# A Robust Method To Estimate Kinematic Earthquake Source Parameters 

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## Zusammenfassung

Ein Ziel bei der Untersuchung von Erdbeben ist die Ermittlung von Herdparametern. Im einfachsten Fall wird der Bruchvorgang als Punktquelle beschrieben. In vielen Fällen ist man nur an den Effekten dieser ersten Näherung interessiert, hierfür existieren bereits verschiedene automatisierte Auswertungsverfahren. Die näherungsweise Betrachtung eines Erdbebens als punktförmiges Ereignis ist für bestimmte Anwendungen, wie z.B. im Rahmen von Tsunami-Frühwarnsystemen und bei der schnellen Abschätzung von Schadens-Szenarien oft unzureichend. Man möchte dort zusätzlich Informationen über Ausdehnung und zeitlichen Ablauf des Bruchprozesses aus den Beobachtungen gewinnen. Diese werden durch die kinematischen Herdparameter beschrieben. Ein Problem bei ihrer Bestimmung ist das häufige Auftreten von mehrdeutigen Lösungen, unter anderem wegen der Überparametrisierung der vorhandenen Modelle.

Ziel dieser Arbeit war es, ein robustes und automatisierbares Verfahren zum Abschätzen von kinematischen Herdparametern aus teleseismischen und regionalen Datensätzen zu entwickeln. Schwerpunkte lagen hierbei im Untersuchen von Mehrdeutigkeiten und in der Quantifizierung der Unsicherheiten der Ergebnisse.

Das Verfahren beruht auf mehreren methodischen Neuerungen: Ein neues vereinfachtes Modell für den Bruchprozess, welches die Gefahr der oben genannten Mehrdeutigkeiten minimiert, wurde vorgestellt, das sogenannte Eikonalmodell. Eine Methode zur adaptiven Gewichtung der seismischen Daten wurde entwickelt, um Fehlgewichtungen zu vermeiden. Im Gegensatz zu anderen Ansätzen wurden hier die Abweichungen zwischen echten und modellierten Daten mit einer $l^{1}$-Norm gemessen. Eine variable Kombination von verschiedenen Suchalgorithmen ermöglicht eine ausreichend vollständige Untersuchung des gesamten Parameterraums.

Um den numerischen Anforderungen dieses nichtlinearen Inversionsproblems gerecht zu werden, habe ich ein Softwarepaket entwickelt, mit dessen Hilfe man synthetische Seismogramme aus im voraus berechneten Greenschen Funktionen für ausgedehnte Herdmodelle effizient berechnen kann. Darauf aufbauend wurde ein flexibles System zur Umsetzung von Inversionsschemata erstellt, welches sich leicht an lokale, regionale und globale Anwendungen anpassen lässt.

Neben der detaillierten Beschreibung der Theorie des Verfahrens wird seine Funktionsfähigkeit mit Hilfe mehrerer Tests gezeigt. Die Anwendung wird anhand des Erdbebens von L'Aquila ( $\mathrm{M}_{\mathrm{W}} 6.3$, 2009) exemplarisch dargestellt. Weitere Erdbeben mit verschiedenen Quellgeometrien werden analysiert und die Resultate mit Referenzergebnissen verglichen.

## Abstract

Automatic methods to determine earthquake source parameters have become essential tools in modern seismology. Currently, most such methods are based on point source (i.e. moment tensor) approximations of earthquake rupture. This simple model presents a restriction for some applications. Especially in the scope of rapid hazard assessment and tsunami early warning, automatic methods revealing more details about extension and temporal evolution of the rupture process (kinematic source parameters) are of great importance. A main problem inherent to many earlier attempts in this direction is their tendency to produce unstable and ambiguous results due to overparameterization.

The aim of the work presented in the following was to investigate the possibilities to robustly determine, based on teleseismic and regional recordings, not only point source but also kinematic earthquake source parameters. The main challenges targeted, were how to identify and prevent ambiguities and how to properly quantify uncertainties of the results.

The methodical requirements were met by a combination of several advances: A new source model has been introduced, the eikonal source, which has been especially designed to avoid overparameterization. An adaptive data weighting scheme has been proposed to gain a robust and balanced procedure with respect to heterogeneous input data. The misfit function used is based on an $l^{1}$-norm between real and synthetic data to reduce the influence of outliers. Large portions of parameter-space are searched in order to detect ambiguities inherent to the specific setup of each investigated event.

To meet the computational demands of this non-linear inverse problem, I have developed a set of tools to efficiently calculate synthetic seismograms for extended earthquake source models based on pre-calculated Green's functions. Upon that, a flexible inversion framework is provided which can be tailored to various application cases on local, regional, and global scales.

In this work I explain the methodical tools which have been developed and used, and present an automatic procedure to estimate point source and kinematic source parameters for global earthquakes. It is exemplified by application
to the $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake of 2009. Details of the method are investigated through test applications to synthetic datasets. Finally, the usability of the method is shown by comparing several test cases with published results.

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## Chapter 1

## Introduction

### 1.1 Motivation

Earthquakes present a continuous threat to people in many regions on our planet. In the 40 year period between 1968 and 2008 there were 1.1 million documented fatalities from earthquakes directly due to shaking-related causes and additional 0.3 million due to secondary effects, mainly tsunami and landslides [Marano et al., 2009].

After every severe earthquake the question is raised: Can earthquakes be predicted? To our present knowledge, it is impossible to give a precise answer to when an earthquake of a given size at a specific place will happen. Regrettably it is unlikely that this will ever change, unless future research will provide us with surprising and substantially new insights into the earthquake mechanism. Fortunately this does not imply that earthquakes are unforeseeable. Many aspects of earthquakes today are well understood, and to some extent even predictable. Over the past decades, seismologists have accumulated thorough statistics on where earthquakes occur, about their strength, and their mechanisms. Our growing knowledge about earth's crustal structure, together with a continuous monitoring of earthquake and tectonical activity allow us to deliver increasingly precise statements about current and future earthquake hazard at any particular site. Although seismology cannot predict the exact timing of an earthquake, we will be able to give rather precise answers on where a major earthquake will occur, how strong it will be, what effects it will have on the surrounding environment and what threat it will be to buildings and people. Near-real-time monitoring even allows us to take action shortly after an earthquake initiates, to deliver tsunami warnings, and to give civil protection forces and governments a picture of the damage to be expected and of secondary risks immediately after an event.

A key role in the framework of earthquake monitoring and hazard assessment falls to methods which can quickly and reliably determine earthquake properties. Nowadays, location, strength, and the overall mechanism of the earthquakes are determined routinely by several national and international services. The earthquake catalogues evolving from these efforts have become indispensable resources for seismological research. However, it is still non-standard to determine other parameters of the earthquake, like its lateral extension, the geometry of the rupture surface or rupture direction and velocity, which are commonly referred to as kinematic ${ }^{1}$ earthquake source parameters.
This work is the thrilling attempt to drill deeper into the problem of earthquake source parameter estimation, to develop a stable, reliable, and automatic method to determine kinematic earthquake source parameters and to ultimately implement this method as a routine application in the German national earthquake data center (BGR), and to provide the first online catalogue for this kind of earthquake parameters on a regional and global scale.
The overall concept of the method is based on various previous studies, but a few innovations have been made and put together, namely in the details on how the earthquake is parameterized, how an efficient forward modeling is done, how misfits are calculated and how errors on the obtained results are quantified.
How can we find out what happened inside earth during a particular earthquake? Of course we cannot observe the earthquake directly. Even if we were able to make a movie of the earthquake, just exactly of how its rupture cut its way through the surface, we would still be unsure of what had happened underneath. In many cases, the best we can get are seismograms, records of the ground motion, taken hundreds or even thousands of kilometers away from the epicenter.

The general challenge is that we know the effects (the seismograms), but are interested in the cause of these (the earthquake). This type of problem, the inverse problem, is in general radically harder to solve than the one we encounter, when we have given a cause and want to predict effects (the forward problem). In many cases, we find it to be so difficult, that we have to use a joker strategy to find solutions: systematically by trial and error.

The trials here are simulations of what could have happened inside earth during the earthquake and a forward modeling of the seismic wave propagation through earth, resulting in a set of synthetic seismograms for the simulated earthquake. The error here is a measure of the mismatch between the simulated and the in reality observed seismograms. Ideally, after some trials, we would eventually

[^0]make a 'perfect guess', in the sense, that the simulated earthquake would exactly match what had happened in nature and the difference between observed and simulated seismograms would vanish, completely.

A number of critical questions must be raised: Couldn't it be, that there are two different hypothetical earthquakes leading to the same observations, so that given only these observations it would be impossible to decide which one just happened? Obviously, there is noise in the observations, ground motion not caused by the earthquake but by other sources, so how do you prevent that you accidentally explain this noise to be due to the earthquake? Also because of noise, it will never be possible to get a perfect match between simulation results and observations - so how can you prove that your best trial is the actual solution to the problem? There are an infinite number of possible hypothetic earthquakes and it is impossible to test them all, so one has to use a simplified earthquake model. How can you be sure that the simplifying assumptions made are justified? Does the model of the earthquake have enough in common with what actually happens in nature? And finally, the modeling of seismic wave propagation cannot be done without error - such a modeling would require not only a knowledge of earth's interior at a yet unreached level of detail, but also a computation facility of yet unavailable power in order to evaluate the thousands of trials needed to find a solution.

