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Realistic modelling of observed seismic motion in complex sedimentary basins

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Abstract

Three applications of a numerical technique are illustrated to model realistically the seismic ground motion for complex two-dimensional structures. First we consider a sedimentary basin in the Friuli region, and we model strong motion records from an aftershock of the 1976 earthquake. Then we simulate the ground motion caused in Rome by the 1915, Fucino (Italy) earthquake, and we compare our modelling with the damage distribution observed in the town. Finally we deal with the interpretation of ground motion recorded in Mexico City, as a consequence of earthquakes in the Mexican subduction zone. The synthetic signals explain the major characteristics (relative amplitudes, spectral amplification, frequency content) of the considered seismograms, and the space distribution of the available macroseismic data. For the sedimentary basin in the Friuli area, parametric studies demonstrate the relevant sensitivity of the computed ground motion to small changes in the subsurface topography of the sedimentary basin, and in the velocity and quality factor of the sediments. The relative Arias Intensity, determined from our numerical simulation in Rome, is in very good agreement with the distribution of damage observed during the Fucino earthquake. For epicentral distances in the range 50 km-100 km, the source location and not only the local soil conditions control the local effects. For Mexico City, the observed ground motion can be explained as resonance effects and as excitation of local surface waves, and the theoretical and the observed maximum spectral amplifications are very similar. In general, our numerical simulations estimate the maximum and average spectral amplification for specific sites, *i.e.* they are a very powerful tool for accurate micro-zonation.

Key words *wave-propagation modelling – seismic strong ground motion – sedimentary basins – seismic micro-zonation*

1. Introduction

The presence of unconsolidated sediments with irregular geotechnical characteristics makes sedimentary basins the zones which are most vulnerable to earthquakes. In fact, when the shear-wave velocity at the surface is low, quite large amplifications of the ground motion are observed, and localized amplification of the signals are often related to lateral irregularities in the subsurface topography (*e.g.* Jackson, 1971). Even smooth variations of the near-surface structure can cause large differential motion, and in relatively close points it is possible to observe signals with significantly different amplitude and duration.

It is well-known that incident plane waves are amplified when the seismic wave travels through an interface from a medium with high rigidity, into a medium with low rigidity. For vertically incident waves, the frequencies of the mechanical resonance which can occur in sedimentary basins, are given by $f_n = (2n+1)\beta/4h$, where β is the shear wave

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velocity in the basin and h is its thickness (Haskell, 1960, 1962). An irregular interface between bedrock and sediments can cause the focusing of waves (*e.g.* Aki and Larner, 1970; Boore *et. al.*, 1971; Sánchez-Sesma *et al.*, 1988), and can excite local surface waves (*e.g.* Trifunac, 1971; Bard and Bouchon, 1980a,b). These local surface waves can be excited not only by body waves but also by the incidence of surface waves (Drake, 1980). Bard and Bouchon (1985) demonstrated that the occurrence of local surface waves is determined primarily by the depth of the basin and by the contrast between the shear-wave velocity of the basin and that of the bedrock. When the wavelength of the incident wave is comparable with the depth of the basin, the local surface waves can have larger amplitudes than the direct signal, and, if the contrast in the elastic parameters between the sediments and the underlying bedrock is high, they can be reflected at the edges of the basin, causing a long duration of the ground motion in the basin. This behavior does not change if a vertical stratification of the sediments, with a large vertical velocity gradient, is considered (Bard and Gariel, 1986).

For preparedness purposes, it is crucial to estimate the seismic ground motion, before an earthquake occurs, and to include these results in the assessment of seismic hazard or in (micro)-zonation studies. A powerful tool to estimate the amplification effects in complex structures are numerical simulations. One major problem which is encountered when doing such simulations is the large number of parameters which have to be specified as input. The choice of these parameters should be based on all available seismological, geological and geotechnical information for the area under consideration. At a specific site, the numerical simulation can predict the seismic response only if the properties of the seismic source and the mechanical (density, velocity, damping, etc.) and geometrical (such as layer thickness) parameters of the path from the source are reasonably well-known. In general this is not the case, and parametric studies are necessary to quantify the variability of the expected ground motion. Whenever possible, such simulations, must be compared with observed ground motion.

The method we use for the modelling of the wave propagation in two-dimensional complex media, is the hybrid technique, which combines modal summation (Panza, 1985; Florsch *et al.*, 1991) and the finite difference technique (Korn and Stöckl, 1982; Virieux, 1986), described by Fäh (1992), Fäh *et al.* (1993a; 1993b), and Fäh *et al.* (1994). The propagation of waves from the source to the sedimentary basin is treated with the mode summation method, applied to plane layered, anelastic structures which represent the average crustal properties along the source-basin path. In our modelling, this structure is used as the reference, bedrock model. The wavefield computed for this bedrock model is used as incident wavefield for the explicit finite difference schemes, which are used to simulate the propagation of seismic waves in the two-dimensional, anelastic model of the sedimentary basin. This hybrid method allows us to take into account the source and propagation effects, including local soil conditions, even when dealing with path lengths of a few hundred kilometers.

In the following, we illustrate a comparison between numerical simulations, and observed strong ground motion or the space distribution of the available macroseismic data. We will focus on the different effects of the source, the path and the local soil conditions. Special emphasis is given to understanding the different features of ground motion in sedimentary basins, and the application of numerical simulations for seismic zonation. The three applications include sites close to the source (a sedimentary basin near to the epicenter of the September 11, 1976, Friuli aftershock at 16 h 35 min 4 s), at an intermediate distance (Rome, about 80 km from the epicenter of the January 13, 1915, Fucino earthquake), and sites that are far from the source (Mexico City, about 400 km from the epicenter of the September 19, 1985, Michoacan earthquake).

2. The sedimentary basins in the Friuli area

The September 11, 1976 Friuli aftershock (16 h 35 min 4 s) was recorded by a few accelerographic stations (CNEN-ENEL, 1977). Records from one of the nearest stations – the three-component records at Buia station (fig. 1) – are considered and compared with the theoretical computations. We focus on the variability of ground motion within the sedimentary basin, and on the sensitivity of the computed ground motion to small changes in the subsurface topography of the sedimentary basin, and in the velocity and quality factor of the sediments. The interpretation of the observed data will here be given in terms of the near-surface geological model by keeping the source and long-distant propagation path unchanged for all numerical simulations.

The bedrock model, describing the average mechanical properties of the path from the epicenter to the sedimentary basin, has been proposed by Fäh *et al.* (1993a), and its uppermost layers are shown in fig. 2. The source-depth used in the numerical modelling is 7.1 km, the angle between the strike of the fault and the epicenter-station line 19°, the dip 28°, the rake 115°, and the source duration is 0.6 s (Florsch *et al.*, 1991).

The area where the station of Buia is located is characterized by terrigenous sediments (Flysch), widely outcropping at Monte Buia (fig. 1), which are covered locally by a thin quaternary layer, forming a sedimentary basin, and which overlap a carbonatic mesozoic sequence. The thickness of the quaternary sediments is well-known (Giorgetti and Stefanini, 1989) and locally can reach 100 m. The aver-

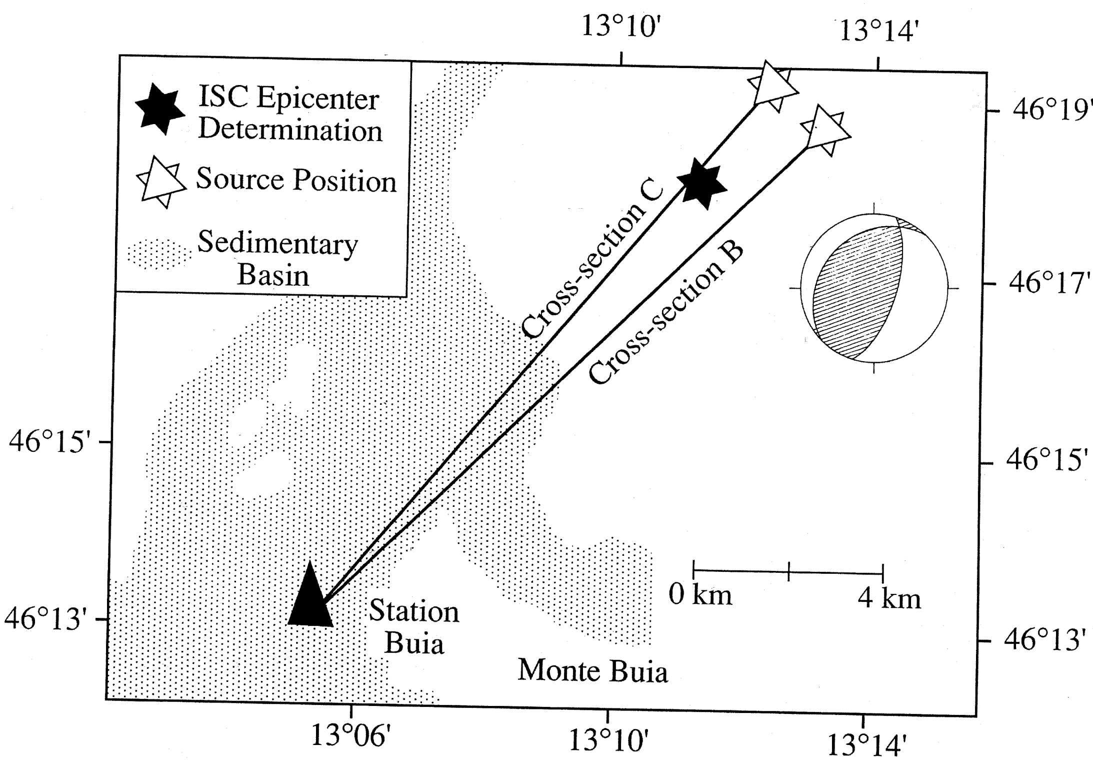

Fig. 1. Overview of the Friuli seismic region, showing the presence of the quaternary basin, the ISC epicenter determination of the September 11, Friuli 1976 aftershock (16 h 35 min 4 s), and the position of Buia station. The solid lines indicate the position of the two cross-sections for which 2D modelling has been performed.

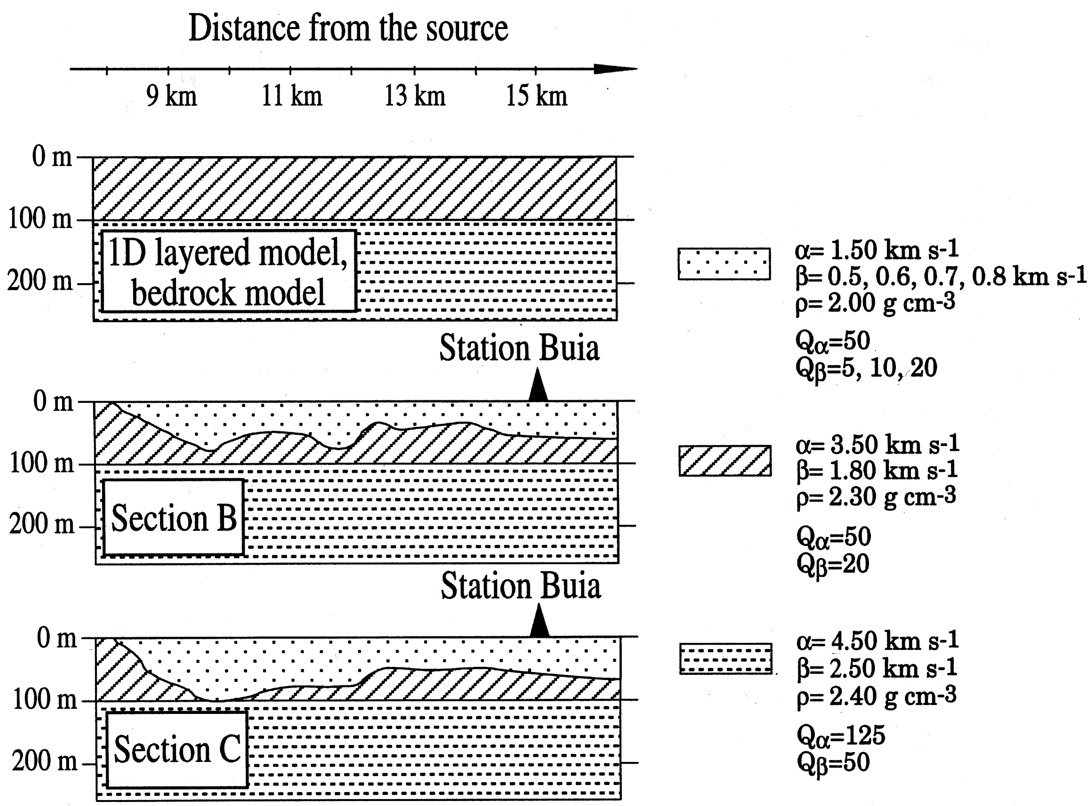

Fig. 2. Surficial layers of the one-dimensional model, describing the propagation of waves from the source position to the sedimentary basin (Fäh *et al.*, 1993a), and of the 2D models corresponding to cross-sections B and C in fig. 1. Only the part near to the surface is shown, where the 2D models are different from the bedrock model.

age model for the sedimentary basin is represented by the cross-section B, while the part of the basin with a thick sedimentary cover is well represented by the cross-section C, both shown in fig. 2.

