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REGIONAL AMPLIFICATION OF GROUND MOTION IN CENTRAL MEXICO. CONSTRAINTS FROM OBSERVATIONS MODELLING

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ABSTRACT

Seismic ground motion in Central Mexico is amplified relative to ground motion observed at the same epicentral distance along the Pacific Coast, in a frequency band that includes destructive ground motion at Mexico City. To date there is no explanation of such amplification. We have analyzed three different datasets with the purpose of increasing our understanding of the spatial distribution of this phenomenon: 1) three-component seismograms obtained during a large scale refraction experiment; 2) acceleration records of the 10.24.93 (Ms=6.6) event that occurred in the subduction zone; and 3) magnitude data reported by Servicio Sismológico Nacional for events recorded during 1993. Our results show that, for distances larger than 100 km, ground motion is amplified during propagation perpendicular to the coast. For periods smaller than 8 s, surface wave dispersion varies significantly from the average model between the coast and Mexico City. The more likely cause for regional amplification is lateral crustal heterogeneity (both in structure and velocity distribution) related to the Transmexican Volcanic Belt.

KEYWORDS

Regional amplification, path effects, velocity data, magnitude residual, dispersion analysis, finite difference, SH waves.

INTRODUCTION

Mexico City may be strongly affected by earthquakes occurring in the subduction zone of the Pacific, more than 300 km away. A major factor is amplification of ground motion generated by a very soft clay layer that covers the lake bed zone. However, an additional significant contribution is ground motion amplification observed on "firm ground" in Mexico City (Singh *et al.*, 1995). This regional amplification was demonstrated by Ordaz and Singh (1992) who computed attenuation curves from acceleration records from 8 large events. They showed that ground motion in central Mexico is amplified by a factor up to 10 in the frequency band 0.2 to 1 Hz, relative to average attenuation curves. Such large amplification was confirmed by similar analysis using data recorded during a large scale refraction experiment (Cárdenas, 1993).

Significant regional amplification is well documented by independent observations. However, no explanation exists, although several hypothesis have been advanced: very large sedimentary valleys (Ordaz and Singh, 1992), lateral heterogeneities in the crustal structure of central Mexico (Chávez-García *et al.*, 1994) and the presence of melt material under Mexico City valley (Rodríguez *et al.*, 1996). The purpose of this paper is

