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On the Duration of Seismic Motion Incident Onto the Valley of Mexico for Subduction Zone Earthquakes

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SUMMARY

We have used finite-difference simulations in two-dimensional (2D) models of the lithosphere to estimate the duration of long-period (> 2 s) ground motion incident onto the Valley of Mexico for subduction zone earthquakes. Our simulations suggest that two heterogeneous structures extend the duration of the ground motion between the subduction zone and Mexico City by more than one minute: (1) the Mexican Volcanic Belt and (2) two low-velocity layers in the coastal region: the accretionary prism and the water layer. The duration generated by a crustal model including these structures is similar to that for earthquake records observed in between the coast and Mexico City. In the Valley of Mexico, our models including only regional-scale heterogeneity reproduce approximately one half of the observed duration. The results suggest that both the regional- and the local-scale low-velocity structures must be taken into account in order to explain the observed extended signal duration in the Valley of Mexico.

Key words: Valley of Mexico, finite difference, diffraction of ground motion, regional propagation

1 INTRODUCTION

For geotechnical purposes the Valley of Mexico is usually divided into the hill zone, the transition zone, and the lake-bed zone (Figure 1). The lake-bed zone consists of a highly compressible clay layer with high water content, underlain by sands (Figure 1). The thickness of the clays are generally between 10 and 100 m, but can be as large as 200 m locally. The anomalous nature of ground motion recorded in the lake-bed zone of the Valley of Mexico is well documented in the literature (e.g., Anderson et al., 1986; Bard et al., 1988; Singh et al., 1988; Campillo et al., 1988; Chavez-Garcia & Bard, 1994). In addition to strong amplification, the motion is characterized by extended coda, represented by a succession of roughly harmonic beats of slowly decaying amplitude.

While the magnitude of the observed ground motion can be relatively well explained as a combination of local and regional amplification, a reasonable explanation of the long signal duration has to our knowledge not yet been published. Shapiro et al. (2001) compile a recent

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detailed classification of the models suggested to explain this duration. Most of these models attribute the origin of the long coda to the resonance of the local sedimentary layers. This point of view has been strongly criticized since it can provide a viable explanation of the long duration only if shear-wave Q in the sedimentary layers is high (200-300). However, both laboratory (Romo & Ovando-Shelley, 1996) and field (Jongmans et al., 1996) measurements yield very low values of Q (10-50) in the lake-bed zone clays. Recently, a new approach has been developed to explain the long coda in Mexico City by the seismic interaction between the soil and buildings (e.g., Wirgin & Bard, 1996; Gueguen et al., 2000). However, this mechanism is unlikely to explain the long ground motion duration observed at sites located inside the lake-bed zone but far from the urbanized areas.

Other studies have provided observational evidence that the long duration of the ground motion can originate outside the lake-bed zone and even outside the Valley of Mexico. Singh & Ordaz (1993) showed that the extended coda is also present in the hill zone of the valley, which is composed of a surface layer of lava flows and volcanic tuffs (Figure 1). They suggested that the cause of the lengthening of the coda lies in the multi-pathing between the source and the site and/or multi-pathing within the larger Valley of Mexico. Importance of the structure deeper than the sedimentary layers has been notified by Chavez-Garcia et al. (1995) and Iida (1999). Shapiro et al. (2001) have analyzed bore-hole records in the lake-bed zone and have shown that most of seismic energy propagates in layers deeper than the lake-bed zone clays. ¿From the analysis of earthquake recordings by a portable array located in the hill zone, Barker et al. (1996) have found strong evidence of multi-pathing for records from several earthquakes. However, no clear evidence of off-azimuth arrivals was found for earthquakes along the Guerrero subduction zone, although the coda duration for these events was equally long. This suggested that, for subduction zone earthquakes, most of the secondary arrivals originate in vicinity of the source.

Figure 2 illustrates the principal characteristics of seismic signals recorded at three stations in Central Mexico for a subduction zone earthquake (October 24, 1993, $M_w = 6.3$, depth = 22 km) whose location is shown in Figure 1b. The accelerograms at Roma, a lake-bed site (Figure 2a,d), show strong amplification of waves and a longer coda duration with monochromatic character as compared to those at the site CUIG which is located in the hill zone (Figure

2b,e). PLIG is located on limestone at a site south of the Valley of Mexico. ¿From Figures 2a and 2b it is clear that the signal duration is much longer at Roma as compared to that at stations CUIG and PLIG. However, Singh & Ordaz (1993) suggested that such simple comparison of records is not appropriate for the analysis of ground motion duration since the clay layer of the lake-bed zone amplifies the ground motion in a very narrow frequency band (0.2-0.5 Hz). Singh & Ordaz (1993) proposed that the long duration and the monochromatic nature of the horizontal component of ground motion is a consequence of this narrow-band filter. Indeed, the bandpass-filtered (0.2-0.5 Hz) CUIG accelerograms in Figure 2d,e have about the same duration as that for the Roma records (Figure 2a,b). It is interesting to note that the bandpass filtered (0.2-0.5 Hz) accelerograms recorded at PLIG, approximately in the middle between the coast and the Valley of Mexico, also show extended duration. Clearly, this extended duration cannot be generated by the local geology of the Valley of Mexico alone.

The observations discussed above suggest that the regional structure has to be taken into account to explain the long duration of seismograms recorded in vicinity of the Valley of Mexico during the subduction-zone earthquakes. The importance of the path effect on the ground motion in Mexico City has been investigated by Campillo et al. (1989) via calculation of synthetic seismograms using 1D crustal models between the Pacific coast and Mexico City. They have shown that an appropriate 1D model can predict the shape of the principal S-wave arrival but not the extended duration. This implies that 2D or 3D heterogeneous structure has to be considered. One of the known regional-scale heterogeneous structures is the Mexican Volcanic Belt (MVB). Shapiro et al. (1997) have used simulations of the wave propagation in a simple 2D model to show that the dispersion and scattering of seismic waves in a low-velocity layer beneath the MVB can significantly increase the signal duration. Furumura & Kennett (1998) have provided numerical simulations in more complicated models of the crust and upper mantle beneath southern Mexico and have obtained similar results.