The accelerograms obtained in correspondence with the cross-section B, with the shear-wave velocity of the unconsolidated sediments equal to 0.6 km/s and the quality factor Q_β equal to 20, are shown in fig. 3. At sites with a thick layer of low-velocity material near the surface, the peak acceleration of the radial component is up to two times larger than the vertical component, while, due to the source radiation pattern, the transverse component is the same size as the radial component.

The heterogeneity, with size comparable to the wavelength of the incident wavefield, causes significant spatial variations in the ground motion (fig. 3). There are three major effects which are caused by the presence of the sedimentary cover: 1) the excitation of surface waves at the edge of the sedimentary basin; 2) the resonance in parts of the basin due to the subsurface topography of the bedrock, and 3) the excitation of very dispersed local surface

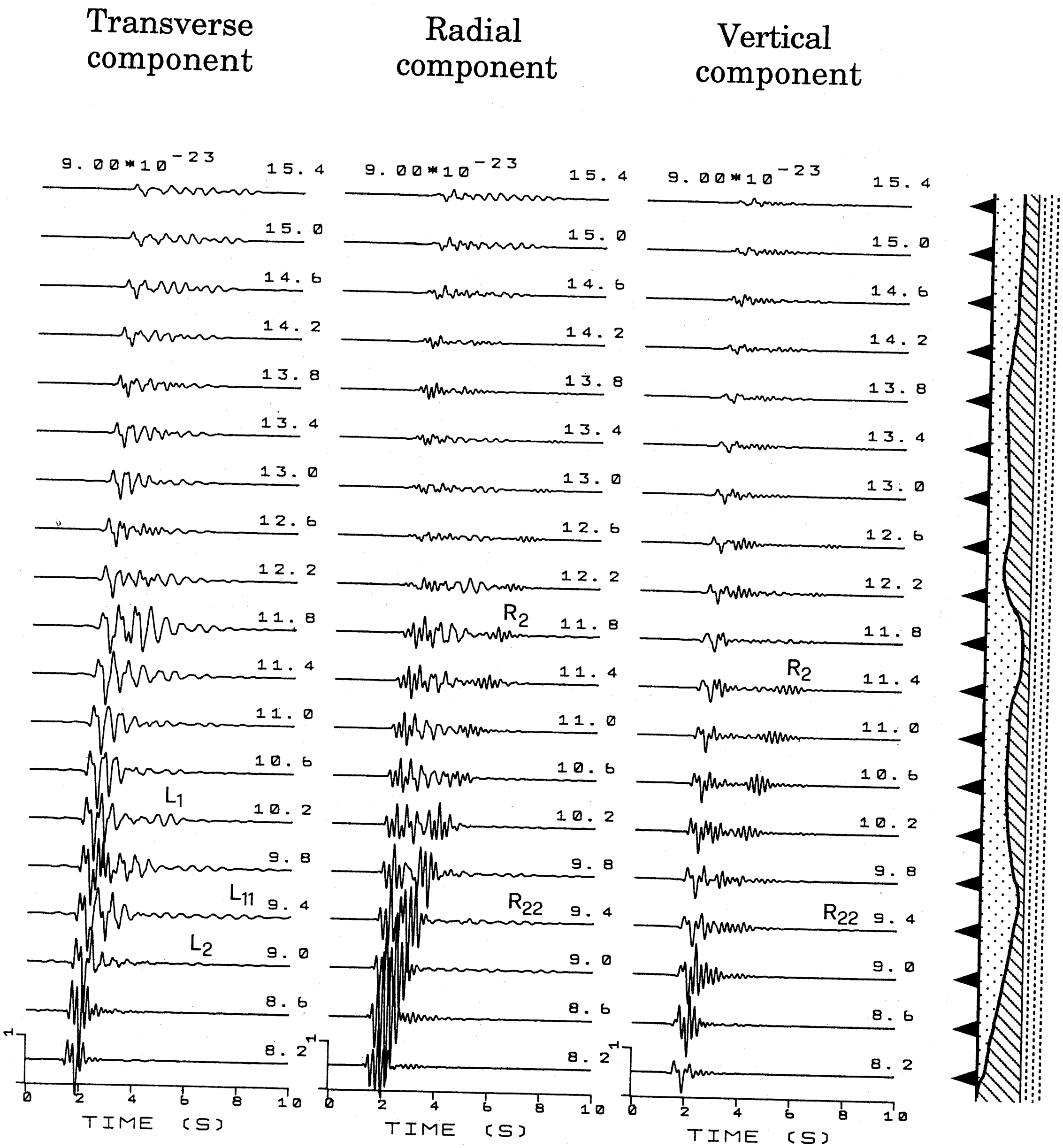

Fig. 3. Acceleration time series for *SH* and *P-SV* waves at an array of receivers, for cross-section B shown in fig. 2 ($\beta = 0.6$ km/s and $Q_\beta = 20$ for the unconsolidated sediments). All amplitudes are related to a source with a seismic moment of 10^{-7} N m. The signals are normalized to the peak acceleration given in units of cm s^{-2}. The distance to the source for each seismogram is given in units of km. The time scale is shifted by 2 s from the origin time (0 s in the figure is really 2 s from the origin time).

waves, with peak energy at about 2 Hz, within the sedimentary basin for epicentral distances larger than 14 km.

Multiple reflections of *SH*-waves can generate local surface waves (phase L_1) – forming the coda of the signals (fig. 3) – which are excited as soon as the fundamental Haskell's frequency of resonance (Haskell, 1960, 1962) for the sedimentary basin is reached. These local surface waves can be reflected inside the basin at places where the sediments become thin. One example is given by the phase L_{11} in fig. 3, which corresponds to the reflected local surface wave L_1. Also for *P-SV*-waves, a dipping layer at the edge of a sedimentary basin gives rise to multiple reflections of body waves and the excitation of local surface waves (phase R_2 in fig. 3), which are characterized by larger amplitudes on the radial than on the vertical component of motion, and can be the dominant part of the wavefield at the edge of the sedimentary basin. The reflections of the local Rayleigh waves inside the sedimentary basin (R_{22} in fig. 3) do not appear as clear as in the case of Love waves, since their amplitudes are small in comparison with the amplitudes of the primary waves.

Resonance occurs in those parts of the basins with smooth variations of the interface between the bedrock and the sediments, and originates from the superposition of forward propagating local surface waves with their reflections, within sub-basins of the sedimentary cover. Examples are seen for the two sub-basins in cross-section B, especially at 9.8 km and at 11.8 km from the source (fig. 3). The resonance is stronger for *SH*- than for *P-SV*-waves, and in general can give rise to very large duration and amplitude, as can be seen from the signal computed at 11.8 km from the source. The excitation of strongly dispersed local surface waves by subsurface lateral heterogeneities can be observed in correspondence with the cross-section B at epicentral distances larger than 14 km, both for *SH*- and *P-SV*-waves.

The synthetic accelerograms, obtained from the numerical modelling, can be used to compute some ground motion related quantities. These quantities are: 1) the peak ground acceleration PGA and 2) the quantity W defined as:

$$W = \lim_{t\to\infty} \int_0^t [\ddot{x}(t)(\tau)]^2 d\tau$$

where $x(t)$ is the time series describing the ground displacement. The Arias Intensity is proportional to the quantity W by a factor $\pi/(2g)$, where g is the acceleration of gravity. To discuss the site effects with respect to the bedrock model it is convenient to consider the quantities PGA(2D), and W(2D), *i.e.* PGA and W obtained from the accelerograms computed for the two-dimensional model, and PGA(bedrock) and W(bedrock), *i.e.* PGA and W obtained from the accelerograms computed for the model representing the average properties of the source-basin path. The ground motion computed for the different sites in the two-dimensional model is normalized with respect to the ground motion obtained for the same source-receiver distance considering the bedrock model.

The spatial distribution of the values of the relative PGA (PGA(2D)/PGA(bedrock)) and relative Arias Intensity (W(2D)/W(bedrock)) for the transverse component of motion, and for different shear-wave velocities and quality factors of the unconsolidated sediments are shown in fig. 4. The relative PGA increases only slightly when the shear-wave velocity of the sediments is reduced, whereas the relative Arias Intensity is very sensitive to small changes in the shear-wave velocity of the sediments. A low shear-wave velocity induces the largest amplitudes and dispersion of the local surface waves. These effects give the dominant contribution to the relative Arias Intensity at sites where resonance effects and excitation of local surface waves are important (for example at 11.8 km from the source). The relative PGA and relative Arias Intensity have been computed for three different quality factors Q_β of the sediments, by keeping the shear-wave velocity fixed at 0.6 km/s (right part of fig. 4). The attenuation of waves increases with decreasing quality factor. This causes the reduction of the duration of the signals at sites

