Numerical modeling for seismic exploration with karstic subsurface structures

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September 21, 2014
Numerical modeling for seismic exploration with karstic subsurface structures

Master of Science Thesis

for the degree of Master of Science in Applied Geophysics at
Delft University of Technology
ETH Zürich
RWTH Aachen University

by

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September 21, 2014
IDEA LEAGUE
JOINT MASTER’S IN APPLIED GEOPHYSICS

Delft University of Technology, The Netherlands
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Dated: September 21, 2014

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Abstract

Seismic data acquired in the presence of karstified limestone formations close the surface is characterized by a low quality. The low quality has an adverse effect on the imaging of the deep subsurface that is a potential target in the oil and gas industry. Karst is an umbrella term encompassing a wide variety of dissolution features: rough topography, dolines, varying soil thickness and caves. All of these features can have a negative impact on the recorded seismic wavefield. Finite difference modeling is used to simulate seismic experiments on a series of 2-D and 3-D models with increasingly more elements associated with karst being incorporated to investigate their effect on the recorded seismic wavefield. Rough topography and fluctuating soil thickness are demonstrated to be major contributors to the deterioration of the quality of the recorded seismic wavefield. Both features coupled with the high acoustic impedance contrast between the soil and the underlying limestone cause most of the seismic energy to be trapped within the soil layer, where it is scattered and diffracted, obscuring reflection events coming from the deeper subsurface. Properly set FK-filters can remove the scattered and diffracted events from the seismic recording, however the reflection events coming from the deep subsurface will not automatically be recovered. In case reflection events are recovered they are not properly aligned, even after refraction statics are applied, creating problems in the stacking. Lateral and vertical velocity variations in the deeper parts of the limestone formation cause the reflection wavefront to diverge from a spherical form, effecting the arrival times.
First of all I want to thank my supervisors, Prof. Maurer, Dr. Manukyan and Dr. Schmelzbach, whose door was always open when I had a question. I would also like to thank Heinrich Horstmeyer for his advise on lowering the amount of working memory my scripts needed to run. Furthermore I would like to thank my parents for financially supporting me in Zurich and last but not least my thanks goes out to my colleagues in FO67.

Swiss Federal Institute of Technology

Willem Ynze Meijer

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Chapter 1

Introduction

1-1 Seismic imaging challenges due to Carbonates in the near-surface

Karstified limestone can have an adverse effect on the quality of the recorded wavefield when the source is situated directly on top of the karst formation. This effect is illustrated by Figure 1-1 and Figure 1-2, both of which are shot gathers from a field campaign in the Middle-East performed in a region with known karstified limestone. Figure 1-1 shows the results of a source located outside of the area containing karst while Figure 1-2 shows a recorded shot gather when the source was placed within the part of the area containing karst. Especially noticeable are the high amplitudes at the small offsets from the source position (a in Figure 1-2), obscuring any potential reflections present. This dataset triggered the CARNEVAL consortium (a collaboration between ETH and industry partners) to investigate the effects of karst on seismic recordings by doing both field tests and numerical modeling of seismic acquisition in areas containing karst. ETH will perform field measurements in September 2014 in the Jura mountains, Switzerland. The Jura mountains are a region with known karstified limestone close to the surface. In this area the consortium will test different geophysical methods to obtain better data quality. This thesis is the numerical modeling part of their investigation into the causes of low seismic data quality. In this thesis the modeling is partially linked to the planned field work campaign.
Figure 1-1: Shot gather from the Middle-East with the source located outside the karstified part of the region (courtesy of the CARNEVAL consortium)
Figure 1-2: Shot gather from the Middle-East with the source located inside the karstified part of the region. a: high-amplitude events potentially obscuring any present reflections. (courtesy of the CARNEVAL consortium)
The high-amplitude events shown in Figure 1-2 (a), that led the CARNEVAL consortium to investigate karst, is a problem that is especially prevalent where high velocity layers are close to the surface, e.g. carbonate or volcanic formations (Pritchett, 1990). In such settings, several problems occur (Taner, 1997; Regone, 1998):

- Only a small portion of the seismic energy is able to penetrate past the boundary between the soil and the high velocity rock
- Strong horizontal ground roll is generated
- Rough topography associated with shallow carbonates can cause scattering of the surface waves further obscuring any reflections

The above mentioned problems do not take into account the possibility that a carbonate formation can be karstified creating features that add to the heterogeneity. Research has been performed on both optimizing the data acquisition and processing to combat these problems. Some research is specifically focused on near-surface carbonates while others try to solve the more general problem of source generated noise, such as ground roll and waves back scattered from rough topography (Strobbia et al., 2014; Edme et al., 2014; Halliday et al., 2010).

Regone (1998) uses geophone arrays in square 2D pattern to identify the 3D coherent noises generated by waves reflecting of an irregular topography and measure its strength compared to the desired signal. The acquired knowledge about the noise can then be used to optimize a proper acquisition setup. In a field study in former Yugoslavic Republic the method was tested to determine the direction of the back scattered waves. Simply by changing the direction of the receiver array they were able to better suppress the undesired back scattered seismic waves.

Taner (1997) suggests a processing flow for seismic data acquired in the presence of shallow subsurface carbonates using conventional processing tools, such as FK-filters and surface consistent deconvolution. Taner’s processing flow manages to deal with a lot of the unwanted events in his shot gathers. However, he does not analyze the effects of back scattered surface waves and if these will cause problems in the processing scheme.

Strobbia et al. (2014) propose a model-based attenuation scheme for scattered and dispersive surface waves. The application of this scheme is not limited to carbonates, but may be used to suppress source-generated noise caused by any high velocity layer close to the subsurface. To start the propagation properties of the surface waves and events caused by refractors are estimated. The properties are then used to build a scattering model to approximate a scattered wave field which is subsequently subtracted from the raw shot gather. Edme et al. (2013) use rotational data for adaptive noise subtraction. The rotational components of the wavefield are estimated using closely spaced conventional vertical geophones. Edme et al. (2014) demonstrate that for a field study in the United Arabian Emirates their method give better results than conventional velocity filtering.

In the fields of hydrogeology and engineering geology a wide variety of geophysical techniques are employed to study karstified areas. The main goal of these studies is to determine groundwater regimes or to locate potential building hazards, e.g. caves and sinkholes (Sumanovac...
Sumanovac and Weisser (2001) evaluate resistivity and seismic methods for hydrological mapping in a karst terrain in Croatia. Both refraction and reflection data are incorporated into the construction of the hydrogeological model. However, there were problems with the reflection data due to the high noise content of the measured data and the very low amplitude of the reflected wave. The analysis of the subsurface relied mostly on the refraction data. De-Giorgi (2014) use seismic refraction tomography (SRT) and ground penetrating radar (GPR) to assess the risk posed by a karstified limestone formation on an existing road in the Apulia region, Italy. The integration of GPR and SRT deliver good results for locating potential geohazards.

A wide range of geophysical methods (GPR, SRT, ERT, etc.) have been used and been proven useful in the investigation of karstified carbonates. However, these investigations are focused on the shallow subsurface. If seismic reflection data was acquired it was ignored due to the bad data quality (Sumanovac and Weisser, 2001). No further inquiries were made into the cause of the low quality of the seismic reflection data. Furthermore the described methods for suppressing the source generated noise work well for the situations described in their respective articles but it remains to be seen if it can be transferred to seismic exploration in the presence of a shallow karstified carbonate formations.

