Basin-related effects on ground motion for earthquake scenarios in the Lower Rhine Embayment

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Abstract

The deterministic calculation of earthquake scenarios using complete waveform modeling plays an increasingly important role in estimating shaking hazard in seismically active regions. Here we apply 3-D numerical modeling of seismic wave propagation to M6+ earthquake scenarios in the area of the Lower Rhine Embayment, one of the seismically most active regions in central Europe. Using a 3-D basin model derived from geology, borehole information and seismic experiments, we aim at demonstrating the strong dependence of ground shaking on hypocenter location and basin structure. The simulations are carried out up to frequencies of ca. 1 Hz. As expected, the basin structure leads to strong lateral variations in peak ground motion, amplification, and shaking duration. Depending on source-basinreceiver geometry, the effects correlate with basin depth and the slope of the basin flanks; yet, the basin also affects peak ground motion and estimated shaking hazard thereof outside the basin. Comparison with measured seismograms for one of the earthquakes shows that some of the main characteristics of the wave motion are reproduced. Cumulating the derived seismic intensities from the three modeled earthquake scenarios leads to a predominantly basin correlated intensity distribution for our study area.

Keywords: earthquake scenarios, deterministic shaking hazard, numerical wave propagation, seismic intensities, site effects

Introduction

The study area, the Lower Rhine Embayment (LRE) is part of the European rift system (Figure 1) and shows moderate intraplate seismicity. On average, several earthquakes of magnitude 5 and higher occur every century (Reamer and Hinzen, 2004). The LRE cuts as a sedimentary basin into the Rhenish Massif with a general NW-SE strike of the major faults. In its western part, the Roer Valley Graben (RVG) sediment layers exceed 1.5 km thickness. The whole basin is densely populated and highly industrialized with large petrochemical and chemical plants. Basins of such structure are known to significantly amplify ground motions during earthquakes. Striking examples are the damages in the Mexico City, M 8.1, Michoacan earthquake (Sing and Ordaz, 1993; Cárdenas and Chávez-Garcia, 2003) and in parts of San Francisco during the M 7 Loma Prieta earthquake of 1989 (Stidham et al., 1999). The variability of ground motion due to lateral variations in sediment thickness has also been studied for various earthquakes in the vicinity of the Los Angeles Basin (e.g., Vidale and Helmberger, 1988; Scrivner and Helmberger, 1994; Olsen et al., 1997; Olsen, 2000).

The LRE was subject to intensive studies on ground motion amplification, especially after the damaging Roermond, 1992, M 5.9 earthquake. Most of these investigations focused on 1D estimation of site amplification based on spectral ratios of local seismicity and ambient noise measurements (e.g., Ibs-von Seht and Wohlenberg, 1999; Scherbaum et al., 2003; Ohrnberger et al., 2004; Hinzen et al., 2004; Parolai et al. 2005). While these techniques give valid estimates on 1D amplification factors, they do not provide information on 3D effects, such as wave front focusing due to low velocity zones or edge-diffracted waves. 3D wave propagation simulations account for such issues and can therefore provide additional information on basin-related seismic hazard. The 1995, M 6.9 Hyogo-Ken Nanbu earthquake near the city of Kobe, Japan, drastically proved the importance of 3D effects on the wave field (e.g., Kawase 1996; Pitarka and Irikura, 1996a). Numerous other studies concentrated on this issue (e.g., Boore, 1972; Olsen et al., 1995a, 1995b; Pitarka and Irikura, 1996b; Graves, 1998; Satoh et al., 2001).

Several recent studies on southern California and the Los Angeles basin showed significant lateral variations of ground motion due to 3D geologic structures (e.g., Olsen et al., 1997; Olsen, 2000; Peyrat et al., 2001; Eisner and Clayton, 2002; Komatitsch et al., 2004). Incorporating complex kinematic finite sources, advantage

is taken of the detailed knowledge of the corresponding earthquake source processes due to dense strong motion observations as well as the high-resolution 3D model of the Los Angeles Basin (Magistrale et al., 1998, 2000; Süss and Shaw, 2003).

In this pilot study of earthquake ground motion in parts of the LRE the basin structure is represented by a 3D model with depth dependent velocity and uniform attenuation upon layered bedrock. The scenario earthquakes are modeled as finite-source moment tensors.

