig.6 and is compared with the obtained in the field shear wave velocity profile.

The modeling process permits to assess the density of each layer, thus by employing the relation

$$G=rVs \qquad [6]$$

we can estimate the shear modulus.

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