1-2 Thesis goals and outline

The goal of this thesis is to determine the effects of features associated with karst on the recorded seismic wavefield and their effect on the detectability of deep reflectors. In order to investigate the effects of karst I used numerical simulations. The method I used is the finite-difference method which is commonly used to investigate seismic wave propagation in the subsurface (Holliger and Robertsson, 1998; Dehghannejad et al., 2012; Greenhalgh and Manukyan, 2013)

In order to perform finite-difference simulations numerical models are needed. I used the work flow as described in Figure 1-3 to create the models. First the different geological and seismic features associated with karstified carbonate formations were determined. The mostly qualitative descriptions of the geological features were subsequently translated into models containing the interfaces between different layers. The interface models utilized to construct either 2-D models or 3-D block models. The interfaces subdivide the model space into different sections to which a number is assigned (Appendix). These numbers are then used to assign parameters (e.g., P-wave velocity, S-wave velocity and density) to the different parts of the model. This results in a numerical models one for each parameter to be used in the simulation. Finally before running a finite-difference simulations source and the receivers locations are assigned. The vertical positioning of the source and receivers is determined by the air-soil interface and will thus follow the topography. The output of the simulations are shot gathers and/or snapshots of the seismic wavefield modeled in the FD simulation.

In Chapter 2 I will review karst and their expression both on the surface and below the surface. In this chapter I discuss the seismic attributes of carbonate rocks and the extent of carbonate deposits. The numerical simulation method I use in this thesis will be discussed in Chapter 3. The chapter will contain a review on the discretization of the model space and the
Figure 1-3: The workflow for creating both 2-D and 3-D models that can be run through a finite-difference simulation.

formulas used in the FD simulation. The 3-D and 2-D models and the results are discussed in Chapter 4 and 5. In Chapter 5 the differences and similarities between 3-D and 2-D are also mentioned. In Chapter 6 I will use basic seismic processing tools on the 2-D dataset (described in Chapter 5) to see if a deep reflector can be retrieved. Finally, in chapter 7 the results are discussed and some conclusions are drawn. In chapter 7 I will give an outlook on possible further investigations into the effect of karst on the seismic wavefield.
2-1 Carbonate Rocks

A rock is considered a carbonate when at least 50% of the rock matrix is consists of carbonate minerals, calcite (CaCO$_3$), dolomite (CaMg(CO$_3$_2) and/or argonite (CaCO$_3$) (Jennings and Jennings, 1985). The mineral argonite is formed by organisms, but is unstable under standard temperature and pressure conditions and transforms into calcite minerals and is thus not often found in carbonate rocks. Carbonate rocks are subdivided in two categories, namely limestones and dolomites. The first consist mostly out of calcite minerals, while the later contains more dolomite minerals. Limestone is a sedimentary rock but unlike siliciclastic formations it is usually not formed by the deposition of weathering material but rather by the precipitation of carbonate minerals. This can be through abiotic or biotic processes. In the first case the Ca$^{2+}$ and CO$_3^{2-}$ go out of solution and precipitate as calcite minerals, because the solubility limit in water is reached. The second and most prevalent situation is biotic precipitation. It can be subdivided into two separate categories, biotically controlled and biotically induced precipitation. In the first case the organism produce carbonate skeletons which will form the rock. In the second case organisms cause calcite to precipitate but they do not create it themselves. There are many environmental parameters that control the precipitation, e.g. light, nutrients and temperature. These parameters control the shape and size of the deposits. The lateral extent of limestone deposits can be up to hundreds of kilometers. Dolomite rocks form through the dolomitization of limestone. In this process calcite minerals are transformed into dolomite minerals, which creates secondary porosity as the dolomite minerals take up less space than calcite.

2-2 Karst

Karst is formed when chemical weathering dominates over mechanical weathering. One form of chemical weathering is the dissolution of rock into water. As calcite is very soluble in water containing acids, this erosion process can become more important than the mechanical
weathering. The major source of acids is dissolved CO$_2$ (Frumkin, 2013). Other factors contributing in the dissolution of calcite are sulfuric acid and various microbiological agents (Frumkin 2013). This is also the case for dolomite. Karst is closely associated with carbonate formations but it is not limited to it. Karstification can occur in evaporite formations and it is also possible that a carbonate formation does not undergo karstification (Jennings and Jennings, 1985).

Karst presents itself both at the surface and in the subsurface. A typical feature of karst is epikarst (Frumkin, 2013). This is the transition zone between the soil or surface and the massive carbonate rock out of which it is formed through weathering. The epikarst forms when carbonate formations crop out and are exposed to the atmosphere. Pre-existing joints are slowly eroded until fine sediments can be trapped, allowing vegetation to grow. The vegetation speeds up the erosion process in a positive feedback loop. The epikarst zone can be up to 30 m thick in some cases and is a mix of soil and large and small blocks of intact rock.

Another of its characteristic features includes dolines (sinkholes) (Frumkin, 2013; Jennings and Jennings, 1985). Sometimes this is considered as part of the epikarst, but as it has some specific features it is treated here. A doline is a closed depression of a roughly circular shape. It can have a diameter of a few meters up to thousands of meters. The depth varies from a few meters to tens of meters. There are several types of dolines that can be distinguished (Kranjc, 2013). The first type is the solution doline (Figure 2-1A), which forms through the dissolution and corrosion of the bedrock. The second type is the collapse doline (Figure 2-1B). This type forms when the ceiling of a cave collapses. The third type is the caprock doline (Figure 2-1C), which forms in the same manner as the previous type. The difference is that
the overlying formation in the case of the collapse doline is a carbonate, while in the case of the caprock doline it is a non-carbonate. The fourth type is the subsidence doline. These can take the form of either a dropout doline (Figure 2-1D) or a suffosion doline (Figure 2-1E), depending on the cohesiveness of the soil overlying the carbonate. Both are formed by soil washing away through fissures or caves in the limestone. Finally the fifth type is a buried doline (Figure 2-1F), which is a solution doline that is covered by sediments.

Caves are another common occurrence in karst. They are formed through the dissolution of carbonates and flowing water although the exact conditions under which these processes start forming caves is a matter of ongoing debate (Frumkin, 2013). The variety in the shape, size and extent of caves and cave systems is large (Jennings and Jennings, 1985). Caves and caves systems can range from tens of meters to kilometers in length. Caves can be air or water filled and some caves are refilled over time by clastic sediments.

2-3 Seismic Properties of Carbonates

Besides structural features realistic P-wave velocities, S-wave velocities and density are important in creating numerical models. Most studies performed on the seismic properties of carbonates focus on applying the findings to the interpretation of seismic data and borehole logs. They use it to identify structural traps and to construct impedance models for synthetic seismic sections. In a study of 295 carbonate samples from different areas and different ages the P-wave and S-wave velocities were measured for different pore fluid and confining pressures (Anselmetti and Eberli, 1997). The measured seismic velocities ranged from 1700 m/s up to 6500 m/s for P-waves and 700 m/s to 3400 m/s for S-waves. The following conclusions were drawn with respect to seismic velocities in carbonates (Anselmetti and Eberli, 1997):

- The type of carbonate mineral of which the rock is made up does not have an significant influence on the seismic velocities
- Carbonate sediments deposited in a shallow water environment have higher seismic velocities than those deposited in deeper water (shelf, slope or basin).
- Neither burial depth nor age are a good predictor of the seismic velocity. Velocity inversion with depth is a possibility.
- Porosity has the most influence. Both P and S-wave velocity increase with decreasing porosity.

In another study, on 173 gas-saturated samples, the relation between bulk density and seismic velocities (Wang, 1997) show a correlation between bulk density and seismic velocity. The velocities for both P and S-waves increase with increasing bulk density (Figure 2-2). However, the spread for the P-wave velocity in respect to bulk density is large.

These findings are used to construct realistic models which for this thesis. Keeping in mind the geometries of karst features and the lateral extent and depth of carbonates platforms. I will use the relations between the seismic properties of carbonates to assign realistic values to the different layers in the numerical models.
Figure 2-2: P and S-wave velocity versus Bulk Density. (Wang, 1997)
Exact calculations of the seismic wave field in a heterogeneous subsurface are not possible. 
There are several numerical methods available for approximating a solution. The most common ones are the finite element method (FEM), the finite difference method (FDM) and the spectral method. In this chapter I will focus on FDM using a standard staggered grid, as this is the approach used in this thesis.