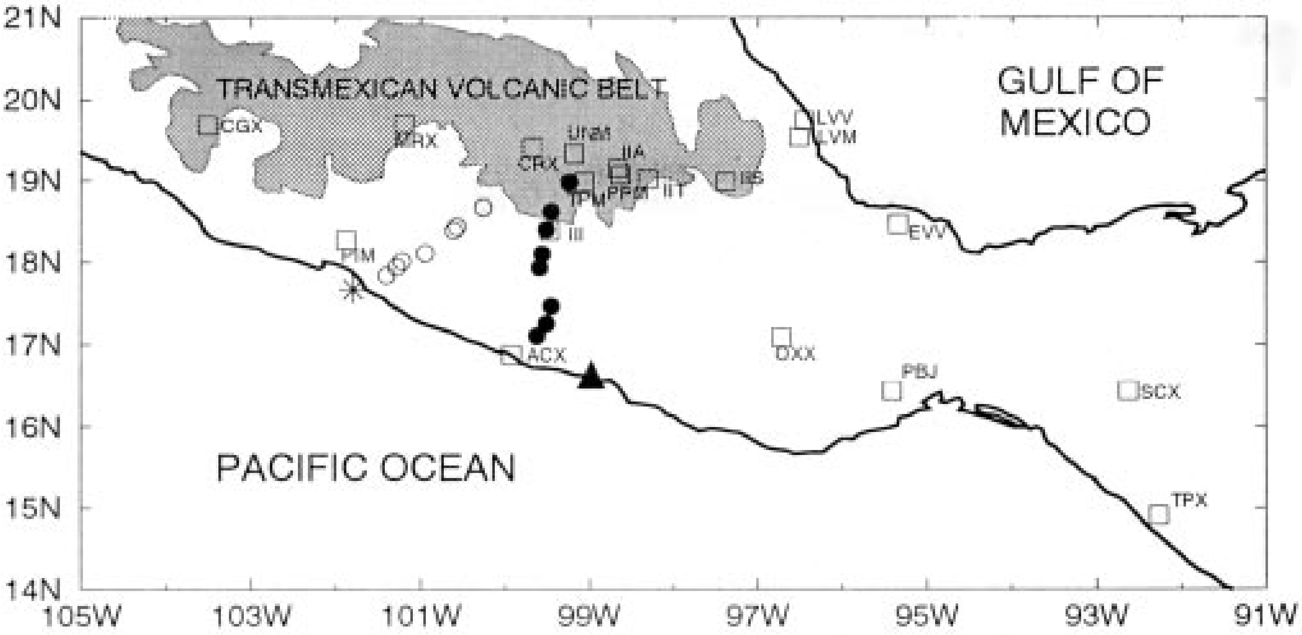

Fig. 1. Data used in this study. Open squares: stations of SSN for which we compiled duration data for 1993. Open circles: location of seismographs that recorded the explosion at sea (shown by a star) in the large scale refraction experiment. Solid circles: location of accelerographs that recorded the 10.24.93 event (shown by solid triangle).

to investigate the spatial distribution of regional amplification in Central Mexico, and to explore which is its likely cause.

DATA

We have analyzed three independent datasets:

1. Velocity, three component seismograms recorded during a large scale refraction experiment (Cárdenas, 1993). Seven stations were used along a profile that extended from Ixtapa, Gro., to Mexico City, spanning 200 km from the coast (Figure 1).

2. Acceleration records of the 10.24.93, Ms=6.6, event that occurred in the subduction zone (16.54°N, 98.98°W). This event was well recorded by eight of the stations that cover the path between the Pacific coast and Mexico City, along a profile 262 km long (Figure 1).

3. Coda duration magnitude data, compiled from the files of Servicio Sismológico Nacional (SSN) for events of 1993. A total of 786 events, recorded by 28 stations were analyzed (Figure 1).

PROCESSING AND ANALYSIS

Velocity and Acceleration Data

The purpose of the analysis of the time series was the identification of the wave propagation mode followed by the seismic energy from the coast inland. As a first step, it was necessary to assign common time to the strong motion records. This was done assuming a constant propagation velocity for the fundamental Rayleigh mode in the period band 8 to 10 s. This was possible because this Rayleigh wave could be unambiguously identified in all the records. As regards velocity data, it was necessary to recover the origin time of the explosion which was not available to us. This was done through a linear regression of P-wave arrival time for the first stations. Additionally, the orientation of the horizontal components of the velocity records had to be corrected imposing the first impulse of the P wave to be in the radial-vertical plane, as it became evident that no systematic orientation of the sensors had been taken.

We tried to identify wavetrains in each one of the records through: a) bandpass filtering and waveshape correlation between neighboring traces, and b) analysis of energy distribution as a function of group velocity and period (Dziewonski *et al.*, 1969; Herrmann, 1987). The latter analysis was intended to identify dispersion curves of group velocity for surface waves. Observed dispersion was compared then with that computed for the average velocity model between the Pacific coast and Mexico City (Campillo *et al.*, 1989). The differences between both sets of dispersion curves can be used as an indication of lateral variations of the velocity structure along the profile.

Coda Magnitude Data

The SSN reports a coda duration magnitude for each event that is the average for all of the stations that recorded it. We compiled the original duration measurements and computed a magnitude residual (MR) for each record. The formula used was that proposed by Havskov y Macías (1983). We analyzed MR for its dependence on four parameters: magnitude and depth of the event, and azimuth and distance between epicenter and station. We also divided the subduction zone in four subzones, and analyzed mean and standard deviation of MR for each station for all of the events in anyone subzone. Each subzone is bounded by fractures of Cocos plate and shows different convergence rates between oceanic and continental plates (Pardo and Suárez, 1995). Subzone 1 (104°W to 102°W) is bounded by Rivera fracture to the NW and by Orozco fracture to the SE. Subzone 2 (102°W to 98°W) is the Guerrero region, the most activeof all, bounded to the SE by O'Gorman fracture. Subzone 3 (98°W to 95°W) is bounded to the SE by Tehuantepec ridge. Finally, subzone 4 entends fom 95°W to 91°W, and is the second most active.