Recent observations show that the low velocity structures located in the vicinity of the subduction zone, i.e. the water column and the accretionary prism, can also significantly increase the signal duration (e.g. Ihmle & Madariaga, 1996; Shapiro et al., 1998). Shapiro et al. (2000) have provided a set of 3D numerical simulations and have shown that the resonance of these layers can generate a strong coda. However, they considered only the wave propagation along the coast and only the signals at relatively long periods (<0.2 Hz). The question still remains whether the reverberations of the near-source low-velocity layers at shorter periods (2-5 s) can affect the signals recorded in the Valley of Mexico, i.e. 300 km away from the Pacific coast.

Our main goal in this study is to examine whether the extended signal duration observed between the Pacific coast and the Valley of Mexico can be reproduced with a realistic heterogeneous model which includes the significant regional scale heterogeneous structures (i.e. the water layer, the accretionary prism and the MVB) but excludes sedimentary layers with extremely low velocity. To address this question we perform a set of numerical simulations of seismic wave propagation using a viscoelastic finite-difference (FD) method. Considering the spatial dimensions of the model and the frequency range of interest, three-dimensional (3D) simulations are presently unfeasible due to limitations in computer memory and cpu-time, even using the largest supercomputers. However, all heterogeneous structures are approximately parallel to the coast, allowing the problem to considered in two dimensions.

2 CRUSTAL MODELS

In order to investigate the origin of the extended duration of the ground motion in the Valley of Mexico we have designed a 2D velocity model of the upper part of the lithosphere below Central Mexico. These models include all or some of the following eight layers (Figure 3): (1) the water layer, (2) a low-velocity layer composed of deposits from the Mexican Volcanic Belt, (3) the accretionary prism, (4-6) three crustal layers, (7) the oceanic crust, and (8) the mantle. The geometry and elastic parameters of the subduction zone are constrained by gravity and seismic data (Shor et al. 1961; Valdes et al., 1986; Kostoglodov et al., 1996), and we use the continental crustal structure determined by Campillo et al. (1996). Following Shapiro et al. (1997) we include a superficial low-velocity volcanic layer with a random irregular basement structure below the Mexican Volcanic Belt. We have used an average wavelength of 10 km with a maximum possible deviation of 1.5 km from the mean value, deduced from the surface topography. We also introduced realistic anelastic attenuation in our model (Table 1). Ordaz and Singh (1992) found evidence of frequency-dependent Q from their study in Guerrero, but it is reasonable to assume constant attenuation within our long-period bandwidth (0.2-0.5 Hz) of interest. The values of Q_s in Table 1 were estimated from the values for the smallest frequencies used by Ordaz and Singh (1992) (1 Hz) and assuming that Q_s increases with depth. Q_p is estimated as 1.5 times Q_s, a relation often used in lack of better constraints.

Our model can be compared with the "subduction model" used by Furumura & Kennett (1998) for their pseudo-spectral simulation of seismic wave propagation in Mexico. There are a number of differences between the two models. One difference is that our model does not include a low-velocity layer in the subducted plate since we have found no observational evidence that such a layer exists in the Cocos plate below Mexico. However, the results by Furumura & Kennett indicate that the presence of this deep low-velocity layer mainly affects the S_n wave but has negligible influence on the total signal duration observed in the Valley of Mexico. The geometry of the subducting slab (uniform dip) used by Furumura & Kennett is taken from Valdes et al. (1986) and corresponds to the Oaxaca region, while the geometry of the slab in our model with a shallow, more gently-dipping and a deeper and steeper-dipping part (Figure 3) is deduced from the Guerrero region towards Mexico City (Kostoglodov et al., 1996). Furumura & Kennett also included the sedimentary basin with a minimum S-wave velocity of 1 km/s below Mexico City. Here, we have excluded these sediments in order to isolate the effects of the volcanics on the signal duration. Another notable difference is that we consider a slightly slower S-wave velocity in the lower crust. Furumura & Kennett (1998) deduced the S-wave velocities indirectly from the P-wave velocities measured by Valdes et al. (1986). In our work, we have used a more recent result by Campillo et al. (1996) where the S-wave velocities were obtained directly from surface wave group velocities inversion.

However, the most important differences between two models occur in the superficial part. Furumura & Kennett (1998) used an S-wave velocity of 2.2 km/s in the volcanic layer. In our model, we have used a lower value (1.7 km/s) which is supported by several observations described in the literature (e.g. Havskov & Singh, 1977-78; Shapiro et al., 1997). Finally, an important element of our model that may significantly affect the signal duration is the accretionary prism (e.g., Shapiro et al., 2000), omitted in the model by Furumura & Kennett.

3 NUMERICAL METHOD

We use a fourth-order staggered-grid finite-difference method to simulate P-SV waves (Levander, 1988) in 2-D models of Central Mexico. A commonly-used rule of thumb requires at least 5 points per minimum wavelength for accurate wave propagation using a fourth-order accurate scheme. However, it is possible that the strong impedance contrast between the water layer and the accretionary prism in our model as well as an abundance of surface waves with relatively slow propagation velocities may introduce an unacceptable amount of numerical dispersion in the results, even when this criterion is met. For these reasons we honor at least 10 points per minimum shear wavelength. The model parameters are listed in Tables 1 and 2. Viscoelasticity is implemented using stress relaxation independently for P and S waves (Robertsson et al., 1994; Blanch et al., 1995) using a standard linear solid model with one relaxation peak. The accuracy, relative to a frequency independent of the attenuation model, of the smallest Q in the model, Q_s=100, is estimated to be less than about 5% for the central 1/3 of the bandwidth of interest (see Blanch et al., 1995, Fig. 4).

The earthquake was simulated as a point source with thrust mechanism and dip according to the slope of the oceanic crust at the source location (Figure 3). We use an isosceles triangular slip rate function with rise time of 0.5 s. The source is implemented in the finite-difference grid by adding $-\Delta t\dot{M}_{ij}(t)/A$ to $\sigma_{ij}(t)$, where $\dot{M}_{ij}(t)$ is the ijth component of the moment rate tensor for the earthquake, A $= dx^2$ and $\sigma_{ij}(t)$ is the ijth component of the stress tensor on the fault at time t. Absorbing boundary conditions (Clayton and Engquist, 1977) are applied to the sides of the computational model. To further reduce artificial reflections the boundaries of the model are padded with a zone of exponential damping (Cerjan et al., 1985).