3-1 Visco-elastic Model

A seismic wave propagating through the subsurface will shortly deform it. The earth will react as both an elastic and a viscous medium under the influence of this deformation. The effect of this behavior on the propagation of a seismic wave is best described using the visco-elastic model, described by Equations 3-1 to 3-5 (Bohlen, 2002).

\[
\frac{\rho \nu_i}{dt} + \nu_i \rho \nabla \nu_i = -\frac{\partial \sigma_{ij}}{\partial x_i} + f_i
\]

In the case of a visco-elastic medium, like the earth, the stress-strain relationship is described by

\[
\dot{\sigma}_{ij} = \left( \frac{\partial \nu_k}{\partial x_k} (M(1 + \tau^s) - 2\mu(1 + \tau^s))) + 2 \frac{\partial \nu_i}{\partial x_i} \mu(1 + tau^s) + \sum_{l=1}^{L} r_{ijl} \right) \text{ for } i=j
\]

\[
\dot{\sigma}_{ij} = \left( \frac{\partial \nu_k}{\partial x_k} + \frac{\partial \nu_i}{\partial x_i} \mu(1 + tau^s) + \sum_{l=1}^{L} r_{ijl} \right) \text{ for } i \neq j
\]
$r_{ijl}$ are the memory variables given by

\begin{equation}
\dot{r}_{ij} = \frac{-1}{t_au_{sl}} = ((M\tau^p - 2\mu\tau^s)\frac{\partial v_k}{\partial x_k} + 2\frac{\partial v_i}{\partial x_i}\mu(1 + t_au^s) + r_{i,j,l}) \quad i=j
\end{equation}

\begin{equation}
\dot{r}_{ij} = \frac{-1}{t_au_{sl}} = (\mu\tau^s(\frac{\partial v_k}{\partial x_k} + \frac{\partial v_i}{\partial x_i}) + r_{i,j,l}) \quad \text{for } i \neq j
\end{equation}

$\rho$ is the density, $i$ and $j$ subscripts indicate spatial directions, $\nu$ is the particle velocity, $t$ is time, $x$ indicates the three spatial directions (x,y,z), $p_{ij}$ denotes the $ij$-th component of the stress tensor, $\tau^p$ and $\tau^s$ define the level of attenuation for P and S-waves respectively, $\sigma_l$ are the stress relaxation times for both P- and S-waves and $f_i$ denotes the components of an external body force. The external body force is present at the source position and is used to simulate a source. $M$ and $\mu$ define the phase velocity models.

### 3-2 Finite Difference Method

To numerically simulate wave propagation a standard staggered grid (SSG) approach, which was proposed by Levander (1988) and later expanded upon by Robertsson et al. (1994). Equations 3-1 to 3-5 must first be discretized in order to be used in a FD scheme. To discretize the equations the derivatives need to be approximated and depending on the required accuracy of the simulation there are different approaches (Moczo et al., 2007). The time derivative, in the SSG method is approximated by a second order centered difference scheme (Equation 3-6). The spatial derivative is approximated using either forward (Equation 3-3) or a backward fourth order staggered scheme (Levander, 1988)(Equation 3-3).

\begin{equation}
\frac{\partial f(x,y,z,t)}{\partial t}_{i,j,k} \approx \frac{f^{n+1}(i,j,k) - f^{n-1}(i,j,k)}{\Delta t}
\end{equation}

\begin{equation}
\frac{\partial f(x)}{\partial t} \approx \frac{1}{24h}(-f(i+2) + 27(f(i+1) - f(i)) + f(i-1))
\end{equation}

\begin{equation}
\frac{\partial f(x)}{\partial t} \approx \frac{1}{24h}(-f(i+1) + 27(f(i) - f(i-1)) + f(i-2))
\end{equation}

Where $f$ is an arbitrary function and $h$ is the grid spacing. For the complete set and description of discretized equations the reader is referred to (Bohlen, 2002)

In the SSG the model parameters are not all located at the same grid positions within the model space but at different location within the grid (Figure 3-1). The SSG approach is explicit, that is the next value is calculated using the present one. The size of the time step...
Finite difference methods suffer from a phenomenon called grid-dispersion, which is a type numerical dispersion of energy and thus does not reflect reality. The effect of the dispersion is cumulative, the longer a wave travels through the model space the more energy will numerically be dispersed. Although this effect can not be removed it can be properly managed by choosing an appropriate grid spacing with respect to the minimum wavelength present in the model. According to (Robertsson et al., 1994) using Equation 3-10 keeps the error caused by grid dispersion below 5% for a SSG approach and a discretization with an accuracy of $O(2,4)$.

\begin{equation}
  h = \frac{\lambda_{\beta}^{\min}}{6}
\end{equation}

$h$ is the grid spacing and $\lambda_{\beta}^{\min}$ is the minimum shear wavelength.

A numerical model consists of a finite number of grid points restricting the model to a certain size. Boundary conditions are imposed on the edges of the model to mimic an infinite extension of the models space. Furthermore, a free surface boundary condition is usually imposed when the top part of the modeled subsurface is in contact with the atmosphere (Moczo et al., 2007). On the bottom and the sides of the model a non reflecting boundary is imposed to ensure that energy is not reflected from the edge back into the model space, as the edge is an

\begin{equation}
  \Delta t = \frac{6h}{7\sqrt{D_{max}}}
\end{equation}

$h$ is the grid spacing, $D$ is the dimension of the simulation (2-D or 3-D) and $v_{max}$ is the maximum P-wave velocity present in the model.

Figure 3-1: The position of the model parameters in a SSG for a 3-D (A) and a 2-D (B) grid (Modified after the Sofi2D and Sofi3D manual)
artificial boundary. One approach is the Kosloff’s method (Moczo et al., 2007). In this method an exponential damping function is applied to the amplitude in a region surrounding the computational domain. The approach works well for normally incident waves but can lead to energy being reflected back into the model space for waves approaching the edge at other angles (Moczo et al., 2007). Most commonly used nowadays are perfectly matched layer (PML) boundaries, which work for a large range of incident angles. This method achieves damping of the incident wavefield by applying a coordinate stretch into the complex domain (Moczo et al., 2007). The stretch is achieved by modifying the spatial differentiator in the PML region (Equation 3-11),

\[
\tilde{\partial}_x = \frac{1}{s_x} \partial_x
\]

\(\partial_x\) is the original spatial differentiator and \(s_x\) is given by,

\[
s_x = 1 + \frac{d_x}{i\omega}
\]

\(\omega\) is the radial frequency and \(d_x\) is the damping profile. The parameter \(d_x\) is larger than zero inside the PML zone. PML’s work well but are not perfect. For very low incident angles it can still reflect energy back into the model space (Bohlen, 2002).

3-3 Parallel Computing

Finite difference schemes are computationally intensive and running a numerical simulation on a serial platform (workstation) severely limits the size of the model that can be investigated. However, the availability of computer clusters and the suitability of FDM for parallel computing make it possible to simulate large models. To do parallel computing the model space is first decomposed into different subsections and the wavefield is computed for each subsection by different processors. After each time step data is exchanged between the sections that compute adjacent parts of the model space. This process is repeated until the end of the simulation. To pass the wavefield information from one section to the other the internal boundaries are padded (Figure 3-2). The width of the padding depends on the order of the derivative approximation. For a fourth order approximation a padding of two grid points is needed (?). The values in the padding layer are used to compute the parameter values in the first gridpoint in the modelspace adjacent to the padding layer.
Figure 3-2: A schematic overview of the data transfer between the different sections (PE). The black dots represent the two grid points closest to the edge being copied to the padding layer of the adjacent PE.
Chapter 4

Numerical Simulations - 3D models and synthetic seismogram generation

4-1 3D models

4-1-1 Design Strategy

As mentioned in the introduction a series of models of increasing complexity is used in this thesis to investigate the effect of karstified limestone close to the subsurface on the recorded wavefield (Figure 4-1). The first is a basic horizontally layered model. In the second model topography was added. For the third model the thickness of the soil layer was made to vary and for the fourth model stochastic fluctuation were added to P-wave velocity, S-wave velocity and the density. Finally, for the fifth model attenuation was incorporated into the model. All the models have the same physical dimension of 500 m by 300 m in the horizontal plane and a height of 200 m.