We present three simulations of damaging earthquakes that occurred in the LRE within the last 250 years using a 3D staggered-grid finite-difference technique (e.g., lgel, 1995; Graves, 1996). The spatial and temporal discretization allows us to simulate the wave field for frequencies up to 1 Hz. The main goal of this study is to investigate the complexity of the resulting wave field and the effects of amplification and prolonged shaking due to 3D effects of the sedimentary basin in the study area. The key questions we want to address are: (1) How does the basin model alter the peak values of ground motion? (2) Where – with respect to the basin geometry – are amplifications to be expected? (3) How does the basin affect hazard estimates derived from peak ground motion?

The results from this study reveal strong lateral heterogeneity in the distribution of peak ground velocities (PGV) and shaking duration due to the basin topography and the source location. Seismogram sections illustrate the different effects of amplified and prolonged shaking and demonstrate the intrinsic three-dimensional nature of the phenomena. Regions dominated by intraplate tectonics are characterized by low deformation rates and long recurrence times for large earthquakes. Consequently, the strong motion database is sparse for such areas and shaking hazard assessment is afflicted with large uncertainties. Our results therefore further emphasize the need for detailed 3D basin models and accurate ground motion modeling in order to assess seismic hazard in the LRE and similar regions of interest.

Study Area – The Lower Rhine Embayment

Location

The LRE is a sedimentary basin located in the northwestern part of Germany, parts of the Netherlands and Belgium. Its shape is roughly a Northwest opening triangle. The northern extension is called West Netherlands Basin ending at the North Sea coast. The close up map in Figure 1 illustrates its location in Central Europe, while the map in Figure 2 gives an impression of the dense urban agglomeration in this area.

Tectonics

The RVG is located within the northern parts of the Cenozoic rift system of central and Western Europe, which spreads over 1100 km from the North Sea coast to the Mediterranean. A detailed discussion about the Cenozoic rift system can be found in Ziegler (1994). Seismicity in this RVG is driven mainly by regional plate interactions leading to a maximum horizontal compressional stress in NW-SE direction. As the RVG is roughly parallel to this direction, it shows a significant active extension (Geluk et al., 1994). The result of this extension regime is the accumulation of thick sedimentary basins as the LRE and a moderate seismicity leading to earthquakes with predominantly dip-slip mechanisms. Arrows in Figure 1 indicate the main directions of extensional stress in this region. Stress inversion from fault plane solutions of 110 earthquakes in the northern Rhine area (Hinzen, 2003) indicates a subvertical direction of the largest principal stress in the LRE. However, in the neighboring Stavelot-Venn Massif and Rhenish Massif the maximum principal stress is subhorizontal. The maximum horizontal compressional stress in the area has a trend of 125°.

Seismicity

Together with the Belgian zone, the RVG belongs to the seismically most active onshore zones in northwestern Europe. Its active tectonic extension regime makes it to one of the few areas in Europe north of the Alps producing damaging earthquakes. From historical records more than 20 damaging earthquakes in the past 300 years are known (Schwarzbach, 1951). Local magnitudes of the strongest historic earthquakes, Düren 1976 and Verviers 1692 were estimated to 6.4 and 6.8 (Hinzen and Oemisch, 2000), respectively. Probabilistic seismic hazard analysis of this region estimates an intensity VII occurrence rate of 10⁻³ per year (Rosenhauer and Ahorner, 1994). The main present day seismic activity is observed in the RVG and its surroundings. Within the last century 38 earthquakes with magnitudes $M_L \ge 4.0$

occurred in the northern Rhine area (Reamer and Hinzen, 2004). Since 1755, the RVG and its extensions have experienced at least nine earthquakes with $M_L \ge 5.0$. The earthquakes of Düren, 1756 M_L 6.1, Euskirchen, 1951 M_L 5.7 and the 1992 Roermond event M_L 5.9 have been chosen as scenarios in this study. The M_L 4.9 Alsdorf, July 22nd, 2002 event is by now the only larger local earthquake recorded by the seismic network in the LRE. Analysis of this earthquake and modeling of the observations allowed the appraisal of the current basin model (Ewald, 2005). It was shown that – despite of local discrepancies – peak ground motions for frequencies up to 1 Hz were reproduced for stations located throughout the basin with a maximum deviation factor of two.

Basin Model

The sediments in the Lower Rhine Embayment have accumulated since the Carboniferous age. This long period results in a complex stratigraphy with locally varying thickness of the different sequences (Geluk et al., 1994). In this study the stratigraphy is represented by a 3D structure of soft sediments embedded by a layered bedrock model. Seismic velocities within the sediments increase smoothly with depth. As velocity depth function we used the relation

 $V_{S}=118^{*}(z+1)^{0.37}$ (Parolai et al., 2002)

for the shear waves, P wave velocities were derived assuming a V_P/V_S ratio of $\sqrt{3}$. For computational reasons the model was limited to a lowest shear wave velocity of 1000 m/s. Although recent research supports lower shear wave velocities in the Cologne Basin (Hinzen et al., 2004) this model was chosen as a compromise, in order to fulfill computational requirements as well as account for the gradient character of sediment velocities with depth. Attenuation within the sedimentary layer is modeled with a uniform quality factor of Q_S=50 for shear waves and Q_P=100 for P waves, using a standard linear solid approach (Robertsson et al., 1994).