RESULTS

Velocity data

We analyzed two time windows, those of the P and S waves. Figure 2a shows the vertical component of the records. Amplitude scale is common to all the traces. We observe a gradually decreasing amplitude of the S wave, although the last three traces show similar amplitude. The P wave amplitude decreases very fast for the first four traces, and then increases at 148 km from the source. This amplification is evident on Figure 2b, which shows, as a function of distance and frequency, amplitude for a window of 4 s duration starting about 1 s before P wave arrival at each record.

In the S wave window, we identified a common pulse of Rayleigh waves between traces 1 and 2 at 0.27 s period. The group velocity for this period between the source and station 1 was 3.4 km/s, whereas that between the source and station 2 was larger than 3.5 km/s. This suggests that lateral variations affect high frequency surface wave propagation at distances larger than 60 km from the coast. Unfortunately, this pulse could not be identified in other stations. In general, it was not possible to correlate the different traces among them. We observed large variations in the group velocity dispersion curves, and no similar trends could be identified between neighboring stations. The spacing between stations was too large for the frequency band that was recorded.

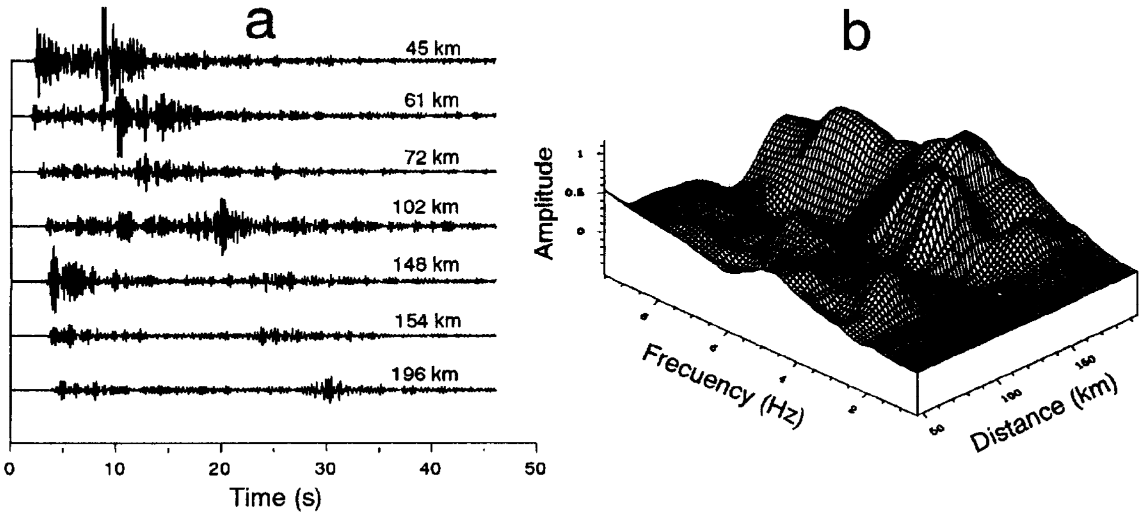

Fig. 2. (a) Vertical velocity recorded during the large scale refraction experiment. Amplitude scale is common to all the traces. The distance to the source is given on each trace. (b) Amplitude of a 4 s window including the first P-wave arrival of signals shown in Fig. 2a. Data was interpolated along the distance axis.

Acceleration Data

A common Rayleigh pulse could be identified in all the acceleration records in the period band 8 to 10 s. This allowed to assign common time to these records by assuming a group velocity of 2.8 km/s for this pulse. We analyzed group velocity dispersion curves in the period band 6 to 8 s. The results are given in Figure 3. Observed dispersion is similar for stations POZU and OCTT, suggesting that the thickness of the layers between the source and these stations could be smaller than those of the average model. Stations CHIL, MEZC, and TNLP indicate consistently that velocity is larger than that of the average model. Stations IGUA and TEAC are very similar and indicate lower than average velocity in the period band about 6 s.