4 RESULTS

The goal of our work is to propose possible mechanisms for the generation of the extended duration of the long-period seismic motion observed at PLIG and subsequently incident onto the Valley of Mexico. Following Singh et al. (1995), we present the results of our simulations only in the frequency range where this coda is observed, i.e., between 0.2 and 0.5 Hz. Figures 4 and 5 show snapshots for crustal models without (left) and with (right) the low-velocity layers in the source region, i.e., the accretionary prism and the water layer, and the corresponding synthetics are shown in Figure 6. Outside the volcanic layer, the snapshots and synthetic accelerograms for the model without the low-velocity layers in the source region reveal a simple wavefield with short duration. Most energetic arrivals are associated with Lg waves and the fundamental Rayleigh-wave mode. However, it can be seen from the snapshots that the energy of the Rayleigh wave is concentrated in the shallow crustal layer while the Lg energy is distributed over all the crust. Contrary to the results of the simulations by Furumura & Kennett (1998), Lg waves are not completely trapped in the upper crust and penetrate in the lower crust. The main reason for this difference is that, in our model, the velocity contrast between the upper and the lower crustal layers is not as strong as in the model considered by Furumura & Kennett.

The signature of the wavefield changes completely in vicinity of the low-velocity volcanic layer. Similar to the results of Shapiro et al. (1997) and Furumura & Kennett (1998), our simulations show that this layer (1) increases the amplitude of the incident waves, especially on the horizontal component, and (2) increases the duration of both Lg and Rayleigh waves significantly. This increase of approximately 30-40 s in the coda duration is due to the dispersion and scattering of the seismic waves inside the low-velocity layer. Therefore, our simulations support that the volcanic layer plays an important role in the generation of the extended signal duration observed in the Valley of Mexico.

However, the volcanic layer cannot explain the existence of the coda outside the Volcanic belt, as discussed earlier. In order to explain this observation, we have have included the low-velocity structures in the source region, i.e. the accretionary prism and the water layer. Separation of the effects from these two features is desirable, but, in reality, virtually impossible to carry out. For example, a model including the prism but without the water layer introduces a boundary between the prism and either air or crustal material, an artificial interface that may generate artifacts in the modeling. A similar situation arises for a model including the water layer but without the prism. For these reasons we prefer to show the combined effects of these two low-velocity layers.

The main effect of the low-velocity layers in the source region is the generation of high-amplitude and long-duration trapped waves (Ihmle & Madariaga, 1996; Shapiro et al., 1998, 2000). These waves dominate the signal at receivers located close to the coast. However, our simulations show that these trapped waves penetrate partially into the continental crust and propagate from the coast up to Mexico City as diffracted waves. As a consequence, a long-duration coda appears on the accelerograms simulated at receivers located outside the Volcanic Belt. It can be clearly seen on the snapshots that the energy of these diffracted waves is distributed from the surface to the lower crust. It means that the coda is composed of both Rayleigh- and Lg-type diffracted waves. The coda is amplified inside the volcanic layer, and as a consequence, the signal duration at receivers located on the volcanic belt is significantly increased.

Figure 7 shows the duration of the synthetic ground motions for models with and without the accretionary prism and the water layer (models 1 and 2, respectively, Table 4). The duration is measured as the cumulative time for which the vector ground acceleration exceeds a threshold value (Tumarkin & Archuleta, 1997) of 10% of the maximum acceleration at PLIG, separately for each earthquake. The durations are compared to the values for accelerograms recorded at stations PLIG, YAIG, and CUIG for seven subduction earthquakes with magnitude **M** 5.5-6.7, thrust focal mechanisms, and epicenters within the area shown in Figure 1b. In addition, we show durations for station YAIG, located just outside the hill zone (see Figure 1a). Figures 8 and 9 show the observed accelerograms for the two events with durations denoted by '+' (event 3) and 'x' (event 7), respectively, in Figure 7. The parameters for the earthquakes are listed in Table 3.

We also tested the sensitivity of the results of our simulations with respect to the choice of the model parameters. In Figure 7, we show the signal durations computed from synthetic accelerograms obtained with perturbations in the seismic velocity inside the accretionary prism (model 3, Table 4), in the source location (model 4, Table 4), and in the source mechanism (model 5, Table 4). It can be seen that these perturbations do not strongly affect the resulting signal duration and that all models including the accretionary prism and the water layer produce significantly larger signal duration compared to the model without these low-velocity layers.

While the durations measured from recorded seismograms show considerable variation, they are all longer than those predicted by the model without the accretionary prism, water layer, and the Mexican Volcanic Belt. The accretionary prism and the water layer extend the duration by about 80 s between the subduction zone and the Valley of Mexico. The model including these two features satisfactorily reproduces the median duration for the accelerations recorded outside the Valley of Mexico, i.e. at PLIG and YAIG, while the model without the features only reproduces about 30% of the duration. However, although the duration is increased by the volcanic layer, the duration of the synthetic accelerograms are still only about one-half of the median of that for the strong-motion data recorded at CUIG. The underprediction of the duration in the Valley of Mexico shows the importance of complex local structure, omitted in the modeling.

¿From the discussion above we conclude that the accretionary prism and the water layer have a strong influence on the regional wavefield. In addition, the the duration of the ground motion is further increased by the volcanic layer before impinging onto the local structure of the Valley of Mexico. Here, local site effects increase the duration of the ground motion by about a factor of two.

5 DISCUSSION AND CONCLUSIONS

We have simulated wave propagation in simple 2D models of the central Mexican lithosphere. We find that the accretionary prism and the water layer must be included in order to reproduce the extended duration of ground motion recorded from subduction zone earthquakes at stations between the coast and the Valley of Mexico. The model including these structures reproduce relatively well the duration of the ground motion recorded outside the Valley of Mexico. In addition to the effects of the low-velocity layers in the coastal region the superficial layer of the Mexican Volcanic Belt further increases the duration of the seismic motion incident onto the Valley of Mexico. Our conclusion that an important part of the extended signal duration originates outside the Valley of Mexico is in agreement with the observations by Barker et al. (1996) who finds that most of the secondary phases in seismograms recorded in the hill zone of Mexico City are incident from the source direction.

In the Valley of Mexico our model generates a duration of about 150 s, approximately one-half of that recorded at station CUIG in the hill zone for subduction zone earthquakes.