A depth-to-basin map was compiled from seismic profiles as well as borehole data with horizontal resolution of about 1500 m (Scherbaum et al., 2003; Hinzen et al., 2004). The model parameters are outlined in Table 1. Figure 2 includes a contour plot of the basin model.

Simulated Earthquake Scenarios

In order to demonstrate combined path and site effects and the resulting complexity in basin amplification the chosen events are all located close to the basin edges but in different epicentral regions. In the following the main features of these events are summarized:

The Düren earthquakes of December 27, 1755 and February 18, 1756 with highest intensities estimated from historical records as $I_{MM} = VIII$, local magnitudes $M_L = 5.8$ and $M_L = 6.1$, respectively, occurred 10 km south-west of the town of Düren between the LRE and the mountain range of the Stavelot-Venn Massif. Their epicenters are close to each other and are most probably correlated with the active Feldbiss fault system, one of the southwestern border faults of the RVG (Meidow and Ahorner, 1994). The earthquake of 1756, which is chosen as one scenario in this study, is the strongest historical event in this region. The Euskirchen earthquake of March 14, 1951, reaching intensities of $I_{MM} = VII-VIII$, and local magnitude $M_L=5.7$ is located at the western border of the LRE near the town of Euskirchen.

The Roermond earthquake of April 13, 1992, with highest intensities I_{MM} = VII and local magnitude M_L = 5.9 is located near the town of Roermond in the southern part of the province Limburg in the Netherlands. It ranges among the strongest instrumentally observed earthquakes in Central Europe. The estimated recurrence interval for an earthquake of this size in the LRE is ca. 200 years (de Crook, 1994). Source parameters for the Roermond event are estimated in various studies (e.g., Ahorner, 1994; Braunmiller et al., 1994; Pelzing, 1994). Scherbaum (1994) discussed the intrinsic variability of the source parameters using a stochastic method. The earthquake scenarios were modeled as circular finite-sources with source radii derived with Brune's model using an assumed stress drop of 10 bars (Brune, 1970). The source parameters of the modeled earthquakes are summarized in Table 2.

Simulation Results

Visco-elastic simulations of the earthquake scenarios described above are employed using a finite-difference scheme on staggered grids (e.g., Igel, 1995, Graves, 1996). The model space was discretized into ca. 417 million grid cells with a uniform horizontal spacing of 100 m and a varying vertical spacing ranging from 50m in the sedimentary layer to 120 m in bedrock. With the numerical scheme used waveform modeling was possible for frequencies up to 1 Hz. A summary of modeling parameters is given in Table 3.

3D simulations of wave propagation provide the complete wave field over the whole model area. This allows gathering different aspects of ground motion, such as peak

values, static deformation and/or rotation, at any point in time and space. In this section, results from simulations of the scenario earthquakes are presented with special emphasis on the lateral variation of the effects at the surface as a function of source location with respect to the basin structure.

In Figure 3 snapshots of the EW velocity component at the surface for the three earthquakes are shown. Basin-related effects on surface ground motion are particularly visible by comparing waves propagating into and away from the basin. Specific details of the wave field are marked with labels (A - D). Especially for the Euskirchen event with its epicenter south of the basin the resulting wave field splits into a nearly undisturbed part south of the basin and the scattered wave field within the basin north of the epicenter. Diffraction and scattering of seismic waves inside the basin is prominent for the Düren earthquake (A). Trapped energy and reverberations inside the basin can be observed for all scenarios (e.g., (B) Roermond earthquake). For wave fronts crossing the deepest parts of the basin focusing effects outside the basin margins are apparent for the Düren simulation (C). This effect also results in amplified PGVs (see Figure 8a). Inside the basin channeling of energy is present (D).