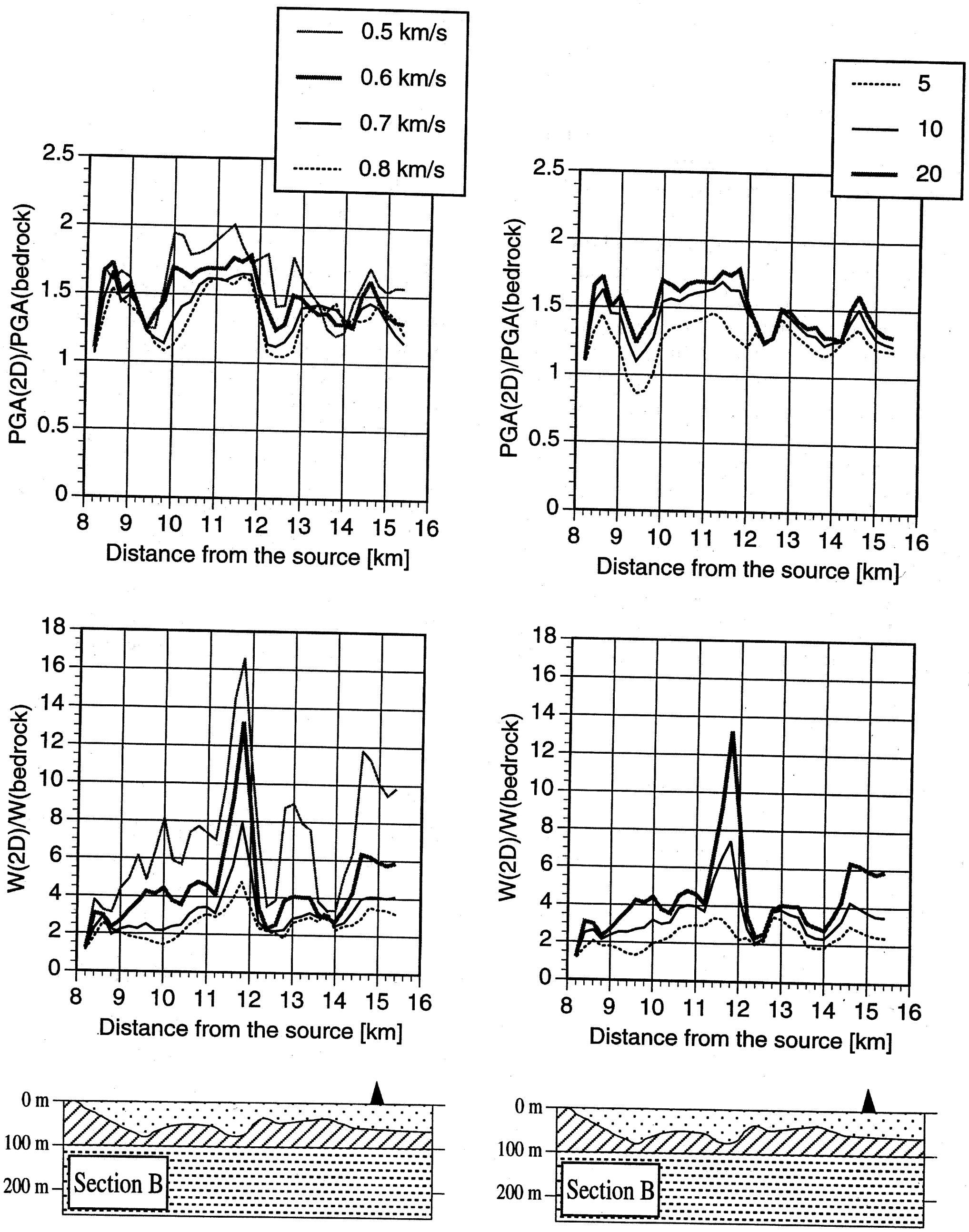

Fig. 4. Relative peak ground acceleration PGA(2D)/PGA (bedrock) and relative Arias Intensity W(2D)/W (bedrock) obtained for cross-section B. The values are shown for four different shear-wave velocities (0.5 km/s, 0.6 km/s, 0.7 km/s and 0.8 km/s), and for three different quality factors ($Q_\beta = 5$, $Q_\beta = 10$, and $Q_\beta = 20$) of the unconsolidated sediments.

where resonance effects and excitation of local surface waves are important phenomena. The smaller the quality factor of the unconsolidated sediments, the shorter is the duration of the resonance, and the smaller the propagation distance of the local surface waves. Consequently, low quality factors strongly reduce the relative Arias Intensity.

In fig. 5a-c, we compare the synthetic signals computed for the two cross-sections B and C, shown in fig. 1, with the accelerograms recorded at Buia station. In correspondence with the recording station, which is at 15 km from the source, the thickness of the sedimentary cover is the same in the two cross-sections.

For the transverse component of motion, the local surface waves have amplitudes that are too large in comparison with the observation. In the choice of the geometry of the bedrock-sediment interface, we have restricted ourselves to that given by Giorgetti and Stefanini (1989). However, to reproduce the observed transverse component, the heterogeneity inside the sedimentary basins, responsible for the excitation of these local surface waves, would have to be different. The considered heterogeneity is either too close to Buia station or the bedrock-sediment interfaces are too close to the free surface. In the radial component the excitation of local surface waves is not very clear. When dealing with *P-SV*-waves, there are two sources of local surface waves: the edge of the sedimentary basin and the places where the bedrock-sediment interface approaches the free surface. From the signals computed for the two cross-sections, it can be concluded that the closer the sediment-bedrock interface is to the free surface, the greater are the amplitudes of the local Rayleigh waves. On the other hand, this lateral heterogeneity can cause the reflection of most of the local Rayleigh waves, generated at the sedimentary basins edge which is closer to the seismic source.

The comparison of the synthetic signals with the observed radial component shows good agreement between the observation and

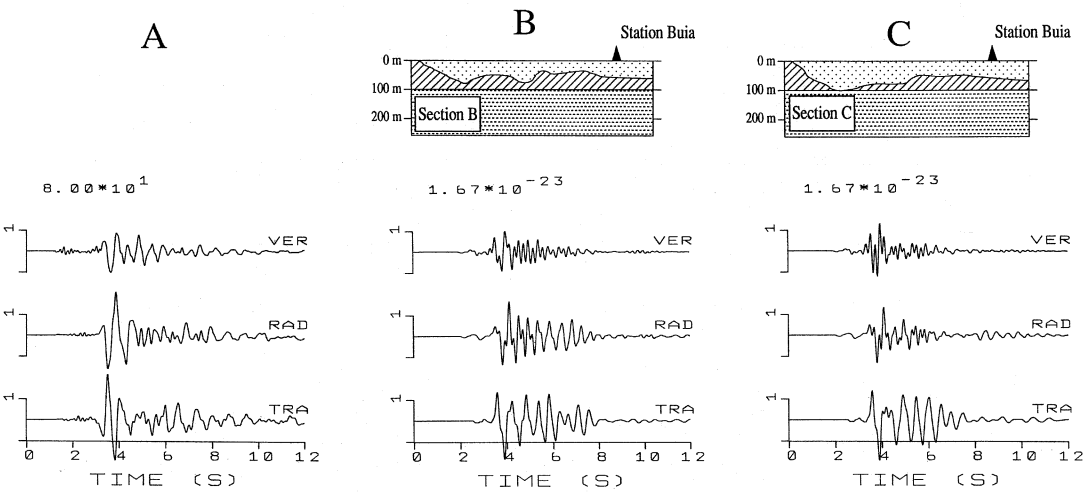

Fig. 5a-c. Comparison between (A) the recorded transverse, radial and vertical components of acceleration, and synthetic signals, computed for (B) the cross-section B and (C) the cross-section C. The time-scale is shifted by 2 s from the origin time. The recorded seismograms are aligned to agree with the synthetic signals. All amplitudes of the synthetic signals correspond to a source with a seismic moment of 10^{-7} N m. The synthetic signals are normalized to the same peak acceleration which is given in units of cm s^{-2}.

the signal obtained for cross-section C. Due to the small amplitudes of the coda in the observation, it can be concluded that the local surface waves have travelled through the deeper parts of the sedimentary basin, and that the lateral heterogeneity within the basin has reflected a relevant part of the local surface waves, which is excited at the edge of the basin. To reproduce the observed signal, the lateral heterogeneity within the basin cannot be strong; a strong heterogeneity, in fact, would excite large-amplitude local surface waves inside the basin, and these are not observed experimentally.

The vertical component of the observed ground motion, especially at low-frequency (below 4 Hz), is quite similar to the synthetic signals, which do not change significantly from one cross-section to the other. The relatively small sensitivity of the vertical component of motion to the lateral variation of sedimentary basins has been observed also at different sites in Mexico City, and the results illustrated in this paper suggest considering this fact a quite general property of sedimentary basins. The high-frequency component, not observed experimentally, but present in the synthetic signals for models B and C, is due to the resonance effects in the shallow part of the sedimentary cover. This difference between the computed and the observed signals indicates once again that, in the modelling, either the shallow parts of the sedimentary cover are too close to the observation point, or the bedrock-sediment interfaces are too close to the free surface.

3. The Rome area and the 1915 Fucino earthquake

The area of Rome, considered here, is characterized by several sedimentary basins of considerable thickness, which, in some parts, are covered by volcanic rocks. The area is very vulnerable to earthquakes, as indicated, for example, by the well documented damage distribution caused by the January 13, 1915, Fucino (Italy) earthquake ($M_L = 6.8$) (fig. 6). Since no strong motion records are available for this event, we have applied the hybrid technique to explain the observed damage distribution (Fäh *et al.*, 1993b). The source parameters, the bedrock model, describing the path from the source to Rome, and the two-dimensional model for the area of Rome are described in Fäh *et al.* (1993b).

To demonstrate that not only the local soil conditions are important to explain a local distribution of damage, but that there are also important regional effects due to the source location, we compare the results of two sets of computations, made by changing only the distance of the seismic source from the city of Rome (fig. 7). The epicenter of the Fucino event is about 85 km east of Rome, and the other hypothetical source, source 2 in fig. 7, is located at a distance of about 65 km, in the same direction.

The synthetic accelerograms are used to compute the quantity W(2D), and the relative Arias Intensity, W(2D)/W (bedrock), of ground motion. The results for the transverse component of motion, corresponding to the two source positions are shown in fig. 8a,b. They are compared with the histogram of the damage distribution, which has been constructed by projecting each observation of damage, shown in fig. 6, on the cross-section used in the numerical modelling (Fäh *et al.*, 1993b). Only those points of the distribution have been used which are located in an area where the geometry of the structure does not differ too much from the geometry of the two-dimensional cross section. This area is delimited in fig. 6 by the two dotted lines. Since neither the type of buildings nor the density of the urbanization can be known in great detail, the histogram shown in fig. 8a,b should be interpreted only in a qualitative manner.

The quantity W, and the relative Arias Intensity, computed for the numerical simulation of the Fucino event are quite well correlated with the damage distribution, as is shown in fig. 8a,b (Fäh *et al.*, 1993b). There are four relative peaks: two at the edges of the Tiber basin, one within the alluvial valley of the Aniene river, and a broad peak where the Sicilian low-velocity zone gets close to the surface. The largest values are observed at the margins

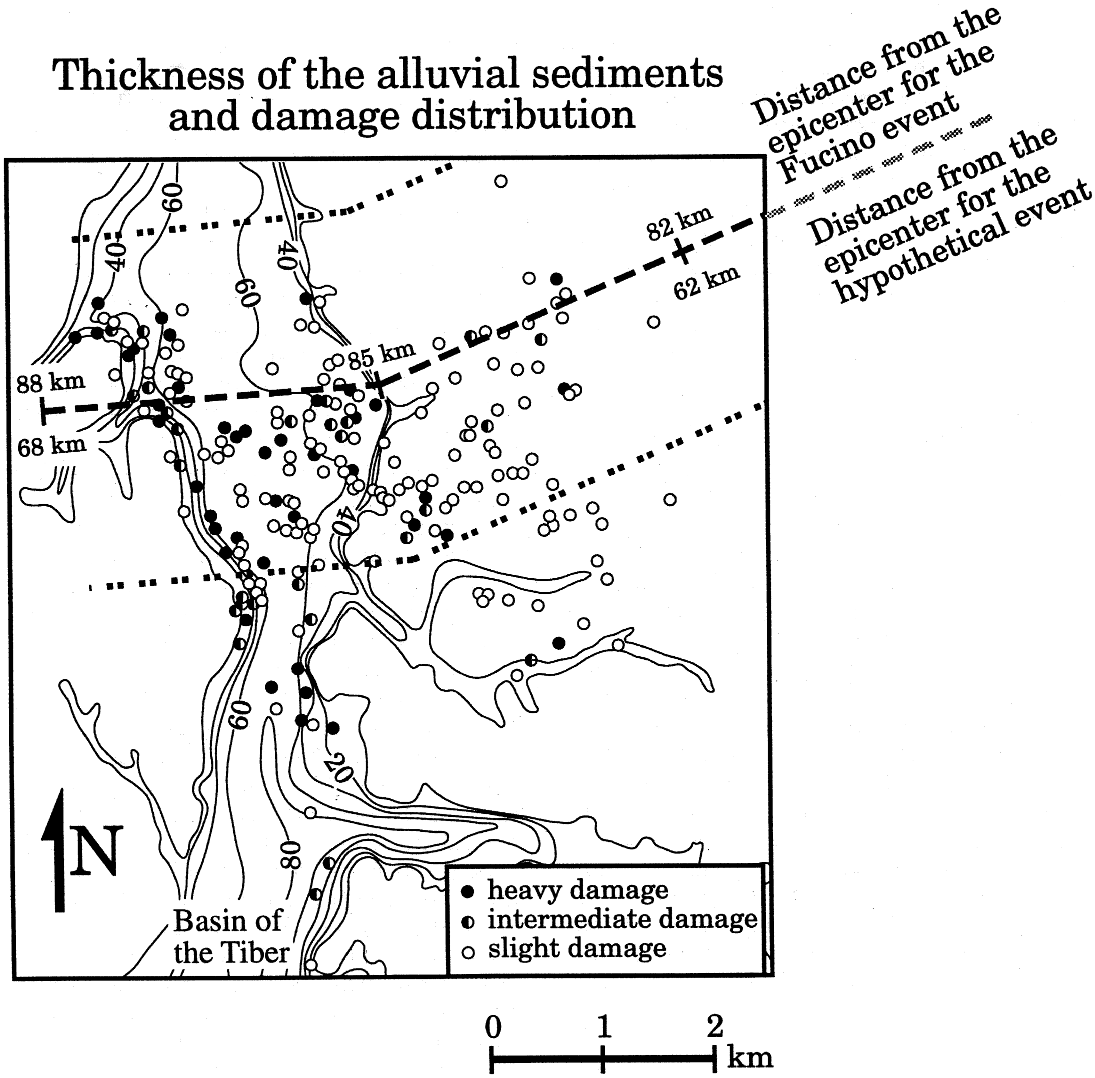

Fig. 6. Damage distribution in Rome caused by the January 13, 1915 Fucino earthquake (after Ambrosini *et al.*, 1986), and thickness of the alluvial sediments (given in meters) (Ventriglia, 1971; Funiciello *et al.*, 1987; Feroci *et al.*, 1990). Three types of damage are distinguished: slight damage (cracking of plaster, the downfall of small pieces of mouldings), intermediate damage (between slight and heavy damage), and heavy damage (deep and diffuse damage of indoor and outdoor walls, downfall of large parts of mouldings and of chimneys). The dashed line indicates the position of the cross section, for which numerical modelling has been performed. The distribution of damage within the area limited by the two dotted lines has been projected on the cross section to construct the histograms, shown in fig. 8a,b.