So how is it possible, with noisy observations, imprecise synthetic seismograms based on inexact earth models, find the true one out of an infinite number of potential realizations of an oversimplified earthquake model in an inherently ambiguous problem?

### 1.2 Historical review

The use of simple models to describe the earthquake rupture and a convenient set of assumptions concerning the simulation of wave propagation and the fitting procedure were extremely helpful in the past to overcome part of these limitations, allowing a first modeling of the earthquake source. The assumption of a point source double couple (DC) model for the earthquake source, joined with the adoption of averaged 1D earth models and the fit of low frequency seismograms, allowed the modeling of the earthquake source in terms of centroid location and a shear crack model, which could be easily related to local faults and tectonic features. The main limitation of the DC model, apart that it gives no information about the extension of the ruptured area and the finiteness of the rupturing process, concerns the intrinsic ambiguity between fault and auxiliary plane, which limits result interpretations in terms of local tectonics. A more complete source model, theoretically described by means of a full moment tensor, could give ev-
idence for deviatoric terms and isotropic component. Moment tensor representation has been extensively used to model the earthquake source, the non-DC component being interpreted both as a source effect, for example the rupturing of sub-events with different focal mechanisms, or as a spurious term arising by mismodeling of the Earth structure and the wave propagation. Moment tensor solutions, for which catalogues are available both for global and regional seismicity, still present those limitations discussed for DC point source models.

Kinematic models have been therefore developed in order to describe the finiteness of the earthquake source. Early models, namely the Haskell [1964, 1966] and Brune [1970] source models, assumed constrained planar rupture areas of rectangular or circular shape. Although introducing important constraints to the rupturing process, their implementation has been successful for the description of several earthquakes. One of the major advantages of using these models to describe the source kinematics is that they are defined by a small number of parameters (e.g. length and width, or radius, are sufficient to describe the rupture area; nucleation point coordinates may take into account directivity effects, etc.). This type of rupture models have the common property, that they are based on a parameterization of the rupture front.

Since the early nineties, a different approach has come into use within the seismological community. Following this modeling approach, the extended source is discretized into a number of point sources, which are allowed to behave more heterogeneously. Typical constraints include the propagation of the rupture along the fault plane given a fixed realistic rupture velocity, and often some limitations concerning slip directions. Results of this modeling can be plotted by a map of slip distribution, indicating the slip (size and direction) along the rupture area, which is interpreted in terms of asperities or slip patches. As the nucleation point is also typically inverted, isochrones of the rupture process are also produced.

This model has the clear advantage of allowing the representation of more complex and, possibly, realistic rupturing processes. However, the number of parameters describing the model is growing critically together with the rupture size and the implementation of the discretized model (the distribution of point sources should be dense enough to allow the retrieval of the desired parameters). As a consequence the inverse problem may result over-parameterized and its solution highly non-unique. A significant case has been described by Beresnev [2003], for the Mw 7.61999 Izmit, Turkey, earthquake. The comparison of slip map results by different authors for this well studied earthquake showed significant inconsistencies, providing a set of very different images of the rupture process. Since all authors could satisfactorily reproduce seismic or other geophysical observations, the ambiguity between different solutions can be hardly solved on the base of the data fit. The result overview simply reflects the non-uniqueness of solutions using this inversion approach. The critical review of kinematic inversion by Beres-
nev [2003] highlighted different weaknesses of typical inversion approaches.
It is worth to discuss these problems more in detail, in view of the proposition of a new kinematic source model. The non-uniqueness of solutions for finite-fault slip inversions, originally discussed by Olson and Apsel [1982], is not the unique source of uncertainties for this inverse problem. According to Beresnev [2003], and to the references mentioned therein, the application of source constraints may be required in order to avoid geologically meaningless solutions, which could still better fit the data. At the same time, specific assumptions concerning the seismological parameters describing the rupture process are often done, in order to simplify the inversion process. The selection of different source constraints and seismological parameterization is a subjective process and synthetic tests [e.g. Olson and Anderson, 1988, Das and Suhadolc, 1996, Das et al., 1996, Saraó et al., 1998, Henry et al., 2000] have clearly shown how specific choices may lead to significantly different images of the rupturing process, with the consequent risk of its misinterpretation. Another source of uncertainty arises by the process of discretization of the finite source. The specific choice will map into the numerical approximation of continuous integral along the rupture area. The safe choice of cell size to discretize the rupture area was discussed in early studies by Hartzell and Helmberger [1982] and Olson and Anderson [1988], but often not considered in following kinematic applications. Furthermore, Beresnev [2003] identifies the limitations concerning slip duration and the source time function shape as additional possible source of errors for standard finite fault slip inversions.

Newer slip inversion techniques use ensemble statistics of inversion results, to extract robust features of a finite fault inversion instead of defining just one single best source model by means of data misfit optimization [Piatanesi et al., 2007], but the general problems still apply.
The overview of kinematic inversion arising from this discussion points out the difficulty of the inversion task, the non-uniqueness of slip inversion results and the high chance that subjective choices affect the final result; on the other side, synthetic tests are suggested as the best approach to evaluate the sensitivity of the inversion process. According to this situation, the interest in the development of a simplified kinematic model becomes evident, which improves the point source model, describes the gross rupturing process along an extended fault, but still allows a more stable inversion, with respect to finite-fault slip inversion techniques. Previous attempts in this direction include the works of Dahm and Krüger [1999], McGuire et al. [2001], and Vallée and Bouchon [2004].

### 1.3 Project environment

The work presented in the following is the scientific outcome of the DFG project KINHERD (see appendix B). The target of this project was to improve methods to estimate kinematic earthquake source parameters. A new source model, inversion and data handling tools, Green's function databases, application sets and different variants of kinematic source inversion strategies have been developed. I frequently use plural forms in this work to express that parts of these results originated in collaboration with project partners. Results of my work and the tools I have developed have vice versa been used and successfuly applied in several cases by collaboration partners (see appendix C).

### 1.4 Novelties of our approach

Here I propose a new model, the eikonal source model, which has been specifically drawn in order to account for some of the problems mentioned in section 1.2. First, the model is intended to require as few parameters as possible to be fully defined. It is also built as the evolution of a point source model in the sense that the parameters describing the point source are also used for the kinematic model, and can be stably defined in advance through a point source inversion, if required. Geology is taken into account, with the attempt to provide a safe, flexible and not subjective implementation, with the final effect of limiting the area of where the rupturing process can take place to the seismogenic region. The problem of the discretization of the extended rupture is also taken into account, allowing an easy and flexible implementation.

In the further work, strategies are developed, to automatically and stably determine the parameters of the eikonal model or other point- and kinematic models based on fitting synthetic and observed seismograms. A non-standard feature is the use of an $l^{1}$-norm to measure the difference between synthetic and observed seismograms or spectra, which makes the described method less sensitive to outliers. Because of the use of an $l^{1}$-norm, the fitting of amplitude spectra, and because the parameters of our model do not enter linearly into the observations, a non-linear inverse problem has to be solved. Due to the presence of local minima in the misfit function, iterative strategies combined with grid searches are applied. Finally, to analyze the stability of the results, and to quantify the uncertainties of the retrieved parameters, I apply a bootstrap technique.

The careful analysis of this non-linear inverse problem requires a vast number of synthetic seismograms to be calculated by our method.

Together with the eikonal model, we provide tools for its handling within inver-
sion algorithms. The set of developed algorithms, unified under the common name of the Kiwi (Kinematic Waveform Inversion) Tools, allow a wide set of tasks, including basic data processing, handling of Green's functions databases, generation of synthetic seismograms, and inversion procedures. Although different point source and kinematic source models may be implemented by the Kiwi tools, these will be here specifically discussed in relation to the implementation of the eikonal source model.

### 1.5 Structure of this work

In chapter 2, I define a rupture model and develope the tools to solve the forward and inverse problems associated with it. In chapter 3, these tools are combined and applied to the $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake of 2009 (section 3.1). A multi-step inversion strategy, suitable to be run without human supervision, is designed. In chapter 4 properties of method and application are studied under the ideal conditions only given by synthetic data. In chapter 5 kinematic source parameters for three additional test cases are compared and a statistical comparison of the point source parameters estimated with our method is given. Conclusions are given in chapter 6.

## Chapter 2

## Method

### 2.1 Overview

In this chapter I describe a rupture model and the tools required solve the forward and inverse problems associated with it.
A flexible kinematic model of earthquake rupture, the eikonal source is introduced in section 2.2. It is based on 13 inversion parameters of which five parameters are needed to define extension and propagation of rupture. The low number of parameters is achieved by the use of geometrical and physical constraints.

The forward modeling problem (i.e. the calculation of synthetic seismograms for extended sources) is solved on the basis of pre-calculated Green's functions (section 2.3). The benefit is two-fold: the inversion is independent of the Green's function generation, and the forward modeling is relatively fast. Rules on how to safely discretize the parameterized source model into point sources are given in section 2.4. The requirements for the Green's function storage are discussed in section 2.5. The interpolation of Green's functions is considered in section 2.5.1.