In order to further illustrate the basin effects, synthetic seismograms are extracted on a profile along the entire basin. In Figures 4 - 6 three component synthetic seismograms recorded at 30 virtual stations along profile A - A' (Figure 2) are shown together with a profile of basin depth. Recordings for each earthquake scenario are scaled individually. To allow the comparison between the three scenarios individual PGV values are given in the top-right boxes. The influence of basin structure on both amplitude and shaking duration is notable on all components and for all scenarios. The most prominent features are marked with labels (A - E):

The most obvious feature of the seismogram sections consists of the higher amplitudes inside the basin (A) compared to those recorded outside the basin (B) at the same distance. However, differences in amplification as a function of source location are notable by comparing the individual scenarios. For the Roermond earthquake the region of strong amplification is much more restricted (D) than for the other two examples. Prolonged shaking is present for the Euskirchen scenario especially in a region between the two deep depressions of the basin (C). The seismogram section for the Düren earthquake shows a single extremely large amplitude above the steep flank of the basin (E). Profiles for the north-south component (figure 5) show similar effects as for the east-west horizontal component.

For the Düren earthquake this component reveals very low frequency content in the northern part of the basin due to superposition of trapped waves that was notable in the snapshots. In figure 7 the same profile sections are shown for the vertical ground motion. This component significantly reveals the dependence of amplitudes on basin depth. Especially for the Euskirchen simulation strong resonant amplification is present.

Earthquake scenario simulations can be calibrated and verified by comparison with observations. Unfortunately, most of the data recorded for the Roermond, 1992 earthquake are clipped for stations located inside the study area. Even though some similarities between observations and simulations are shown below, these should not be overemphasized. Figure 7 compares the three component synthetics with observations for the station TGA at an epicentral distance of 56 km near the city of Bergheim (Figure 2). Synthetics and observations show similarities in four important features: (1) matching P and S wave onsets in all components; (2) amplitudes are reproduced with a maximum difference of ca. 50 %; (3) the amplitude ratio between P and S waves is similar. For both horizontal components the synthetic ratio (peak-to-peak amplitudes used) is within 20 % of the observed values; (4) envelopes that are influenced by the source mechanism and the underground structure look similar except for the EW component. An excellent match in waveforms is found for the P and S-wave arrivals on the north-south component and also late arriving energy on the vertical component is reproduced.

Numerical simulations allow the mapping of ground motion parameters such as PGV. From these PGV maps, seismic intensity can be estimated and compared with observed intensities. For historical earthquakes macroseismic intensities are usually the only source of information. Seismic intensities can be estimated from PGV through empirical relations according to the following equations:

 I_{MM} =2.10 log (PGV) +3.40 for intensities below V, and

 I_{MM} =3.47 log (PGV) +2.35 for intensity V and above (Wald et al., 1999).

Figure 8 shows intensity maps for the earthquake scenarios. These maps clearly reveal amplification due to the three-dimensional basin effects, which were apparent also in the snapshots. For the Düren simulation high intensities outside the basin are visible and can be related to the shadow effect of waves passing through the deepest parts of the basin (labeled B in Figure 8a). However, this simulation reveals in general a strong correlation of intensity with basin depth (A). Channeling of wave energy identified for the Euskirchen simulation leads to amplification in the

corresponding regions (C in Figure 8b). For the Roermond earthquake strong correlation of intensity with basin depth is notable, especially on the small depressions southeast of the main basin (D in Figure 8c). The most notable effect is a strong amplification of ground motion in a comparably shallow part of the basin due to the refraction of wave energy at the deeper basins flanks (E in Figure 8c). The waves generating these amplitudes are also prominent in the snapshots (Figure 3). Comparing this result to the Düren simulation, where these flanks worked as a boundary for the wave energy (F), the sensitivity of basin response to source location with respect to prominent features of basin topography is demonstrated. Generally, the simulations produce overestimated peak intensities in the epicentral areas.

For the Roermond, 1992, earthquake a comparison of seismic intensities derived from synthetic peak ground velocities according to the formulae given above with the observed intensity distribution is presented in Figure 9. Area wide synthetic seismic intensities are plotted in the same color scale as in previous figures. Black lines indicate the isoseismals derived from observations with seismic intensity levels as indicated. These isoseismals are derived by Meidow and Ahorner (1994) from a set of 2000 macroseismic reports at 600 different locations. From this set of observations, also the isoseismal radii shown in figure 8 were derived. The results are discussed in detail in the following section.

Three dimensional basin effects can significantly increase durations of ground motion. Several studies have found surface wave generation at basin edges particularly responsible for this phenomenon (e.g., Vidale and Helmberger, 1988; Olsen et al., 1995b). The basin effects on the duration of shaking are addressed with a simple approach: For each grid point at the surface the time is recorded at which each component of the ground velocity exceeds a certain threshold value for the first time. The last time this value is again exceeded is also recorded. As threshold value a velocity of 4 cm/s is chosen (modified after Olsen, 2000), which corresponds to an intensity of IV-V according to the relations given above. The length of the determined time window is called the shaking duration and mapped for each scenario individually. In Figure 10 isolines of shaking duration for the three simulated earthquakes are shown, colors indicating duration in seconds. Solid black lines depict the outline of the sedimentary basin.