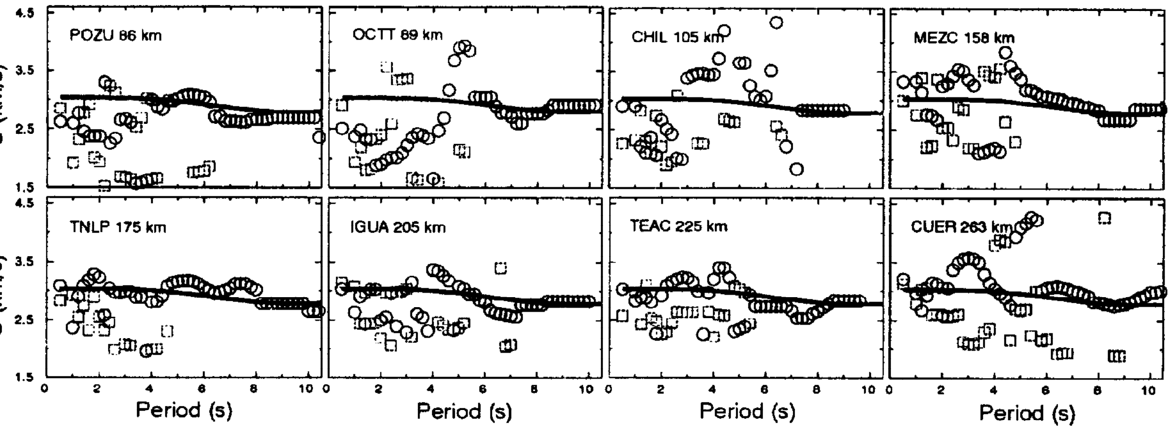

Fig. 3. Multiple filter analysis for the acceleration records of the 10.24.93 event. On each diagram the epicentral distance is given. Each symbol indicates where maximum amplitudes occur in the U-T plane. Solid line: dispersion curve of the fundamental Rayleigh mode computed for the average velocity model.

Coda Duration Magnitude Data

We analyzed MR for its dependence on four parameters: magnitude and depth of the event, azimuth and distance between epicenter and station. The magnitude of the event affects dispersion of MR for magnitudes below 4. Larger magnitude events show less dispersion of MR. This effect is related to the fact that the smaller events are less well recorded. It is difficult to separate the effects of distance and azimuth given the epicenter distribution (most of them occur in a narrow band along the coast). However, average MR at CGX is negative for events in subzone 2 and positive for all others. MR at OXX also shows complicated variations depending on the subzone considered. If we consider the stations on the Transmexican Volcanic Belt (TVB), we observe that MR is positive for epicentral distances larger than 550 km. For smaller distances, MR is close to zero or negative. The depth of the event does not affect MR. An example of the results is given in Figure 4, where we show average MR for events of subzone 4. Significant MR is evident in the central part of TVB, contrasting with very small or negative values for all the other stations. This suggests that: regional amplification is clearly detected by MR, that it may be related to the structure under the central part of TVB, and, given that the more important effects appear at large epicentral distances, that it is likely related to surface wave propagation