We expect that the remaining part of the duration is generated by a combination of several features. The most important feature is likely the local structure. Indeed, the influence of the very low-velocity sedimentary layers of the Valley of Mexico has been documented in numerous studies. Topographic scattering may account for another part of the unmodeled duration. However, we expect that the long-period wavefield (smallest wavelength 6 km) is relatively insensitive to the variation in topographic relief of about 1 km about the average from the coast to the Valley of Mexico. Clearly, the topographic scattering becomes more important for higher frequencies. Finally, it is possible that three-dimensional effects contribute to the duration.

Our main conclusion is that the complete explanation of the extended signal duration observed in Mexico City requires models including low-velocity structures at both regional and local scales. Therefore, future studies should combine simulations at regional (e.g., Furumura & Kennett, 1998; this work) and local (e.g., Kawase & Aki, 1988; Sánchez-Sesma et al., 1988) scales. In particular, the lake-bed sediments must be considered. Despite the extremely low impedance of these sediments, they can be included in the FD models by certain techniques, such as variable-grid procedures proposed by Faeh et al. (1994), Moczo et al. (1997) and Olsen et al. (2000). However, such modeling is beyond the scope of this study.

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Table 1 Modeling Parameters

Spatial discretization (km)	0.3
Temporal discretization (sec)	0.0225
Number of grid points laterally	5500
Number of grid points vertically	1000
Lateral extent of model (km)	1650
Vertical extent of model (km)	300
Minimum source frequency (Hz)	0.0
Maximum source frequency (Hz)	0.5
Number of timesteps	10000
Simulation time (sec)	225

Table 2 Elastic Model Parameters

		P velocity (km/s)	S velocity (km/s)	ρ (g/cm^3)	Q_p	Q_s
1	Water	1.5	0	1.0	1500	1000
2	Volcanics	2.9	1.7	2.1	150	100
3	Accretionary Prism	2.6	1.5	2.1	75	50
4	Continental Crust 1	5.2	3.0	2.5	300	200
5	Continental Crust 2	5.9	3.4	2.7	450	300
6	Continental Crust 3	6.4	3.7	3.0	750	500
7	Oceanic Crust	6.8	3.9	2.9	1050	500
8	Mantle	7.8	4.5	3.3	1500	1000

Table 3 Source Parameters of the Earthquakes Used in This Study (From Harvard CMT Catalog)

	yyyy/mm/dd	Longitude	Latitude	Depth (km)	M_W
1	1993/10/24	16.77	-98.61	22	6.6
2	1996/03/27	16.44	-97.95	21	5.5
3	1996/07/15	17.50	-101.12	22	6.6
4	1997/01/21	16.49	-97.99	40	5.5
5*	1997/12/22	17.25	-100.90	10	5.6
6	1998/07/11	17.28	-101.17	24	5.4
7	1998/07/12	16.78	-99.91	15	5.5

* the location and the magnitude for event 5 are from the Mexican Seismological Service

Table 4 Model Descriptions

	Source Position*	Source Dip (^{o})	Prism+Water	V_s of Prism (km/s)
1	1	11	Yes	1.5
2	1	11	No	-
3	1	11	Yes	2.0
4	2	11	Yes	1.5
5	1	45	Yes	1.5

* see Figure 3

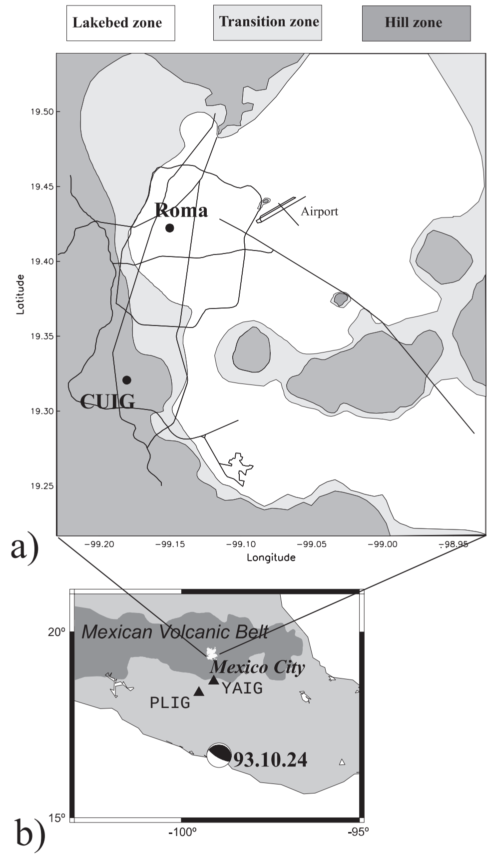

Figure 1. (a) Map showing the geotechnical division of the Valley of Mexico in the hill zone, the transition zone, the lake-bed zone, and the location of strong motion stations Roma and CUIG. (b) Map of Central Mexico showing the locations of stations PLIG and YAIG and the epicenter of event 1 (Table 3).

