INVESTIGATIONS ON SITE EFFECTS IN ITALY

A. Rovelli [1], L. Malagnini [2] and M. Di Bona[2]

ABSTRACT

Amplitude variations among strong ground motions recorded at close sites and anomalies of the macroseismic fields were observed in Italy on the occasion of recent earthquakes. We present results for three sites where amplifications of ground motion larger than a factor of ten were experienced, or a difference of one intensity degree compared with nearby sites was observed. Moreover, some experiments were carried out on sediment bodies in order to investigate site response, based on measurements of weak motions produced by teleseisms, local earthquakes and explosions. For all these study cases, 1- and 2-D numerical techniques and microtremor Fourier spectra were used to compare theoretical and empirical estimations. Generally, numerical modelling allowed a good prediction of the frequency band where amplifications occur, but a satisfactory fit of amplifications observed during earthquakes was obtained in some cases only. In the cases where strong motion records were available at nearby "firm" and "soft" sites from the same earthquake, the spectral ratios from microtremor measurements in conditions of minimum cultural noise showed a trend similar to the corresponding acceleration spectral ratios computed from the earthquake recordings. But the possibility that local sources of cultural noise do affect the different site spectra cannot be excluded a priori, and this represents a significant limitation for a widespread application of this method to the prediction of amplifications during earthquakes.

At this moment, the state-of-art suggests that weak motions (e.g. small local earthquakes or regional events) are the most suitable means for predictions of strong amplifications. Column shots, when exploded with proper techniques generating shear waves, give results consistent with earthquake data.

INTRODUCTION

After earthquakes, different levels of damage are often observed within small distances. Such phenomena can be very local, and are currently named site effects. The cause is due to the small-scale heterogeneities of the subsurface geology or irregularities of morphology. Starting from the Michoacan earthquake of September 19, 1985, seismologists and engineers became more and more aware of the predominant role that these effects can play in controlling the behaviour of ground motion and the distribution of damages during earthquakes. The need emerged for developing methodologies able to predict where and to what extent site properties can produce

[1] Research Director, Instituto Nazionale di Geofisica, Via di Villa Ricotti 42,00161 Rome, Italia

[2] Research, Instituto Nazionale di Geofisica, Via di Villa Ricotti 42,00161 Rome, Italia

amplifications of seismic waves. In Italy, a systematic study of the most significant cases of large amplifications recorded by strong-motion accelerograms was developed. These investigations are important because the collected data offer a unique opportunity to check the different methodologies proposed so far for the prediction of site response. In this framework, the Istituto Nazionale di Geofisica (ING) activated some research projects devoted to test theoretical and empirical methodologies for the prediction of local seismic response. This paper describes the activities carried out so far.

TEST CASES

In order to study source properties of recent Italian earthquakes, Rovelli et al. (1988) and Cocco and Rovelli (1989) analyzed strong-motion accelerograms of the most active areas in Italy (north-eastern Alps, and central and central-southern Apennines). They found that for many of the accelerograms, ground motion spectra scale quite regularly and are satisfactorily fit by a theoretical curve including source scaling and attenuation laws. On the contrary, a few sites showed significant deviations in amplitude in particular frequency bands. Such a kind of amplification had a repetitive character, independently of azimuth, magnitude and focal distance. The cases where the spectral amplitudes were significantly higher than the theoretical values were interpreted as site effects. We selected three sites as those showing the largest amplification during earthquakes. In all these sites, amplification of ground motion occurs in concomitance with a local magnification of the macroseismic intensity. We tried a numerical approach to model the seismic response of these sites. For two of them, subsurface geology was sufficiently known based on well log data. Peter Moczo of the Geophysical Institut of the Slovak Academy of Science GISAS performed a finite-difference scheme with irregular grid spacing (Moczo, 1989). In the framework of a cooperation between ING and GISAS, 1- and 2-D finite-difference codes were implemented at ING in order to model observed amplifications. Moreover, also the microtremor method (see Kanai and Tanaka, 1961) was tested to check the feasibility of using ambient noise to predict site response. In the old lake-bed zone of Mexico City microtremors resulted to yield the same dominant frequencies as strong motion recordings (Lermo et al., 1988). Together with Shri K. Singh of the Universidad Nacional Autonoma de Mexico UNAM we measured microtremor in the "soft" sites under study comparing their spectra with nearby "firm" reference stations (Rovelli et al., 1991). The goal was to seek whether the soft-to-hard site microtremor spectral ratios agree with the corresponding strong-motion spectral ratios.

For the selected cases of study, the results obtained from numerical modelling experiments and microtremor measurements are here described.

(1) The Forgaria Site (Friuli).

The Friuli mainshock of May 5, 1976 ($M_L = 6.5$) produced very different levels of damage at two sites, namely San Rocco and Forgaria, which are 650 m apart (Figure 1a) and about 20 km from the epicenter. The former, located on Mesozoic limestone, suffered light damage while the latter, located on an alluvial deposit roughly 20 m thick, was severely damaged. In order to investigate the cause of this difference in damage, ENEA-ENEL Joint Commission installed accelerographs at these sites a few days after the mainshock.

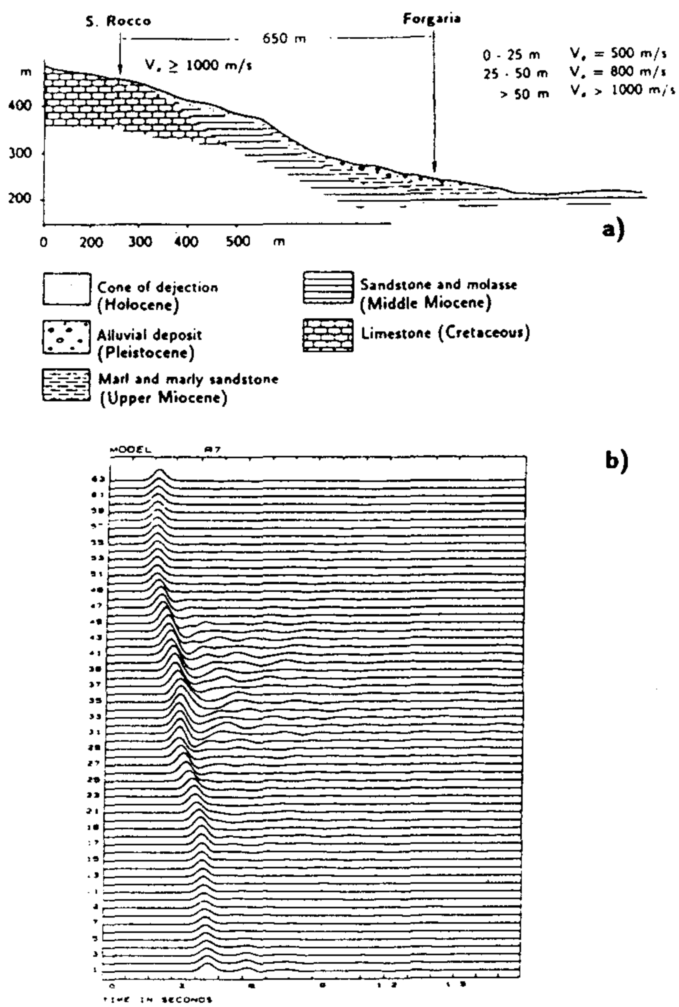

FIGURE 1. (a) Geological section crossing the sites of San Rocco and Forgaria. (b) Synthetic seismograms obtained along the profile using a finite-difference code: n° 13 is relative to San Rocco, and 37 to Forgaria.

Recordings of several aftershocks up to a magnitude of 6 showed that the difference in the level of damage at the two sites was related to the difference in the ground motion characteristics (Muzzi and Pugliese, 1977). The acceleration time histories recorded at Forgaria had significantly higher amplitudes: the spectral ratio is shown in Figure 2a. The shaded area represents $\pm$ one standard deviation around the mean value obtained by averaging spectra from 5 available aftershocks. We have used the finite-difference computer code written by Moczo (1989) to model seismic response along the profile shown in Figure 1a. We have simulated a vertically incident SH-wave (a displacement Gabor-function pulse, chosen to give a flat spectrum between 0.1 and 10 Hz). Synthetic seismograms generated along the profile are shown in Figure 1b (details of modelling are described in a paper by Moczo et al. , in preparation). Theoretical transfer functions of the San Rocco and Forgaria sites, and the theoretical spectral ratio were then computed from synthetics. In Figure 2a, the Forgaria/SanRocco spectral ratio obtained from the accelerograms of 5 Friuli aftershocks is compared with the spectral ratio obtained from the synthetic seismograms simulating the response of the Forgaria and San Rocco sites.