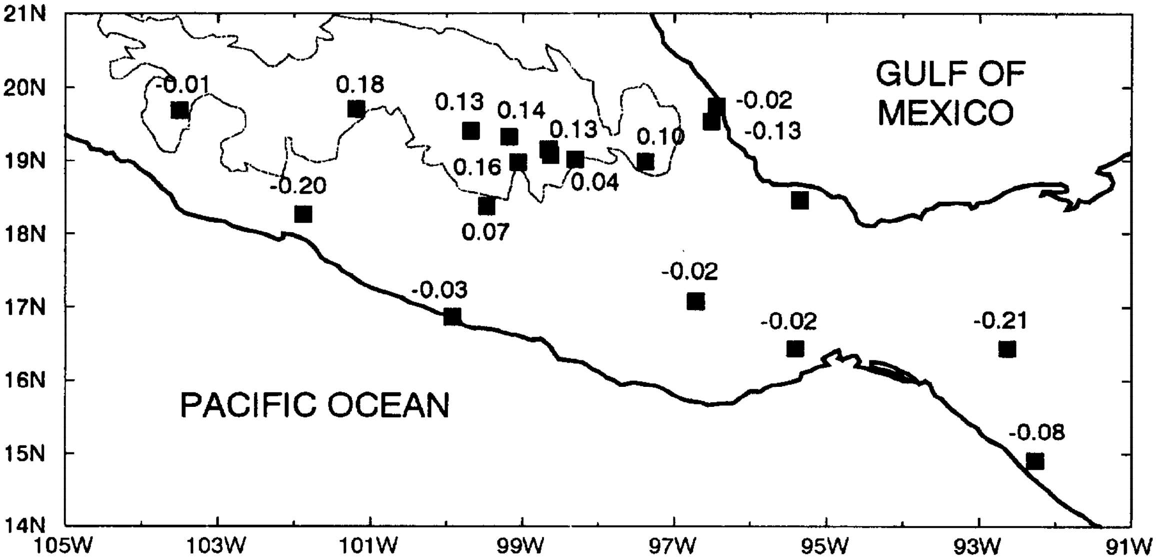

Fig. 4. Mean MR computed for each of the SSN stations for 1993 events of subzone 4 (dashed rectangle). Dotted line delineates the TVB.

PRELIMINARY WAVE PROPAGATION MODELLING

Our results shed some light on the spatial distribution of regional amplification, but are insufficient to propose a model that could explain the observations. They do allow to discard the hypothesis of large alluvial valleys (which should be too large), and suggest that regional amplification is more likely to be related to the crustal heterogeneity caused by the presence and the geometry of the TVB. In spite of the indeterminacy of the model, we have explored different preliminary 2D models to investigate the effects on ground motion of two factors: a horizontal velocity gradient between the Pacific coast and Mexico City, and an irregular interface within the first 30 km of the crust. The basic idea was proposed by A. Gusev (pers. comm.). We have used the finite difference method in the version given by Moczo (1989) for SH waves. The model was designed to include frequencies up to 1 Hz. The signal used as the source was a Ricker wavelet of 2.5 s dominant period. Figure 5 shows an example of the models. The coast (and the point source) is assumed to be located at distance 0 km. Mexico City would be at about 300 km. S-wave velocity (β) varies from 2.9 at 0 distance to 2.3 km/s at 300 km in the first layer, and from 4.3 km/s at 0 distance to 3.6 km/s at 300 km in the second layer. Below 30 km, the model is homogeneous with a β of 4.6 km/s. Density varies from 2.8 to 2.3 gm/cm^3 between 0 and 300 km distance, and is the same for the two irregular layers. In the halfspace density is 3.0 gm/cm^3. Anelastic attenuation was not considered. We have computed results for additional models changing the shape of the interface and the amount of lateral velocity variation. The results discussed here are representative of all of the models studied sofar.

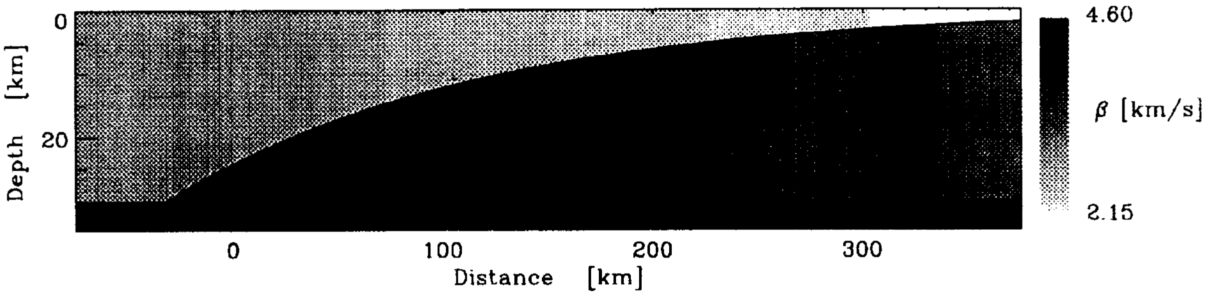

Fig. 5. Example of the model used in the numerical modelling. Gray scale gives S-wave velocity distribution in km/s. The star indicates the position of the source. Mexico City would be at about 300 km distance.

