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Three-dimensional simulation of earthquakes in the Grenoble's basin

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ABSTRACT: The town of Grenoble (Western Alps) is characterized by a location on a deep sediment-filled basin and the proximity of active faults. We have used a 3D finite difference method to simulate ground motion in the Grenoble area. Simulation are carried out for hypothetical earthquakes on potential active faults in the Grenoble area. Results show that sites with the largest particle velocities and duration time of motion overlie the deepest part of the basin (Grenoble downtown). Peak velocity and motion duration in the Gresivaudan and Voreppe Valleys are strongly dependent on the location of the earthquake. To check the validity of our simulations we compare our 3D synthetic seismograms to observed ground motion of the magnitude 2.5 Lancey earthquake located at 5 km of the basins border.

1. INTRODUCTION

The town of Grenoble (Western Alps) is characterized by a location on a deep sediment-filled basin and the proximity of active faults. We have used a 3D finite difference method to simulate ground motion in the Grenoble area. Simulation are carried out for hypothetical earthquakes on potential active faults in the Grenoble area. To check the validity of our simulations we compare our 3D synthetic seismograms to observed ground motion of the magnitude 2.5 Lancey earthquake located at 5 km of the basins border.

2. SEISMIC POTENTIAL AND LOCATION OF ACTIVE FAULTS IN THE GRENOBLE'S AREA

Our study of historical seismicity in this region shows that the Grenoble region is characterized by moderate seismicity (Figure 1). The strongest shock in the last 100 years (1962 April 25) occurred near Corrençon, in the Vercors area; its magnitude was estimated as 5.3 (Levret et al., 1994). The depth of its focus is shallow (between 5 and 10 km). Gamond (1994) states that the earthquake occurred on an overthrusting structure (incline) in this subalpine mountain range. The structure responsible of the 1962 earthquake could potentially generate a similar earthquake under the city of Grenoble at a depth around 5 km.

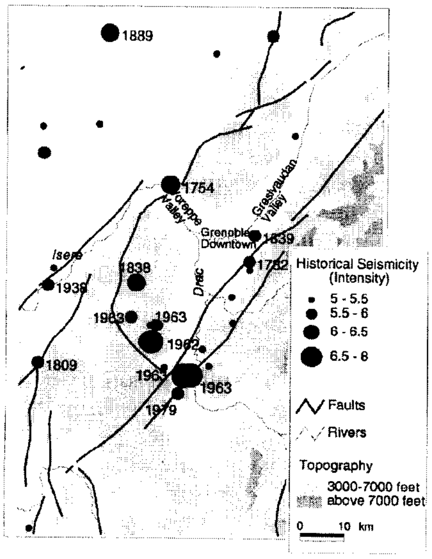

Figure 1. Historical seismicity in the Grenoble area

Instrumental seismicity studies (Thouvenot, 1996) also suggest a strike slip motion of a fault located at the border of the Belledonne Massif. These results suggest that two accidents could potentially generate earthquakes of significant magnitude in the Grenoble area: a thrust faulting located between 5 km and 10 km under the Grenoble area and a dextral strike slip faulting located between 12 and 20 kilometers east of the Grenoble area.

3. SITE EFFECTS EVIDENCE IN THE GRENOBLE'S BASIN

The Isère Valley is a NW-SE structural accident aggravated by glacial and fluviatile erosion. The substratum consists of Jurassic marls and marly limestone. Sedimentary fill consists of coarse fluvio-glacial deposits at the base, overlined by a layer of clays and a layer of sand and gravel and finally a silt-layer near the surface. There are few data about mechanical characteristics (shear wave velocity, density, damping) of the alluvial filling in the Isère Valley. Shear waves velocity in the first ten meters are estimated to be between 200 and 300 m/s (Gariel, oral communication).

Macroseismic observations of the Corrençon earthquake (1962) show that the effect of the quaternary post-glacial fillings appears to have quite considerable effects in the Grenoble basin (Figure 2). Levret et al. (1994) show the distribution of isoseismals and quaternary superficial formations. In the Isère, Gresivaudan, Romanche and Drac Valleys isoseismals are clearly elongated along the valleys. Through a quantitative assessment of the relationships between macroseismic observations and surface geology Cushing et al. (1995) have shown that, on average, quaternary formations increase the intensity by one degree. In 1995, 10 stations have been installed for 8 months in the city of Grenoble to record earthquakes and ambient noise (Lebrun, 1997). Spectral ratio to a reference station and horizontal-to-vertical (H/V) spectral ratios using noise recordings have shown that the resonance frequency of the basin is around 0.3 Hz. The spectral ratios show that amplifications in the basin reach values between 5 and 10.