Except for the Düren simulation the observation of enduring strong ground motion is restricted to the basin area. Besides this general prolongation of shaking duration different spots of maximum duration can be identified from the contour plots. Some of

these spots appear to be related to corners in basin topography wherein wave energy got trapped (A in Figure 10). Spots labeled (B) appear above strong gradients in basin topography, whereas another group of extended duration spots is due to the channeling effects notable also on snapshots and intensity maps (C).

Although the number of simulated earthquakes is not sufficient for a representative seismic hazard evaluation a first attempt is made to combine results of the individual scenario simulations. Occurrence frequencies for the three events are not considered and therefore the seismic intensity values from the simulations are equally weighted and stacked. The resulting intensity distributions are called cumulative intensity maps in the following.

Figure 11 shows the evolution of this cumulative intensity map throughout the series of the simulated earthquakes. The series starts with the intensity map from the simulation of the Düren, 1756 earthquake as shown before (Figure 11a). The Euskirchen, 1951 earthquake does not significantly change the cumulative energy map. This is reasonable because epicenters are located relatively close to each other and the magnitude of the 1951 event is distinctly lower than the one from 1756 (Figure 11b). After the Roermond, 1992 earthquake the northwestern part of the cumulative intensity map is dominated by this event. However, after this series of only three earthquakes the source related patterns of the cumulative intensity starts to change over into a more basin correlated shape (Figure 11c). The probabilistic seismic hazard model for the LRE presented by Rosenhauer and Ahorner (1994) is compared to the cumulative intensity map (Figure 11d). The black lines with numbers indicate the probabilistic earthquake intensity isolines with an occurrence rate of 0.001 per year. Except for the region near Roermond, which is dominated by the 1992 earthquake, the shape of the isolines roughly matches. The values can only be compared qualitatively because in the cumulative intensity map the occurrence rate is not taken into account.

Discussion

The simulated ground motions show distinct differences in amplification on the three components of ground motion with varying source location.

For the Düren, 1756, earthquake the maximum intensity from the simulation is about one unit higher than the maximum intensity of VIII as estimated from historical records (Meidow, 1994). In general, the simulation results overestimate intensities in

the near source area. For large parts of the basin, very large amplitudes and consequently high seismic intensities are produced above the basin edges. Two explanations have to be considered: 1) Although the sedimentary structure is modeled with a velocity gradient an overestimation of the velocity contrast at the basin margin is still possible. Smoother horizontal velocity gradients or scattering may reduce focusing effects above basin edges that are responsible for the large amplitudes. 2) The number of observations for the Düren, 1756, earthquake is limited and the observations themselves may be uncertain. Information on local disproportional high amplitudes might have been lost. However, the mean radii for $I_{MM} = VI$, and $I_{MM} = V$ are better matched than the maximum intensities near the epicenter. Amplified ground motions in a shadow zone north east of the basin are visible both, in the snapshots and intensity maps. This effect is matched by macroseismic observations from historical reports (Meidow, 1994).

The intensity map of the Euskirchen, 1951, earthquake shows a sharper distribution as it is the case for the other two scenarios due to lower seismic moment and focal depth. The black circle (Figure 8) demarks the mean isoseismal radius for intensity V with 51 km, as it is derived from macroseismic maps (Meidow and Ahorner, 1994). This intensity level is well matched by the simulation result. The observed epicentral intensities of I_{MM} = VII-VIII are again significantly lower than the ones obtained by the simulations.

The intensity maps for the Roermond, 1992, earthquake shows strong correlation of high intensities with the contours of the sedimentary basin. The peak intensities from the simulations at the epicenter (IX) are higher than the reported ones, which are estimated after correction between VII and VIII (Meidow, 1994). The mean isoseismal radii for $I_{MM} = VI$ (approximately 45 km) and $I_{MM} = V$ (100 km) are demarked as black circles (Meidow, 1994) and correspond roughly with expected radii derived from the synthetic results. Meidow and Ahorner (1994) used a set of 2000 macroseismic reports to derive isoseismals for the Roermond, 1992 earthquake. In Figure 9 these isoseismals (solid black lines with indicated intensity levels) are compared with intensity maps derived from synthetic peak ground motion.