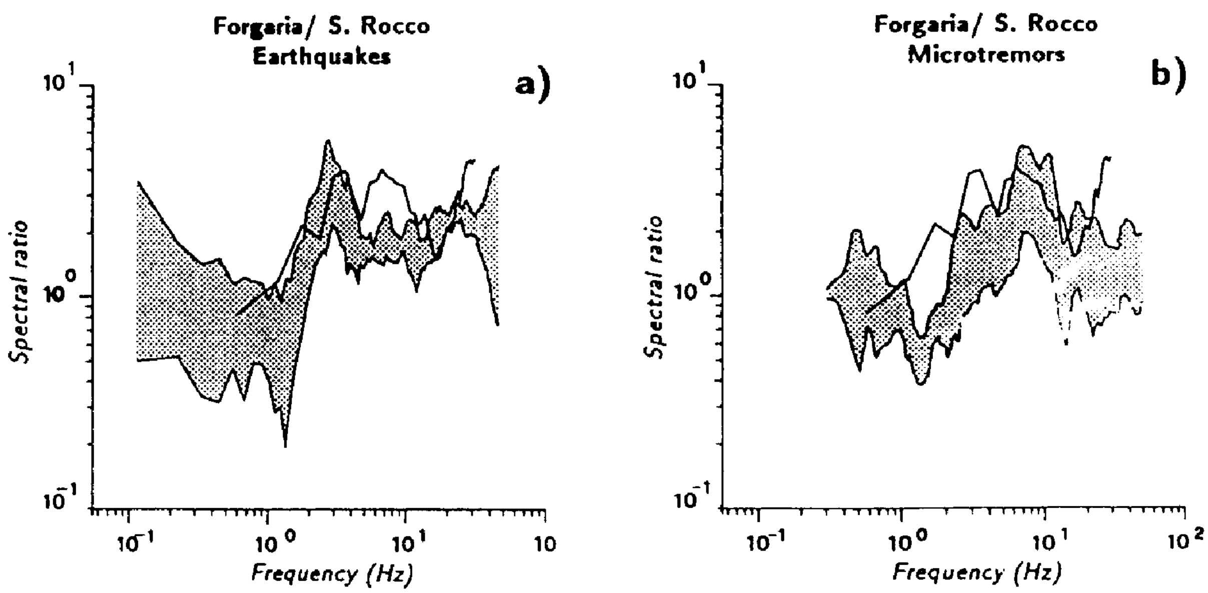

FIGURE 2. (a) Acceleration and (b) microtremor spectral ratio between the Forgaria and San Rocco sites. The shaded areas represent $\pm$ standard deviation around the mean value; the spectral ratio between synthetic seismograms derived from 2-D numerical modeling is overimposed on both acceleration and microtremor spectral ratios.

The agreement is particularly satisfactory: consistingly, the displacement time histories recorded at the Forgaria site are well fit by a synthetic time history computed from the San Rocco displacement deconvolved to give the input at the basement, and convolved then for the Forgaria transfer function (see Figure 3 for an example). The quality of the fit confirms the reliability of numerical techniques when details of local subsurface structures are sufficiently well known. Also microtremors were recorded at the sites of Forgaria and San Rocco. Figure 2b shows the microtremor spectral ratios at Forgaria with respect to San Rocco: the shaded area in Figure 2b represents ± 1 standard deviation around the mean value obtained by averaging spectra from 12 microtremor windows 20 sec large.

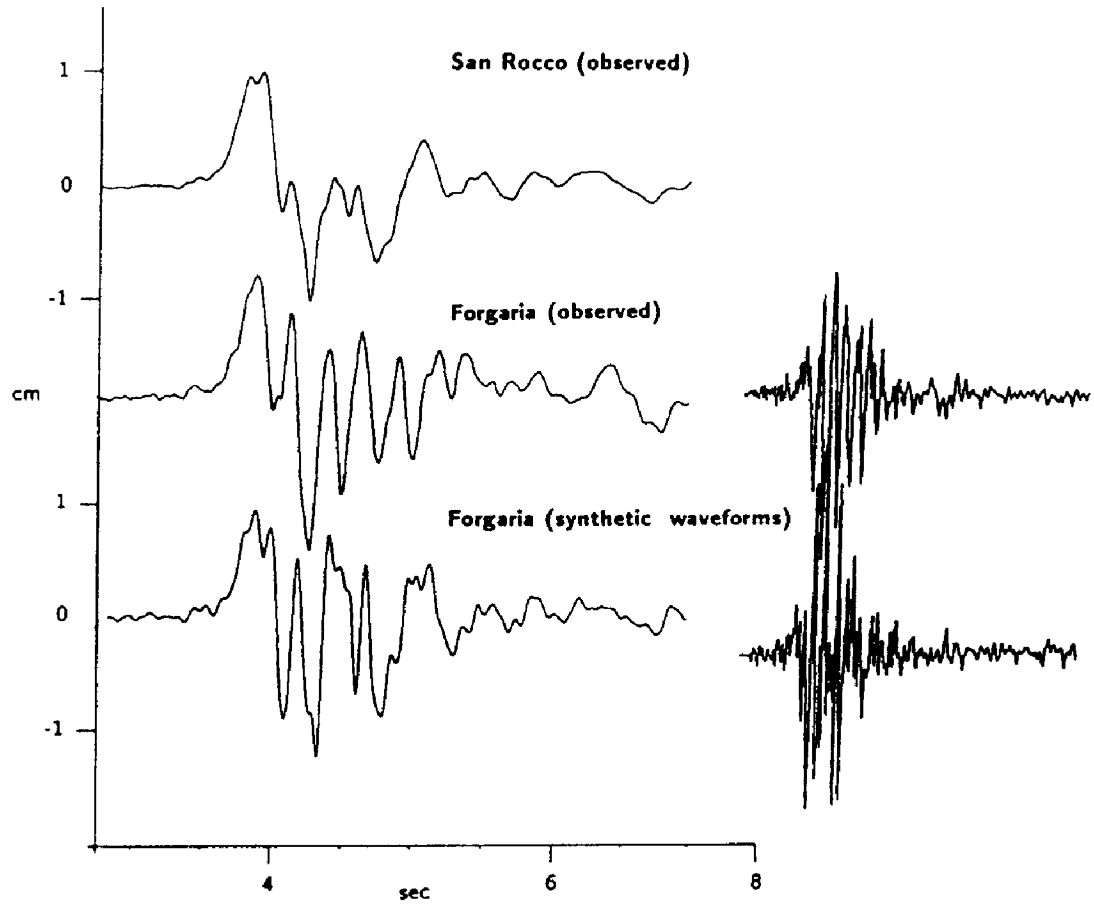

FIGURE 3. Displacement time history observed at San Rocco (at the top) and Forgaria (in the middle). At the bottom a synthetic ground displacement is shown. These synthetic waveforms were obtained from the San Rocco displacement after deconvolution for the San Rocco transfer function and convolution for the Forgaria transfer function. Observed and synthetized ground accelerations at Forgaria are also shown.

We note that the acceleration spectral ratio has a peak at about 2.5 Hz whereas the peak in the microtremor spectral ratio occurs between 6 to 8 Hz (Figure 2, a and b). Although the two spectral ratio curves differ in other details as well, the general trend of the site amplification at Forgaria with respect to San Rocco could be inferred from the microtremor measurements.

However, the level of ground motion observed for this site seems to be better explained by numerical modelling than by the microtremor spectral ratio.