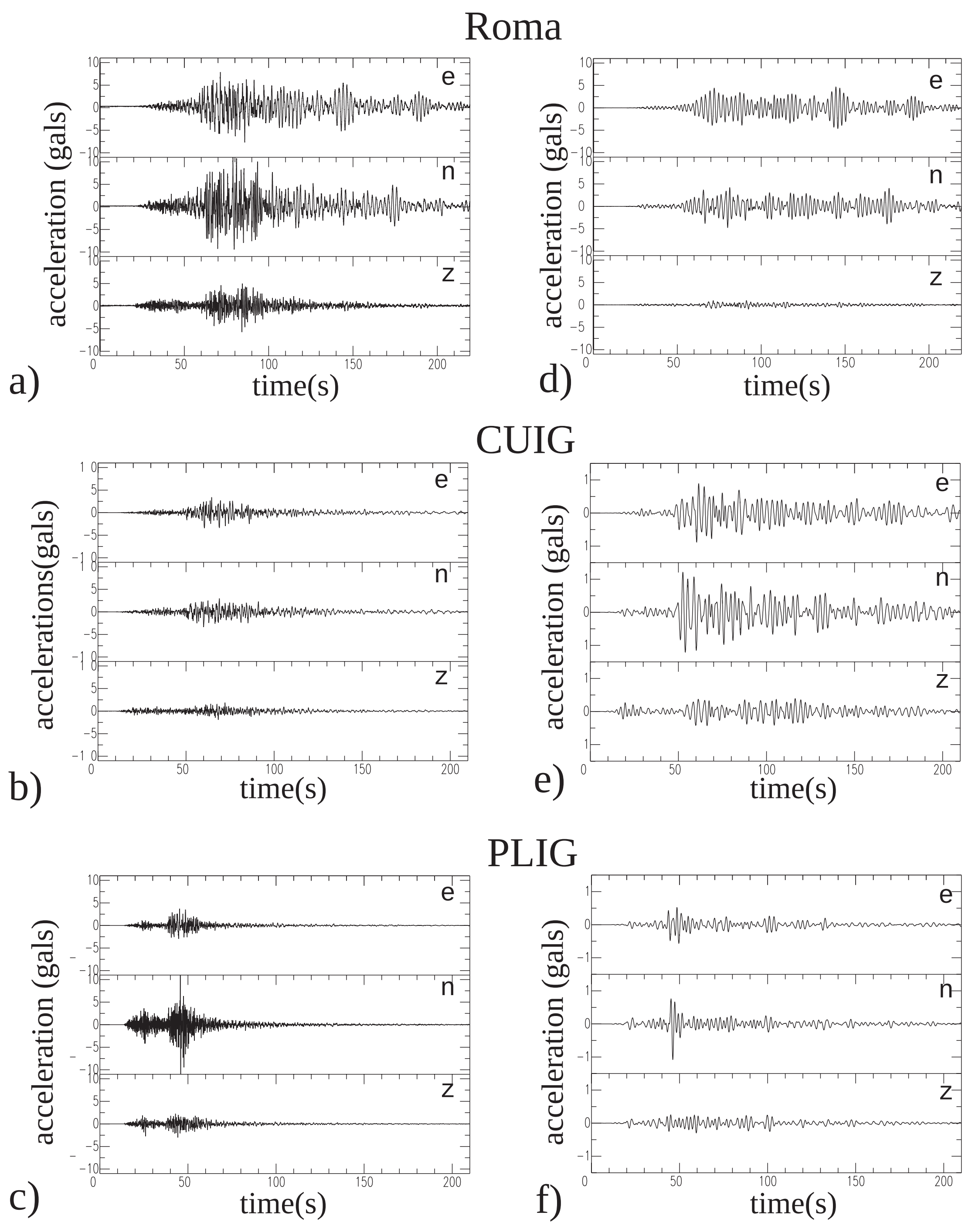

Figure 2. Observations illustrating the variation in seismic response during event 1 for the regions shown in Figure 1: accelerograms at Roma (a,d), CUIG (b,e), and PLIG (c,f), unfiltered (left) and bandpass filtered between 0.2 and 0.5 Hz (right).

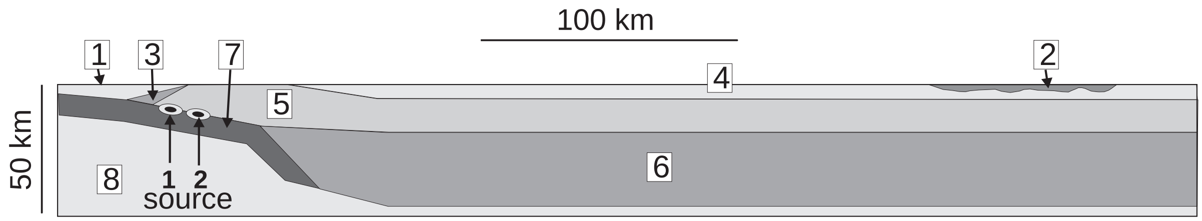

Figure 3. 2D model used in our simulations.