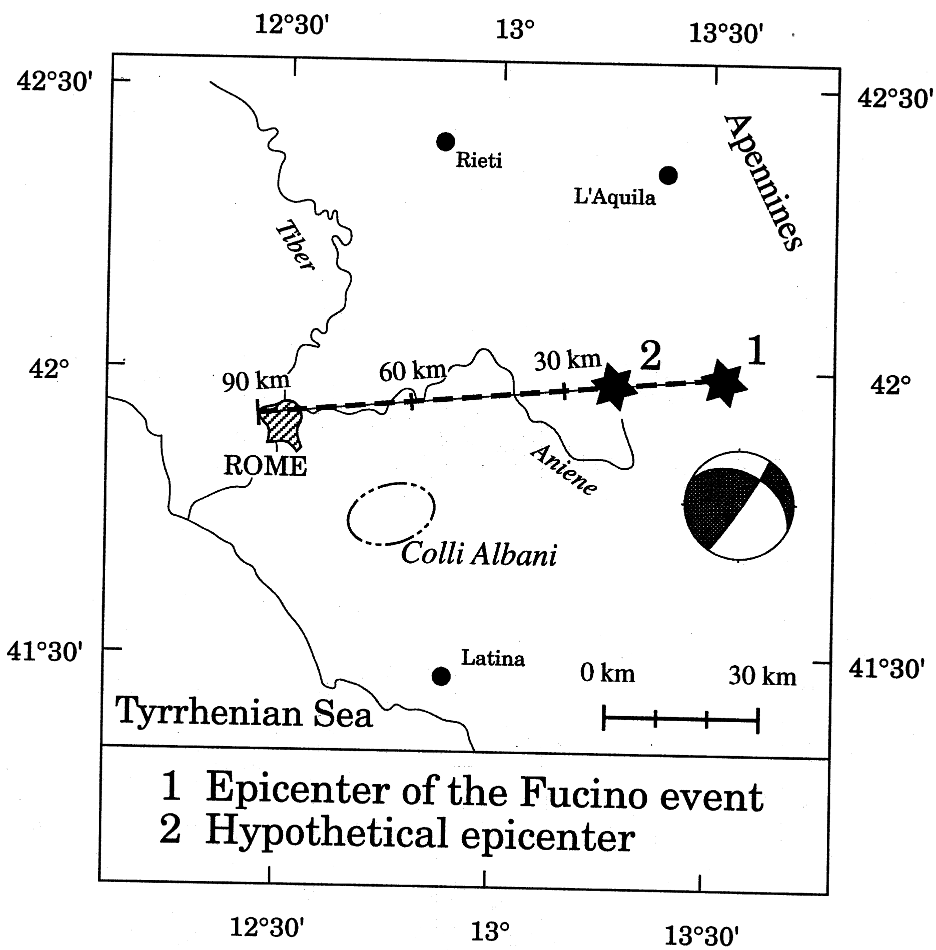

Fig. 7. Approximate epicenter location of the January 13, 1915 earthquake in the Fucino valley (1), and of a hypothetical event (2) located at about 65 km from Rome. The dashed line indicates the cross section along which the numerical modelling has been performed. The source depth is 8 km, the angle between the strike of the fault and the epicenter-station line is 38°, the fault dip 39°, and the rake with respect to the strike 172°. The seismic moment is 10^{19} N m (Fäh *et al.* 1993b).

of the Tiber bed. There the signals have the largest amplitudes and duration, due to the low impedance of the alluvial sediments, the excitation of local surface waves, and resonance effects. Minimum values can be observed for sites placed above the volcanic layer overlying the Paleotiber basin, which acts as a shield, and reflects part of the incoming energy: the thicker the volcanic layer, the smaller are the W values observed at the surface. As is shown in Fäh *et al.*, (1993b), similar conclusions can be drawn for the radial and the transverse components of acceleration, with the difference that the amplitudes of the transverse component are about half the size of those of the radial component. This is due to the *SH* and *P-SV* radiation patterns of the source.

The results of our modelling, obtained considering source 2, are quite different if compared with the modelling of the Fucino event.

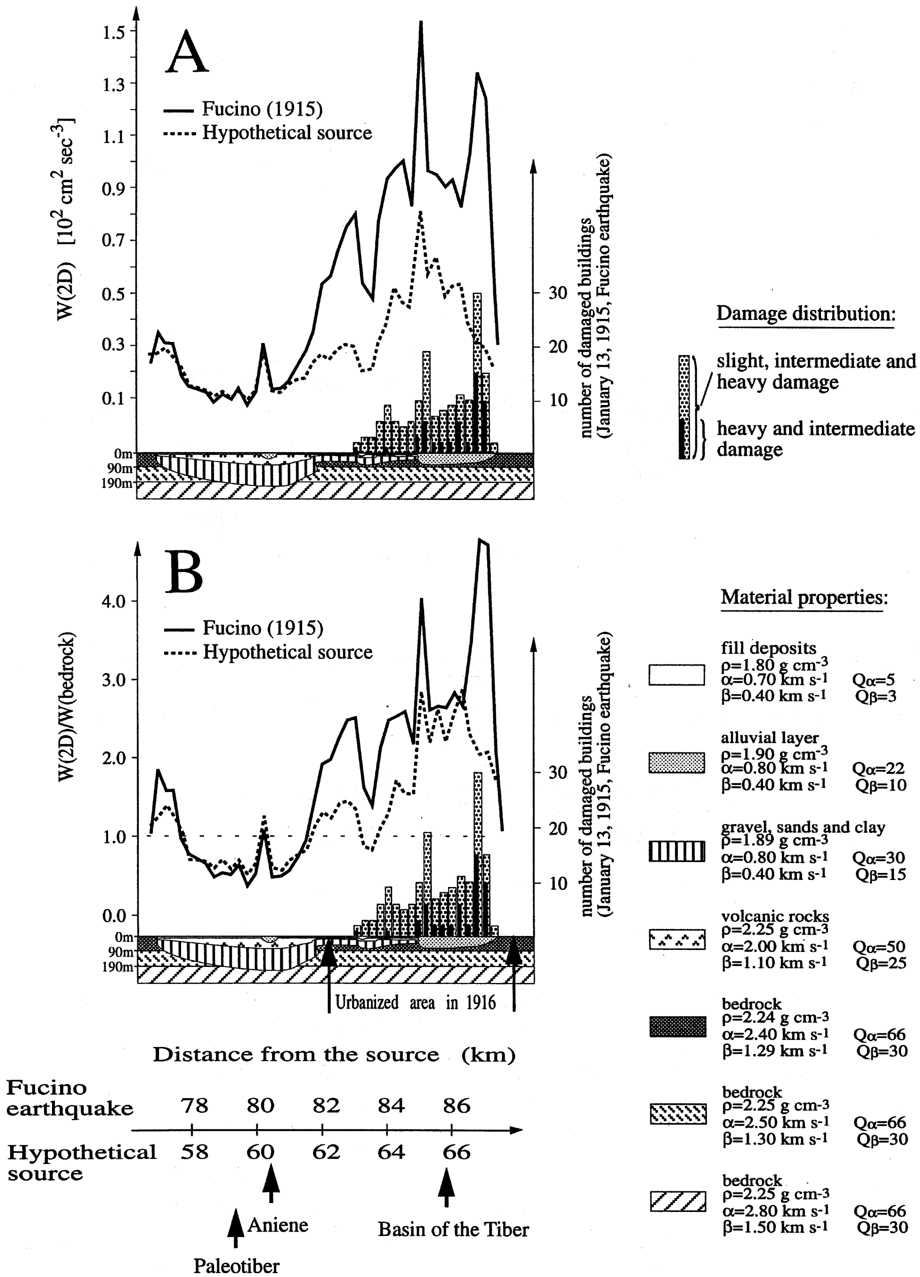

Fig. 8a,b. (A) Quantity W(2D) obtained for the two-dimensional model and the two source positions given in fig. 7, and (B) corresponding relative Arias Intensity W(2D)/W(bedrock). The results are compared with the histogram of the damage distribution caused by the January 13, 1915 Fucino earthquake. The part of the structure near to the surface, where the 2D model deviates from the bedrock model, is given at the bottom of both parts of the figure.

The quantity W and the relative Arias Intensity are similar in the area of the Paleotiber basin, whereas in the other areas the values obtained considering source 2 are considerably smaller, and the two peaks at the margin of the Tiber basin are significantly reduced. A comparison of W, determined from the two numerical experiments, shows that the total energy associated with the Fucino event is significantly larger than the one carried by source 2, even if the former event is 20 km farther from Rome than the latter.

These differences can be explained by the attenuation of PGA and W with distance. The example concerning the transverse component of motion, computed for the bedrock model, is shown in fig. 9a,b. The PGA and W are not monotonically decreasing with distance. This behavior is due to the fact that for epicentral distances less than 50 km, the PGA and W are essentially controlled by the crustal *Sg* phase, while at greater distances they are increasing owing to the contribution of several *S*-wave phases reflected mostly at the Moho (Suhadolc and Chiaruttini, 1985), which gradually become a part of the *Lg* waves and further increase PGA and W.

At distances of the order of 80 km (average distance of the Paleotiber from the Fucino event) and 60 km (average distance of the Paleotiber from the hypothetical source), the W values in our modelling are about the same. This explains the similar values of W in the area of the Paleotiber (fig. 8a,b) computed considering the two events. Similarly, it is possible to explain the values of W observed at distances in the range 61 km-67 km from source 2 (fig. 8a,b), which are smaller than the ones corresponding to the Fucino event in the epicentral distance range 81 km-87 km. Finally from fig. 9a,b, it is evident that, when dealing with earthquakes occurring in the Apennines, for epicentral distances between 50 and 100 km, the largest damage, for a given seismic moment tensor, can be expected from an event as far as 90 km from Rome.