The topic of how seismograms and simulation results are compared is layed out in section 2.6. It is subject to a definition of misfit (section 2.6.1), data weighting (section 2.6.2), and data selection with tapering and filtering (section 2.6.4).

Section 2.7 has been dedicated to the finding of solutions to the inverse problem. It is necessary to search misfit space for a global minimum, i.e. to find the choice of model parameters which produces the best fit in the data (section 2.7.1). To quantify errors or probabilities on the retrieved results a bootstrap technique has been adapted (section 2.7.2).
The chapter is completed with some considerations on the automatic processing of earthquake event data in section 2.8. A procedure to evaluate a priori station
qualities is developed (section 2.8.1), which can be used in combination with a special station selection algorithm (section 2.8.2).

The tools developed in this chapter enable us to automatically estimate values and uncertainties for the following earthquake parameters: centroid time, location, and depth, scalar moment, orientation of the fault plane, slip direction, rise-time, relative position of the nucleation point with respect to the centroid location, size of the rupture surface, and rupture velocity.

### 2.2 A kinematic earthquake model based on rupture fronts: The eikonal source

In order to gain a robust inversion procedure, the goal is to design a description of the earthquake rupture using as few parameters as possible but which is still able to cover the main features of the rupture process. This is a compromise that may lead to some restrictions: if we try to image rupture at a very detailed level, the methods quickly become unstable, leading to ambiguous results.
We employ a generalized rupture-front/healing-front description of the rupture process, that is based on a variable rupture velocity, does not assume planar rupture fronts and considers natural boundaries of a rupture plane. Prior assumptions, such as the geometry of the seismogenic zone can be specified to reduce the number of free parameters. The propagating curved rupture or healing front is modeled by equations usually applied to ray propagation problems, i.e. by the eikonal equation [e.g., Aki and Richards, 2002, page 87]. This assumes that an approximate analogy between rupture fronts and wave fronts can be used. Such an analogy is indicated by field data for tensile cracks [e.g., Müller and Dahm, 2000], but has not been applied to shear crack rupture so far. The analogy is useful for our purpose to derive a simple representation of rupture and healing front, as we show below, but it is not strictly valid in general. For instance, rupture fronts may be continuous or discontinuous at a material interface, depending on the difference in strength, while a wave front of a first arrival is always continuous at such an interface. As long as the rupture velocity field is smooth, the analogy is a convenient approximation of the true behavior of earthquake rupture.

A point source model for a pure shear crack serves as a starting point for investigations on the extension of the rupture. This base model is described by nine parameters: time $t$, location and depth $(x, y, z)$, scalar moment of the event $M_{0}$, orientation of the fault plane (strike $\phi, \operatorname{dip} \delta$ ), direction of slip on the fault plane (slip-rake $\lambda$ ) and the event duration $T$ (see figure 2.1 a).

Location and depth are measured in a local Cartesian coordinate system with its


Figure 2.1: Source model parameterization. The eikonal source model is defined by 13 parameters: time, location, and depth $(t, x, y, z)$ of center point (relative to a fixed origin), scalar moment $M_{0}$, orientation of the fault plane (strike $\phi, \operatorname{dip} \delta$ ), slip direction (slip-rake $\lambda$ ), border radius $R$, relative location of the nucleation center $\left(n_{s}, n_{d}\right)$, relative rupture velocity $v_{r} / v_{s}$, and rise-time $\tau$. Constraints introduced are: shear wave velocities $v_{s}(x, y, z)$ and geometry of the seismogenic zone (surface, lower bound).
principal axes pointing north, east, and downward. The local coordinate system's origin is at the surface, at an arbitrary preliminary location in the source region. The usual conventions for strike, dip, and slip-rake are used [Jost and Herrmann, 1989].

Extension and time history of the rupture, are defined by five additional model parameters and a fixed set of geometrical constraints. The geometrical constraints are chosen so that they represent the seismogenic zone wherein earthquake rupture shall be possible. E.g. for an application to a crustal earthquake, where rupture happens within the brittle part of the crust, we would use two constraining planes: one for the surface and one for the bottom boundary of the brittle zone. Rupture, which is initially allowed at any point on the infinite fault plane, is thus first restricted to this zone. Secondly, we allow rupture only to happen within a certain radius $R$ away from the previously defined point source location labeled center point in figure 2.1. This radius $R$ is the only variable in our model controlling the size of the of the rupture surface.

A circular length $R$ to define the size of the rupture, together with possible geometrical boundaries, is justified if the slip distribution on the rupture plane is
smooth, and if the geometrical boundaries do not shift the centroid point significantly. The radius parameter then ensures that the centroid location of the point source and the extended source model are comparable. This feature is beneficial in the multi-step inversion approach as described later. Note, that the center point of the construction, given with the point source location does not necessarily have to coincide with the mean centroid of the earthquake.

Two further parameters are required to locate the nucleation point within the rupture surface. These are two coordinates measuring the offset between center point and nucleation point along strike $n_{s}$ and down dip $n_{d}$.
For the temporal evolution of the rupture we assume, that slip at individual points on the rupture surface is triggered by a rupture front, which is propagating from the nucleation point outward until the complete surface has ruptured. The rupture front is followed by a healing front, which stops slip at individual points on the rupture surface. Each point on the rupture surface can only rupture once. Given the rupture velocities $v_{r}\left(x_{s}, x_{d}\right)$ within the rupture surface, the time at which each point is reached by a front $t_{r}\left(x_{s}, x_{d}\right)$ is calculated by solving the eikonal equation

$$
\begin{equation*}
\left|\nabla t_{r}\left(x_{s}, x_{d}\right)\right|=\frac{1}{v_{r}\left(x_{s}, x_{d}\right)} \tag{2.1}
\end{equation*}
$$

where $x_{s}$ and $x_{d}$ are rupture coordinates measuring along strike, and down dip, respectively. Similarly, the propagation of the healing front is derived from the same equation when interchanging the rupture velocity by a healing velocity field $v_{h}$, and additionally introducing a retardation (time shift) of the onset of the healing front at the nucleation point (e.g. defined by the rise-time at the nucleation point by $t_{r}-t_{h}=\tau$ ).

$$
\begin{equation*}
\left|\nabla t_{h}\left(x_{s}, x_{d}\right)\right|=\frac{1}{v_{h}\left(x_{s}, x_{d}\right)}, \tag{2.2}
\end{equation*}
$$

As a simplifying assumption, we choose rupture velocity to be proportional to the shear wave velocity on the rupture surface. The constant factor between rupture velocity and shear wave velocity $v_{r} / v_{s}$ will be our next model parameter. Such a relation is justified by different field experiments and theoretical arguments for rupture in homogeneous media [Broberg, 1996, Berezovski and Maugin, 2007]. Often, for earthquakes, a rupture velocity in the range between $50 \%$ and $90 \%$ of the shear wave velocity is found [Geller, 1976]. Geller [1976] give 72\% as a mean value. McGuire [2004] finds rupture velocities of $80 \%$ of Rayleigh velocity for M 2.7 earthquakes. Park and Mori [2008] estimate values of $20 \%-40 \%$ of the shear wave velocity for deep-focus earthquakes. An overview and further references are given in Park and Mori [2008].

Independent determination of rupture velocity for earthquakes is difficult, because to first order, both, lower rupture velocity and larger rupture length, lead to wider pulses in the observed seismograms.

Also for simplicity, we choose a constant time delay between rupture front and healing front and assume the healing velocity equal to the rupture velocity. This time delay is commonly referred to as rise-time $\tau$. The assumption leads to slip pulses of equal length over the whole rupture plane. However, other healing velocity models may be easily considered in future applications.
Note, that if the extension of our model is set to zero $(R=0)$, the model is effectively reduced to a point source. In this case, the event duration $T$ coincides with the rise-time $\tau$ of the single remaining rupture point. We use the term pointsource rise-time when referring to this type of event duration. If we approximate an extended source using a point source model, the rise-time of the extended source model is typically shorter than that of the point source approximation.
The displacement, as seen when looking at an individual point of the rupture surface, is modeled as a linear function of time in the time interval given by the incidents the two fronts pass by.

The amount of final- or static displacement $u_{s}$ is connected to the scalar moment $M_{0}$ released by a portion of the fault by

$$
\begin{equation*}
\mathrm{d} M_{0}=\mu u_{s} \mathrm{~d} A \tag{2.3}
\end{equation*}
$$

where $\mu$ is the shear modulus and $A$ is the fault area [e.g. Aki and Richards, 2002]. Using (2.3), we choose to distribute the scalar moment evenly over the fault area as another simplifying assumption (constancy of moment tensor density). Similarly one could also demand for constancy of slip, which would lead to a moment tensor density increasing with depth.
Note that all the simplifying assumptions introduced in the last section are not required by the eikonal model. They are introduced to reduce the number of independent parameters when the resolution of source parameters is limited. If better guesses for rupture velocity, healing velocity, rise time and slip can be made, these may be included in the parameterization of the eikonal source model.
In any case, in addition to the eight point source parameters $\left(t, x, y, z, M_{0}, \phi\right.$, $\delta, \lambda)$, we have five inversion parameters which describe the extension and time evolution of the rupture in our model: $R, n_{s}, n_{d}, v_{r} / v_{s}$ and $\tau$.