Each model contains four different layers, namely: air, soil and 2 carbonate layers. Only four layers are used in the models to keep the complexity of the deeper subsurface to a minimum. This does not represent reality but as the focus of the research is to determine the effect of the shallow karstified limestone on a deep reflector adding more layers would unnecessarily complicate the model. The fourth layer is put into the model to create a reflective interface which causes the reflection I am trying to retrieve. The carbonate formation is present over the entire lateral extent of the model space. As the horizontal extent of the model is only 500m by 300m this is a realistic assumption as carbonates can have a lateral extend of kilometers (Chapter 2). For simplicity it is assumed there are only carbonate formations until the bottom of the model space.

The P-wave velocities for the soil and the first carbonate rock layer (Table 4-1) come from a walkaway test preformed for the CARNEVAL project in the Jura, Switzerland. These values were taken to more closely relate the constructed models to the field study that is to be performed in September 2014. Using empirical relations given in literature (Figure 2-2) realistic values were assigned to these two layers. The model parameters for the second
carbonate layer were taken from literature (Chapter 2). These values were taken to contrast the first carbonate layer and cause a reflection.

Table 4-1: The mean P-wave velocity, S-wave velocity and density for each layer used in the 3D models

<table>
<thead>
<tr>
<th>Layer</th>
<th>( v_p ) [m/s]</th>
<th>( v_s ) [m/s]</th>
<th>( \rho ) [kg/m(^3)]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air</td>
<td>340</td>
<td>10</td>
<td>1.2</td>
</tr>
<tr>
<td>Soil</td>
<td>1600</td>
<td>600</td>
<td>1700</td>
</tr>
<tr>
<td>Carbonate 1</td>
<td>3500</td>
<td>2000</td>
<td>2500</td>
</tr>
<tr>
<td>Carbonate 2</td>
<td>5000</td>
<td>3000</td>
<td>2600</td>
</tr>
</tbody>
</table>

4-1-2 Model 1.1 - Horizontally Layered Earth

The first model in the series (Figure 4-2) is a horizontally layered earth model. All the interfaces between the different layers are horizontal. It was constructed as a basic reference model to determine the effect of the different layers and layer boundaries on the recorded seismic wavefield without the effect of any complex geometrical features.

4-1-3 Model 1.2 - Topography

In model 2 (Figure 4-3) topography from the Jura, Switzerland, was added to the model (Appendix A). This was done to tie it into the CARNEVAL project. The other reason for taking existing topography from a region with known karstified limestone is the effect karst can have on the roughness of the topography. Taking existing topography is both easier and more realistic than trying to model the impact of karst on the topography. The altitude data was taken from the Swiss ALTI3D database, which has a lateral resolution of 2m. To use this data in the model a 2-D linear interpolation was performed to have a height value at each node in accordance with the grid spacing used in the numerical simulation. The vertical
Figure 4-2: Model 1.1, the three interfaces present in the horizontally layered earth model (a) and the P-wave velocity profile along the line AA' (b).
Figure 4-3: The three different interfaces present in Model 1.2 (a) and the P-wave velocity profile along the line AA’ (b).
resolution of the Swiss ALTI3D data set was higher than the grid spacing used in the models, thus the height values were rounded to the nearest node. The height difference introduced by the topography is around 20 m between the highest and the lowest point. The soil thickness was kept at a constant 10 m over the entire model space, making the soil-carbonate interface undulate together with the topography.

4-1-4 Model 1.3 - Fluctuating Soil Thickness

In reality the soil thickness will not be constant and might fluctuate over an area of 500m by 300m. In some places there will be no soil cover and the limestone will outcrop while in other places the soil cover might be thicker than 10 meters. To simulate this changing soil thickness was varied over the entire model space using stochastic fluctuations. The median of the fluctuations is kept a depth of 10 m below the surface and the depth of the soil-carbonate interface fluctuates around that value (Figure 4-5) and the correlation length was put at 30m, in both the x and the y direction. The fluctuating soil thickness will also introduce some of the heterogeneities associated with buried dolines (Chapter 2) into the model without explicitly modeling them, as this type of dolines are filled with soil and can thus be seen as a thickening of the soil layer.

4-1-5 Model 1.4 - Stochastic Fluctuations

All previously described models contain layers with constant seismic parameters. However, normally the seismic parameters will not be constant throughout a given formation. To mimic the variation in seismic parameters in a layer stochastic fluctuations were added to the model parameters. Using the same method as was used for the soil thickness fluctuations. The correlation length used in the x and y direction was 30 m and 10 m in the z direction. I assumed that there is a larger correlation length in the lateral directions than in the vertical. The geometry of the model is the same as for Model 1.3.

4-1-6 Model 1.5 - Attenuation

It is known that attenuation can have a strong influence on the wave propagation, especially in the soil layer. To make the simulation more realistic attenuation was added to model 5. The attenuation in the soil layer was put at Q=25 for both P and S-waves. The attenuation for the carbonate layers was set at Q=1000. In reality this will most likely not be the case but the few studies performed (Assefa et al., 1999; Hackert and Parra, 2003) on the attenuation levels in carbonate formations focus on the high frequency range, kHz to MHz, and is as such not directly applicable to the frequency range in which seismic exploration is performed. As the karstic features I want to investigate are close to the surface I decided to keep Q high for the carbonate formations, so the reflection coming back from the lowest boundary will not be affected by the attenuation lowering the amplitude of the event.
Figure 4-4: The soil-carbonate boundary and the Carbonate-Bottom Layer boundary in model 1.3 (a) and the P-wave velocity profile along the line AA' (b). The surface topography remains the same as in model 2, which makes the soil layer fluctuate in thickness over the model space.
4-2 Numerical Simulation

For the numerical simulation the program SOFI3D was used, which was developed at the Karlsruhe Institute of Technology. The most important model parameters are given in Table 4-1. As mentioned previously the simulation performed in this thesis were done using a SSG and a PML boundary condition (Chapter 3). The PML boundary I used was 40 grid points in width. The grid spacing was 0.5m and the time step $4.5 \cdot 10^{-5}$s for models 1.1, 1.2 and 1.3. Models 1.4 and 1.5 needed a smaller time step as these two models contain higher velocity due to the stochastic fluctuations. The time step was set at $3.75 \cdot 10^{-5}$s to ensure numerical stability according to the Courant criterion (Equation 3-9). To ensure a lot of the noise was recorded the total recording time simulated was 1.08s.

Each of the five models are simulated five times with the source at a different locations 100 meters apart (Figure 4-6). The first was placed at $x=50m$ and the last one at $x=450m$, the y coordinate for all shots are the same, $y=150m$. The sources were placed at a depth of 2m below the surface. The source type used was a pressure source and the source wavelet a Ricker wavelet with a central frequency of 50Hz (Figure 4-7). The pressure source simulates an explosive source (e.g. dynamite) and the central frequency is within the range normally used for exploration seismics.

Receiver positions were defined along one line in the computational domain (Figure 4-6). The spacing is 1m and the receiver line consists of 481 receivers. For all models the receivers’ vertical placement follows the topography. The lines starts at $x=10m$ and continues to $x=490m$, the y coordinate stayed the same and is at the center of the width of the model space ($y=150m$). The receivers record the parameters every 5 time steps for Model 1.1 to 1.3 and every 7 time steps Model 1.4 and 1.5. The output from the simulations were shot gathers of the particle velocity in the x, y and z direction, pressure, curl and divergence. Out of these five parameters only the particle velocities will be directly measured in a field campaign.