The Fourier-spectrum of the signal computed at 85 km from the source, corresponding to the epicentral distance of the margin of the Tiber basin from the Fucino event, is quite

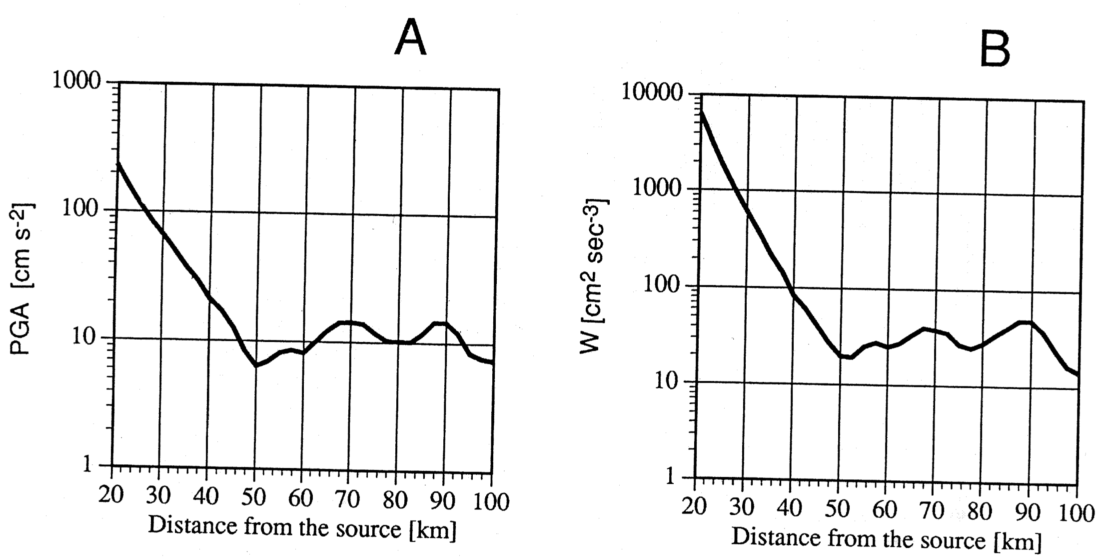

Fig. 9a,b. (A) Attenuation of PGA(bedrock), and (B) attenuation of W(bedrock) with distance from the seismic source for the one-dimensional layered model (bedrock model).

large for frequencies below 2.5 Hz if compared with the spectrum of the signal obtained at 65 km from the source, which corresponds to the epicentral distance of the margin of the Tiber from source 2 (fig. 10). As we will see later, for frequencies below 2.5 Hz strong amplifications occur at the margins of the Tiber basin. Since the incident wavefield computed for source 2 contains relatively small energy at frequencies below 2.5 Hz, resonance effects and excitation of local surface waves are not the dominant phenomena, and this justifies the absence of the peak in W at the margin of the Tiber basin computed for source 2. Thus, for epicentral distances in the range 50 km-100 km, the source location and not only the local soil conditions control the local site effects.

All the quantities used to measure strong ground motion like the maximum amplitude,

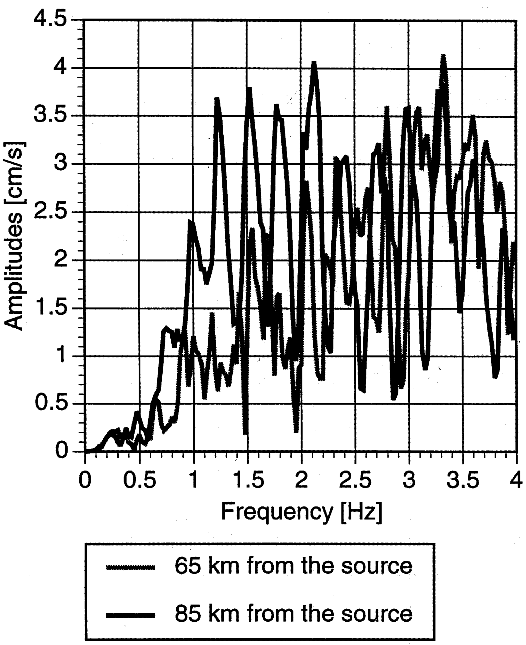

Fig. 10. Fourier-spectra of the signals obtained at 65 km and 85 km from the source for the one-dimensional layered model. The distances correspond to the distance of the eastern margin of the Tiber basin from source 2 and from the Fucino earthquake.

the duration and the Fourier spectrum provide only a very limited description of the ground motion and certainly do not quantify its damage producing potential. A better quantity is the spectral acceleration Sa of the earthquake ground motion. A representation of the local soil effects is given by the spectral amplification $Sa(2D)/Sa(\text{bedrock})$ computed from the spectral accelerations obtained for the two-dimensional and the bedrock models. This procedure allows us to identify the frequency bands and sites at which amplification and attenuation effects occur. For SH-waves, the spectral amplification for zero damping are shown in fig. 11a,b for the two simulated events, as a function of frequency and of the spatial location along the section. The darker an area is, the stronger the amplifications due to the two-dimensional effects. The greatest amplification is observed for the numerical simulation of the Fucino event at the western edge of the sedimentary basin of the Tiber river (about 87 km from the source), for frequencies around 2 Hz. The maximum amplification is of the order of 5-6, and it is due to the combination of resonance effects and the excitation of local surface waves. This amplification effect is responsible for the relative peak in the quantity W at the margin of the Tiber basin (fig. 8a,b).

The global distribution of the shaded areas can be related to the geometry of the structural model. The results are similar for the Fucino event and for source 2, except at the margins of the Tiber basin. An amplification over almost the entire frequency band is observed outside the Paleotiber basin (82-87 km from the source for the Fucino event, 62-67 km from source 2). Some amplification occurs in the Aniene basin, for frequencies above 2 Hz. For frequencies above 0.8 Hz, in the Paleotiber basin, the volcanic layer acts as a shield reflecting part of the incoming energy, and the values of the spectral amplification are smaller than 1. The underlying sedimentary complex (Sicilian) causes spectral amplification of the order of 2-3, due to resonances, which are most pronounced at frequencies around 0.4 Hz, where the fundamental resonance of this low-velocity zone is excited. In this part of the Paleotiber, in the frequency band 1.5-2.0 Hz,

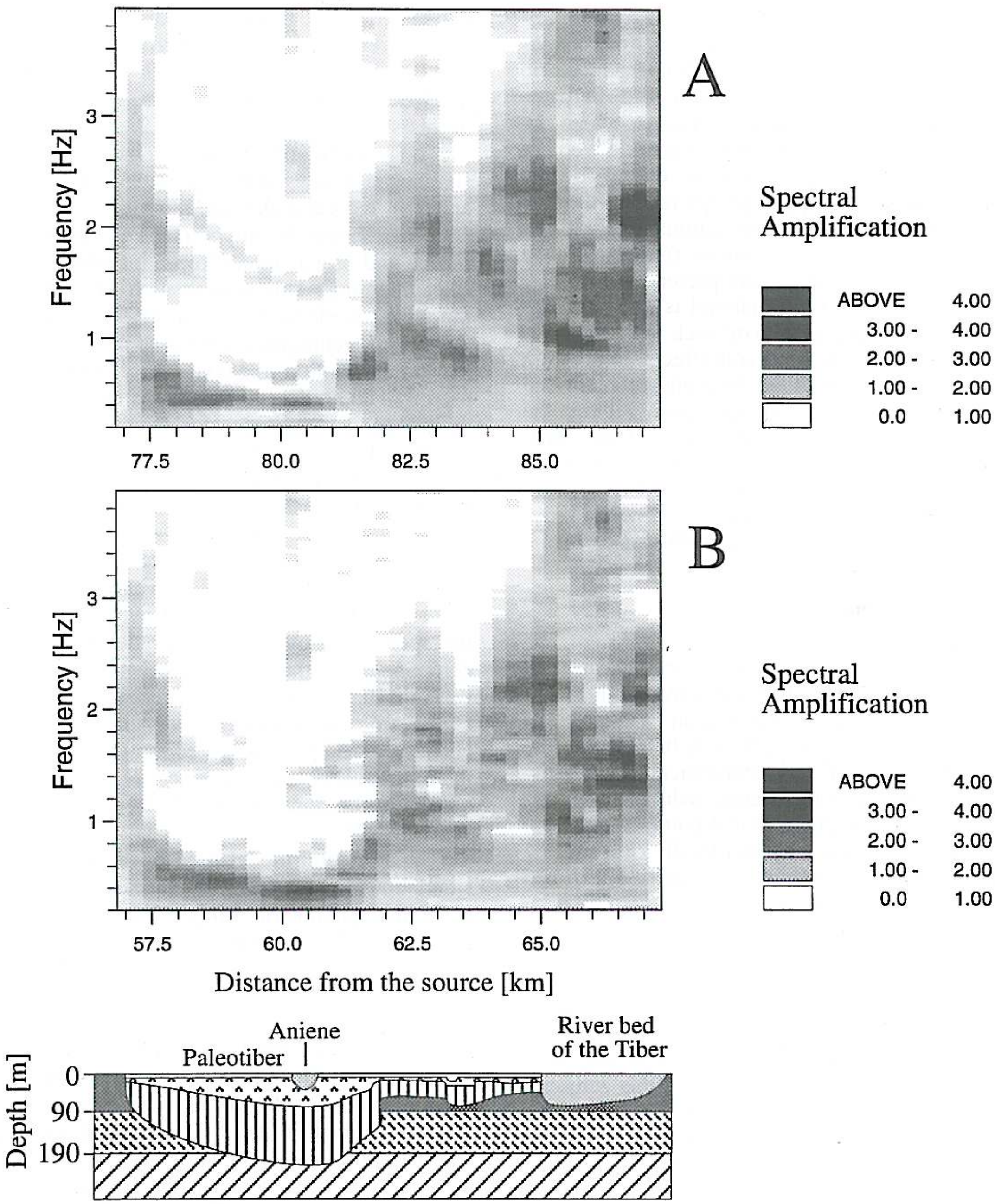

Fig. 11a,b. Spectral amplification for the transverse component of motion over the entire cross section, (A) for the Fucino event and (B) for the hypothetical earthquake. The reference signals are the one obtained for the one-dimensional layered model. At the bottom of the figure, the geometry of the two-dimensional cross-section is given, while its mechanical parameters are given in fig. 8a,b.

there is also evidence for the excitation of some higher modes of resonance. At distances of the order of 82-83 km from the Fucino event, and 62-63 km from source 2, between the Paleotiber and the Tiber basins, where the wave guide and overlying volcanic cover are thinning, focusing of seismic energy occurs and most of the trapped energy reaches the surface. This leads to amplifications, of the order of 2, over almost the entire frequency band considered. Therefore, the presence of a near-surface layer of rigid material is not sufficient to classify a site as a «hard-rock site». Reliable determinations of local soil effects, in addition to the knowledge of the frequency content and direction of the incoming signal, require the knowledge of both the thickness of the surficial layer and of the deeper parts of the structure, down to the real bedrock. This is especially important in volcanic areas, where pyroclastic material often covers alluvial basins.

4. Mexico City

Mexico City is an area of particular interest since extensive damage occurred in the lake-bed zone during the Michoacan earthquake of September 19, 1985. This can be attributed to the geotechnical and geometrical characteristics of the unconsolidated sediments in this zone. From the geotechnical point of view, the valley of Mexico City can be divided into the hill zone, the transition zone, and the lake-bed zone (fig. 12). The hill zone is formed by alluvial and glacial deposits, and by lava flows. The transition zone is mainly composed of sandy and silty layers of alluvial origin. The surficial layers in the lake-bed zone consist mainly of clays. These deposits are poorly consolidated, with a high water content and very low rigidity. The thickness of this surficial layer varies between 10 m and 70 m, and increases regularly towards the east (Suarez *et al.*, 1987). The clay layer is overlying the so-called «deep sediments» found below 10-70 m. These deeper deposits reach depths of the order of 700 m, with an uncertainty which may be as large as a few hundred meters (*e.g.* Bard *et al.*, 1988). There are three outcrops of the basement: at Chapultepec, Peñon, and Cerro de la Estrella (fig. 12). In the last years, a strong motion network has been operating in the valley of Mexico City (*e.g.* Mena *et al.*, 1986; Espinosa *et al.*, 1990), and the positions of the stations used in this study are shown in fig. 12.

Today's research, to understand the extensive damage caused by the 1985 earthquake and the recorded ground motion in Mexico City, has shown the importance of considering source and propagation effects, including local soil conditions. This is dealt with explicitly in the recent work by Fäh *et al.* (1994) in which the hybrid technique has been applied to study the ground motion in Mexico City. The model used by Fäh *et al.* (1994) explains the observed difference in amplitudes for receivers located inside and outside the lake bed zone. The ratio between the computed, horizontal peak ground displacements inside and outside the lake-bed zone reaches values ranging from 5 to 7, and about the same ratio is obtained for the observed ground motion. The validity of the modelling is further confirmed by the fact that the spectral ratios obtained for the horizontal components of the synthetic seismograms are very similar to comparable spectral ratios obtained from observations. For the 1985 Michoacan event, the energy contributions of the three subevents are important to explain the observed durations in the lake-bed zone (Fäh *et al.*, 1994).