Despite the manageable number of parameters, a variety of simple earthquake ruptures can be approximated by this source model. Figure 2.2 illustrates some examples. The simplifying assumptions introduced in our eikonal source model are not helpful to describe earthquakes consisting of multiple sub-events, when the rupture takes place on a curved fault, when more than one fault is involved


Figure 2.2: Examples of different rupture models covered by our source parameterization: (a) circular, (b) unilateral, (c) bilateral. Given additional constraints, somewhat more complicated source geometries can be studied, e.g. ruptures in a subduction slab (d). Filled white circles represent the nucleation point. Black contours are isochrones at constant time intervals. This figure additionally illustrates the possibility of using different rupture velocity profiles: (a) homogeneous, (b) positive depth gradient, (c) negative depth gradient, and (d) sharp discontinuity at a fixed depth.
or when slip direction or amount or rupture velocity are strongly heterogeneous. However, if shear wave, rupture and healing velocity would be decoupled, the eikonal source model is able to simulate more complex ruptures, for instance to consider the first order effects of barriers and asperities (see figure 2.3).

### 2.3 Forward modeling

In this section, the formulas to calculate synthetic seismograms from pre-calculated Green's functions for extended sources are given for the important case of spherical/cylindrical symmetry. It is shown, that ten Green's function components are required to compose seismograms for any moment tensor point source in this case ${ }^{1}$.

In Cartesian coordinates, the displacement $u_{n}$ due to a moment tensor density $m_{p q}$ distributed on a rupture surface $\Sigma$ is

$$
\begin{equation*}
u_{n}(\boldsymbol{x}, t)=\iint_{\Sigma} m_{p q}\left(\boldsymbol{x}^{\prime}, t\right) * G_{n p, q}\left(\boldsymbol{x}, \boldsymbol{x}^{\prime}, t\right) \mathrm{d} \Sigma \quad \boldsymbol{x}^{\prime} \in \Sigma, \tag{2.4}
\end{equation*}
$$

with $n, p, q \in\{x, y, z\}$ and where $G_{n p, q}$ is the Green's tensor and the asterisk denotes temporal convolution [e.g. Aki and Richards, 2002].

For a spherically symmetric earth model, the Green's tensor depends only on the depths of source element and receiver, the surface distance $r^{\prime}\left(\boldsymbol{x}, \boldsymbol{x}^{\prime}\right)$ between them and the azimuth $\varphi^{\prime}\left(\boldsymbol{x}, \boldsymbol{x}^{\prime}\right)$ of the receiver as seen at the source element. This can be used to reduce the number of independent Green's tensor components from eighteen to ten [e.g. Müller, 1985].

A fixed point at the surface above the source serves as the origin for two coordinate systems: First, a Cartesian coordinate system in which we specify the source's moment tensor density $\boldsymbol{m}\left(\boldsymbol{x}^{\prime}, t\right)$ (the axes of this system are chosen so that they point north $\hat{\boldsymbol{e}}_{x}$, east $\hat{\boldsymbol{e}}_{y}$, and downward $\hat{\boldsymbol{e}}_{z}$ ). Second, a curvilinear system is used to locate the receiver using surface distance $r$, azimuth $\varphi$ and depth $z$. The components of displacement at the receiver will be measured radially along $\hat{\boldsymbol{e}}_{r}$, transversely along $\hat{\boldsymbol{e}}_{\varphi}$ and downward along $\hat{\boldsymbol{e}}_{z}$. In these coordinates, the displacement (2.4) may be written as

$$
\begin{equation*}
u_{n}(r, \varphi, z, t)=\iint_{\Sigma} R_{n n^{\prime}}\left(\lambda-\lambda^{\prime}\right) \cdot\left[m_{p q}\left(\boldsymbol{x}^{\prime}, t\right) * G_{n^{\prime} p, q}\left(z, r^{\prime}, \varphi^{\prime}, z^{\prime}, t\right)\right] \mathrm{d} \Sigma \tag{2.5}
\end{equation*}
$$

[^1]

Figure 2.3: Rupture evolution under the presence of heterogeneities in the rupture velocity. Isochrones of the rupture front are shown as black lines. The star indicates the nucleation center. Except in the area indicated by a gray circle, the rupture velocity is $v_{0}$. Within this part of the fault it has been lowered to $0.67 v_{0}$ in (a), and raised to $1.5 v_{0}$ in (b).
with $n, n^{\prime} \in\{r, \varphi, z\}$ and $p, q \in\{x, y, z\} . \boldsymbol{R}$ is a matrix describing rotation around $\hat{\boldsymbol{e}}_{z}$, accounting for the differing back-azimuths $\lambda$ and $\lambda^{\prime}$ to origin point and source element, respectively.

For media invariant to rotation around $\hat{e}_{z}$ at the source element, the Green's tensor $\boldsymbol{G}=\boldsymbol{G}\left(z, r^{\prime}, \varphi^{\prime}, z^{\prime}, t\right)$ can be given as a rotated version of the special case Green's tensor $\boldsymbol{G}^{\mathbf{0}}=\boldsymbol{G}\left(z, r^{\prime}, 0, z^{\prime}, t\right)$ at zero azimuth:

$$
\begin{equation*}
G_{n^{\prime} p, q}=R_{p q^{\prime}}\left(\varphi^{\prime}\right) G_{n^{\prime} p^{\prime}, q^{\prime}}^{0} R_{q p^{\prime}}\left(\varphi^{\prime}\right) \tag{2.6}
\end{equation*}
$$

In the following, the geometry is reduced to this case. At zero azimuth, P-SV motion can only be excited by moment tensor components $m_{x x}, m_{y y}, m_{z z}, m_{x z}$, and $m_{z x}$, due to the symmetries taken into account. Similarly, SH motion can only be exited by moment tensor components $m_{x y}$ and $m_{y x}$ there. This reduces $G^{0}$ to the shape

$$
\begin{aligned}
& \boldsymbol{G}_{r}^{\mathbf{0}}=\left(\begin{array}{ccc}
G_{r x, x}^{0} & 0 & G_{r x, z}^{0} \\
0 & G_{r y, y}^{0} & 0 \\
G_{r z, x}^{0} & 0 & G_{r z, z}^{0}
\end{array}\right) \\
& \boldsymbol{G}_{\varphi}^{\mathbf{0}}=\left(\begin{array}{ccc}
0 & G_{\varphi x, y}^{0} & 0 \\
G_{\varphi y, x}^{0} & 0 & G_{\varphi y, z}^{0} \\
0 & G_{\varphi z, y}^{0} & 0
\end{array}\right) \\
& \boldsymbol{G}_{z}^{\mathbf{0}}=\left(\begin{array}{ccc}
G_{z x, x}^{0} & 0 & G_{z x, z}^{0} \\
0 & G_{z y, y}^{0} & 0 \\
G_{z z, x}^{0} & 0 & G_{z z, z}^{0}
\end{array}\right) .
\end{aligned}
$$

Thus, with (2.6) the convolution of $\boldsymbol{m}$ with $\boldsymbol{G}$, as bracketed in (2.5), reduces to

$$
\begin{align*}
{\left[m_{p q} * G_{r^{\prime} p, q}\right]_{r^{\prime}}=} & \left(m_{x x} \cos ^{2} \varphi+m_{y y} \sin ^{2} \varphi+m_{x y} \sin 2 \varphi\right) * G_{r^{\prime} x, x}^{0} \\
& +\left(m_{x x} \sin ^{2} \varphi+m_{y y} \cos ^{2} \varphi-m_{x y} \sin 2 \varphi\right) * G_{r^{\prime} y, y}^{0} \\
& +\left(m_{x z} \cos \varphi+m_{y z} \sin \varphi\right) *\left[G_{r^{\prime} x, z}^{0}+G_{r^{\prime} z, x}^{0}\right] \\
& +m_{z z} * G_{r^{\prime} z, z}^{0} \\
{\left[m_{p q} * G_{\varphi^{\prime} p, q}\right]_{\varphi^{\prime}}=} & \left(\frac{1}{2}\left(m_{y y}-m_{x x}\right) \sin 2 \varphi+m_{x y} \cos 2 \varphi\right) *\left[G_{\varphi^{\prime} x, y}^{0}+G_{\varphi^{\prime} y, x}^{0}\right]  \tag{2.7}\\
& +\left(m_{y z} \cos \varphi-m_{x z} \sin \varphi\right) *\left[G_{\varphi^{\prime} y, z}^{0}+G_{\varphi^{\prime} z, y}^{0}\right] \\
{\left[m_{p q} * G_{z^{\prime} p, q}\right]_{z^{\prime}}=} & \left(m_{x x} \cos ^{2} \varphi+m_{y y} \sin ^{2} \varphi+m_{x y} \sin 2 \varphi\right) * G_{z^{\prime} x, x}^{0} \\
& +\left(m_{x x} \sin ^{2} \varphi+m_{y y} \cos ^{2} \varphi-m_{x y} \sin 2 \varphi\right) * G_{z^{\prime} y, y}^{0} \\
& +\left(m_{x z} \cos \varphi+m_{y z} \sin \varphi\right) *\left[G_{z^{\prime} x, z}^{0}+G_{z^{\prime} z, x}^{0}\right] \\
& +m_{z z} * G_{z^{\prime} z, z}^{0} .
\end{align*}
$$