Calculations of 1-D transfer function using various hypothetical 1-D velocity models of the Grenoble's basin have been performed by Cushing et al. (1995) and Lebrun (1997). Both authors calculated the transfer function using horizontal layers and a vertically incident S wave. Cushing et al. (1995) have performed several calculations with various velocity models. They show that dependent on models (depth of the basin between 380 and 620 meters, S velocity of the sediments between 900 and 300 m/sec and shear wave velocity of limestones around 2500 m/s) the most significant amplification occur between 0.3 and 0.7 Hz and is between 2 and 6. Lebrun (1997) shows similar numerical results using a 1D transfer function. These results are coherent with experimental results of Lebrun (1997).

4. DESCRIPTION OF THE BASIN

Using geological, geotechnical and ten years of gravity measurements a rough 3D model of the Grenoble area has been constructed. Few informations about the depth of the Grenoble basin do exist. A deep borehole had been carried out in the city of Grenoble in 1944, in order to explore the hydrocarbon evidences of the filling of the glacial valley (Gignoux, 1944). Two main formations had been pointed out from the rough logging (from 0 to 72 m : coarse glacial material ; from 72 to 400m : alternance of marl and micaceous sands corresponding to lacustrine deposits at the crossing of the main valleys). This borehole of 400 meters depth did not reach the basement. Bouguer anomaly analysis of ten years of gravity measurements (Vallon et al., 1996) provided us the underground bedrock's topography (Figure 2).

Based on these measurements, the maximum thickness of the sediments is 900 m. The map of the basin depth obtained by Vallon et al. (1996)

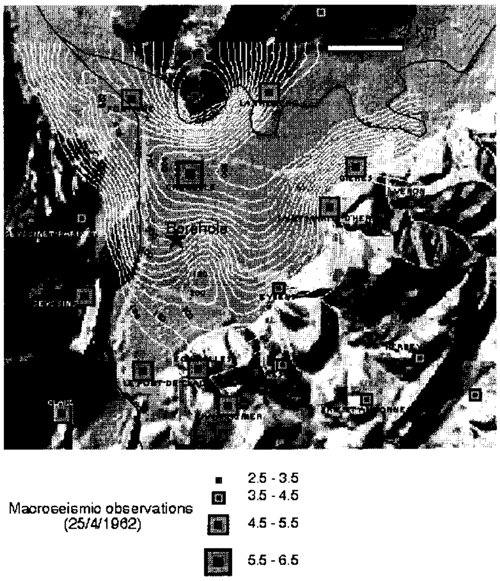

Figure 2: Digital elevation model of the Grenoble basin and 3D contour map of the depth of the basement (thick black lines) of the Grenoble basin model (Vallon et al., 1996). The star shows the location of the 1944 borehole which did not reach the basin at 400 meters depth (Gignoux, 1944). The macroseismic observations (black circles) of the 1962 Corrençon earthquake in the Grenoble basin (SIRENE, 1998) shows that the Grenoble basin increases the intensity

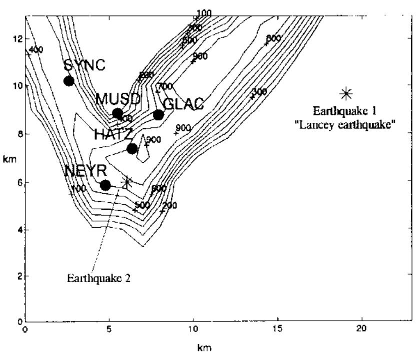

Figure 3. 3D contour map of the depth of the basement of the Grenoble basin model used in this study. The stars show the location of the earthquakes simulated in this study (earthquake 1: Lancey earthquake located at a depth of 2 km and recorded in 1995 by Lebrun (1997), earthquake 2: hypothetical earthquake located on a thrust at 5 kilometers depth). Circles: stations used by Lebrun (1997) to record the Lancey earthquake.

indicates a basement depth value of 475 meters at the borehole location (Figure 2).

This result confirms the high basin depth obtained trough the interpretation of the gravity measurements. The velocity model chosen in the 3 D simulation is described in Table 1 and Figure 3. The crustal structure of the Grenoble area is modeled by 4 layers. The topography above the sea level was not included.