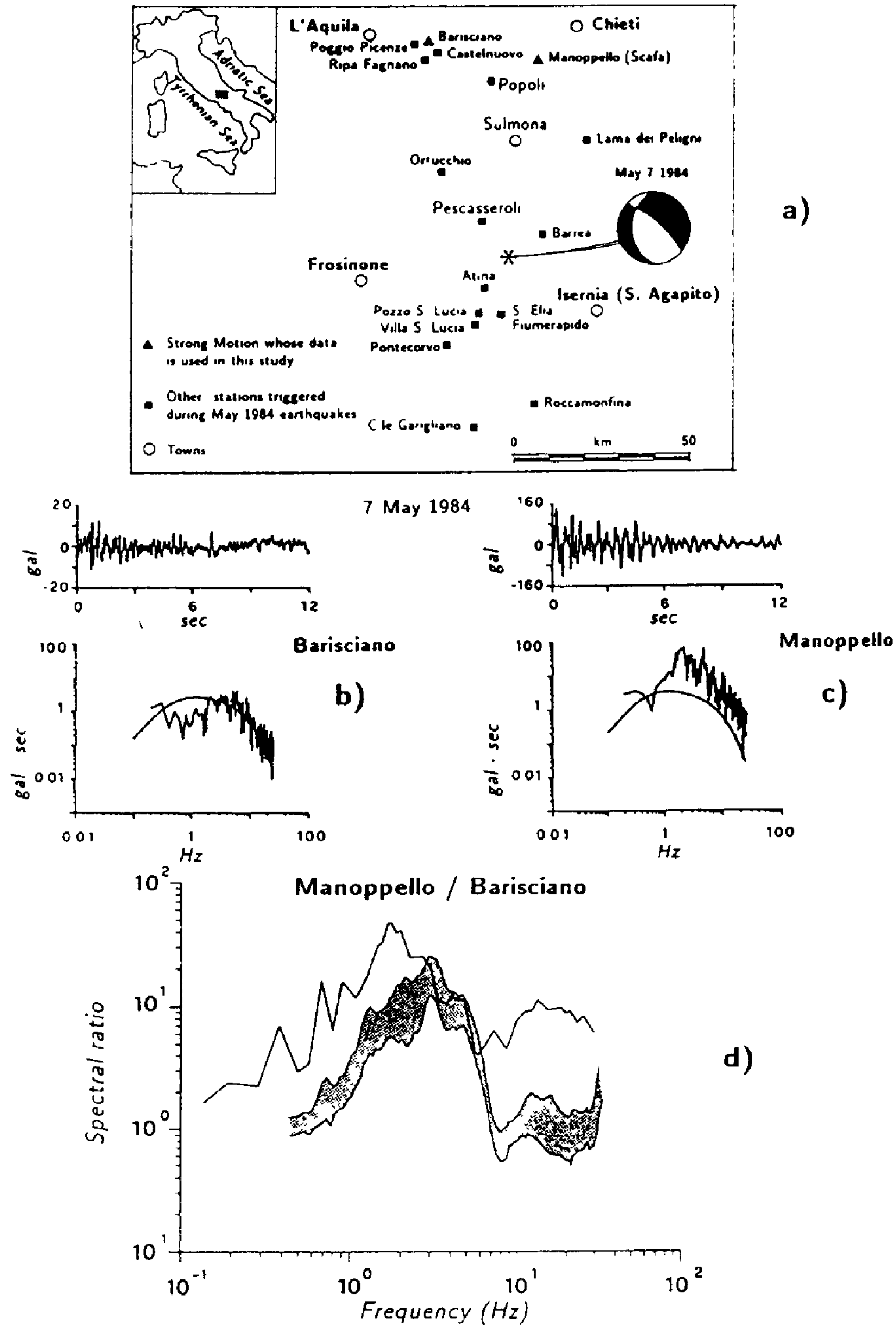

FIGURE 4. (a) Location map of Manoppello and Barisciano sites and the epicenter of the May 7, 1984, earthquake. (b) and (c) are the *NS* components of accelerogram at the two sites, and their relative spectra. (d) Acceleration and microtremor spectral ratios between Manoppello and Barisciano. The shaded area represents $\pm$ 1 standard deviation around the mean value of microtremor spectral ratio, continuous line is acceleration spectral ratio computed from the Manoppello and Barisciano accelerograms.

(2) Manoppello Site (Central Italy).

The area around the town of Manoppello systematically suffered higher macroseismic intensities compared with nearby towns (Carrara et al. , 1987). Rovelli et al. , 1988, (see Figure 13a to 13c of that paper) found that for all the events recorded at Manoppello, the observed Fourier acceleration spectra were roughly 10 times higher than the predicted spectra in the frequency band from 0.3 to 20 Hz. One of these events (May 7, 1984, M_L = 5.4) was also recorded at Barisciano, a hard rock site about 32 km from Manoppello (Figure 4a). The epicenter of the earthquake was 68 and 60 km from Manoppello and Barisciano, respectively (Figure 4a). Although the azimuths to the two sites from the epicenter differ by about 30 °, the S-wave radiaton pattern values are roughly the same and, therefore, the acceleration spectral ratio, in the absence of site response difference, is expected to be close to one. Figures 4b and 4c show acceleration time histories and spectra recorded at these two sites. The acceleration and microtremor spectral ratios are given in Figure 4d. The observed relative amplification from accelerograms occurs over a wider frequency band than that obtained from microtremors. The peak spectral ratio is 45 at f=1.6 Hz from the acceleration data whereas the corresponding value is 17 at f=3.2 Hz from the microtremors. In spite of these differences the qualitative agreement between strong ground motions and microtremors suggests that the microtremor spectral ratio provides at least a rough estimate of the site response at Manoppello. The absence of data on the velocity structure at Manoppello does not permit a theoretical modelling of the site response.

(3) The Garigliano Site (Southern Italy).

The Garigliano site is located on a 20 km wide sedimentary basin filled by a sequence of Mio-Pliocene sandstones and clays, with a 80 m thick alluvial upper layer (for details see Figure 5a). Two earthquakes were recorded at this site, namely the Irpinia main shock of November 23, 1980 (epicentral distance 113 km), and the Lazio-Abruzzo main shock of May 7, 1984 (epicentral distance 54 km). A comparison of observed acceleration with the predicted spectra suggests an amplification of seismic waves at this site (Rovelli et al. , 1988). The predicted spectra were computed on the basis of a theoretical model including an omega-square source scaling and attenuation (geometrical spreading and dissipation) whose parameters were estimated from the strong-motion data bank available for Central Italy (see Rovelli et al. , 1988). Unfortunately, there was no nearby hard-rock site where these earthquakes were recorded. We took the predicted acceleration spectrum at Garigliano for these two earthquakes as the hard rock site spectrum and divided the smoothed observed spectra by the predicted curve to obtain an estimate of the local amplification. The recorded waveforms are dominated by body waves for the Lazio-Abruzzo earthquake, and by surface waves for the Irpinia earthquake. This probably explains the slightly different response of the site during the two earthquakes (Figure 5, b and c). Also for this site we computed the response to a vertically incident SH-wave using the finite-difference algorythm by Moczo (1989). Because of the geometry of the basin (Figure 5a) we adopted a 1-D model, either with constant velocity layers or with a constant velocity gradient in each layer (see Figure 5e). Both models produce an amplification level that qualitatively agrees with the strong-motion results (Figure 5d), and confirm the role played by the local geology in the observed amplifications.

Microtremors were measured at Garigliano and a nearby hard rock site, at a distance of 10 km, approximately. The estimated local amplification and microtremor spectral ratios are shown in Figure 5d.

Microtremor spectral ratio is somewhat greater than the acceleration spectral ratio but by less than a factor of two for $0.8 \leq f \leq 14$ Hz. Previous measurements of the microtremor spectral level at Garigliano were carried out by Martini and Milana (1988): that investigation pointed out an amplification of about 10 near $f \sim 1$ Hz using a different hard-rock reference site. Their result is in agreement with that shown in Figure 5d. We conclude that both microtremor measurements and numerical modelling can give a good estimate of site response at Garigliano.