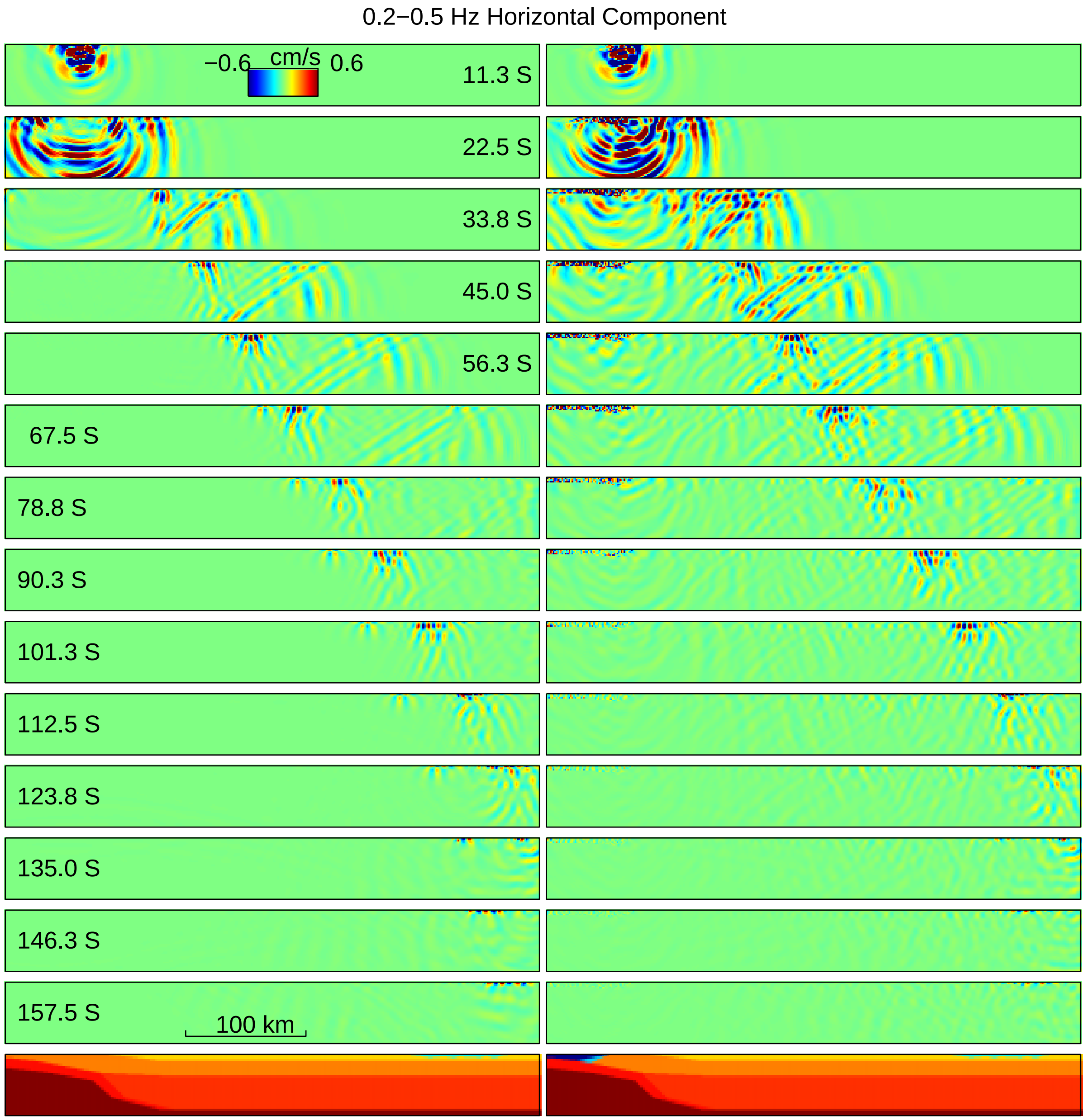

Figure 4. Horizontal-component snapshots of the particle velocity for the model without (left) and with (right) the accretionary prism and water. For both models we used source position 1 (Figure 3) and thrust faulting source mechanism with a dip angle of 11°.

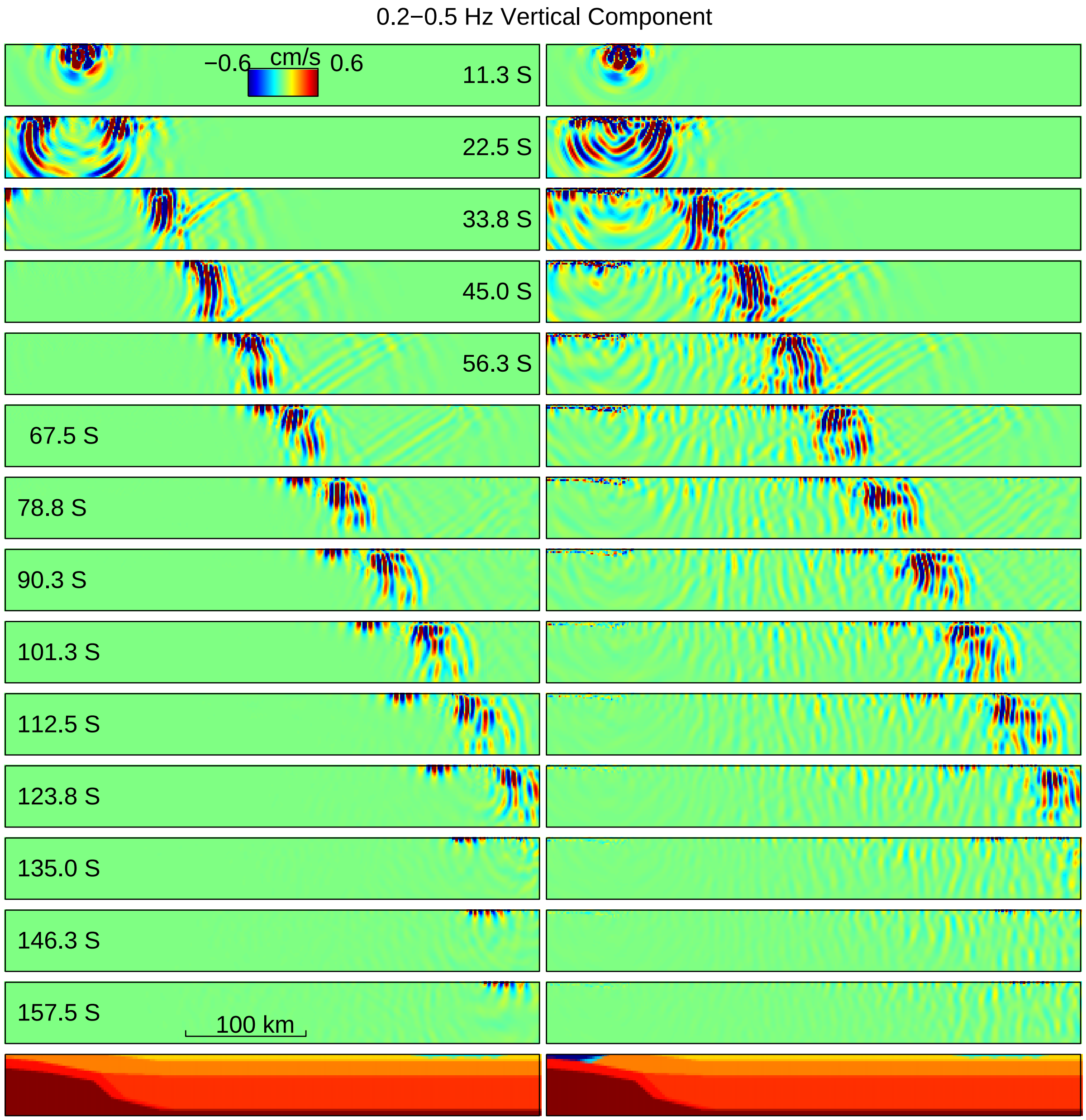

Figure 5. Same as Figure 4, but for the vertical component.