Figure 4-5: The P-wave velocity profile along the line $y=150$ m for Model 1.4 containing stochastic velocity fluctuations.
Figure 4-6: Receiver line and source locations for the models containing topography.

Figure 4-7: Shape of the source wavelet in the time domain (a) and the corresponding frequency amplitude spectrum (b).
Table 4-2: The most important parameters used in the FD-simulation of the 3-D models.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gridspacing, dh</td>
<td>0.5m</td>
</tr>
<tr>
<td>Timestep, dt (Models 1.1, 1.2, 1.3)</td>
<td>$4.5 \times 10^{-5}$ s</td>
</tr>
<tr>
<td>Timestep, dt (Models 1.4 and 1.5)</td>
<td>$3.75 \times 10^{-5}$ s</td>
</tr>
<tr>
<td>Total simulation time</td>
<td>1.08s</td>
</tr>
<tr>
<td>Receiver spacing</td>
<td>1m</td>
</tr>
<tr>
<td>Number of receivers</td>
<td>481</td>
</tr>
<tr>
<td>Source spacing</td>
<td>100m</td>
</tr>
<tr>
<td>Number of source positions</td>
<td>5</td>
</tr>
<tr>
<td>Receiver recording interval (Models 1.1, 1.2, 1.3)</td>
<td>$2.25 \times 10^{-4}$s</td>
</tr>
<tr>
<td>Receiver recording interval (Models 1.4 and 1.5)</td>
<td>$2.63 \times 10^{-4}$s</td>
</tr>
<tr>
<td>Source type</td>
<td>Pressure</td>
</tr>
</tbody>
</table>

4-3 Results

As previously mentioned for each model the particle velocity in the x, y and z direction were measured. The particle velocity in the z direction is considered the vertical particle velocity while in the x and y direction are the radial and transverse respectively. Not all the shot gathers can be found in this chapter as some did not contribute to the analysis.

4-3-1 Model 1.1 - Horizontally Layered Earth

I identified 5 different zones in the vertical particle velocity shot gather of Model 1.1 (Figure 4-8) that clearly show different seismic events. As ambient noise is not included in the numerical modeling the first arrivals are clearly visible (Figure 4-8, Zone 1). The events in Zone 1 show two different apparent velocities. Up to an offset of 15 m the first arrivals travel with an apparent velocity of 1600 m/s. This is the same velocity as the P-wave velocity given to the soil layer. This means that these first arrivals are the direct P-wave traveling from the source to the receivers. From an offset of 15 m (in both directions) until the largest offset an event is present with an apparent velocity of roughly 3500 m/s. This value corresponds with the P-wave velocity assigned to the first carbonate rock layer and would make this event the refracted P-wave from the soil-carbonate rock interface. This is backed up by the recorded radial particle velocity shot gather of the wavefield which shows no energy in this region (Figure 4-9, Zone 1). The refracted wave arrives at the receivers at a near vertical angle.

The second region (Figure 4-8, Zone 2) is defined by not containing the first arrivals but not yet showing the strong amplitudes observed in Zone 3 (Figure 4-8). The events at earlier times travel (2a in Figure 4-8) with the same apparent velocity as the refracted first arrivals while the later events (2b in Figure 4-8) show a lower apparent velocity of 1200 m/s. The apparent velocity is the same as for the events visible in zone 3 (Figure 4-8). This zone consists of a number of parallel events, with an apparent velocity of 1200 m/s. Noticeable is that at earlier times (3a in Figure 4-8) the amplitude decreases with offset and the earliest ones fade out before the events reach the furthest recorded offset. Another noticeable occurrence is the higher frequency with which the event appears at later times (3b in Figure 4-8), at the same
receiver location. These characteristics are consistent with those of guided waves. The same events are also visible in the radial component shot gather (Figure 4-9, Zone 3).

Zone 4 (Figure 4-8) contains linear events originating at the source position and running all the way to the edges of the shot gather. Its appearance indicates that it is the ground roll having an apparent velocity of 510 m/s, which is close to the S-wave velocity given to the soil layer.

Zone 5 (Figure 4-8) shows a repetition of slightly curved events. The events become steeper, indicating a decrease in apparent velocity, the further away from the source they are. The amplitude of the events decrease with time and are not constantly present in all the traces. The events disappear first in the traces closest to the source position. Different events are clearly distinguishable in the FK domain (Figure 4-10). The events are not linear but seem to curve slightly.

4-3-2 Model 1.2 - Topography

Zone 1 (Figure 4-11) is not significantly affected by the addition of topography. The first arrivals are still clearly distinguishable. Zone 2 (Figure 4-11) still shows the same events as in Model 1.1. At earlier times there are events that run parallel to the first arrival (2a in Figure 4-11) and at later times events with a slower apparent velocity are visible (2b in Figure 4-11). However, zone 3 (Figure 4-11) is significantly effected by the topography. In model 1.1 (Figure 4-8, Zone 3) there were clear parallel events, which diminished in amplitude with increasing offset and had a rise in the frequency with which they appeared. However, for Model 1.2 these events are (3a in Figure 4-11) still visible at the smaller offsets, but for the larger offsets the measured wavefield becomes incoherent (3b in Figure 4-11). The incoherency only occurs in the recorded vertical particle velocity (Figure 4-11, Zone 3). In zone 3 there are also events that appear to be diffractor patterns (3c in Figure 4-11); these events are more dominant on the right side of the source location than on the left side. The topography on the right side is also rougher than on the left side of the source position. The events that appear to be diffractors are not present in the radial component shot gather (Figure 4-12) but are visible in the transverse particle velocity shot gather (3a in Figure 4-13). The particle velocities measured in the radial direction Zone 2 and 3 (Figure 4-12) are unaffected by the addition of topography, the recorded wavefield in this direction still shows a shingling effect taking place (Figure 4-12, Zone 3).

The ground roll (Figure 4-11, zone 4) to the left of the source location in the vertical component shot gather is unaffected by the addition of topography. However, to the right side of the source location it disappears at an offset of 75 m, just after the drop in elevation.

The most noticeable changes in the recorded wavefield between Model 1.2 (Figure 4-11) and Model 1.1 (Figure 4-11) are in Zone 5. In Model 1.1 Zone 5 contained a number of curved parallel events mirrored around the source location, while in the recorded wavefield for model 1.2 there are events that run parallel to the ones in zone 4 (Figure 4-11), on either side of the source position. The parallel events are dominant close to Zone 4 but form a crisscross pattern in the middle of the Zone 5 (Figure 4-11). The straight events appear to be back scatter. The events closely resemble the patterns found by Wang et al. (2001) and Duan and Fu (2001), who did a study on the effects of topography on the recorded wavefield. Both articles relate the strong scattering to the rough boundaries and the high velocity contrast
Figure 4-8: The vertical particle velocity shot gather recorded in the numerical simulation of Model 1.1
Figure 4-9: The radial particle velocity shot gather recorded in the numerical simulation of Model 1.1
between the air and soil. The back scatter pattern is visible in all the recorded shot gathers (Figure 4-11, Figure 4-12 and Figure 4-13), indicating that the back scattered waves come from all directions. However, the events do not have the same apparent velocity in each of the recording directions. This results in the zones defined in the vertical particle velocity shot gather (Figure 4-11) not containing the same events as the shot gathers in the radial and transverse direction (Figure 4-12 and Figure 4-13).

There is a significant change when looking at the FK-spectrum, whereas in the Model 1.1 energy was concentrated in clearly distinguishable events it has become blurred with the addition of topography (Figure 4-14). Some events are still visible at frequencies above 110 Hz (a in Figure 4-14), but these are also less pronounced than in Model 1.1.