The Green's tensor components which are needed can be calculated with any method capable of generating synthetic seismograms on a regional or global scale [e.g. Wang, 1999, Friederich and Dalkolmo, 1995].

$$
\begin{array}{lll}
g_{1}=G_{r^{\prime} x, x}^{0} & g_{2}=G_{r^{\prime} x, z}^{0}+G_{r^{\prime} z, x}^{0} & g_{3}=G_{r^{\prime} z, z}^{0} \\
g_{4}=G_{\varphi^{\prime} x, y}^{0}+G_{\varphi^{\prime} y, x}^{0} & g_{5}=G_{\varphi^{\prime} y, z}^{0}+G_{\varphi^{\prime} z, y}^{0} &  \tag{2.8}\\
g_{6}=G_{z^{\prime} x, x}^{0} & g_{7}=G_{z^{\prime} x, z}^{0}+G_{z^{\prime} z, x}^{0} & g_{8}=G_{z^{\prime} z, z}^{0} \\
g_{9}=G_{r^{\prime} y, y}^{0} & g_{10}=G_{z^{\prime} y, y}^{0} &
\end{array}
$$

An arbitrary seismic trace for azimuth $\varphi=0$ can be set up as a linear combination of $g_{1}-g_{10}$, with the entries of the moment tensor as weighting factors. For this reason, they can be interpreted as elementary seismograms.

The components $g_{9}$ and $g_{10}$ contain only near field terms [Müller, 1985], so they can be neglected for far field applications, as in our case.

### 2.4 Discretization of the source model

Lets assume for the moment that the centroid location, a possible orientation of the rupture plane, and the moment of the point source are known. The problem
then is to properly discretize the eikonal source and other rupture front models on the given rupture plane. A practical approach to calculate the displacement with equation (2.5) numerically is to replace the continuous moment tensor density by a sufficiently large number of moment tensor point sources. This converts the integrations and the convolution into a sum. ${ }^{2}$

As the number of required points enters linearly into computation time, it would be nice to know how far the point sources can be taken apart without introducing large errors. The discretization can be regarded as a two step process. First, the rupture surface is divided into sub-faults of homogeneous moment tensor density wherein each point radiates with a common, but time shifted, source time function, and then secondly, it is pretended that all sub-fault motion happens at the centers of these sub-faults. This results in a convolution of the source time function with the sub-fault's rupture history. The errors introduced by this approximation can be grouped into three different categories:

- Each point in the sub-fault incorrectly radiates the same Green's function waveform,
- The temporal shifts of the signal radiated by different parts of the sub-fault are neglected,
- Suppression of local inhomogeneities of the moment tensor density inside the sub-fault.

On the small scale limit, where the Green's function waveforms involved do not change significantly over the sub-fault, the second kind of errors will always dominate and we can find a simple rule on how large the sub-faults can be made without introducing large errors due to aliasing, given a certain frequency limit.

$$
\begin{equation*}
\Delta x<\frac{1}{2}\left(\frac{1}{v_{\min }}+\frac{1}{\beta_{\min }}\right)^{-1} \frac{1}{f_{\max }} \tag{2.9}
\end{equation*}
$$

where $\Delta x$ is the extension of the sub-fault, $v_{\text {min }}$ and $\beta_{\text {min }}$ are the minimal rupture and shear wave velocities within the fault, and $f_{\max }$ is the maximal frequency where the approximation shall hold.

Going to larger scales the other categories additionally become important making it very difficult to give such a rule without the use of forward modeling for error estimation. Also it is important to note, that the first kind of errors can affect even the lowest frequencies in the observed waveforms.

[^2]
### 2.5 Green's function databases

The calculation of synthetic seismograms with (the summation equivalent to) equation (2.5) is simply a summation of many weighted and time-shifted Green's tensor components. Because the same Green's tensor components are used over and over again during a typical application, it is advisable to calculate and store the special case Green's function components $g_{1} \ldots g_{10}$ from (2.8) in advance.

Although it would be possible to calculate only exactly the parts of the Green's function which are needed for a given application, we find it more practical to broadly calculate vast sets of Greens functions for large epicentral distance ranges in advance. This is convenient for several reasons. First, it makes the inversion code completely independent of the code which generates the Green's functions. Secondly, once such a Green's function "database" has been created, it can be reused for other purposes and shared with other researchers.

As seen in section 2.3, if we only consider receivers at the surface of a spherically symmetric earth, the Green's function depends on source depth, surface distance between source element and receiver, and time. Thus, it is necessary to evaluate time traces of the ten required Green's tensor components (2.8) on a grid of source depths and surface distances.

The spacing $\Delta x$ of this grid has to be dense enough, that pulses traveling with the largest horizontal slownesses $s_{\max }$ in the wave-field do not get separated by more than half a period $T_{\min }=1 / f_{\max }$ between two neighboring grid nodes.

$$
\begin{equation*}
\Delta x \ll \frac{1}{f_{\max } s_{\max }} \tag{2.10}
\end{equation*}
$$

A similar condition should be enforced for the vertical spacing of source depths $\Delta z$. If these conditions are met, we do not introduce large errors, when using bilinear interpolation to evaluate the Green's function between grid nodes of the pre-calculated Green's function.

Considering that trace length grows approximately with the difference between maximum and minimum slowness over distance, the amount of disk space needed to store the ten Green's function components is about

$$
10 \times \frac{x_{\max }-x_{\min }}{\Delta x} \times \frac{z_{\max }-z_{\min }}{\Delta z} \times \frac{\left(x_{\max }+x_{\min }\right)\left(s_{\max }-s_{\min }\right)}{2 \Delta t} \times \nu,
$$

where $x$ corresponds to distance, $z$ to depth, $s$ to slowness, $t$ to time and $\nu$ is the space needed to store a single data sample.

To get a feeling on how much space is needed in a typical application let us examine the following example. I intend to use frequencies of up to 0.3 Hz , so that I can expect to get a rough picture of rupture structure on the order of 10 km .

For teleseismic body-waves, the slowest apparent velocities are about $3.5 \mathrm{~km} / \mathrm{s}$, so grid spacing should stay below 6 km . If I include surface waves which have a slower apparent velocity of maybe $2.5 \mathrm{~km} / \mathrm{s}$ at the frequencies in question, grid spacing should be even below 4 km . So 1 km grid spacing seems like a safe choice for this setup. I use the same lateral spacing also in depth. If I then pick a reasonable sampling interval of 2 Hz , so that the Nyquist frequency is well above the interesting frequencies, an uncompressed database with 4 byte per sample floating point numbers will require in total about 545 GB for complete epicentral distance coverage of earth and source depths of up to 100 km .

In our implementation in the Kiwi Codes (A.1) the Green's function traces are stored in a file format based on HDF5 [The HDF Group, 2009], which provides a stable platform independent binary storage mechanism. A Fortran 95 module providing a simple interface to store and retrieve Green's function traces into and from such databases has been implemented. Each Green's function database can be split into several files. Each file has a trace lookup index associated with it allowing fast random access to the traces. In a typical application, only traces clustering approximately plus-minus source length around the source-to-receiver distances in use, are employed. For moderate earthquakes in a global application, this is only a small fraction of the database, but these traces are used over and over again. To encourage this behavior and speed up repetitive access, traces are cached in RAM, once accessed ${ }^{3}$.

Finally, a set of command line programs is provided with the Kiwi Tools to create such Green's function databases, to fill traces into and to extract traces from them. An overview on the available tools is given in appendix A.1.

### 2.5.1 Green's function interpolation

If the Green's functions have been evaluated on a reasonably fine grid, bilinear interpolation can be safely used to evaluate the Green's functions between precalculated nodes: Given the neighboring node values (or Green's function traces) $g(0,0), g(0,1), g(1,0)$, and $g(1,1)$ the function $g(x, z)$ is approximated by

$$
\begin{align*}
g(x, z) & \approx g(0,0)(1-x)(1-z) \\
& +g(0,1)(1-x) z  \tag{2.11}\\
& +g(1,0) x(1-z) \\
& +g(1,1) x z \quad \text { with } \quad x, z \in[0,1] .
\end{align*}
$$

The required grid spacing depends on the the highest frequencies to be modeled and the smallest apparent velocities of the seismic waves to be considered to

[^3]avoid aliasing effects. By using more sophisticated interpolation techniques, it can be possible to partly overcome the spacial aliasing limit; one example is the $f-k$ or $f-k_{x}-k_{y}$ interpolation method for spatially aliased data by Gülünay [2003] ${ }^{4}$

### 2.6 Comparing seismograms: misfit definitions

To ultimately search for a "best fitting" earthquake model by comparing simulation results with observations, we need a way to quantify the difference between synthetic and observed seismograms. We use the term misfit to identify this difference. The misfit function should evaluate to zero when the synthetic data exactly match the observations and give positive values for disagreement. Its exact definition however, is application dependent. The misfit function is affected by the choice of norm, by the tapers and filters applied to the traces, by weighting and by the choice to compare the data in the time domain, spectral domain, or other ${ }^{5}$. Misfit function design is a subjective task, guided by the need to gain a good signal-to-noise ratio and to suppress the influence of inaccuracies in the forward modeling. Sometimes it is also desired to increase the sensitivity of the misfit function to the particular set of parameters to be investigated, whilst making it less sensitive to the influence of others.