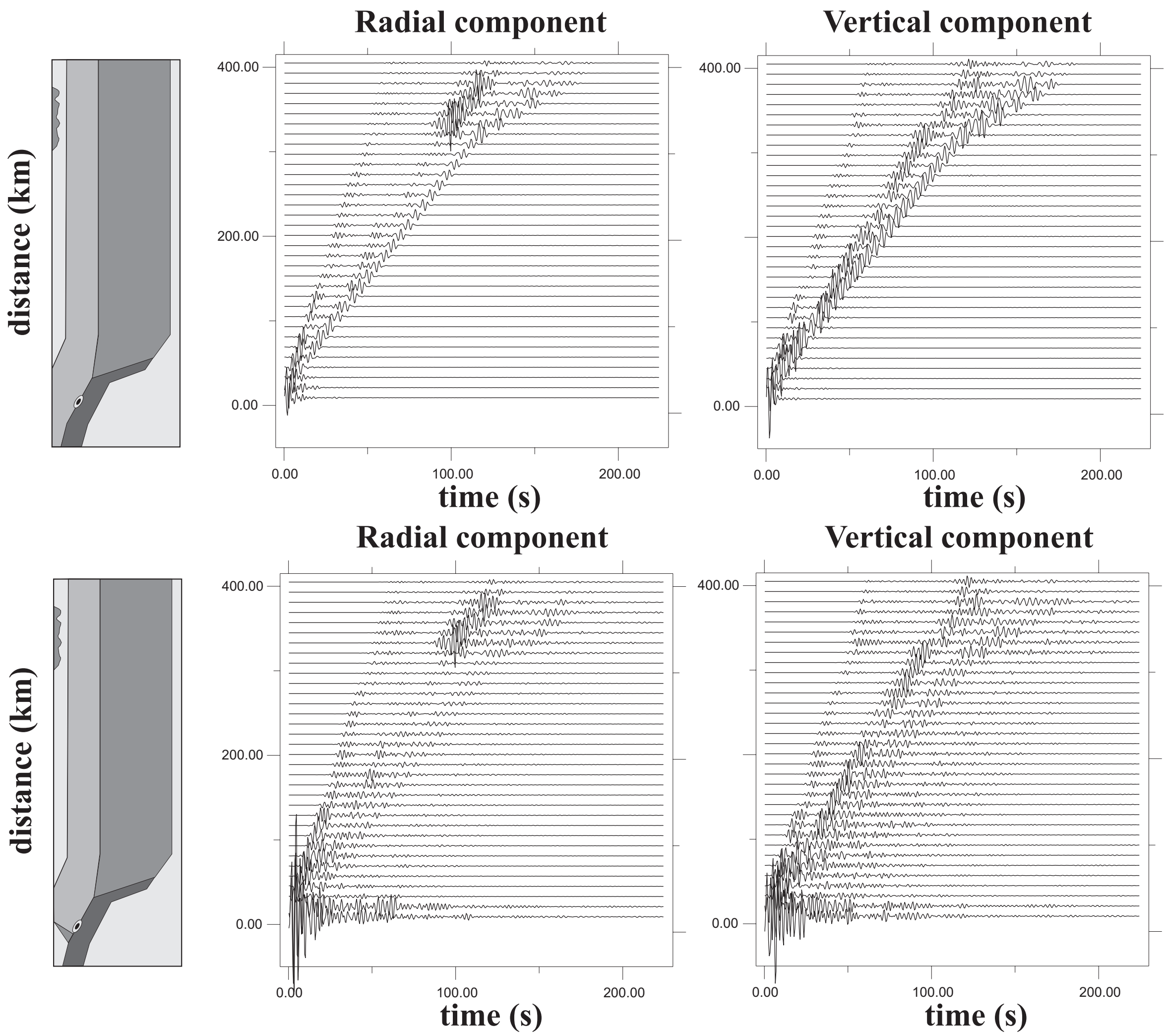

Figure 6. Synthetic accelerograms for crustal models without (top) and with (bottom) the accretionary prism and the water, bandpass filtered between 0.2 and 0.5 Hz. For both models we used source position 1 (Figure 3) and thrust faulting source mechanism with a dip angle of 11°.

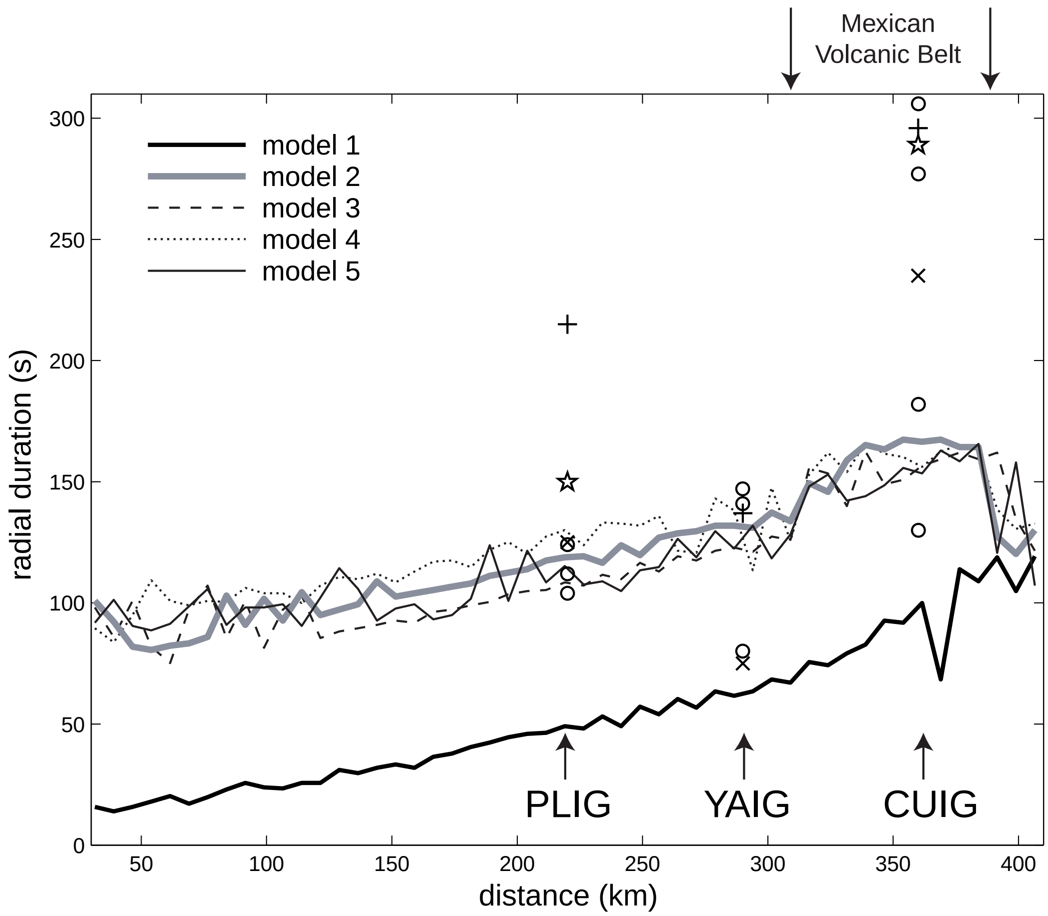

Figure 7. Duration of the radial ground motion for models 1-5 (see Table 4). The duration for the earthquake for which we displayed accelerograms in Figures 2, 8 and 9 (events 1, 3, and 7, respectively) are shown by a pentagon, a '+' and an 'x', respectively, and those for the remaining earthquakes are depicted by 'o'. Both synthetic and observed accelerograms were bandpassed between 0.2 and 0.5 Hz before calculation of the duration.