4-3-3 Model 1.3 - Fluctuating Soil Thickness

The events in Zone 1 (Figure 4-15) are not linear anymore but fluctuate, this happens on both sides of the source position. The first arrival at an offset of between 50m and 100m to the left of the source position (a in Figure 4-15) are at earlier times than in Model 1.2. This phenomenon is visible until the end of the recording time and also influences Zones 2 and 3 (Figure 4-15). Another noticeable difference the behavior of the groundroll in Zone 4 (Figure 4-15) to the left of the source location after an offset of 175 m, whereas in Model 1.2 the groundroll in this zone reached the last receiver location. The disappearance of the groundroll corresponds with the rapid thinning of the soil layer at an offset of 175m to the left of the source location. At one point the soil layer has almost completely vanished (b in Figure 4-15).

The parallel event in Zone 5 (Figure 4-15) are still present in the shot gather of Model
Figure 4-11: The vertical particle velocity shot gather recorded in the numerical simulation of Model 1.2
Figure 4-12: The radial particle velocity shot gather recorded in the numerical simulation of Model 1.2
Figure 4-13: The transverse particle velocity shot gather recorded in the numerical simulation of Model 1.2
4-3 Results

1.3, although less pronounced than in the previous model. Another difference between the recorded wavefield for Model 1.2 and Model 1.3 are the high amplitudes recorded at an offset around 180m to the left of the source location (a in Figure 4-15). The receivers measure higher amplitudes until zone 4 is reached. After that the amplitudes are consistent with the other traces in the shot gather, which might be an effect of the trace normalization and taking a cut off value.

For Model 1.1 and 1.2 there were some events in the FK-domain (Figure 4-10 and Figure 4-14) that could still be distinguished, but these have completely vanished in the result of Model 1.3 (Figure 4-16). The only clear event still present in the FK-domain is the groundroll.

4-3-4 Model 1.4 - Stochastic Fluctuations

Comparing Model 1.4 (Figure 4-17) with Model 1.3 (Figure 4-15) there is no obvious difference, the same events occur. Similarly the FK-domain (Figure 4-18) exhibits no major changes. It appears that stochastic velocity fluctuations do not have a significant impact on the recorded wavefield when topography and fluctuations in the soil thickness have already been included in the model.

4-3-5 Model 1.5 - Attenuation

Including an attenuation of $Q=25$ to Model 1.5 does not result in an obvious difference between this model and Model 1.3 and 1.4. I would expect that the attenuation would have
an influence on the ground roll but looking at the shot gathers these events (Figure 4-19, Zone 4) are still as pronounced as in Model 1.3 and 1.4. The FK-spectrum Figure 4-20) are unchanged from the FK-spectrum of Model 1.4 (Figure 4-18, other than slightly lower energy levels at higher frequencies. This is in line with the expectation that higher frequencies are attenuated faster than lower frequencies.
Figure 4-15: The vertical particle velocity shot gather recorded in the numerical simulation of Model 1.3.
Figure 4-16: The FK-domain of the vertical particle velocity data from Model 1.3.
Figure 4-17: The vertical particle velocity shot gather recorded in the numerical simulation of Model 1.4.
Figure 4-18: The FK-domain of the vertical particle velocity data from Model 1.4.
Figure 4-19: The vertical particle velocity shot gather recorded in the numerical simulation of Model 1.5.
Figure 4-20: The FK-domain of the vertical particle velocity data from Model 1.5.
5-1 2D models

5-1-1 Design strategy

In the previous chapter the construction and the simulation of 3-D models with a reflector at a depth of approximately 150 m below the surface was discussed. However, the depth of the reflector in the 3-D models is shallow compared to the usual targets in exploration seismic acquisition. Due to limited computer power it was not possible to extend the size of the 3-D models. To investigate the effects of karstified limestone on a deeper reflector than the one present in the 3-D models I constructed a set of three 2-D models to test the detectibility of a deep reflector and to compare 2-D with 3-D modeling.

All models are 2 km wide and 1 km in height. The 2-D models are built by extending \( y = 150 \) m section of the corresponding 3-D models. The line \( y = 150 \) m is the same line along which the wavefield was recorded for the 3-D simulations. The models contain either 4 or 5 layers depending on whether the deep reflector is present. The first 4 layers are the same as the ones used in the 3-D models, namely: air, soil and two carbonate layers (Table 4-1). The fifth layer is in the model to have a reflective interface at the bottom of the modeling space and has the same seismic parameters as the first carbonate layer (Table 5-1). All the models contain stochastic velocity and density fluctuations.

5-1-2 Model 2.1 and 2.2 - Deep Reflector

The fluctuating soil thickness in Model 2.1 and 2.2 is not exactly the same as for the 3-D model since the script for adding 1-D stochastic fluctuations result in slightly different thicknesses than the 2-D one even when the same parameters are used. However, the same kind of fluctuation is still present in the soil thickness, but not exactly at the same locations.
Table 5-1: The mean P-wave velocity, S-wave velocity and density for each layer used in the 2-D models

<table>
<thead>
<tr>
<th>Layer</th>
<th>$v_p$ [m/s]</th>
<th>$v_s$ [m/s]</th>
<th>$\rho$ [kg/m$^3$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air</td>
<td>340</td>
<td>10</td>
<td>1.2</td>
</tr>
<tr>
<td>Soil</td>
<td>1600</td>
<td>600</td>
<td>1700</td>
</tr>
<tr>
<td>Carbonate 1</td>
<td>3500</td>
<td>2000</td>
<td>2500</td>
</tr>
<tr>
<td>Carbonate 2</td>
<td>5000</td>
<td>3000</td>
<td>2600</td>
</tr>
<tr>
<td>Carbonate 3</td>
<td>3500</td>
<td>2000</td>
<td>2500</td>
</tr>
</tbody>
</table>

Both Model 2.1 (Figure 5-1) and 2.2 contain all 5 layers and have the same geometrical design. The difference between Model 2.1 and Model 2.2 is the addition of attenuation. This is done to make the wave propagation through the model more realistic. Another reason to add attenuation is to investigate the impact the attenuation has on the recorded wavefield compared to the model without it. In the 3-D case (Chapter 4) the impact of attenuation appeared to be minimal, but this might change with a greater vertical and horizontal extent of the model space. The attenuation values in Model 2.2 are the same as in Model 1.5 (Chapter 4), $Q=25$ for the soil layer and $Q=1000$ for the carbonate layers.

Figure 5-1: The P-wave velocities in Model 2.1. The model contains five layers with the deepest layer being present at a depth of 980m until 1000m.

5-1-3 Model 2.3 - Without the Deep Reflector

Model 2.3 (Figure 5-2) is the same as Model 2.2. The difference between Model 2.2 and Model 2.3 is that in Model 2.3 the third carbonate layer was left out. This means that there are no reflection coming back from the lowest part of the model space. The results from the FD simulation are subtracted from the results of Model 2.2. The recorded seismic wavefields until the arrival of the reflection from the deepest boundary will cancel out so the effect of the shallow subsurface on the reflection from the deepest boundary can be investigated.
5-2 Results

5-1-4 Numerical Simulation

For the numerical simulation of the 2-D models I used the SOFI2D code developed by the Karlsruhe Institute of Technology. The most important modeling parameters are given in Table 5-2. The simulation was done using a SSG discretization of the model space and a 30 grid point absorbing boundary on all the edges. Although a PML boundary will work better this was not an option in the software package. The grid spacing was 0.5m and the time steps were $4.5 \cdot 10^{-5}$s. The total recording time was 1.5s.

Each of the models was simulated 102 times. Once at the position corresponding to source position the central source location in the 3-D models, to investigate the difference between doing 2-D and 3-D simulations. The 101 other locations are a line of sources placed from $x=500$ until $x=1500$ at an interval 10 m, to produce a dataset which can be used for processing. For the 2-D models I use the same source type and frequency content as for the 3-D models (Chapter 4).