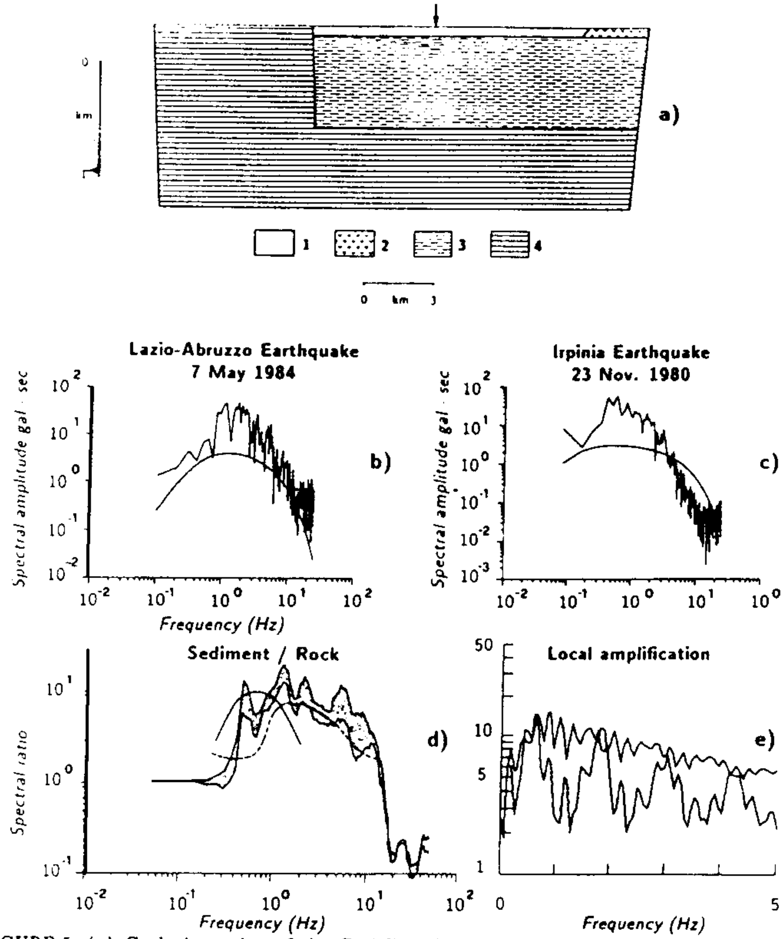

FIGURE 5. (a) Geologic section of the Garigliano basin (1: Quaternary deposits, 2: Volcanic products, 3: Pliocene clay and sand; 4: Miocene marl and marly sandstone). The arrow points out the location of the site where the accelerograph is installed. Observed acceleration spectra at Garigliano site are shown in (b) for May 7, 1984, earthquake, and in (c) for November 23, 1980, earthquake. The predicted spectrum (Rovelli *et al.*, 1988) is also drawn (smooth curve). (d) Microtremor spectral ratio between Garigliano and a rock reference site is shown by the shaded area which represents ± 1 standard deviation around the mean value. The two curves represents the spectral amplification obtained from (b) (dashed line) and from (c) (continuous line) after visual smoothing of the acceleration spectra. (e) Theoretical site transfer function estimated for *SH*-waves using a 1-*D* finite-difference scheme, assuming horizontal layers as shown in (a), with a constant velocity (lower curve) or a constant velocity gradient (upper curve) in each layer.

FURTHER EXPERIMENTS

In Italy, the collection of strong motions is relatively recent (since 1972, in practice). Only a few usable strong-motion recordings showing clear amplification exist. In the previous section, these recordings have been shown and an explanation for the reason of amplification has been tried for some of them. However, the relevance of this issue makes mandatory to collect further data in situations suggesting potential amplification of ground motion during earthquakes. To this purpose, in 1990 two experiments were carried out by ING. The first one consisted in studying seismic response of Quaternary sediments of the Tiber flood plain. This research is of particular importance because these sediments characterize a large part of ground of the City of Rome. The second experiment was devoted to study amplifications (i) occurring close to a steep rock-to-sediment contact, and (ii) due to topographical variations. Details concerning these two experiments can be found in the papers by De Cesare et al. (1990) and Malagnini et al. (1990), respectively. The most relevant features and results are here summarized.

The Tiber flood plain Experiment.

The city of Rome felt the effects of local and regional earthquakes several times during its long history, with intensities up to VII degree. The local geology is characterized by a continuous bedrock made up of highly consolidated Pliocene clays overlaid by softer Pleistocene and Holocene deposits along the Tiber valley. The Tiber river passes through the oldest part of the city, and the distribution of damages occurred during historical earthquakes shows high correlation with the maximum thickness of the alluvium fill of the Tiber palaeovalley (Ambrosini et al. , 1986). At the moment, seismic recordings from local or regional events are not available for the urban area; their acquisition has been planned but requires special cares because of the high cultural noise. For this reason, we performed the experiment outside the city, on the same alluvial sediments that characterize a large part of the urban area. During this experiment, two 5-sec S-1 seismometers were deployed for two weeks in the Tiber flood plain, 30 km north of Rome (Hough et al. , 1991). One of them, installed on a valley site, was deployed in a vacant field. The other one was located on a limestone outcrop $\gtrsim$ 15 km away. Because of the higher cultural noise in the valley, a telemetry link between the two recorders was established, so that the station on the sediments would record only when externally triggered from the reference rock station. A teleseism of a South Alaska (m_b = 5.6) and a local event ($M_L \sim 1.5$) were recorded during the experiment (data are shown in the paper by Hough et al. , 1991) at both stations. Although seismic noise within the valley is high and earthquake waveforms are masked by noise for the soft soil site, the response of sediments could be inferred from a moving window spectral analysis (see Figure 6). The small local earthquake was particularly useful to this purpose: while the rock site recording shows that this earthquake was dominated by energy in the band 4-20 Hz, the response of Quaternary sediments was characterized by an increase of the spectral amplitude in a narrow band centered at 2.5 Hz. The same frequency is also sharply peaked in the spectra of ambient noise in the valley. These observations suggest that the spectral response of these sediments is controlled by resonance phenomena related to the geological and mechanical properties of the subsurface materials. Therefore, geometry and geotechnical parameters of sediments were taken

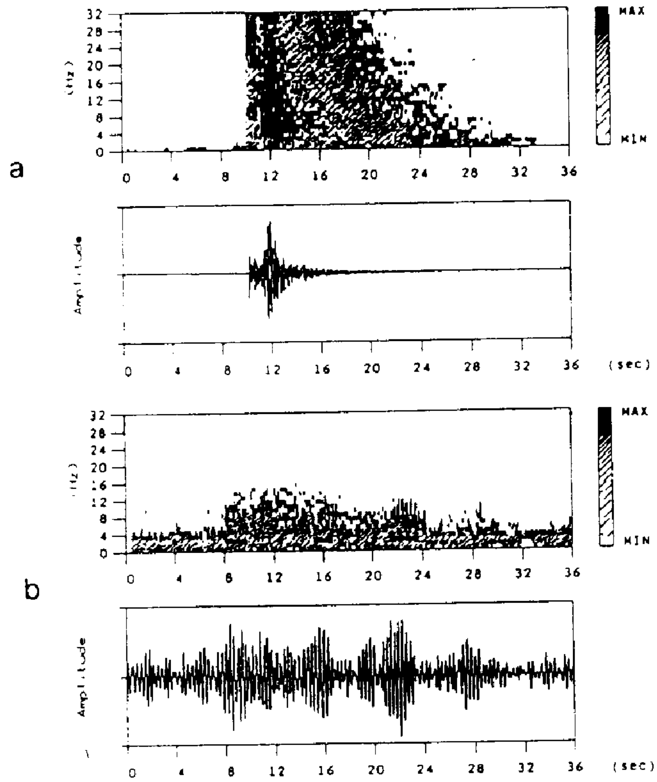

FIGURE 6. The same local earthquake was recorded in the Tiber valley area 30 km north of Rome, on Quaternary sediments (a) and on Mesozoic limestones (b). Velocity time history and dynamic amplitude spectrum are shown for both sites. While in (a) the spectral content ic controlled by the source spectrum, in (b) the local earthquake produces the maximin amplitude in the band 2.5-3 Hz, that is also significantly peaked in the ambient noise spectra within the valley.

into account. It has been demonstrated (Aki, 1957 and 1965) that phase velocity can be determined for upper sediment layers using a statistical analysis of the ambient noise. This approach consists in recording the ambient noise with a circular array, having a station in the centre of the circle and calculating the azimuthally averaged correlation function between the centre and the external recordings. Seismic noise is assumed to be stationary in time and space and the incoming waves must be statistically independent. In other words, there should be neither preferential direction the recorded noise wavefield is coming from, nor recognizable transients during the recording windows; moreover, the propaganting waves must have the same phase velocity for a given frequency. A detailed description of this methods can be found in the previously quoted papers by Aki (1957 and 1965). Given a circular array of stations with a station in the middle, a space correlation function can be defined as

$$\phi(r,\lambda) = < \mathrm{u}(x,y,t) \cdot \mathrm{u}(x + r\cos\lambda, y + r\sin\lambda, t) >$$

where r is the station separation and brackets denote the ensamble average. The azimuthal average is given by

$$\phi(r) = \frac{1}{\pi} \int_0^{\pi} \phi(r, \lambda) d\lambda$$

This correlation function is related to the power spectrum $\Phi(\varpi)$ via the Hankel Transform

$$\phi(r) = \frac{1}{\pi} \int_0^{\infty} \Phi(\omega) J_o\left(\frac{\omega r}{c(\omega)}\right) d\omega$$

where J_0 is the zero-order Bessel function. If noise array recordings are bandpass filtered over a narrow frequency band centered on ω_0 and $\phi(r, \omega 0)$ is the correlation function computed for the filtered series, we have

$$\phi(r, \omega_o) = P(\omega_o) J_o\left(\frac{\omega_o r}{c(\omega_o)}\right)$$

where $P(w_0)$ is the power spectral density at frequency w_0. Thus, Rayleigh-wave phase velocity can be estimated from correlation functions of vertical motion recorded by a circular array.