5. SIMULATION METHOD

One of the most difficult problem with predicting ground motion in this region is the three dimensional nature of the velocity structure, especially the deep sedimentary basin which has a significant impact on the progression of the seismic waves. Until recently, the prediction of ground-amplification in alluvial basins was limited to 1D and 2D modelings. Such modeling was not completely adapted to the study of the Grenoble basin effect because of the complicated 3D geometry. In this study, we use a fourth order staggered grid finite-difference method. The numerical implementation of the 3D scheme is described by Pitarka et al. (1998). The anelasticity and free surface condition in the FD calculation is included using the technique proposed by Graves (1996). The attenuation is approximated using a reference frequency f_0=0.5 Hz. Based on Grave's (1996) Q model, we expect that, energies for frequencies greater than 0.5 Hz will be underattenuated and energies lower than 0.5 Hz will be overattenuated. The same attenuation factor is applied to S and P waves. The absorbing boundary conditions are described in Clayton and Engquist (1977). The basin model (23 x 13 x 8 km^3)is discretized with a grid spacing of 0.140 km equivalent to 5 nodes per minimal wavelength (Graves 1996). Simulations were carried out on DEC computer using approximately 80 Mbytes of physical memory. All the synthetics are band-pass filtered in the frequency range of 0.2-0.8 Hz to exclude the numerical errors in at other frequencies.

Table 1. Velocity model

depth (m)	Vp (m/sec)	Vs (m/sec)	density (g/cm^3)	Q
3D	1000	500	1.6	150
3D	1200	600	1.8	150
5000	4250	2700	2.5	200
	5500	3600	3.2	500

To evaluate the effect of the 3D geometry of the basin on the ground motion, two earthquakes are simulated. The first one is similar to the « Lancey » event recorded by Lebrun (1997). This earthquake is a magnitude 2.5 earthquake which occurred north east of the basin near the village of Lancey. Secondly, we simulate an earthquake with a similar magnitude than the Lancey earthquake with a thrust mechanisms located at a depth of 5 km under the Grenoble's basin.

6. EARTHQUAKE SIMULATION

6.1 *Effects of a strike slip event located east of the Grenoble's basin (Lancey earthquake)*

We choose the source to be a point source with a bell shaped function (0.6 second duration). In Figure 4 we show snapshots of the horizontal amplitude of the ground motion velocity. At 1.0 second the wave field is well developed and controlled by the radiation pattern of the source. The classical double couple radiation pattern for a strike slip mechanism

Table 2. Source caracteristics of the simulated earthquakes

Event	strike	dip	rake	depth (km)	source duration (sec)	moment (dyne-cm)
Earthquake 1	39°	80°	-5°	2.0	0.6	10^{20}
Earthquake 2	90°	10°	90°	5.0	0.6	10^{20}

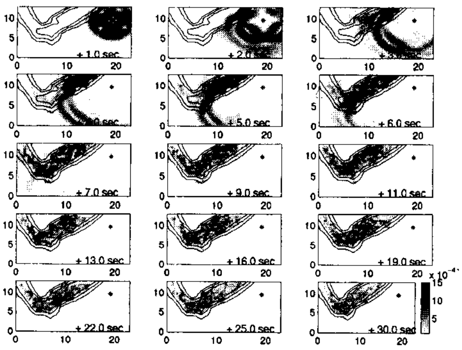

Figure 4. Snapshots of simulated wave propagation in the Grenoble area for an earthquake located near the village of Lancey. The snapshots depict the horizontal velocity amplitude from 1.0 to 30.0 seconds after the origin time of the earthquake. The thin line depicts the 100, 400 and 700 meters depth of the basin (Figure 3).

is evident. The snapshots at 2.0 seconds and 3.0 seconds show the P wave front followed by S wave front. The S wave amplitude dominates the frame. By 3.0 sec, the Grenoble basin is affecting the S waves. Waves radiated in the rock maintain circular front whereas waves propagating in the basin are distorted. Thereafter the seismic waves are characterized by a more complex pattern. Intense shaking throughout the basin lasts for about 40 seconds.

Several phenomena control the propagation in the basin and amplification of waves in such a 3D medium (Bard, 1983). First, there is an amplification of the displacement which occurs when the seismic waves travel trough an interface of media with high and low rigidity. The second phenomena is the focusing of seismic rays which are reflected at the free surface and at the interfaces of the basin. The third phenomena is the generation of surface waves by diffraction on the edge of the basin and their subsequent trapping in the basin. These phenomena lead the long duration and high amplitude of the motion in the basin. Visual identification of the phases in the snapshots is complicated by complex 3D wave interference pattern and the vast amount of digital information generated by the 3D simulation. Present work was limited to the localization of high amplification and long duration ground motion area in the Grenoble's basin for hypothetical earthquakes. Further work will include an identification of the various phases propagating in the basin.