During the simulation the vertical and horizontal particle velocity, pressure, curl and divergence were recorded. A total of 1989 receivers were placed at an interval spacing of 1 m, starting from $x=5$ m until $x=1994$ m. For all models the receivers' vertical placement follow the topography. The receivers record the wavefield every $2.25 \cdot 10^{-4}$s.

5-2 Results

5-2-1 Comparison 2D and 3D Results

As in Chapter 4 the particle velocity in the $z$-direction and the particle velocity in the $x$-direction are referred to as the vertical particle velocity and the radial particle velocity respectively. Unlike in Chapter 4 the shot gathers are not divided in zones, the description of the shot gathers will be limited to the major similarities and differences between the results of the 3-D simulation (Chapter 4) and the results of the 2-D simulations.
Table 5-2: The most important parameters used in the FD-simulation of the 3-D models.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gridspacing, dh</td>
<td>0.5m</td>
</tr>
<tr>
<td>Timestep, $dt$ (Models 1.1, 1.2, 1.3)</td>
<td>$2.0 \cdot 10^{-5}$ s</td>
</tr>
<tr>
<td>Total simulation time</td>
<td>1.5s</td>
</tr>
<tr>
<td>Receiver spacing</td>
<td>1 m</td>
</tr>
<tr>
<td>Number of receivers</td>
<td>1989</td>
</tr>
<tr>
<td>Source spacing</td>
<td>10 m</td>
</tr>
<tr>
<td>Number of source positions</td>
<td>102</td>
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<tr>
<td>Receiver recording interval</td>
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</tr>
<tr>
<td>Source type</td>
<td>Pressure</td>
</tr>
</tbody>
</table>

Comparing the results from Model 2.1 (Figure 5-3) and Model 1.4 (Figure 4-17) there are some differences and similarities. The longer receiver line allows the refraction from the first carbonate-carbonate interface (a in Figure 5-3) to be recorded. The snapshots (a Figure 5-4a and b) confirm that the event (a in Figure 5-3) is the refraction from the first carbonate-carbonate boundary. A hyperbolically shaped event has also been recorded starting from an offset of 380 m to the right of the source location (b in Figure 5-3). It arrives a little later than the theoretical arrival of the reflection (red dashed line in Figure 5-3). However, the measured event does follow the same trend as the theoretical arrival of the reflection. This can be due to the stochastic fluctuations of the velocities and irregular soil thickness present in the model, both of which were not taken into account in computing the theoretical reflection. At the moment the reflected wave (b in Figure 5-4a and b) arrives at a receiver with an offset larger than 380 m the only clear event that has reach the same point is the refracted wave (a in Figure 5-4). The reflection arrives at smaller offset earlier than the refracted wave from the carbonate-carbonate interface, but around these receiver locations there is still a lot of energy trapped (Figure 5-4), obscuring any arrivals. The trapped energy does not appear to diminish in amplitude over time (c in Figure 5-4b and c). In parts of the soil further away from the source location there is still relatively little energy present (d in Figure 5-4b and c), even after the refracted and reflected wave from the first carbonate-carbonate boundary have passed. The reflection from the deepest boundary is not visible in Figure 5-3 but is visible in Figure 5-4b (e). However, when looking at a later time (Figure 5-4c) the reflection from the deepest boundary is obscured.

As in the results of Model 1.4 (Figure 4-17) the groundroll in Model 2.1 (c in Figure 5-3) vanishes from the recording before it can reach the end of the receiver line. However, there is a difference when comparing the 3-D (Model 1.4) and the 2-D simulation (Model 2.1) concerning the groundroll. In the 2-D case the groundroll dominates the shotgather less than in the 3-D case. This effect is most apparent when comparing the first arrivals associated with the refraction from the soil-carbonate interface (Zone 1 in Figure 4-17 and d in Figure 5-3), which is more clearly visible than in the results of Model 2.1 (d in Figure 5-3). The refractor patterns visible in the results of Model 1.4 (Figure 4-17) are also present in the results of Model 2.1 (e in Figure 5-3).

The most striking difference between the 3D (Model 1.4) and the 2D (Model 2.1) FD simulation is in the FK domain. Whereas in the case of Model 1.4 hardly any separate events were
distinguishable (Figure 4-18), this is not the case for Model 2.1 (Figure 5-5). A horizontal event is present in the FK domain of Model 2.1 (a in Figure 5-5). The events does not start out horizontal but gradually bends towards the 50 Hz line.

5-2-2 Deep Reflection

From the previous results it is unclear what exactly the effects of the irregular soil thickness and the topography are on the reflection from the deepest boundary as it is obscured by all the near surface effects in both Figure 5-3 and in Figure 5-4. To get a clear image of the recorded deep reflection the results from Model 2.2 and Model 2.3 are subtracted from each other to create a shotgather that shows the difference in the recorded wavefield (Figure 5-6). Noticeable is the up going reflection does not arrive first at a zero offset from the source location but at an offset of 350 m to the right of the source location (a in Figure 5-6). When the seismic wavefront is traveling down, the first part of the wavefront to reach the deepest reflector is still underneath the source location (Figure 5-4a). However, when looking at the upgoing seismic waves it appears that the wavefront to the right of the source location is traveling faster (Figure 5-7). The earlier arrival does not appear to be an effect of the irregular soil thickness or the topography but rather of the stochastic fluctuations in the model.

Some parts of the reflection have a lower amplitude (a and b in Figure 5-6) than other parts (c Figure 5-6). The low amplitude around an offset of 600 m appear to be an effect of the fluctuating soil-carbonate boundary. The reflection arriving at this offset appears not to be transmitted through the soil-carbonate boundary (a in Figure 5-8a). At further offsets, more than 1000 m, it appears that the low amplitudes are due to spherical spreading, as the amplitude of the up going reflection is lower than around the source location (a in Figure 5-8). The up going wavefield from the deep reflector is also diffracted once it reaches the surface. The diffractor patterns are more pronounced at the offsets where the reflection has a high amplitude than in the parts where it has a low amplitude, but nevertheless diffraction patterns are recorded over the entire receiver spread. Events that appear to be multiples are also present in Figure 5-6 (e). The multiples are most likely due to the high impedance contrast causing the seismic waves that gets into the soil layer to be trapped and bounced around between the air-soil and the soil-carbonate interface. However, the multiple is only visible from an offset of 200m to the left up to an offset of 100m to the right of the source location. This event is the last one that can be clearly identified. The rest of the difference between the recorded seismic wavefields of Model 2.2 and Model 2.3 show the same chaotic behaviour as the original recording of Model 2.2 (Figure 5-3).
Figure 5-3: The vertical particle velocity shot gather recorded in the numerical simulation of Model 2.1
Figure 5-4: Snap shots of the pressure field present in the FD simulation of Model 2.1 at 0.15s (a), 0.27s (b) and 0.36s (c)
Figure 5-5: The FK-spectrum of the common shot gather of the source located at x=250m
Figure 5-6: The difference between the vertical particle velocity shot gather recorded in the numerical simulation of Model 2.2 and 2.3.

Figure 5-7: Snap shots of the difference in the pressure field for Model 2.2 and 2.3 present in the FD simulation of Model 2.1 at 0.25s.
Figure 5-8: Snap shots of the difference in the pressure field for Model 2.2 and 2.3 present in the FD simulation of Model 2.1 at 0.44s (a) and 0.52s (b).
In Chapter 5 I showed that the reflection from the deepest boundary is obscured by near surface events in the raw data. However, the reflection itself is not badly affected and retrieving an image of the subsurface using some basic processing should be possible. First I will use an FK-filter to remove the groundroll from the data and apply an amplitude correction. Secondly I will use deconvolution the sharpen the image of the events. Lastly I will apply refraction statics and do a constant velocity stack to determine if the deep reflector is retrievable. For the processing of the data I used PROMAX. I use the dataset from Model 2.2 in my processing.