Following this approach, we recorded ambient noise at three stations equispaced along a circle with a radius $r = 50$ m. Another station was installed in the centre of the circle. Averaged correlation functions were computed for a set of narrow frequency bands between 1 and 3 Hz: estimated phase velocity are shown in Figure 7a. The error bounds indicate the mean value $\pm$ 1 standard deviation as computed over data from 5 microtremor windows. Assuming that noise consists predominantly in surface waves generated by incident body-wave energy and that the vertical components reflect Rayleigh wave motion, a shear wave velocity of 440-530 m/sec was estimated for the sediments of the valley. Details about measurements and computations are described in the paper by Hough et al. (1991). In a different way, given the inferred fundamental resonance frequency of 2.5 Hz and assuming a layer thickness of 40 m, which corresponds to the known thickness of the Holocene sediments, we derive a shear-velocity of 400 m/sec on the basis of the well known quarter-wavelength law. These two estimates are in satisfactory agreement.

Being the Tiber valley very large with respect to the thickness of the alluvial fill, a one-dimensional model was sufficiently realistic to try reproducing its seismic response (see Hough et al. , 1991). In Figure 7b the calculated transfer function is shown together with the scaled spectrum of the recorded noise, a shear-wave velocity of 450 m/sec having been assumed for the sediments. The interpretation of the frequency peaked in microtremor spectra in terms of a sediment resonance is consistent with all the other observations.

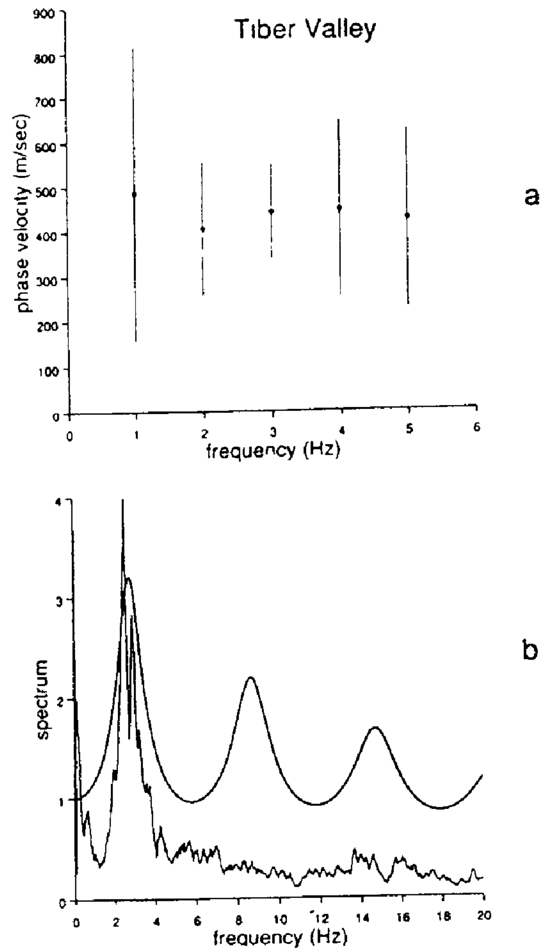

FIGURE 7. (a) Dispersion of microtremor surface waves recorded by a circular array. (b) Theoretical transfer function and observed microtremor spectrum on Quaternary sediments. Velocity estimated in (a) and thickness derived from well log data have been used in computing the theoretical response of the sediment fill.

In Mexico City, for the lake-bed zone resonant periods identified in ambient noise have been shown to match dominant resonances observed in strong motions (Celebi et al., 1987; Lermo et al., 1988; Singh et al., 1988). Although transfer functions are difficult to be estimated from ambient noise studies because of the presence of very local sources of cultural noise, sediment-induced resonances can be detected in cases of impedance contrast strong enough to trap large amount of energy in the sediment upper layers. For ancient river valleys, previous studies suggest that sediment-induced resonances can be detected by analyzing ambient noise: Flushing Meadows, New York (Field et al , 1990) and Quinnipiac River valley in New Haven, Connecticut (Hajnal et al., 1990). For the Tiber valley, this study points out a peak at 2.5 Hz in the ambient noise spectra that represents a quarter-wavelength resonance of the sediments: a small local earthquake having a spectral content dominated by energy in the band 4--20 Hz is shown to excite the same resonance within the valley as the noise.

The Mount Cetona Experiment.

Our goal was to study the effects of topography and strong acoustic impedance contrast on ground motion in some simple structures when artificial sources are used (column shots for refraction exploration). In the Vulsini volcanic complex, the Consiglio Nazionale delle Ricerche (CNR) and the Istituto Nazionale di Geofisica (ING) planned a refraction experiment to study the deep structure of the area. In the CNR--ING joint-project, also a site-effect study was realized using one of the refraction shots and comparing seismic response in sites characterized by different geology and variations of topography. The area of the experiment had to be chosen sufficiently close to the shot, in order to have a good signal-to-noise ratio, but far enough from it in order to have the same entire-path propagation effects. A short distance between the stations with respect to the mean distance from the source allows to obtain the optimal site response estimates, in the linearity domain and far field conditions. The most suitable situation was the Monte Cetona chain and the Pliocene clays-filled valley, which is eastward bounded by the Monte Cetona limestones, along a normal fault which runs approximately North-South for several kilometers. A simplified geological section of the Monte Cetona is sketched in Figure 8*a*.

The Monte Cetona is a Mesozoic limestones recumbent anticline, eastern turning with its axis in North-South direction, cutted by a normal fault along the entire western limb. Along its axis, the anticline is ~ 11 Km long and ~ 3 Km wide and reaches 1148 m on the sea level. There is a topographic gap of 450-500 m between the top and the western side valley, where the Pliocene clays outcrop. The dip of the western limb is ~ 35^{o}, whereas the dip of the eastern one is ~ 45^{o}. The argillous complex which outcrops in the valley, in conformity with a Miocene polygenic conglomerate, is more than 1000 m thick. From the geologic profile shown in Figure 8a it can be seen that the argillous complex thickness on the eastern side of the Monte Cetona is less than 200 m. To justify a 1000 m thick sedimentary fill in the "Celle sul Rigo" drill data, it is necessary to introduce a graben-like geometry with one or more normal faults that break the continuity of the argillous complex bed.