Before discussing the considerable differences, it must be considered that seismic intensities compared herein are achieved with completely different approaches resulting in a number of limitations, which are briefly discussed in the following. Whereas the synthetic seismic intensities are derived from PGVs of the simulated wavefield at each point, the observed isoseismals are derived from macroseismic

reports, which intrinsically include effects like the local building conditions. In fact, the distribution of macroseismic reports reflects strongly the urban areas of the region, in particular south of the Roermond epicenter. As discussed above the numerical model for these simulations does not contain shallow structures particularly in the region close to the river Rhine. In consequence, individual site conditions in these areas are not included in the synthetics. Proper validation of the modeling results should be done with observations in the same frequency band as carried out by Ewald (2005). Due to the lack of observational data this cannot be done for the earthquakes investigated in this study. The macroseismic reports do not allow determining the period causing the reported shaking level.

Maximum observed intensity VII and above is concentrated in a relatively small area of 92km² shifted eastward with respect to the instrumental epicenter (Meidow and Ahorner, 1994). This area is elongated in NW-SE direction and it coincides with the inner flank of the Peel boundary fault. The areas of highest intensity produced by the simulations match this observation and the geometry supports the interpretation that these high levels of ground shaking are due to geometric amplification at the steep basin edges below this area. The simulations overestimate absolute maximal values of seismic intensity. It must be stated that the observed maximum levels of seismic intensity are considered as significantly too low, resulting in an estimation of hypocenter depth as deep as 26 km from these observations compared to the instrumentally calculated depth of 17 km which was used for the simulation setup (Meidow and Ahorner, 1994). Isoseismals for intensity VI are roughly matched by the simulation results except for the area with lower intensity north-east of the epicenter. The abrupt change in the shape of the related isoseismal in that area is constructed from only three macroseismic reports and is not supported by the simulation results. Islands of intensity VI south of the basins margins also refer to singular observations and are not matched in particular by the synthetics. The generally NW-SE trending shape of isoseismals for intensity V-VI are not recovered by the simulation results and the same accounts for the area of intensity VI south of the city of Bonn (see Figure 1 and 2). The observation of elevated ground motion in this area may be related to the presence of shallow soft sediments along the river Rhine, which are not included in the velocity model used in this study.

From a comparison of the seismogram sections discussed above (Figures 4 - 6) it can be understood that basin related amplification effects show strong variability with source location. This is especially significant for the east-west component

seismograms shown in Figure 4. This variability has also a clear expression in above discussed intensity maps derived from synthetic peak ground motion. The intensity maps in Figure 8 show a stronger correlation between ground motion amplitude and basin depth for the Roermond event than for the other simulated earthquakes. Seismogram sections of the horizontal east-west component further demonstrate this behavior. While the amplitudes of the Roermond earthquake show the highest values above the deep basin depressions the ground motion of the Düren and Euskirchen earthquakes have maximum amplitudes above the steep basin edges. Scattering of wave energy is the dominant effect for these recordings. Reverberations from basin edge reflections dominate the duration of strong ground motion for the Euskirchen event. Because the amplitudes are generally lower for this earthquake, the influences of the basin are more significant than for the other earthquakes. Reflected phases can be identified for both surface and body waves. For the Düren earthquake with its greater magnitude and deeper hypocenter the basin influence is more visible in amplitudes than in duration. Due to the small epicentral distance, most of the effects in the northwestern part of the basin are invisible. Strong influence can be seen at the southeastern depression of the basin.

The NS component (Figure 5) of ground velocities in general display similar effects than the EW component. However, some interesting differences appear. For the Düren earthquake effects can be seen in amplitude as well as in duration of shaking. Reflected phases from the basin margins can be identified. Although the duration of shaking is significantly prolonged, the wave field is much more restricted in time compared to the Euskirchen event. The largest amplitudes are present 20 to 30 s after the first arrivals. For the Euskirchen event, strong reverberations are present and single phases cannot easily be identified from the NS component. A dominant onset on a single station at receiver offset 60 km is notable (labeled (A) in Figure 5). With respect to snapshots and amplification maps, this can be explained by focusing effects of the steep depression. Basin effects on the seismograms from the Roermond earthquake show strong correlation between amplitude and basin depth. The effects in shaking duration are not as dominant as for the other two earthquake scenarios. Note the difference in amplitudes of shear and surface waves inside and outside of the basin.

Correlation of amplitudes with basin depth is not so obvious for the vertical component ground motion as it is the case for the horizontal components (see Figure 6). The main effects consist of longer shaking duration and generation of surface

waves with high amplitudes. The vertical seismograms from the Düren event show longer shaking durations and less amplitude variations than the horizontal component seismograms.