The structural model used in the numerical simulations for the 1985 Michoacan event, and a detailed parametric study of the effects of different soil properties in Mexico City is given in Fäh *et al.* (1994). The flat-layered structure in table I describes the path from the seismic source to the valley of Mexico City (Campillo *et al.*, 1989), and was deduced directly from refraction measurements in the Oaxaca, Southern Mexico region (Valdes *et. al.*, 1986). The depth of the Moho is about 45 km, and the upper five kilometers are composed of low-velocity material. The two-dimensional structure for the Mexico City valley, modelling the Chapultepec-Peñon cross-section (solid line in fig. 12), is shown in fig. 13. The structure is rather simple, in agreement with the resolving power of the available data.

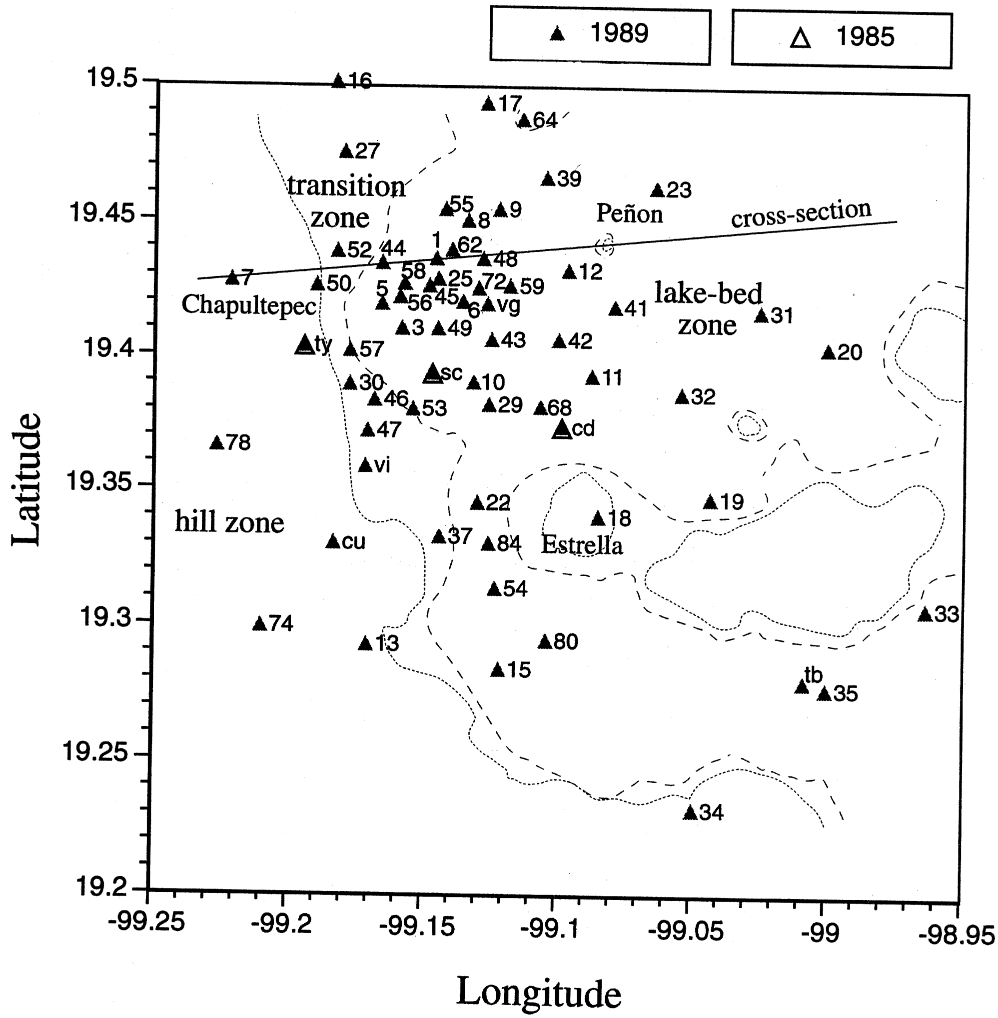

Fig. 12. Map of the area of Mexico City showing the locations of strong motion accelerometric stations. The stations represented by large triangles were operating during the 1985 Michoacan event, while the small gray triangles represent the stations which recorded the April 25, 1989 event. The solid line indicates the position of the cross-section, for which the 2D modelling was performed.

As in Fäh *et al.* (1994), to keep the source model as simple as possible and to avoid any, *a priori*, enhancement of resonance effects at the longer periods (above 2 s), we consider first a constant, frequency independent, seismic moment rate spectrum. The focal mechanism is the one proposed by Campillo *et al.* (1989), based on the results of Houston and Kanamori (1986) and Riedesel *et al.* (1986). The distance from the source to the valley of Mexico City is 400 km, the angle between the strike of the fault and the epicenter-station line is 220°, the source depth is 10 km, the dip 15°, and the rake is 76°.

Table I. Numerical parameters for the schematic crustal model describing the path from the source in the Michoacan subduction zone, to Mexico City (Campillo *et al.*, 1989).

Layer	Thickness (km)	ρ (g/cm³)	α (km/s)	β (km/s)	Q_α	Q_β
1	5.0	2.67	4.30	2.53	800	500
2	10.0	2.77	5.70	3.30	800	500
3	15.0	3.09	6.80	4.03	800	500
4	15.0	3.09	7.00	4.10	800	500
5	∞	3.30	8.20	4.82	800	500

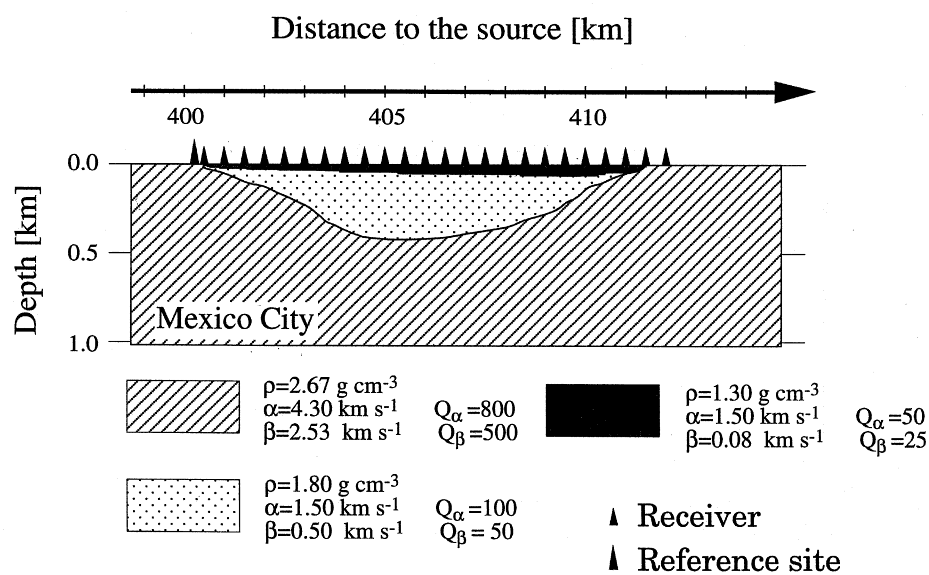

Fig. 13. Two-dimensional model of the Chapultepec-Peñon cross-section. Only the part of the structure near to the surface is shown, where the 2D model deviates from the bedrock model (table I).

The use of an instantaneous time-function gives rise to synthetic signals that contain too much energy at high frequency (above 0.6 Hz), and we can remove this drawback applying the ω^2 scaling law for the seismic moment rate spectrum, proposed by Kanamori *et al.* (1993) for the events occuring in the Mexican subduction zone:

$$\hat{M} = M_0\, \omega_c^2\, (\omega^2 + \omega_c^2)^{-1} \qquad (4.1)$$

where $\hat{M}$ is the seismic moment rate spectrum, M_0 is the seismic moment, and ω_c is the corner angular frequency. Following Kanamori *et al.* (1993), $\omega_c = 0.196\ s^{-1}$ and $M_0 = 0.5\ 10^{21}$ Nm. With these values we obtain a good reproduction of the shape of the observed signals at station CD (fig. 14a-c), especially for the radial and vertical component of motion (fig. 14b). The absolute observed accelerations are underestimated (fig. 14b). This discrepancy can be

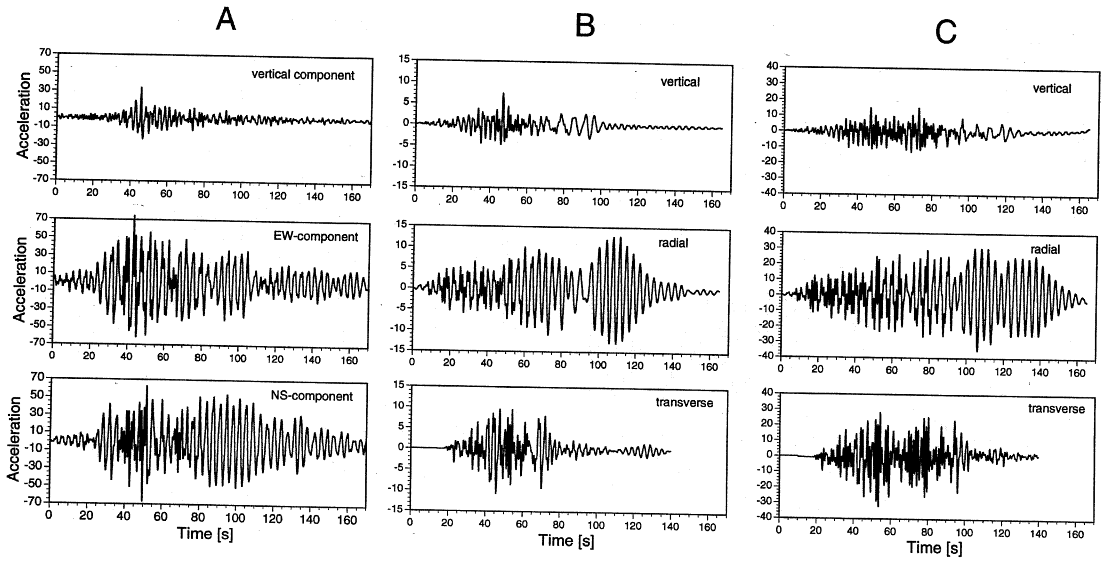

Fig. 14a-c. Comparison between (A) the recorded ground motion at station CD for the 1985 Michoacan earthquake with (B, C) synthetic signals, computed at the eastern edge of the basin, 409.5 km from the source (see fig. 13). All signals are low-pass filtered with a 5-pole butterworth filter having its corner frequency at 1 Hz. The accelerations are given in units of cm s^{-2}. A) Observed horizontal and vertical components of acceleration at station CD. Both horizontal components of motion are characterized by *SH*-waves as well as *P-SV* waves (see text); B) synthetic accelerations due to one point source. The synthetic signals are scaled assuming the seismic moment rate spectrum proposed by Kanamori *et al.* (1993); C) synthetic accelerations due to three point sources, all located at a depth of 10 km and the same distance. The three point sources have different weights and time shifts (1.0, 1.0, 0.2 and 0 s, 26 s, 47 s). The synthetic signals are scaled assuming the seismic moment rate spectrum proposed by Singh *et al.* (1990) for the 1985 Michoacan event.

reconciled considering the errors affecting the estimates of ω_c and M_0, and taking into account the subsequent rupture episodes of the Michoacan event, mainly the one occurring about 26 s after the origin time. If we assume a seismic source that is composed of three subevents, as proposed by Houston and Kanamori (1986), with the seismic moment rate spectrum proposed by Singh *et al.* (1990), the durations increase by about 45 s (fig. 14c) compared to those relative to a single event (fig. 14b). The computed radial component of motion (in fig. 14c) is very similar to the observed horizontal ground motion (fig. 14a). The underestimate in amplitude of the synthetic signals, and the remaining difference between synthetics and observations, is due to the simple source model in our modelling, the absence of dynamic irregularities in the rupture process, and the fact that only one cross-section of the sedimentary basin is considered.