Figure 6 shows some results for this model. We have plotted maximum velocity (MV) observed in the synthetic seismograms as a function of distance for different cases. MGA corresponds to the model depicted on Figure 5, with the source at a depth of 10 km. MGC is equal to MGA but with the source at 20 km depth. MGB has the source at 10 km depth but β was constant within each layer (3.05 km/s for the first and 4.3 km/s for the second layer). MGD considered an irregular interface of similar shape, but with smoother lateral variation, and a source at 10 km depth. Finally, MCH is the result for flat layers (each one 15 km thick) including the horizontal velocity gradient, and with the source at 10 km depth. At 0 distance, peak velocity is maximum for all the cases. In all the models, MV decreases rapidly within the first 100 km, due to geometric expansion of S-waves. At a distance of about 125 km, MV is taken over by surface waves. We observe large differences at large distances between models MGA and MGC (almost a factor 3) due only to the depth of the source. Snapshots of the motion (not shown) make clear that the difference comes from the energy distribution between the fundamental and higher Love modes. If the source is close to the irregular interface (models MGC and MGD) higher modes are excited and energy is channelled by the irregular interface. If the source is shallow, most of the energy is channelled by the free surface. Results for models MCH and MGB show that both an irregular lateral structure and a horizontal velocity gradient are necesary to amplify significantly these surface waves.

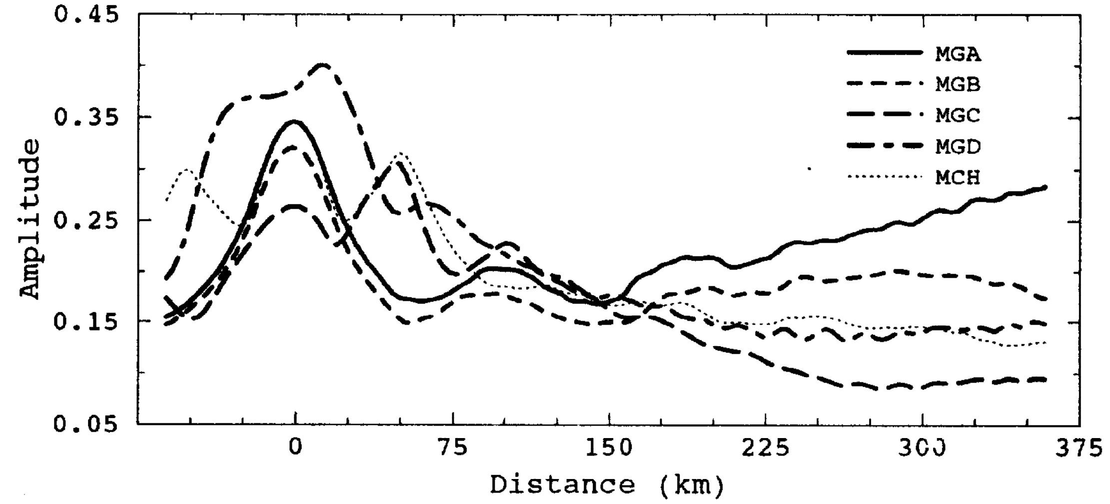

Fig. 6. Peak horizontal velocity as a function of distance for a series of 2D models like the one shown on Fig. 5. See text for explanation.

CONCLUSIONS

We have analyzed three independent datasets in order to investigate regional amplification in central Mexico. Our results show that, for distances larger than 100 km, ground motion is amplified during propagation perpendicular to the coast. For periods smaller than 8 s, surface wave dispersion varies significantly from the average model between the coast and Mexico City. Regional amplification is clearly detected by MR. Results for MR indicate that regional amplification is present in a large part of central Mexico, and that very probably it is related to surface wave propagation. Our results show that the most likely cause for regional amplification is lateral crustal heterogeneity (both in structure and velocity distribution) related to the Transmexican Volcanic Belt. Results from preliminary modelling for SH waves indicate that source depth affects in a very significant way the amount of regional amplification. This had already been hypothesized by Singh *et al.* (1988) concerning the observed differences in ground motion between 19.09 and 21.09, 1985, events. Finally, our preliminar results suggest that both an irregular lateral structure and a horizontal velocity gradient are necesary to explain regional amplification.

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