### 2.6.1 Misfit definition and normalization

Depending on the application, we use different kinds of misfit definitions. They are either based on an $l^{1}$-norm or $l^{2}$-norm and are applied to either the difference between the seismograms or the amplitude spectra of the seismograms. I use the term trace here interchangeably to identify either the tapered and filtered time series at one receiver component or the amplitude spectrum of such a time series. This section only covers the application of an $l^{1}$-norm. Each trace $j$ of the total of $N$ traces in a dataset contributes a trace misfit $m_{j}$ to the global misfit $M$ :

$$
\begin{equation*}
M=\frac{\sum_{j=1}^{N} w_{j} m_{j}}{\sum_{j=1}^{N} w_{j} n_{j}} \tag{2.12}
\end{equation*}
$$

[^4]with
\[

$$
\begin{equation*}
m_{j}=\Delta_{j} \sum_{i=1}^{S_{j}}\left|s_{i j}-r_{i j}\right| \quad \text { and } \quad n_{j}=\Delta_{j} \sum_{i=1}^{S_{j}}\left|r_{i j}\right| \tag{2.13}
\end{equation*}
$$

\]

where $n_{j}$ is called trace misfit normalization factor, $s_{i j}$ is synthetic sample $i$ of trace $j, r_{i j}$ is the corresponding observed sample, $\Delta_{j}$ is the sampling interval of trace $j$, and $w_{j}$ is a weight factor. The global misfit is normalized, so that a misfit value of $M=1$ corresponds to the case, where the synthetic traces are set to zero, i.e. $s_{i j}=0$.

The terms station misfit and station misfit normalization factor will be used in the following work to identify the misfit value resulting when combining the trace misfits for multiple components of a station.

### 2.6.2 Weighting

When the weighting factors $w_{j}$ in (2.12) are chosen to be equal, trace misfits from receivers near to the source may strongly dominate the global misfit. This is especially true, if the time windows which are used do not grow with distance. To avoid this, the weights should be designed, at least, so that they compensate for amplitude decay with distance, (but also taking into account that the time windows might grow with distance). Additionally, it might be useful to compensate for the overall radiation pattern, when one is interested not in absolute amplitudes, but in the shape of the waveform. In this extreme case, the weights are chosen as

$$
\begin{equation*}
w_{j}=1 / n_{j} . \tag{2.14}
\end{equation*}
$$

These weights have the advantage that very noisy traces down-weight themselves automatically. A difficulty with this approach is, that traces with almost no signal are over-weighted.

### 2.6.3 Adaptive station weighting

A very appealing possibility would be, to set the weights equal to the inverse of the mean expected trace misfits. These can be calculated by forward modeling synthetic seismograms for a large ensemble of sources representing a given a priori expectation on the source parameters. A simplified but practical application of this is to create weights that compensate for the amplitude decay with distance, by forward modeling point sources of various strike, dip, rake combinations. This can be done by calculating unweighted misfits of synthetic seismograms compared against observed seismograms set to zero for a sufficiently large ensemble of sources, while applying the same tapering and filtering setup
(see 2.6.4) as for the following inversion. We use the inverse of the mean values of the trace misfits obtained in this manner as trace weights,

$$
\begin{equation*}
w_{j}^{-1}=\left\langle\Delta_{j} \sum_{i=1}^{S_{j}}\right| s_{i j}| \rangle_{\text {sources }} . \tag{2.15}
\end{equation*}
$$

We refer to this kind of weighting as adaptive station weighting, because it does not only automatically consider amplitude decay with distance, but also the different amplitudes of different phases, the effect of time window tapering which might not be of the same length at different distances, and the definition of misfit which is used.

### 2.6.4 Tapering and filtering

Before comparing observed and synthetic seismograms by calculating a misfit with (2.12), we filter and taper the traces to be compared. This is done to extract specific phase arrivals from the seismogram and to restrict the comparison to the frequency band which matters in a particular application. The tapering and filtering is applied to both sets of seismograms in exactly the same way, so that any artifacts of the processing appear symmetrically in both, the synthetic and the observed trace to be compared. Tapering is applied before filtering, so that no energy from nearby stronger phases leaks into the window to be analyzed due to filter ringing effects. We use simple cosine flanked windows, with

$$
w(t)=\left\{\begin{array}{cl}
0 & t<t_{1}  \tag{2.16}\\
\frac{1}{2}-\frac{1}{2} \cos \left(\pi \frac{t-t_{1}}{t_{2}-t 1}\right) & t_{1}<=t<t_{2} \\
1 & t_{2}<=t<t_{3} \\
\frac{1}{2}+\frac{1}{2} \cos \left(\pi \frac{t-t_{3}}{t_{4}-t_{3}}\right) & t_{3}<=t<t_{4} \\
0 & t_{4}<=t
\end{array}\right.
$$

to extract the region of interest from the signal (see figure 2.4). The timings $t_{1}$ to $t_{4}$ are usually specified as offsets to theoretical arrival times of the phases to be compared.

The same kind of taper is also used for filtering the spectrum of the seismogram. Here, four frequencies $f_{1}$ to $f_{4}$ must be specified to define the frequency band to be analyzed. This is an acausal filter which does not produce any phase shifts. Its frequency response is unity between $f_{2}$ and $f_{3}$.


Figure 2.4: Simple cosine flanked window function, as defined in (2.16), used to extract specific phases from the seismograms.

### 2.7 Solution and error estimation

### 2.7.1 Searching misfit space

With the tools developed in the previous sections we can set up a misfit function depending on 13 parameters. To find a solution to our problem, it is necessary to find the particular choice of the 13 parameters which minimizes this misfit. Many different strategies are commonly used to solve this kind of problem [see Press et al., 1992, for a collection of algorithms]. Typically there is a trade-off between speed of convergence, e.g. the number of function evaluations required to find a minimum, and its robustness in the presence of local minima.

To choose an appropriate minimization method for a given problem, one typically has to consider further points: e.g. the number of free parameters, the computational or memory cost of the minimization method itself, its ability to deal with constraints, and the smoothness of the misfit function.

A grid search is a brute force approach to solve this kind of minimization problem. The misfit function is simply evaluated on a grid of (all) possible parameter combinations. Its main advantages are that, if the grid spacing is chosen properly, it does not only always find the global minimum, it can also map local minima, alternative solutions, and map ambiguities. Also it is easy to constrain the grid search to a region of reasonable parameter choices. Its disadvantage is, that it usually requires more function evaluations than any other minimization method. This is in most cases not feasible so this outweighs all its advantages. In contrast, gradient methods are good at quickly finding a minimum, but may easily get trapped in local minima.

In our case where we have a 13-dimensional parameter-space it is currently not possible to employ a full grid search on all parameters. We can however separate the search space into (partly overlapping) subspaces, which can be searched for a minimum one by one using combinations of grid and gradient searches. How to set up the hierarchy of searches, how to set parameter ranges and grid spac-
ings for the grid searches, and how to combine the minima found in different subspaces is application dependent. A detailed example is given in chapter 3. Finally, to investigate the stability of a solution, we apply a grid search to the neighborhood of a previously found minimum (see also section 2.7.2).
The separation into subspaces is possible here, because the misfit function can be designed in a way, which makes it independent of some of the parameters. For example, if we base the misfit calculation on amplitude spectra, it becomes insensitive to the exact location and time of the centroid, or by low-pass filtering the seismograms, it becomes less dependent of the parameters dealing with the extension of the fault. We exploit this in a way that, first, point source parameters are determined from the low frequency content of the seismograms. Then, the point source parameters are fixed and the remaining parameters are determined by including higher frequencies.

### 2.7.2 Bootstrap test and confidence intervals

In order to check the stability of the results with respect to data selection, a bootstrap test is applied. The bootstrap test works by repeatedly solving the minimization problem, each time using a different subset of the available data. In particular, if we have a measurement with N samples, (one station is one sample in our case), we randomly draw N samples from these, while allowing that the individual samples may be chosen more than once. Doing so leads to a distribution of results, which is linked to the probability density function of the solution. This is the so-called "Quick-and-Dirty Monte Carlo" method [as described for example in Press et al., 1992, page 691], which is especially cheap when applied to grid search results, because station selection can be seen as a special case of weighting (by choosing the weights $w_{j}$ in equation (2.12) to be zero when the station should be excluded). So the intermediate results for the trace misfits $m_{j}$ and trace misfit normalization factors $n_{j}$ of a complete grid search can be reused without repeating the forward modeling. I typically use 1000 bootstrap iterations to get result probabilities for multidimensional grid searches.

The distribution of the bootstrap results can be reported graphically as 1D histograms, one for each parameter searched in a multidimensional grid search (see figure 3.9 for an example), or combined, for example as 2D histograms revealing the joint probability of finding specific combinations of two of the parameters (see figure 3.10 for an example). From the 1D parameter result probabilities, we report the lower and upper margin of wherein $68 \%$ of the bootstrap results fall, as error margins.
The joint probabilities from the bootstrap test allow to identify ambiguities in the retrieved parameters. If, for example, two parameters cannot be resolved inde-
pendently, but their ratio is constant, the one-dimensional bootstrap result probabilities would give broad distributions for each of the parameters and the ratio between the two parameters could be identified as a line of high probabilities in the joint probability density (see figures 3.10 and 3.20 for examples).