For the Euskirchen event the change in frequency content mentioned above for waves outside the basin is also visible (Figure 4, middle panel D). For the Düren event a strong correlation of wave amplitude with basin depth is notable. Scattering effects are not as prominent as they were on the horizontal components. For the Roermond simulation, most of the effects are invisible due to the spherical correction used in the profile plots. This indicates them as being weaker compared to horizontal component seismograms.

For the single station TGA where a direct comparison of synthetic and observed seismograms is possible we achieve some significant similarities. First, the onsets of P and S waves are in good agreement. Secondly, the amplitude ratio of P and S waves is matched to within 20 %. The maximum difference of about 50 % in absolute peak amplitudes is expected to be due to uncertainties on earthquake source parameters (see Table 4), as well as the fact that our rheology model lacks the uppermost low velocity layers. The fact, that peak amplitudes as well as some features such as the characteristics of late arriving energy on the vertical component can be roughly reproduced demonstrate the potential of 3D waveform modeling for hazard assessment in regions of moderate seismicity, where the lack of observations complicates shaking hazard estimates. Deriving the shaking duration from the observed and synthetic seismograms according to the approach used in this study results in a value of 11.5 seconds for the observation and only 2.7 seconds for the synthetics. However, it must be stated that this approach is designed in order to provide an area wide rough estimate on this quantity at run time of the simulation. Inevitably, results are subject to large spatial heterogeneities and are also depending in a nonlinear fashion on the chosen threshold value.

Visual comparison of the waveforms of the corresponding traces in Figure 7 reveals that a small reduction of the threshold value would result in significantly better agreement between observations and synthetics. From area wide mapping of shaking duration shown in Figure 10 it can be understood that for most locations close to the station TGA later arrivals contribute to the effective shaking duration and the estimates on this parameter are between 15 and 25 seconds. This is in much better agreement with the observed value.

Conclusion and Outlook

In this study basin related effects on waveform, PGV and shaking duration from earthquake ground motion in the Lower Rhine Embayment are investigated. We employed an approach incorporating a fairly simple 3D basin model and moment tensor finite source to be capable of simulating both recent and historical earthquake scenarios, the latter limiting the knowledge on source parameters. Even with the simple physical model used, we obtain a high level of wave field complexity caused by the basin structure with strong dependence on the location of the individual events. This emphasizes the importance of accounting for both site and path effects in seismic hazard studies of basin regions. Furthermore, it can be concluded that the use of station correction factors obtained by 1D approaches may lead to errors in estimating earthquake source parameters, such as local magnitudes.

Although we found some agreement between synthetic and observed waveforms for the one station where a direct comparison was possible, strong motion records for further events from various stations are necessary to check our simulations against observations. On July $22^{nd} 2002$ an M_L 4.9 earthquake occurred within the study area near the town of Alsdorf. Strong motion recordings of this event on a recently installed seismic network in the Lower Rhine Embayment were modeled with the same approach used in this study and allowed a detailed appraisal of the velocity model (Ewald, 2005). Another limitation is the small number of three earthquake scenarios, which makes it possible only to give a flavor of the variability of ground motion with different earthquake locations. Our results, especially the strong dependence of amplified ground motion on source location demonstrates the intrinsic three-dimensional nature of such effects and therefore confirm the need for 3D simulations to assess shaking hazard in our study area.

Acknowledgements

We thank the Leibniz Rechenzentrum in Munich for providing access to their computational resources, especially the Hitachi SR8000 supercomputer. We thank R. Wang (GFZ Potsdam) for providing his reflectivity code that we used for testing the accuracy of our finite-difference schemes. This work was funded partly by the MunichRE Reinsurance Company and through the KONWIHR project (NBW).

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Layer	Depth	P-wave velocity	S-wave velocity	Density
	[km]	[m/s]	[m/s]	[kg/m ³]
Sediment	0 – 1.9	1730-3110	1000-1800	2200
Bedrock	0-1	5500	3175	2800
	1-2	5750	3320	2800
	2-4	5980	3452	2800
	4-10	6120	3533	2800
	10-18	6300	3637	2800
	18-26	6360	3672	2800
	26-30	7350	4244	2800

Table 1: Model parameters

	Düren	Euskirchen	Roermond
Date	February 18, 1756	March 14, 1951	April 13, 1992
Latitude	50°45'	50°38'	51 °10'
Longitude	6°21'	6°44'	5°56'
Depth [km]	14	9	17
Strike [`]	135	110	120
Dip [`]	70	80	70
Rake [`]	270	270	260
Rise Time [s]	1.0	1.0	1.0
M ₀ [10 ¹⁶ Nm]	14.0	3.7	7.5
Source Radius [m]	3900	2500	3200