For the first event, the largest amplification is observed in the center part of the basin (Grenoble downtown) and east of the Gresivaudan Valley where the basin is close to the earthquake (Figure 5). In these areas, the ground motion peak amplitude is equal and even larger than the one observed close to the epicenter. Long durations are observed in the central part of the basin and west of the Grésivaudan Valley.

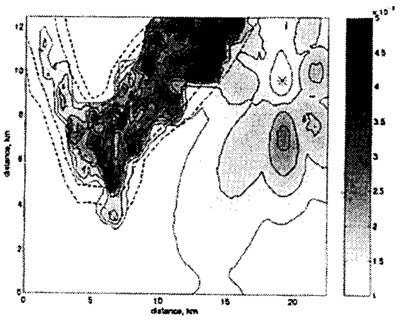

Figure 5. Distribution of horizontal ground motion peak velocity simulated for the « Lancey » earthquake (earthquake 1).

6.2 *Effects of a thrust event located under the Grenoble's basin.*

In the second step, we simulate a magnitude 2.5 earthquake with a thrust mechanism located at 5 kilometers depth under the basin. As already observed in the previous simulation, the structure of the basin and the low seismic velocities in the basin amplify and extend the duration of the ground motion significantly (Figure 6).

The largest peak ground motions (Figure 7) are observed in the central part of the basin (Grenoble downtown). Long durations are also observed in this

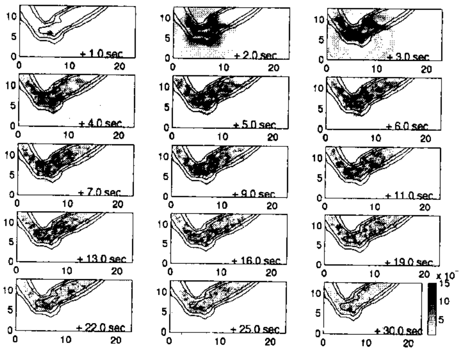

Figure 6. Snapshots of simulated wave propagation in the Grenoble area for earthquake 2 (horizontal velocity, cm/sec).

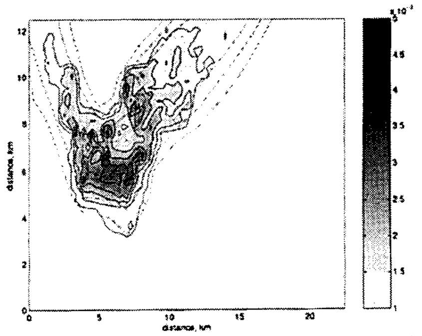

Figure 7. Distribution of horizontal ground motion peak velocity simulated for earthquake 2.

area. The Gresivaudan and the Voreppe valleys are less affected by long duration and high amplification than Grenoble downtown.

7. COMPARISON BETWEEN SIMULATIONS AND OSERVATIONS.

To test the validity and limits of these simulations we compare ground motion data from the Lancey earthquake recorded by Lebrun (1997) with synthetic time histories. Data and synthetics have been band-pass filtered for a frequency band 0.1 to 0.8 Hz. In Figure 8 we show a comparison between the data and synthetics.

The station MUSD located on the rock shows a good agreement between observations and simulations: the duration, the amplitude and the frequency content of motion are in fair agreement. This shows that our representation of the source-moment and source function are correct. The ground motion at MUSD which is located on the border of the basin is also affected by the basin structure and low amplitude late arrivals as seen in the data and predicted by the simulation. Our simulation predict the amplification between the station located on the rock (MUSD) and the stations (HATZ, NEYR, GLAC, SYNC) located in the basin. The peak amplitude of the data are fit within a factor 2 for all the stations. Our simulations reproduces some of the observed late arrivals. Among the stations located in the basin, the best result is obtained for station NEYR. Ground motion simulated at HATZ, SYNC