Two linear arrays were set up in the area: one of them was put across the mountain chain, whereas the other one was developed on the Pliocene clays, eastward from a hard rock reference site on the Cetona limestones, orthogonally to the surface evidence of the fault plane (Figure 8, *b* and *c*). The Profile to Study the Topographic Effect (PSTE, hereafter) had a length of 1 km, approximately, with a mean source-receiver distance of ~ 19 km, and the Profile to Study the Seismic Impedance Contrast Effect (PSSICE) was ~ 1.1 km long, with a mean source-receiver distance of ~ 14 km. We used a homogeneous set of twenty Mars-88 digital recorders, each of which was equipped with a three-component Mark L-4 C seismometer (1-sec period) Seven of those instruments were placed on the PSTE (one reference site), and the remaining thirteen on the PSSICE (two reference sites to prevent instrumental failures). Unfortunately, one PSSICE

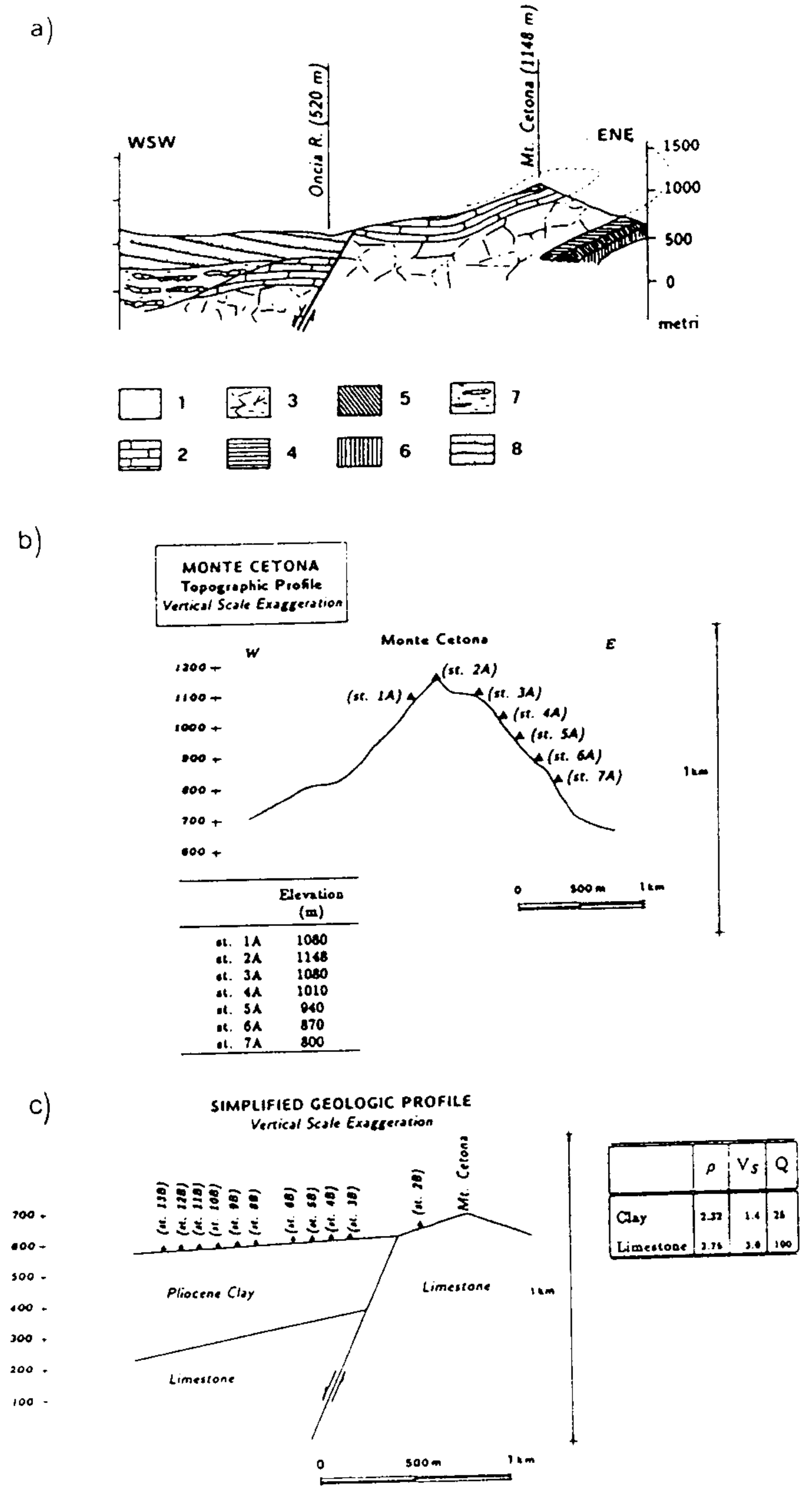

FIGURE 8. (a) Geological section on the Monte Cetona area (1: Magnesian limestone; 2: Limestones and marls with *Avicula Contorta*; 3: Massive limestone; 4: Cherty limestones; 5: Marls with *Posidonia*; 6: Radiolarites; 7: Miocene polygenic conglomerate; 8: Pliocene clays). Profiles for the study of the effects of topograpy variations and impedance contrast are shown in (b) and (c), respectively.

station did not work, so there was a lack of regularity in the array, as it can be seen in Figure 8c. The PSTE seismometers were put on fracturated and weathered outcropping limestones, whereas the PSSICE reference station (hereafter we will consider only one of the two reference stations for this profile) was placed on a proper "firm" site, the other PSSICE instruments were set up on the Pliocene sedimentary units. The distance between two recording sites on the profiles has been taken constant, approximately 100 m from each others for the PSSICE and ~ 150 m for the PSTE. The column shot was performed using $\sim$ 400 kg of Seismic-2 explosive, which was put in column in a 50 m deep hole. Using this technique shear waves are generated (Biella et al., 1989), and a more realistic simulation of the earthquake waves can be carried out. Also incidence angle has been estimated to be between 0 and 5 degrees, very similar to an earthquake wavefield incidence.

In Figures 9a and 10a, EW components of seismograms recorded along the two seismic profiles are shown. They have been plotted using the same scale, so these figures immediately stress differences in site response.

It is notewhorty that on the PSSICE there are very strong amplifications, up to a factor of 20, whereas for the PSTE recorded seismograms the amplification values reaches a factor of 4 (the maximum peak of amplitude increasing is calculated with respect to the reference station). As far as duration is concerned, we see that significant signal duration does not vary in the PSTE recordings, whereas in the PSSICE ones the presence of sediment upper layers produces a significant increase of it, moving from the reference site toward the most external one. This fact is basically due to energy reverberations in the sediment body and to surface waves generated at the surface intersection between different geological units.

One way to quantify site transfer functions is to use spectral ratios (see Tucker and King, 1984; King and Tucker, 1984). In order to choose the time interval of data to be windowed for spectral analysis, we performed a polarization study on the recorded waveforms. Spectral ratios for *S*-wave windows 2.5-sec large are shown in Figure 9*b* for the PSSICE profile, and in Figure 10*b* for the PSTE one. In both cases, data were tapered by using 10% cosine windows. For PSSICE recordings we used also 30-sec windows beginning with direct shear-waves and including also the coda waves. Spectral ratios for these longer widows of data do not differ from the previous ones.

The behaviour of the signal-to-noise ratio as a function of frequency is important to evaluate the reliability of frequency-domain results. Spectra of explosion recording compared to noise show that the frequency band where signal-to-noise ratio is higher than a factor of 6 is between 0.5 and 5 Hz. The lowest frequency bound is mainly determined by the instrumental response, whereas upon the higher one the signal content tends to be comparable to the ambient noise level. When