Table 2: Earthquake Source Parameters

Spatial discretization [m] horizontal	100
Spatial discretization [m] vertical	50 120
Temporal discretization [s]	0.0198
Lowest S-wave velocity [m/s]	1000
Grid size (physical model)	1400 x 1400 x 180
Grid size (computational model)	1444 x 1444 x 200
Model size (physical)	140km x 140km x 30km
Number of time steps	16960
Simulation time [s]	60
Memory usage [GB]	110
Computation time [h]	12
Using 32 nodes of the Hitachi SR8000 at	
Leibniz Rechenzentrum, München	

Table 3: Summary of simulation parameters

Author	Lat	Lon	Depth	Strike	Dip	Rake	M ₀ [10 ¹⁶ Nm]
Braunmiller et al., 1994	51.17	5.97	13	139±2	58±1	263±2	9.4±0.5
Braunmiller et al., 1994	51.17	5.97	18	138±1	58±1	262±2	9.2±0.4
Ahorner, 1994	51.170	5.925	14.6	124	68	270	9.8
Paulssen et al., 1992	51.17	5.97	20	124	70	270	-
Dziewonski et al., 1993	51.17	5.94	15	143	68	273	13.3

Table 4: Summary of source parameters for the Roermond, 1992, earthquake fromdifferent studies.



Figure 1: Geographic location of the study area. The Lower Rhine Embayment developed inside the Lower Rhine Graben, the northwestern part of the Rhine Graben System. Direction of regional maximum compressive crustal stress is NW-SE, resulting in a zone of tension oriented as indicated by black arrows (Hinzen, 2003). The main tectonic blocks are divided by normal faults. The location of several large earthquakes within the study area is shown (asterisks).



Figure 2: Detailed map of the study area. Red stars indicate epicenters of the simulated earthquakes. Grayscale shading depicts regional topography whereas urban areas are colored in dark gray. The sedimentary basin depth is indicated by dashed isolines in 100 m intervals. The basin reaches a maximum depth of about 1900 m. Seismic network stations are marked as stars and by their station codes. A synthetic profile A-A' with 30 receivers is shown as black dots.



Figure 3: Snapshots of the simulated wave field (EW component) for three earthquake scenarios in the Lower Rhine Embayment (Düren, 1756, Euskirchen, 1951, and Roermond 1992, see text for details) at 6, 14, 22, 30, and 38 s simulation time. The covered region is the same as in Figure 2. Amplitudes are scaled individually for each scenario. Complex wave front distortions due to the low velocity structure of the basin are observable for all events. Note in particular the different extent of prolonged shaking due to reverberations inside the basin. Some regions discussed in text are indicated by capital letters.



Figure 4: Velocity seismograms of the EW component along profile A-A' (see Figure 2) for the three simulated scenarios. The black line below zero indicates basin topography along the profile. Recordings for each earthquake scenario are scaled individually. Peak ground velocities along the profile are given for the individual scenarios in the top-right corner to allow comparison between the scenarios.



Figure 5: Same as Figure 4 for the NS component velocity seismograms.



Figure 6:Same as Figure 4 for the vertical component ground velocity seismograms.



Figure 7: Comparison of simulated (red lines) and observed (solid lines) seismograms at the station TGA (near Bergheim) for the ML5.9 Roermond, 1992 earthquake.



Figure 8: Modified Mercalli Intensity for the Düren, Euskirchen and Roermond earthquake calculated from simulated PGV values. Mean isoseismal radii from observations are marked as black circles (see text for details).



Figure 9: Comparison of synthetic and observed seismic intensities for the M 5.9, 1992 Roermond earthquake. Seismic intensities derived from synthetic peak ground velocities (see text for details) are color-coded. Black lines indicate isoseismals derived from observations (Meidow, 1994).



Figure 10: Isolines of shaking duration for the simulated earthquakes. Colors indicate shaking duration in seconds. Solid black lines depict the shape of the sedimentary basin. Note the strong correlation between prolonged shaking and basin depth.



Figure 11: Evolution of combined intensities throughout the sequence of the simulated three scenarios (a - c) and comparison with the current seismic hazard map for the study area (d). The black lines with numbers indicate the probabilistic earthquake intensity isolines with an occurrence rate of 0.001 per year (Rosenhauer and Ahorner, 1994).