### 2.8 Automatic processing

Most difficulties with the automatic processing of earthquake data arise from data quality, which may vary strongly from one station to the other. An automatic procedure must be aware of gaps in the data, missing meta-information (station responses), incorrect meta-information, inhomogeneous station distributions, artifacts in the data, noise, and more. Some of these problems, like gaps in the data, can be detected during preprocessing of the data. Some can be solved during processing by using a norm which is less sensitive to outliers or by removal of traces which badly fit. However, some problems are not easily detected by such means. Furthermore, todays seismic networks provide so much data, that data selection becomes more and more important. Though for some applications it is useful to include as much data as possible, for others it is not feasible to use all available data, by means of the computational cost. This is especially the case, if a solution is required quickly. In these cases, the process of data selection becomes important.
This section describes two tools used in our automatic processing to firstly evaluate and quantify a priori station quality and secondly to select stations in networks with inhomogeneous distributions, based on azimuthal and distantial coverage, taking into account expected station quality.

### 2.8.1 A station quality evaluation procedure

For some applications, it is necessary to quantify a kind of a priori quality of all stations in a seismographic network, for example to pre-select stations which are known to be useful for the application in question. This can be done by an investigation on how well the observed and synthetic seismograms at each station agreed over a number of previous events. There are of course numerous different ways to do this. The following recipe has been applied successfully to the GEOFON and GEOFON partner networks for our purposes. The method has been set up to operate fully automatic.

All broadband data which are available through the network are acquired for all $n$ events with magnitudes in the range $\mathrm{M}_{\mathrm{W}} 6.3$ to $\mathrm{M}_{\mathrm{W}} 7.6$ for a given time period. Data are preprocessed with a similar procedure as described in section 3.2. Then,
synthetic seismograms for these events are computed. An $l^{1}$-norm applied to the difference between the seismograms filtered in the frequency band between 0.005 and 0.02 Hz , is used to compare synthetic seismograms and observations. Adaptive station weighting (section 2.6.3) is applied and tapering is set up so that the complete seismograms enter into the analysis. Centroid time and rise-time are adjusted through a grid search, while all other point source parameters are set according to a double-couple point-source based on the Global CMT (gCMT) catalog [Ekström and Nettles, 1982] solutions for the events.

For the final analysis, trace misfits as defined in section 2.6.1 are calculated for variations to the best fitting centroid time. The misfit at station j component k for event i and a time shift of $l \Delta T$ is $m_{i j k l}=m_{i j k}\left(T_{i}+l \Delta T\right)$. For a station component, with a good match between synthetics and observations, one would expect to find the the minimum in misfit exactly at the true centroid time of the event (at $l=0$ ). The obtained misfit values are then averaged over all events

$$
\bar{m}_{j k l}=\frac{1}{n} \sum_{i=1}^{n} m_{i j k l}
$$

because we are interested in the average performance of each component at each station. From the averaged misfit curves several common problems can be identified. For example, if the component has a flipped sign, it will show a clear maximum at $l=0$. If the clock of the data-logger at a station is going wrong, all components of the affected station have the minimum time-shifted by a common offset. If a station is always noisy, no clear minimum can be identified. Errors in the meta-data for the instrument correction sometimes result in a distorted shape of the misfit curve. If the gain at a station is incorrectly handled, the misfit has anomalously high values.

For practical purposes, a simple automatic criterion is used to define a stream badness based on these misfit curves. If the temporal offset of the minimum of the trace misfits is less than a certain threshold, the misfit value at the minimum is reported as what I call stream badness, otherwise it is set to a large value. These badness values can be used in other applications for example, to define thresholds for data selection.

The stream badnesses obtained using this procedure have to be taken with caution, because they are biased by the choice of the events which are used for their determination. The number of the events entering into the procedure as well as the distribution of their locations may introduce systematic changes in the stream badnesses. Nevertheless, despite its weaknesses, it is a very useful concept to rate station quality in large seismographic networks.

### 2.8.2 Station weeding

This section describes a simple data selection algorithm, used to reduce the number of stations which are included into the analysis. The goal is to reduce the number of stations first, where the station density is highest, so that only redundant data is removed. This should ideally lead to a more uniform distribution of stations. Another priority is, that stations which are known to deliver data of poor quality should preferably be removed. Again, there is no perfect solution to this problem, so the following pragmatic procedure is used.
Let's assume, we have $N$ stations in total, from which we would like to choose a subset of $M$ stations. A special norm is used to measure the distance between two stations. The distance between station $i$ and $j$ is here defined as

$$
\begin{equation*}
D_{i j}=\sqrt{(\Delta \alpha)^{2}+(\Delta \beta)^{2}} \tag{2.17}
\end{equation*}
$$

based on the difference between the azimuths $\Delta \alpha$ and difference between the angular distances $\Delta \beta$, as measured from the earthquakes preliminary origin.
Three tuning parameters are introduced in the following steps of the algorithm: $L$ is the number of nearest neighbors to consider, $X$ is the fraction of stations to be subject to removal in each iteration of the procedure, and $a$ is a clearance distance around removed stations preventing that two close-by stations are simultaneously removed in a single iteration. (Using $L=3, X=4$, and $a=3$ is a good starting point.)

By the following steps, a choice of stations is carried out:

1. For each of the $N$ (remaining) stations calculate the mean distance $\bar{D}_{i}$ to its $L$ next neighbors.
2. Take the $N / X$ stations with the lowest $\bar{D}_{i}$, and sort these according to their station badness, if these are available.
3. Iterate over all stations of the subset, start with worst. Mark station for deletion, if no station closer than $a \bar{D}_{i}$ to it has already been marked in this round. End the procedure, if enough stations have been marked for deletion.
4. Continue at 1 .

This procedure does a pretty fair job in solving our data selection problem. The station density is reduced where it is highest, and it preferably removes bad stations. Through the use of the special distance definition (2.17), the algorithm tends to produce a more uniform station distribution in azimuth and distance.

## Chapter 3

## Application

In this chapter I show how the method outlined in the preceding chapter can be used in an automated way to reliably derive earthquake parameters and importantly, uncertainties of these. As an example of how seismic data is collected, preprocessed and analyzed, a detailed description of the analysis of the L'Aquila earthquake of 2009 is given. I derive parameters in the framework of the eikonal source model. Special attention is given to the estimation of uncertainties and systematic errors. The ability that the eikonal source model can represent the seismic source at different levels of detail is consequently exploited by splitting the inversion procedure into a series of sub-steps. With each step, the model is refined. Point source as well as kinematic parameters are derived consecutively within a consistent framework.

Inputting faulty data in the inversion would introduce systematic errors in the obtained model of the earthquake. We show that by introducing some simple quality control schemes and by basing our misfit calculation on an $l^{1}$-norm, reliable point and kinematic source parameters can be derived without human supervision and interaction.
To estimate values for the parameters of an earthquake model, our goal is to find the choice of source parameters which minimizes the global misfit between the modeling results and the observations. The earthquake model we use is the eikonal source model as given in section 2.2. The kind of misfit function we use, has been described in section 2.6.1.

Our misfit function in general does not contain only a single minimum, so action has to be taken that one does not mistake a local minimum for the global one. It is not feasible to evaluate the misfit function for the complete parameter space of reasonable source models when using the eikonal source model because forward modeling for too many source models would have to be done. Fortunately though, the parameter-space can be decomposed into several only partially en-
tangled subspaces. In each of them, the respective global minimum is obtained by gradient or grid search methods and the minima are combined to a final solution.

This decomposition opens the possibility to set up a kind of multi-step inversion: first all point source parameters except for the event duration are determined from the low frequency content of the seismograms (periods longer than the event's duration). Then, the extended source parameters are examined at higher frequencies, while keeping the point source parameters fixed. It is crucial to first determine a best fitting point source model before inverting for extended rupture properties, because wrong assumptions about the point source parameters may introduce systematic errors into the further analysis.

This chapter describes the overall procedure which can be used to routinely determine the extended fault properties for global earthquakes in the magnitude range $M_{W} 6-7.5$. The consecutive steps are illustrated by application to the $M_{W} 6.3$ L'Aquila earthquake of 2009.

This event has been chosen because there are detailed studies based on reliable near field data [Cirella et al., 2009] and there exists a rich historical record of the seismicity in the source region.
Compared to other existing methods to determine point source approximations of earthquake rupture, the use of an $l^{1}$-norm in our method is non-standard, as well as that there is a direct search for strike, dip, and slip-rake, so that only plain double-couple sources are considered. As a consequence, this procedure can only be used to analyze plain shear cracks.
This chapter is structured as follows: In section 3.1 published results for the L'Aquila Earthquake are summarized. In 3.2 I describe the data selection and preparation. Section 3.3 is dedicated to the design of the misfit function for this application. Finally, in section 3.4 point source and in section 3.5 kinematic source parameters are derived and investigated. In section 3.6 I atempt to estimate the rupture velocity of the event.