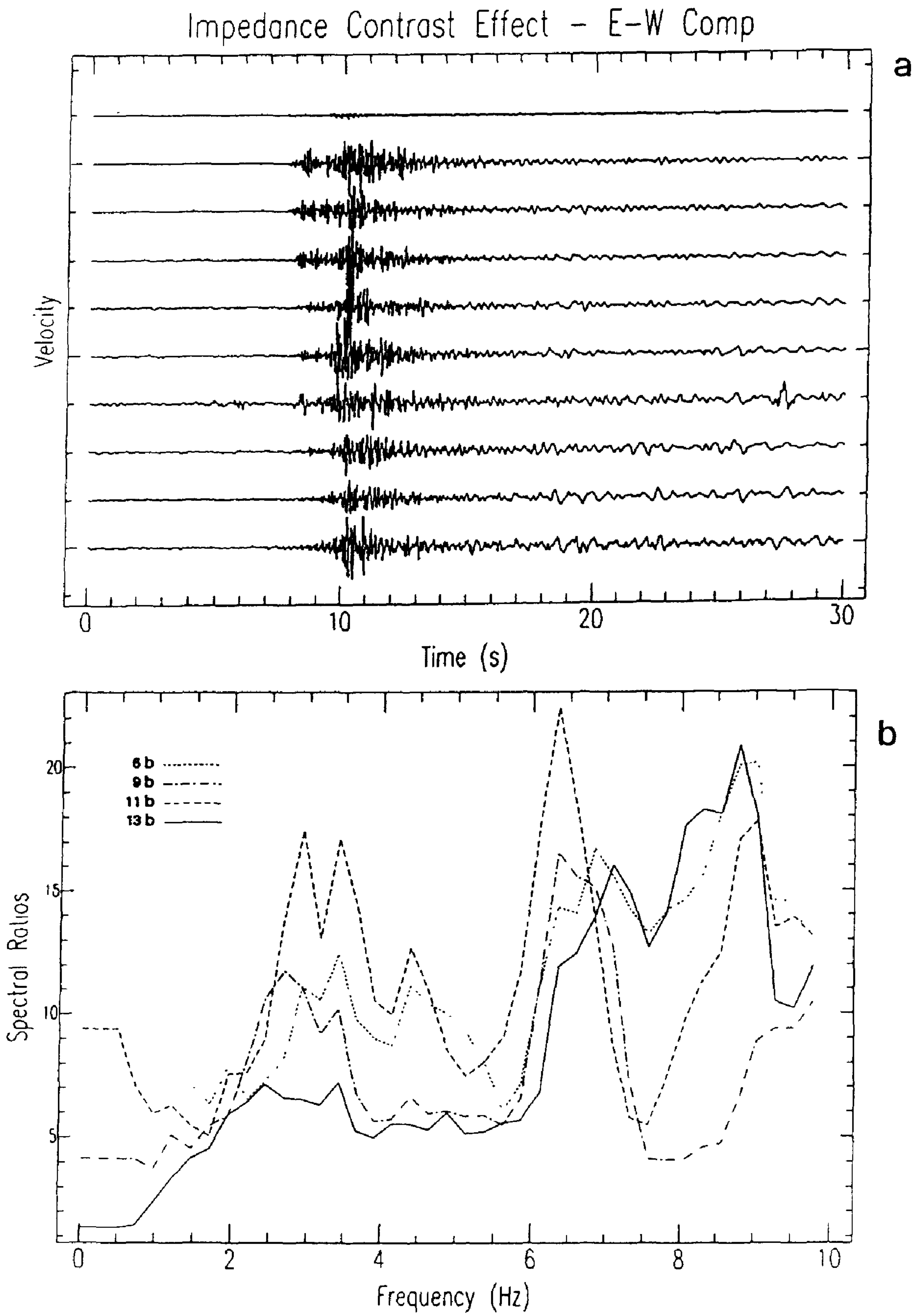

FIGURE 9. (a) Seismograms showing explosion recording along PSSICE, and (b) spectral ratios computed using limestone station as a reference site.

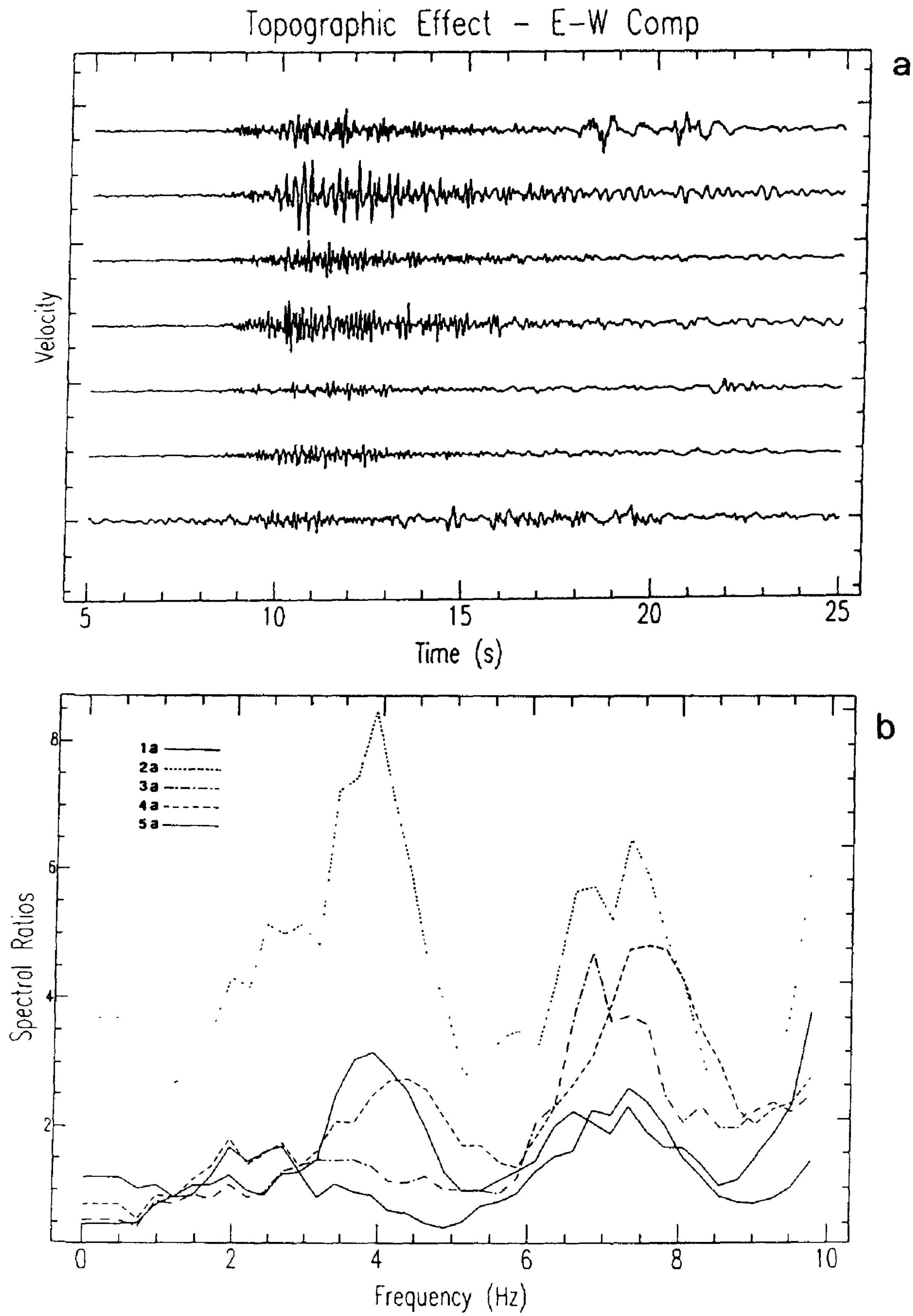

FIGURE 10. (a) Seismograms showing explosion recording along PSTE, and (b) spectral ratios computed using bottom station as a reference site.

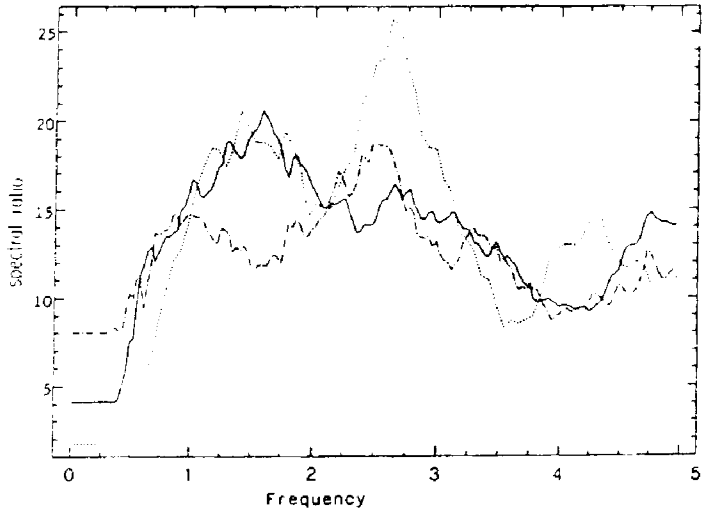

FIGURE 11. Spectral ratios computed from explosion recordings, and from data of local and regional earthquakes.

the band 0-10 Hz is considered (Figures 9b and 10b), it has to be taken into account that at 10 Hz explosion amplitude differs only by a factor of 2 than noise level.

For both profiles, observed amplifications are much higher than those predicted on the basis of theoretical estimations (e.g., using SH-wave modelling). The choice of the refence station may be responsible for this difference. This explanation is probably partially true for PSTE, where local fluctuations of the degree of surface fracturation can bias the estimation of amplification due to variations of topography. On the contrary, the reference station for PSSICE was installed on unfractured and unweathered materials where anomalous responses are not likely to occur. Excluding a significant role of the reference site for PSSICE, we interpret the observed large amplification as the effects of energy reverberation in the sediment body and local generation ofsurface waves.