Spectral amplification at the site of interest computed with respect to a reference site gives a good representation for micro-zoning purposes, especially from the engineering point of view (Fäh and Suhadolc, 1994). We compute the spectral amplification, *i.e.* the relative spectral velocities $Sv(2D)/Sv(Ref)$, for zero damping and 5% damping. $Sv(2D)$ is the spectral velocity obtained for the receivers in the two-dimensional structural model shown in fig. 13. $Sv(Ref)$ is the value obtained for the reference station shown in fig. 13. For all the receivers, we have computed the relative spectral velocities for one hundred frequencies of the oscillator in the range 0.1-1.0 Hz. From these values

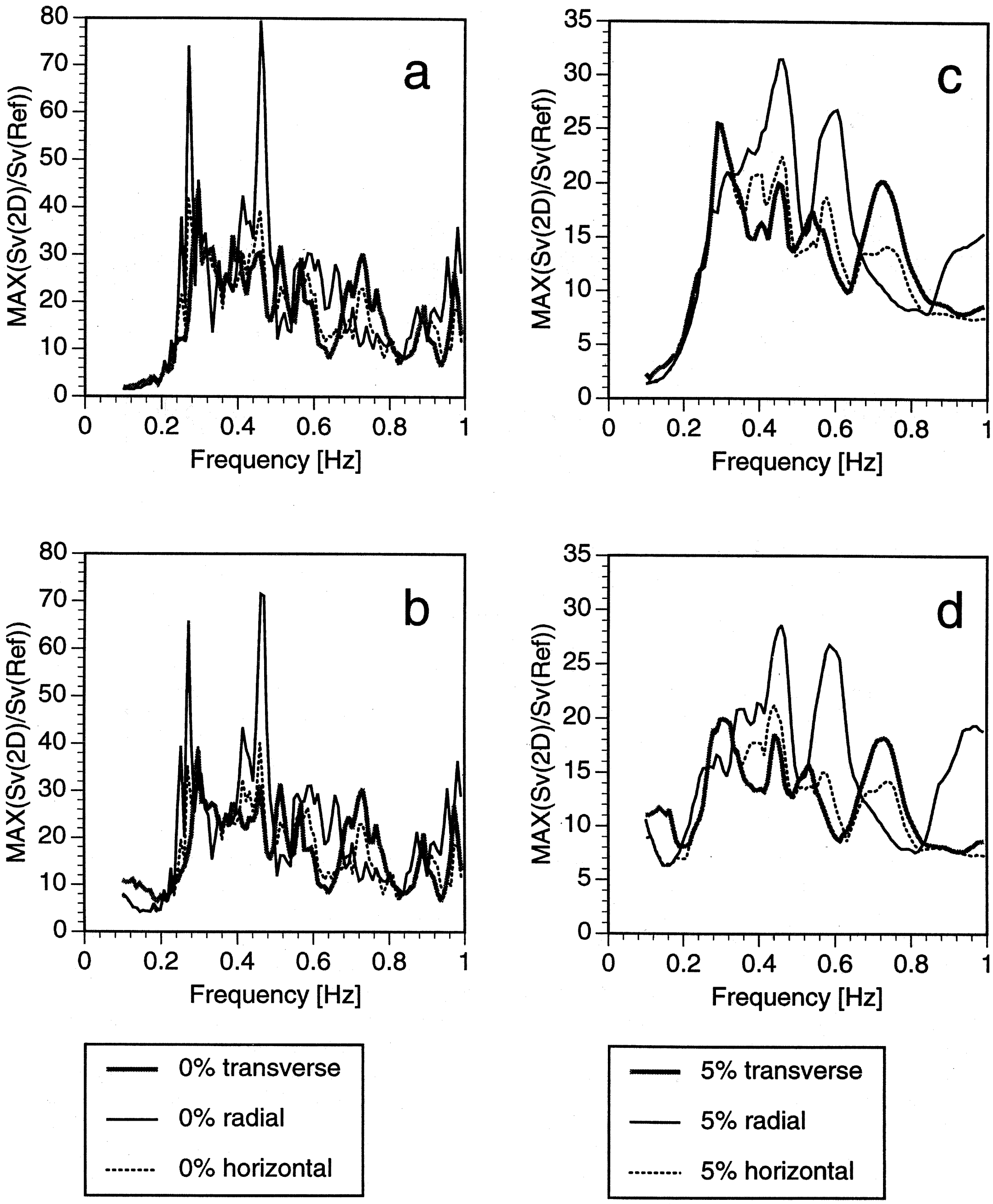

Fig. 15a-d. Maximum relative spectral velocities *Sv*(2D)/*Sv*(Ref), for zero damping (a,b) and 5% damping (c,d), obtained with the synthetic signals, which are scaled by assuming the seismic moment rate spectrum proposed by Kanamori *et al.* (1993) (a,c), and by assuming a constant, frequency independent, spectrum (b,d).

we then computed the maximum spectral amplification (MSA) for the entire sedimentary basin. The MSA obtained from our numerical simulations are shown in fig. 15a-d.

The MSA obtained from the numerical simulation is rather independent from the shape of the seismic moment rate spectrum, as can be seen in fig. 15a-d, where the results obtained with the seismic moment rate spectrum proposed by Kanamori *et al.* (1993) are compared with the results obtained for a frequency-independent spectrum.

In two-dimensional modelling, the *SH*-waves (transverse component of motion) and *P-SV*-waves (radial component of motion) are two independent wave fields. The direction of propagation of the local surface waves in the sedimentary basin is therefore well correlated with the source-receiver direction. In Mexico City, the local surface waves are excited at different locations of the interface between the bedrock and the sediments, and the direction of propagation of the local surface waves in the lake-bed zone no longer corresponds to the source-receiver direction (F.J. Sánchez-Sesma, personal communication). Therefore, both horizontal components of motion at station CD (fig. 14a) are characterized by *SH*-waves as well as *P-SV* waves. To justify the comparison between the observed ground motion and the synthetic signals, we have to compute the spectral amplification for the complete horizontal ground motion. For this purpose, we have applied the horizontal ground motion to an oscillator with two-degrees of freedom. The results for the synthetic signals are also given in fig. 15a-d. The maximum peaks in the MSA in fig. 15a-d can be attributed to sites where a strong interaction between the deep sediments and the surficial clay layer occurs. At such sites, the resonance frequencies of the two layers are almost the same.

The seismograms observed during the 1985 Michoacan earthquake, have been convolved with a high-pass Ormsby filter (Mena *et al.*, 1986), and this filtering allows us an estimate of MSA only for frequencies above 0.1-0.2 Hz. For the Michoacan earthquake we have usable signals only from the stations CD, SC and TY, which are not located on a hard rock site (*e.g.* Ordaz and Faccioli, 1994). Therefore, we have to test our theoretical results using observations from other events, such as the April 25, 1989 earthquake.

The epicenter of the 1989 event is located in the Guerrero gap, *i.e.* south of the epicenter of the 1985 Michoacan earthquake. Therefore, to justify a comparison between the numerical results based on the modelling of the 1985 event and the ground motion observed during the 1989 earthquake, we first have to compare the results obtained from the analysis of the records of the 1985 and 1989 earthquakes. The spectral amplification is shown in fig. 16a-d, for station CD with respect to TY, both for the 1985 and for the 1989 earthquake. For the two events, the amplification effects at station CD occur in about the same frequency ranges. There is some difference in the maximum value, which is larger for the 1985 earthquake. The strong peaks in the spectral amplification observed at the station CD, during the 1985 earthquake, can be due to the fact that a relevant amount of energy of the incident wave field is present in the frequency band close to the resonance frequency of the clay layer at the site, and, therefore, local surface waves and resonance effects may dominate the horizontal components of motion. The two peaks in the MSA can be attributed to the deep sediments (peak at about 0.8 Hz) and the surficial clay layer (peak at about 0.25 Hz) (Fäh *et al.*, 1994).

The observed MSA for the 1989 earthquake is shown in fig. 17a-d. The MSA is determined for all the stations shown in fig. 12 using as reference station TY and station 78. The MSA obtained with the reference station TY (fig. 17a,c) has lower values than those computed theoretically (shown in fig. 15a-d). This is due to the fact that the site of station TY is not a hard rock site (*e.g.* Ordaz and Faccioli, 1994). The MSA obtained with the reference station 78 (fig. 17b,d), on the other hand, is very similar to the theoretical curves shown in fig. 15a-d. The peaks, for zero damping of the oscillator, occur when the energy of the incident wave field is relevant in the frequency band close to the resonance frequency of the clay layer at the site of interest. This happens in dif-

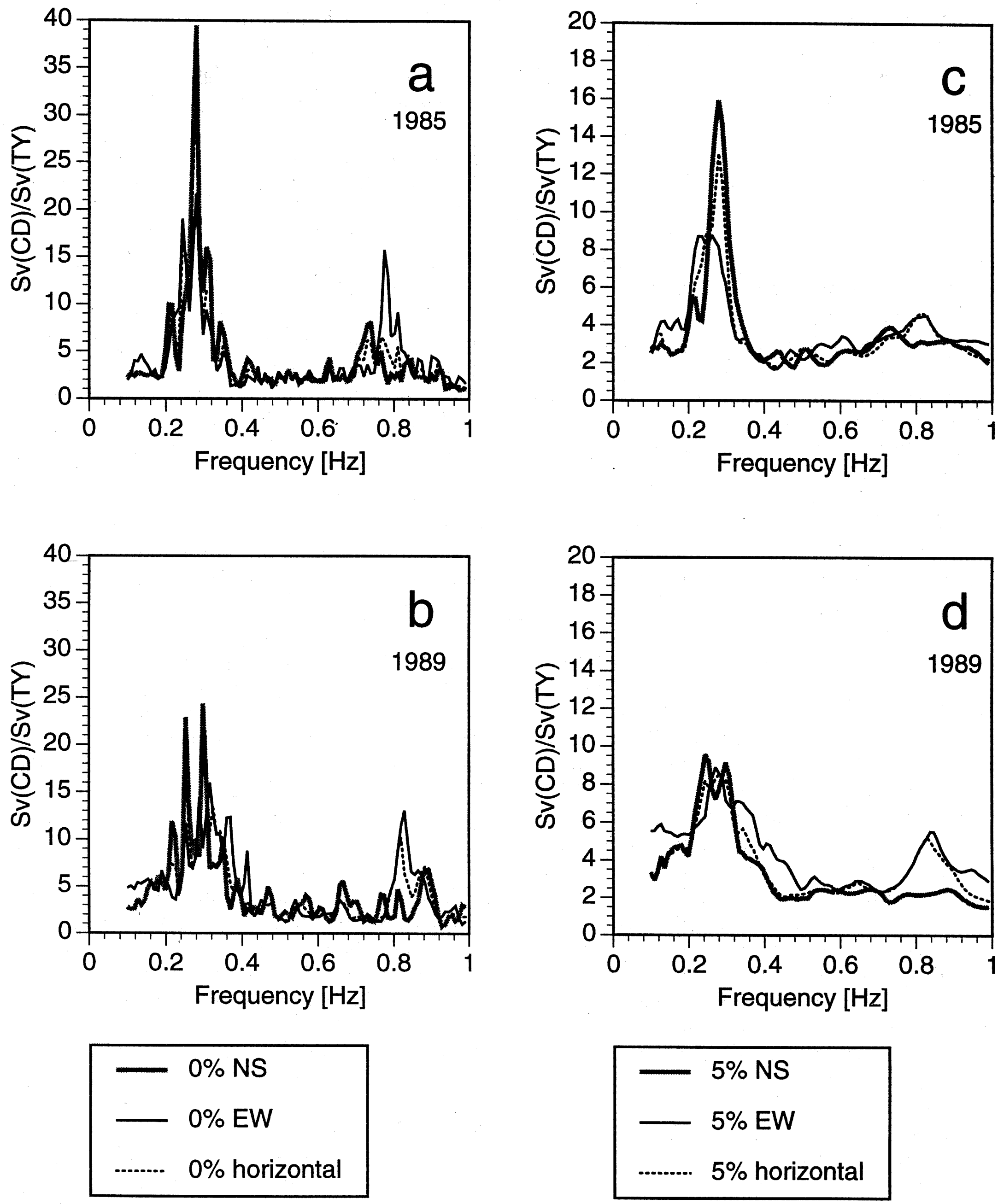

Fig. 16a-d. Relative spectral velocities Sv(CD)/Sv(TY), for zero damping (a,b) and 5% damping (c,d). They are determined from the stations CD using TY as reference. Results are shown for the 1985 Michoacan earthquake (a,c) and the April 25, 1989 event (b,d).