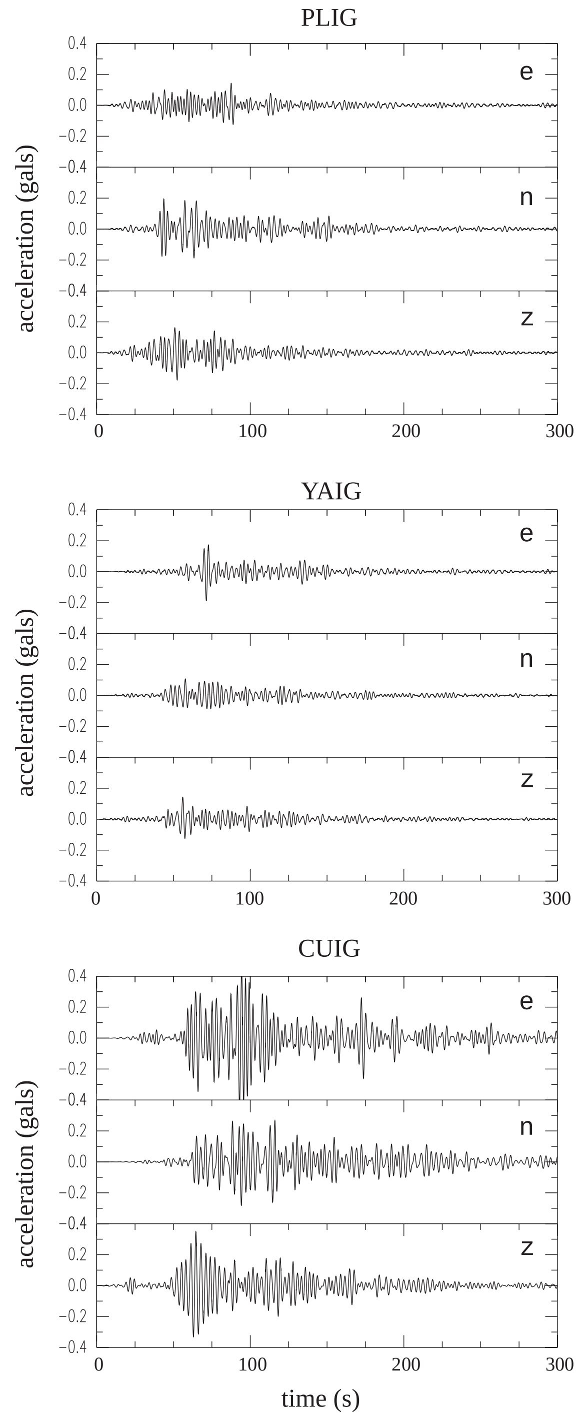

Figure 8. Accelerograms observed at stations PLIG, YAIG, and CUIG for event 3 (denoted by '+' in Figure 7). Signals are bandpassed between 0.2 and 0.5 Hz.

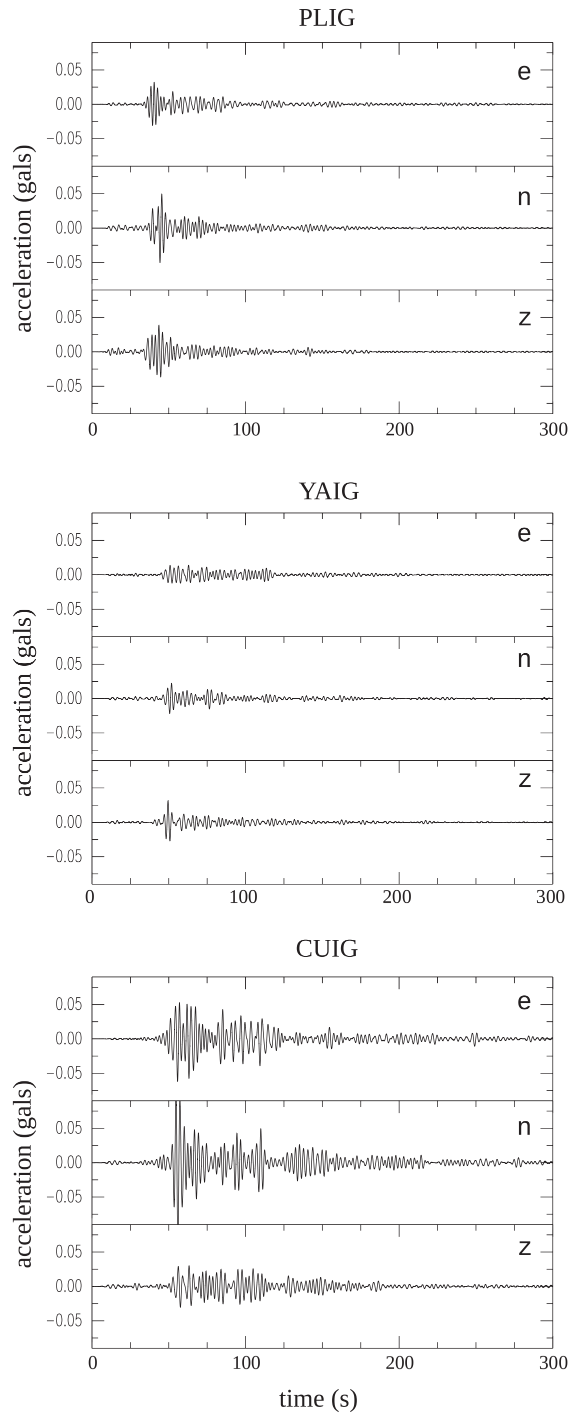

Figure 9. Accelerograms observed at stations PLIG, YAIG, and CUIG for event 7 (denoted by 'x' in Figure 7). Signals are bandpassed between 0.2 and 0.5 Hz.