In order to compare explosion results with amplifications observed during earthquakes, a couple of recording stations were deployed on the PSSICE, one of them on the reference site and the other one very close to the end of the profile (site n° 13).

The instruments worked for approximately one month, recording local, regional and teleseismic events. A telemetry link was established between the recording stations, so the sediment station would start recording only when externally triggered. Among the data set of 12 simoultaneously recorded events, two earthquakes (one local and one regional event) were selected on the basis of the most favourable signal-to-noise ratio. The computed spectral ratios are shown in Figure 11, together with the explosion results for site n° 13.

The significant agreement between explosion and earthquake data confirms the suitability of recordings of artificial sources to investigate site effects.

CONCLUSIONS

The aim of these investigations has been to test the applicability of different methodologies to estimate site transfer functions during earthquakes. In Italy only a few strong-motion accelerograms showing clear amplification are available. We have used some of these recordings to check the suitability of numerical modelling techniques and microtremor method in reproducing ground motion amplifications. Some concluding remarks emerge from these (few) test cases. First, efficiency of the numerical modelling is generally limited by poor knowledge of details of the subsurface structures and their geotechnical parameters. When reliable information is available, results can be very satisfactory. Second, comparison between amplifications observed from strong motions and from microtremors gave qualitatively consistent results. However, the spectral details of amplifications observed during earthquakes are not always well reproduced by the microtremor data. This means that predictions based on ambient noise measurements can be successful only under particular circumstances: a widespread use of this technique cannot be recommended yet.

At this moment, the state-of-art suggests that weak motions (e.g. small local earthquakes or regional events) are the most suitable means for predictions of strong amplifications. Two experiments were devoted to measure weak motions on sediment bodies where ground motion amplifications are suspected. The first one was concerning the Quaternary sediments of Tiber flood plain. Within the Tiber valley, ambient noise resulted to be sharply peaked near 2.5 Hz; a small local earthquake simultaneously recorded on sediments and at a reference rock site, although dominated by energy in the band 4-20 Hz, was shown to excite the same narrow frequency band as the noise. Available well log data and phase velocity estimated from dispersion of microtremors recorded by a circular array allowed to interpret the peaked frequency as the quarter-wavelength resonance of the sedimentary fill. In the second experiment, shots for explorations were used to measure amplifications of ground motion on Pliocene clays compared to a nearby "firm" reference site. Amplitudes on sediments resulted to be larger than those predicted by theory. Amplifications estimated from explosion recordings were consistent with results obtained from data of local and regional earthquakes.

ACKNOWLEDGMENTS

The review presented in this paper describes investigations developed in the recent years in the framework of many cooperations between the Istituto Nazionale di Geofisica (ING) and other national and international Institutions: the Italian Council of Research (CNR), the Agency for Electric Power Supply (ENEL), the Universidad Nacional Autonoma de Mexico (UNAM), the Lamont-Doherty Geological Observatory (LDGO), the Geophysical Institute of Slovak Academy of Sciences GISAS of Bratislava. We thank all the collegues from these Institutions who partecipated to the works described in the text. Particular thanks are due to Peter Moczo whose

experience in using finite-difference methods was resolutive for many problems, and to Shri K. Singh for many discussions and suggestions.

REFERENCES

Aki, K. (1957). Space and time spectra of stationary stochastic waves, with special reference to microtremors, Bull. Earthq. Res. Inst. 35, 415--457.

Aki, K. (1965). A note on the use microseisms in determining the shallow structure of the Earth's crust, Geophysics 30, 665--666.

Ambrosini, S., S. Casteneto, F. Cevolani, E. Di Loreto, R. Funiciello, L. Liperi, and D. Molin (1986). Risposta sismica dell'area urbana di Roma in occasione del terremoto del Fucino del 13/1/1915. Risultati preliminari, Mem. Soc. Geol. It. 35 , 445-452.

Biella, G., R. de Franco, E. Lentini, and R. Masotti (1989). Analisi delle onde S nella sismica a rifrazione profonda, 8-th Annual Meeting of Gruppo Nazionale di Geofisica della Terra Solida, Rome, Italy.

Carrara, C., M. Frezzotti, C. Margottini, D. Molin, and B. Narcisi (1987). Indagini sismotettoniche sui terremoti laziali-abbruzzesi del 7 e 11 maggio 1984, E.N.E.A. Technical Report, Rome.

Celebi, M., C. Dietel, J. Prince, M. Onate, and G. Chavez (1987). Site amplification in Mexico City (determined from 19 September strong-motion records and from recordings of weak motions, in Ground Motion and Engineering Seismology, ed. A.S. Cakmak, 141-152, Elsevier.

Cocco, M. and A. Rovelli (1989). Evidence for the variation of stress drop between normal and thrust faulting earthquakes in Italy, JGR 94, 9399-9416.

De Cesare, S.E. Hough, L. Malagnini, A. Rovelli, L. Seeber, and G. Selvaggi (1990). Rumore sismico e risonanza su sedimenti prodotta da terremoti: un esperimento nella valle del Tevere, 9-th Annual Meeting of Gruppo Nazionale di Geofisica della Terra Solida, Rome, Italy.

Field, E. F., S. E. Hough, and K. Jacob (1990). On the use of microtremors to assess potential earthquake site response: A case study in Flushing Meadow, New York, Bull. Seism. Soc. Am, 80, 1456-1480.

Hajnal, Z., J. Park, and S.E. Hough (1990). Ambient noise and sediment resonances around New Haven, CT, (abstract), EOS Trans. AGU}.

Hough, S.E., L. Seeber, A. Rovelli, L. Malagnini, A. De Cesare, A. Lerner-Lam, and G. Selvaggi (1991). Ambient noise and weak motion excitation of sediment resonances: results from the Tiber valley, Italy, to be submitted to Bull. Seism. Soc. Am

Kanai, K. and T. Tanaka} (1961). On microtremors. VIII, Bull. Earthq. Res. Inst., Tokio Univ., f 39}, 97-115.

King, J. L. and B. E. Tucker (1984). Observed variations of earthquake motion across a sediment-filled valley, Bull. Seism. Soc. Am., 74}, 137-151.

Lermo, J., M. Rodriguez, and S. K. Singh (1988). The Mexico earthquake of September 19, 1985 - Natural period of sites in the valley of Mexico from microtremor measurements and strong motion data, Earthquake Spectra, 4, 805-814.

Malagnini, L., A. Amato, G. Biella, A. Caserta, A. De Cesare, R. De Franco, M. Di Bona, A. Rovelli, and G. Selvaggi (1990). Determinazione della risposta sismica locale mediante esplosioni in colonna: un esperimento sul Monte Cetona, 9-th Annual Meeting of Gruppo Nazionale di Geofisica della Terra Solida, Rome, Italy.

Martini, M. and G. Milana (1988). Uso dei microtremori per caratterizzare fenomemi di amplificazione locale, 7-th Annual Meeting of Gruppo Nazionale di Geofisica della Terra Solida, Rome, Italy.

Moczo, P. (1989). Finite difference technique for SH-waves in 2-D media using irregular grids. Application to the seismic response problem, Geophys. J. Int., 99}, 321-329.

Muzzi, F. and A. Pugliese (1977). Analysis of the dynamic response of a soil deposit in locality "Ca' Dant" (Cornino - Forgaria), Proceedings of Specialist Meeting on the 1976 Friuli earthquake and the antiseismic design of nuclear installations, Rome, 11 - 13 October 1977.

Rovelli, A., O. Bonamassa, M. Cocco, M. Di Bona, and S. Mazza (1988). Scaling laws and spectral parameters of the ground motion in active extensional areas in Italy, Bull. Seism. Soc. Am, 78, 530-560.

Rovelli, A., S.K. Singh, L. Malagnini, A. Amato, and M. Cocco (1991). Feasibility of the use of microtremors in estimating site response during earthquakes: some test cases in Italy, submitted to Earthquake Spectra.

Singh, S. K., J. Lermo, T. Dominguez, M. Ordaz, J. M. Espinosa, E. Mena, and R. Quass (1988). A study of amplification of seismic waves in the valley of Mexico with respect to a hill zone site (C.U.), Earthquake Spectra, 4, 653-673.

Tucker, B. E. and J. L. King (1984). Dependence of sediment-filled valley response on input amplitude and valley properties, Bull. Seism. Soc. Am., 74, 153-165.