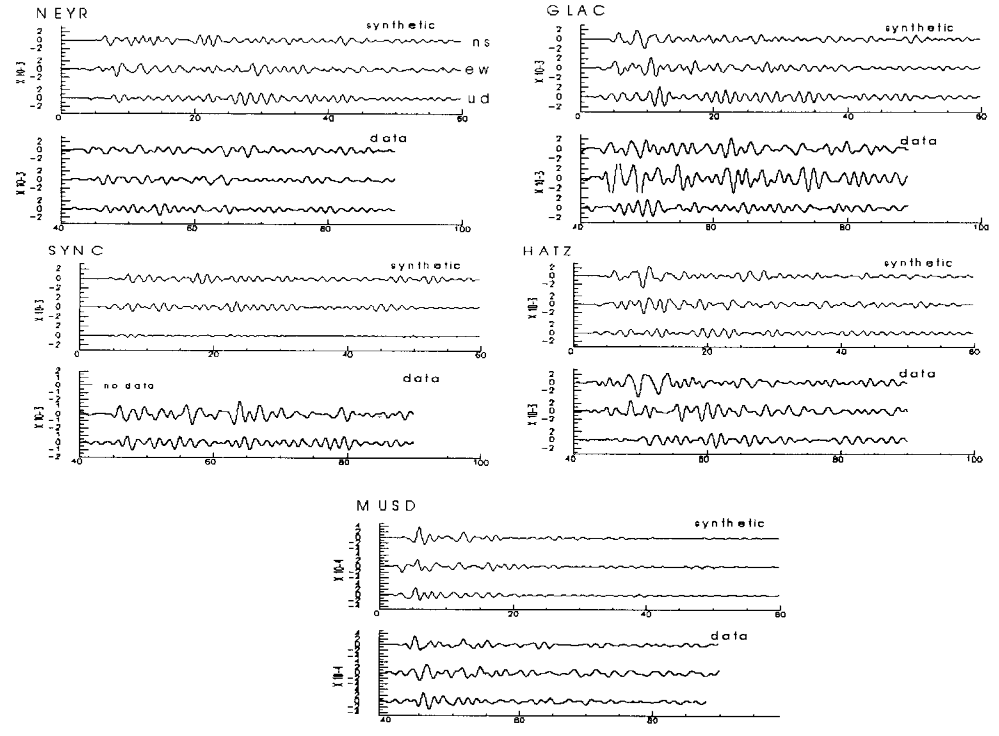

Figure 8. Comparison of simulated and observed 0.2 to 0.8 Hz velocity seismograms for the Lancey earthquake at 5 recording sites.

and GLAC have higher frequency than the observed one. This suggests that the mean velocity of the sediment in the basin could have been overestimated in our model or that we overestimate the attenuation between 0.3 and 0.5 Hz. The worst fits occurs at station GLAC and SYNC where the velocity model on the 3D structure should be improved.

8. CONCLUSION

To evaluate the effect of the 3D geometry of the basin on the ground motion, two earthquakes have been simulated. The first one is the « Lancey » earthquake recorded by several stations (Lebrun, 1997). In a second step we have simulated an earthquake with a similar magnitude than the Lancey earthquake but a thrust mechanisms located at a depth of 5 km under the Grenoble's basin. In both simulations the central part of the basin (Grenoble downtown) increases the amplitude of shaking, traps the seismic waves and prolongs the signal duration. However the 3D structure coupled with earthquakes of different location highlight the large variability of the ground motion for earthquakes with the same seismic moment in the Gresivaudan and Voreppe Valleys. The simulation of the Lancey earthquake matches the peak velocities and observed durations. However the observed duration in the basin strongly depend on the attenuation factors which are poorly constrained by actual data. The observed duration could also depend on several factors that are not taken into account in the actual simulations (high slope of the edge of the basin or scattering from structural boundaries like the border of the Belledone chain) .

REFERENCES

Clayton, R., and B. Enquist, Absorbing boundary conditions for acoustic and elastic wave equations, *Bull. Seism. Soc. Am.*, *67*, 1529-1540, 1977.

Cushing, M., J.C. Gariel, and A. Levret, Contribution of macroseismicity to the evaluation of site effects., in *Fifth International Conference on Seismic Zonation*, edited by O. Editions, pp. 1490-1497, Nice, France, 1995.

Gamond, J.F., Normal faulting and tectonic inversion driven by gravity in a thrusting regime, *J. of Structural Geology, 16*, 1-9, 1994.

Gignoux, M., L'épaisseur des dépôts quaternaires dans la plaine de Grenoble, *CR de la Soc. Geol. de France*, *8*, p.77, 1944.

Graves, R.W., Simulating seismic wave propagation in 3D elastic media using staggered-grid finite differences, *Bull. Seism. Soc Am.*, *86*, 1091-1106, 1996.

Lebrun, B., Les effets de site: étude expérimentale et simulation de trois configurations, 194 pp., Université Joseph Fourier, Grenoble, 1997.

Levret, A., J.C. Backe, and M. Cushing, Atlas of macroseismic maps for French earthquakes with their principal characteristics, Natural Hazard, 10, 19-46, 1994.

Pitarka, A., K. Irikura, T. Iwata, and H. Sekigushi, Three dimentional simulation of the near fault ground motion for the 1995 Hyogo-ken Nanbu (Kobe) , Japan, earthquake, Bull. Seism. Soc. Am., 88, 428-440 1998.