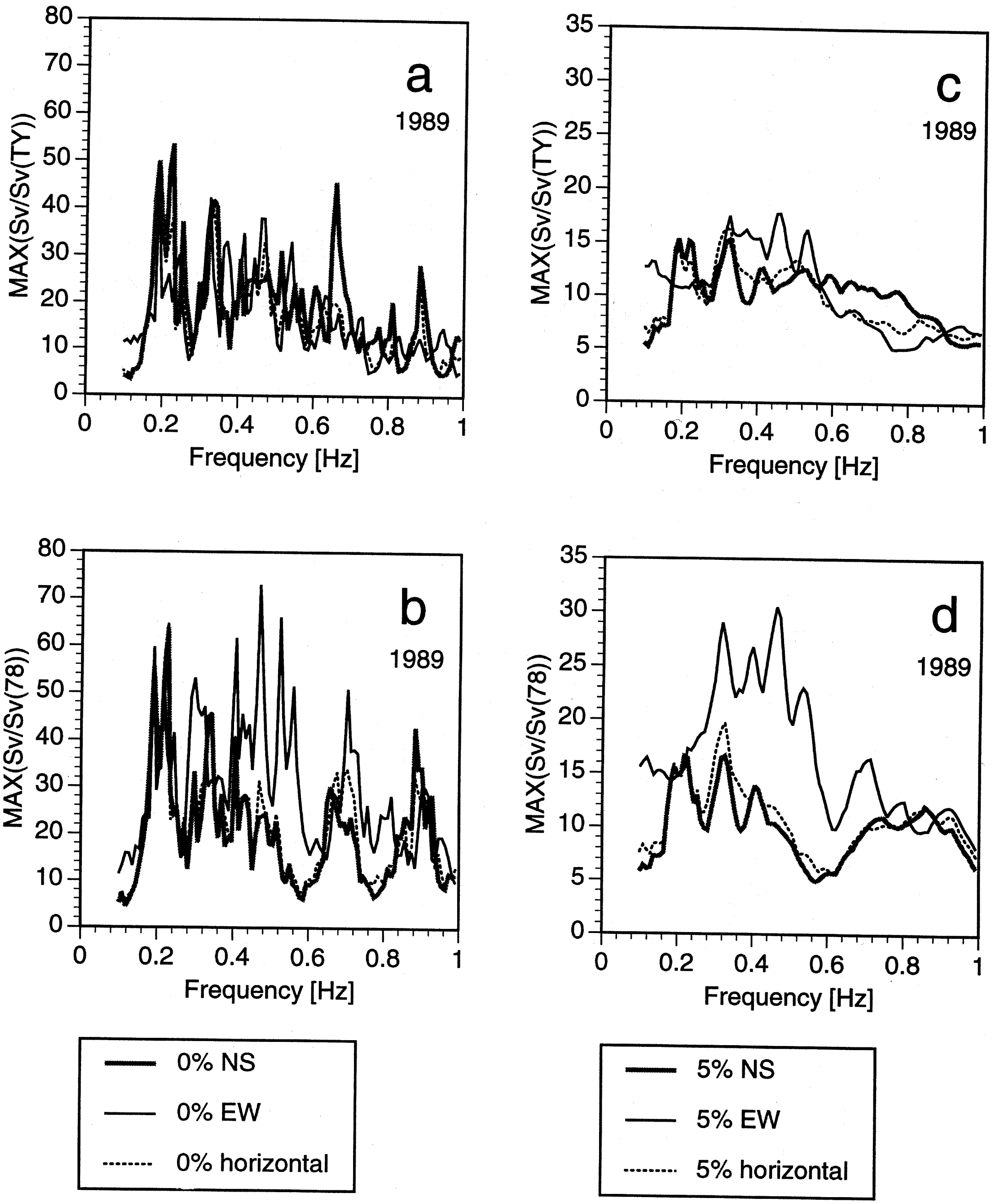

Fig. 17a-d. Maximum relative spectral velocities *Sv*/*Sv*(TY) and *Sv*/*Sv*(78) observed during the 1989 event, for zero damping (a,b) and 5% damping (c,d), obtained for all stations shown in fig. 12, using as reference station TY (a,c) and station 78 (b,d), respectively.

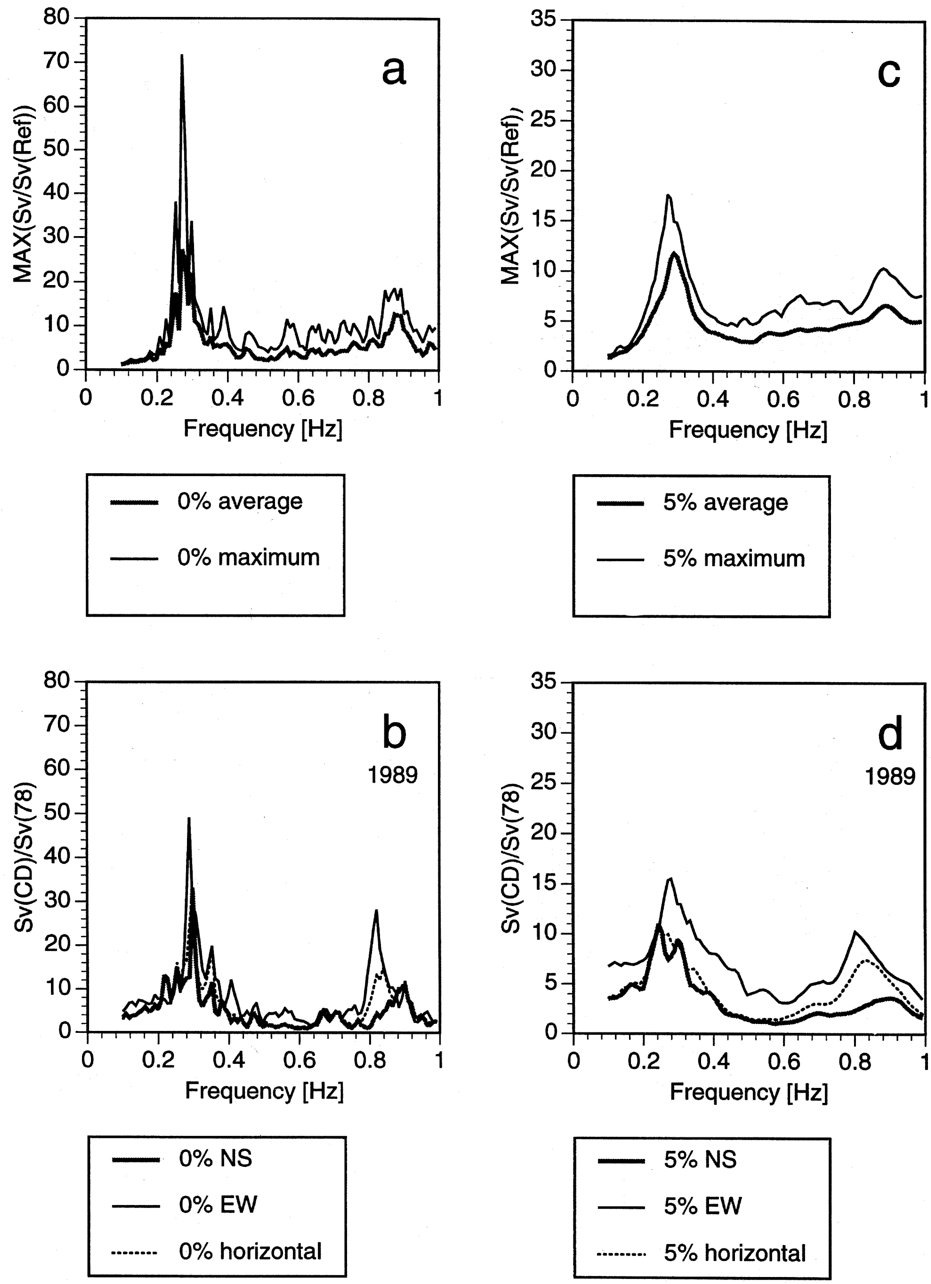

Fig. 18a-d. Average and maximum relative spectral velocities for zero damping (a), and 5% damping (c), computed with the synthetic signals obtained for the receivers at distances between 408.5 and 410 km from the source. They are compared with the observed relative spectral velocities *Sv*(CD)/*Sv*(78) for the 1989 earthquake, for zero damping (b), and 5% damping (d), determined from the station CD using station 78 as reference.

ferent sites within the lake-bed zone, and therefore each peak in the MSA (fig. 17b) is obtained from a different receiver. The maximum values obtained theoretically and from the observations are about the same, and the greater complexity in the observed MSA for zero damping, with respect to the theoretical one, which has only two peaks around 0.27 Hz and 0.47 Hz (fig. 15a,b), is not surprising since our numerical model is restricted to one cross-section. In our model the maximum clay thickness is 65 m and, therefore, the MSA shown in fig. 15a-d is valid only for sites in the lake-bed zone that are characterized by a clay layer with a thickness not exceeding this value.

For 5% damping, the theoretical (fig. 15c,d) and the observed MSA (fig. 17d) are very similar, both in shape and in maximum values, and the two-dimensional model under study can be considered representative for the general geological situation in Mexico City. This is a valuable contribution for micro-zoning purposes, and it permits for each site, where the stratigraphy is reasonably well-known, a realistic estimate of the maximum and average spectral amplification with respect to a bedrock site. An example of the theoretical prediction, based on our results, appropriate for a site similar to the one of station CD (see fig. 12), is shown in fig. 18a,c: the theoretical MSA values are quite satisfactorily compared with the observations made at the station CD in connection with the 1989 event (fig. 18b,d).

5. Conclusions

The hybrid technique presented in this study makes it possible: 1) to study local effects even at large distances (hundreds of kilometers) from the source; 2) to include the seismic source, and 3) the propagation path. This technique can assist in the interpretation and prediction of ground motion at a given site. It can be applied in (micro-)zonation studies, and provides realistic estimates of spectral amplifications for detailed two-dimensional, anelastic models.

The synthetic signals explain the major characteristics (relative amplitudes, spectral amplification, frequency content) of the considered seismograms, and the space distribution of the available macroseismic data, even when quite simple source models are assumed. The theoretical computations show that waveforms and frequency content of seismograms are sensitive to small changes in the subsurface topography of the sedimentary basin, the velocity and quality factor of the sediments. The absence of dynamic irregularities in the rupture process in our simulations, in general, causes only an underestimate of the absolute accelerations.

To achieve a realistic simulation of seismic ground motion, it is necessary to include source, path and local soil effects, to study both *SH* and *P-SV* wave propagation, and to consider anelastic absorption. The reasons for the damage caused by the Michoacan earthquake and the Fucino event can be found not simply in the local site conditions, but also in the effects of the seismic source and of the long-distance propagation path in the crust. The dynamic irregularities of the rupture process and the properties of wave propagation in the crust control the frequency content of the incident wavefield; if a relevant amount of energy is present in the frequency band close to the resonance frequency of the unconsolidated sediments, local surface waves and resonance effects may dominate the horizontal components of motion within the sedimentary basin.

One aspect which is not included in our discussion is the influence of surface topography on ground motion. This approximation can be justified for sites inside sedimentary basins, where topographic features are in general small. However, topography can become important at the edges of sedimentary basins and near outcrops, especially in mountainous regions (Fäh and Suhadolc, 1994).

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