6-1 FK-filtering

The first step in the processing is to apply a FK-filter in combination with a TAR (True Amplitude Recovery). Several FK-filters were tested, but the best results were obtained using the FK-filter shown in Figure 6-1. Figure 6-2 shows the results of the applied FK-filter for shot 14 (x=640m). The filter worked in removing the groundroll, but only part of the reflection from the deepest boundary is visible (a in Figure 6-2). In some other shotgathers the reflection is not visible at all (Figure 6-3).

6-2 Surface Consistent Deconvolution

Taner (1997) suggests to use surface consistent deconvolution when limestone is present in the shallow subsurface to get the best results. Surface consistency implies that effects of the source, receiver, offset and the earth’s impulse response are all taken into account during the deconvolution. As I am dealing with synthetic data created using a Ricker wavelet, which is zero phase, the deconvolution will be limited to spectral whitening. The best results were obtained when using an operator length of 40ms. Figure dd shows the result of applying surface consistent deconvolution to shot 14. The events in Figure 6-2 are more clearly defined than in Figure 6-4.
Figure 6-1: The FK spectrum for the common shot gather with the source located at $x=640\text{m}$. Everything outside the black triangle is filtered out when the FK-domain is transformed back to the $xt$-domain.

Figure 6-2: Common shot gather for the source located at $x=640\text{m}$ in Model 2.2 after the FK filter is applied.
Figure 6-3: Common shot gather for the source located at x=570m in Model 2.2 after the FK filter is applied
6-3 Statics

I tested both the Diminishing Residual Matrices (DRM) and the Delay-time Method (DLT) refraction static schemes. Both DRM and DLT give similar corrections and I chose apply DRM refraction statics. However, applying refraction statics does not appear to help in aligning of the deep reflection (Figure 6-5). In some parts the alignment improves (a in Figure 6-5), while in others it appears to shift two parts of the reflector away from each other (b in Figure 6-5).

6-4 Constant Velocity Stacking

Normally one would do a velocity analysis to create a velocity model of the subsurface. I chose to use a constant velocity stack as a proof of concept that the deep reflector is recoverable from the dataset. The deep reflector (red dotted line in Figure 6-6) is most clearly visible when a constant velocity of 5302 is used (Figure 6-6). However, the reflector is only partly retrieved (a in Figure 6-6). The unrecoverable parts are most likely due to the reflection not appearing over the entire receiver line. Another contributing factor might be the misalignment of the reflection events causing destructive interference.

Figure 6-4: Common shot gather for the source located at x=640m in Model 2.2 after surface consistent deconvolution
Figure 6-5: Common shot gather for the source located at x=640m in Model 2.2 after DRM refraction statics have been applied
Figure 6-6: Constant velocity stack, with a velocity of 5302 m/s, applied to the dataset of Model 2.2
Chapter 7

Discussion and Conclusions

The results show that topography, fluctuating soil thickness and stochastic fluctuations of the seismic parameters, all of which are features associated with karstified limestone, have an impact on the quality of the recorded wavefield. The groundroll is back scattered by the rough topography. The back scattered events obscure the reflection within an area whose edges are defined by the groundroll. Both topography and the fluctuating soil thickness contribute to the appearance of diffractor patterns, further deteriorating the quality of the recorded seismogram. However, at the far offset from the source location the reflection from the first carbonate-carbonate boundary is still visible. This part of the shot gather is not affected by the back scattered events and only slightly by the diffractor patterns. The groundroll, the back scattered events and parts of the diffractor patterns can be removed from the seismogram by applying an FK-filter. However, the removal of the unwanted events does not automatically result in the reflection from the deepest boundary becoming visible in the shot gather. Even in the best cases the reflection is only partially retrieved. The unrecovered parts of the reflection are most likely a results of the fluctuating soil-carbonate boundary.

At first glance the effect of the stochastic fluctuations in the seismic velocities and density is not obvious, as the back scattered energy and the diffractor patterns are very dominant and the stochastic fluctuations do not significantly contribute to these events. However, the stochastic fluctuations do have an impact on the arrival time of the reflection from the deepest boundary in the 2-D model. The fluctuations in seismic velocities result in the reflection to travel at different speeds through the model space, which cause the reflection to loose its spherical form. The stochastic velocity fluctuations and the topography result in the non hyperbolic recorded reflection event from the deepest boundary. The effects of the topography on the arrival times can be dealt with by using refraction statics. However, statics do not adjust for the effects of the velocity fluctuations. The non hyperbolic form of the reflection leads to destructive interference when stacking the seismic data, resulting in a poorly imaged deep reflector.

Furthermore, there are differences in the recorded wavefield between the results of the 3-D and the 2-D models. The most obvious is the difference in the amplitudes of the first arrivals, which are a lot higher in the 2-D case than in the 3-D case. The other noticeable difference
is visible when comparing the FK-spectrum of Model 1.5 (3-D) and Model 2.3 (2-D). For Model 1.5 there are no clearly distinguishable events except for the groundroll, whereas in the FK-spectrum of Model 2.2 the energy is more clearly concentrated in a couple of events. One of the events in the FK-spectra of the 2-D models is a horizontal event with a frequency of 50 Hz. It remains unclear what in the seismograms causes these horizontal events to appear in the FK-spectrum.

A comparison between the particle velocity shot gathers in the three recorded directions of Model 1.2 shows that similar events are recorded in all three directions. However, the apparent velocities observed in the shot gathers for similar events are not necessarily the same. The difference is most apparent when comparing the vertical particle velocity shot gather (Figure 4-11) with the transverse particle velocity (Figure 4-11). Both shot gathers that have back scattered events, but the apparent velocity of the events recorded in the transverse particle velocity shot gather are higher. The higher measured velocity is most likely because the recorded events in the transverse shot gather originate to either side of the receiver line. The events will thus reach the receiver line at an angle in the horizontal plane, making the velocity of the events appear higher.
In this thesis I have shown that in the presence of topography, a fluctuating soil thickness and sharp impedance contrasts a deep reflection can partly be retrieved. However, these factors are not the only ones that play a role when doing seismic exploration in the presence of near surface karst. In this thesis only buried dolines, which can be seen as a change in soil thickness, are taken into consideration, while there are more types of dolines (Chapter 2). Caprock dolines for example can have steep sides and are filled with air and rock debris, which create a high seismic impedance contrast. Further research on the effects of dolines can be done by more explicitly incorporating them into the modeling, with more control over the shape, size and distribution. Caves are another feature of karst that was not included in this thesis. However, when present caves would also represent an element in the model with a strong seismic impedance contrast with the surrounding subsurface.

It is unclear whether the methods to remove scattered events and groundroll mentioned in the Introduction (Strobbia et al., 2014; Edme et al., 2014; Halliday et al., 2010) work in the presence of shallow karstified limestone. The effectiveness of these methods could first be tested on the synthetic data created for this thesis, to determine if there are any potential problems, before testing the methods in the field.

During the model building I discovered that there is very little information available about seismic wave attenuation in carbonate rocks at an exploration frequency. Most of the available information is about attenuation at a kHz or MHz frequency range, while seismic exploration is performed in the range of 10’s of Hz. It is unclear whether the Q-factors found for limestone in the kHz to MHz range can be directly translated to the frequency range used in seismic exploration. However, in order to realistically model limestone this would be an interesting topic to investigate.

Running a FD simulation requires a lot of time, even when using computer clusters. In order to reduce computational time it would be worthwhile to investigate the possibility of using local grid refinement. The grid spacing used for the models in this thesis was a concession between the amount of detail in the model and computational time. However, there are large parts in the models where very little changes, in which case a gridspacing of 0.5 m is too fine and could be increased.


Appendix A

Location of the altitude data used in this thesis

Location of the topographic data