### 3.1 The $M_{W}$ 6.3 L'Aquila earthquake of 2009

At 01:32 UTC on April 6, 2009, an $\mathrm{M}_{\mathrm{W}} 6.3$ Earthquake struck central Italy with its epicenter near to the town of L'Aquila in the Abruzzo region. INGV [INGV] located the hypocenter of the earthquake to $42.3476^{\circ} \mathrm{N}, 13.3800^{\circ} \mathrm{E}$. It caused 300 casualties and heavy damage to L'Aquila town and the surrounding villages.

Moment tensor solutions from different agencies are summarized in table 3.1. The event showed normal faulting behaviour with a strike of $120^{\circ}-144^{\circ}$ and a

|  | $\mathrm{M}_{W}$ | Depth [km] | Strike | Dip | Rake |
| :--- | :---: | :---: | :---: | :---: | :---: |
| gCMT | 6.3 | 12 | $127^{\circ}$ | $50^{\circ}$ | $-109^{\circ}$ |
| USGS CMT | 6.3 | 10 | $122^{\circ}$ | $53^{\circ}$ | $-112^{\circ}$ |
| USGS BW MT | 6.2 | 2 | $113^{\circ}$ | $60^{\circ}$ | $-118^{\circ}$ |
| InSAR-u | 6.2 | 7 | $144^{\circ}$ | $54^{\circ}$ | $-105^{\circ}$ |
| This study | 6.2 | 5 | $138^{\circ}$ | $48^{\circ}$ | $-95^{\circ}$ |

Table 3.1: Best double-couple solutions as reported by different sources [Ekström and Nettles, 1982, U.S. Geological Survey, National Earthquake Information Center, 2009, Walters et al., 2009].
dip of about $50^{\circ}$ (InSAR data and geomorphology indicate a strike in the range $140^{\circ}-145^{\circ}$ [Walters et al., 2009]). The south-west dipping Paganica fault, has been identified as the active fault of the main shock [Walters et al., 2009, and references therein].

A detailed study of the L'Aquila earthquake sequence is given by Chiarabba et al. [2009]: The main shock was preceded by two small foreshocks ( $\mathrm{M}_{\mathrm{L}} 4.1$ and $\mathrm{M}_{\mathrm{L}}$ 3.9), close to the location of the main shock. The main shock occurred at 01:32 UTC on April 6, 2009 and had a Magnitude of $\mathrm{M}_{\mathrm{W}} 6.3$ (gCMT, [Ekström and Nettles, 1982]). The two strongest aftershocks occurred about 15 km more to the north $\left(\mathrm{M}_{\mathrm{W}} 5.6\right)$ and about 15 km to the south-east $\left(\mathrm{M}_{\mathrm{W}} 5.4\right)$ of the area of the main event, each followed by an own series of smaller aftershocks, so that at the end of the sequence three separate clusters of aftershocks accumulated (figure 3.1), see also Chiarabba et al. [2009]). The aftershocks following directly after the main event indicate a rupture length between 15 and 18 km .

The rupture length estimated from aftershocks is confirmed by Walters et al. [2009] who determined from InSAR data a rupture length of 12 km or 19 km depending on whether they assume uniform or non-uniform slip.

A joint inversion of strong motion and GPS data has been done by Cirella et al. [2009]. They derive a heterogeneous slip distribution with a shallow slip patch located up-dip of the hypocenter and a large deeper patch located southeastward, together accounting for a rupture extension of about $18 \times 12 \mathrm{~km}^{2}$. Rupture is characterized slightly bilateral, propagating more to the southeast ( $\approx 13 \mathrm{~km}$ ) than to the northwest $(\approx 5 \mathrm{~km})$. For rupture velocity they derive values of $2.2-2.8 \mathrm{~km} / \mathrm{s}$ up-dip and about $2 \mathrm{~km} / \mathrm{s}$ along strike.

Italy is frequently hit by normal faulting events with sizes comparable to the L'Aquila event. Large earthquakes occur mainly in a narrow belt in the central Appenines. This region is characterized by a NE-SW oriented extensional stress regime. The extension rate is about $3 \mathrm{~mm} / \mathrm{yr}$ [D'Agostino et al., 2008]. The extensional stresses are caused by retreat of the subduction slab of the adriatic


Figure 3.1: Figure from [Chiarabba et al., 2009] showing aftershocks of the L'Aquila 2009 earthquake series. Original caption: "Map of the 3200 relocated events ... showing that aftershocks originate around three main patches that ruptured during the MW 6.3". The large aftershocks with $\mathrm{M}_{\mathrm{W}} 5.6$ and $\mathrm{M}_{\mathrm{W}} 5.4$ are each surrounded by an own cluster of aftershocks, in areas which were not covered by aftershocks immediatly after the main shock.
microplate [Collettini et al., 2006]. In response to the extensional stresses northeastward dipping low angle normal faults (LANFs) develop. The LANFs are weak faults not causing earthquakes. Deformation and the extensional stresses in the crust above the LANFs is released in a system of sub-parallel southwestward dipping normal faults. The larger normal-faulting earthquakes, typical for this region, occur on these faults.

### 3.2 Data selection and preprocessing

For this example, I restrict the procedure to use broad-band seismograms from the Global Seismographic Network (GSN), [Berger et al., 2009] because the stations from this network have a good global coverage and data is provided free of charge in real-time. Using this set of stations enables us to obtain results of comparable quality anywhere on the globe. This data is available event-based via the WILBER II [IRIS, Incorporated Research Institutions for Seismology, 2009] web interface. Using this web interface, is one possible option, of how seismographic data can be acquired by an automatized procedure. For a given event, we include


Figure 3.2: Broadband data from 81 stations (red triangles) of the Global Seismographic Network GSN within an epicentral distance range of up to $90^{\circ}$ is used for my analysis of the $\mathrm{M}_{W}$ 6.3 L'Aquila earthquake of 2009.
seismograms from stations with a distance of up to $90^{\circ}$ distance. Broadband data are requested for time windows spanning from 5 minutes before until 40 minutes after the expected P phase arrival for the event. The data is downloaded as a SEED volume, which is convenient, because all required meta-information, like station locations and instrument responses are included. An example station distribution is given in figure 3.2 for the L'Aquila 2009 earthquake.

To be used by the procedure, some data preparation has to be done. Our preprocessing involves removal of incomplete traces, restitution to ground displacement, downsampling, and rotation of the horizontal components of ground displacement into radial and transversal directions with respect to the earthquake origin. Ground displacement is calculated by deconvolving the instrument re-
sponses from the observed seismograms in the frequency-domain using

$$
\begin{equation*}
U=W T^{-1} X \tag{3.1}
\end{equation*}
$$

where $X$ is the complex spectrum of the recorded seismogram, $T$ is the transfer function of the recording instrument, and $W$ is a taper used to restrict the restitution to a given frequency band ${ }^{1}$. To carry this out numerically, the spectrum of the seismogram is calculated using an FFT after tapering the raw data trace. The flanks of the taper applied to the raw data trace are set so that they are longer than the period corresponding to the lower corner frequency of the taper used in the deconvolution. This procedure has the practical consequence, that data should be requested for a time window which is, on both sides, roughly $1 / f_{\text {min }}$ times longer than the time window needed afterwards for the inversion.
The downsampling in the preprocessing is required because the sampling rate of the seismograms has to match the sampling rate of the pre-calculated Green's functions.

The principal intention of the preprocessing is to gain a good approximation of the displacement seismograms in a frequency band which is larger than what is later used in the inversion (where additional filtering is done).

### 3.3 Misfit function design

### 3.3.1 Tapering and filtering

Tapering and filtering can be seen as a main part of the misfit function design. The kind of tapers and filters used are described in section 2.6.4.

Only P and SH phases are used in this procedure, because these are best matched by our modeling. To extract these phases, we use time-window tapers based on the expected arrival times of the phases. The same tapers are used throughout the whole inversion procedure for global earthquakes. For P phases, only the vertical component of displacement is used, because it has the highest signal to noise ratio. Though it is possible to use $S$ phases on both horizontal or all three components of the seismograms, we use only SH waves on the transversal component in the automatic procedure, because their modeling suffers least from the bad earth model approximation. The tapers are set up so that the main pulse of the phase is fully covered. For the automatic procedure which focuses on events with magnitudes of 6 to 7.5 and which have rupture durations of up to 20 s , a window length of $50 \mathrm{~s}+20 \mathrm{~s}$ flanks has shown to be useful. For retrieval of the

[^5]point source parameters, a low pass filter is used (frequency taper falling as a cosine flank from 0.05 to 0.1 Hz ). For retrieval of the extended source parameters, higher frequencies are included.

### 3.3.2 Station weighting

The misfit function is also affected by station weighting. For the automatic procedure, adaptive station weighting as described in section 2.6.3 is used to counteract amplitude decay with distance and to equalize the impact of the different phases included.

### 3.3.3 Norm

Due to the imperfect forward modeling, noise in the data, and the simplistic source model, no good fit between synthetics and observations can be expected in general. Thus, I prefer to use an $l^{1}$-norm to measure the difference between the observed and the synthetic seismograms or spectra, which makes the misfit function less sensitive to outliers. The misfit is evaluated on the difference of the time traces of the seismograms or on the difference of the amplitude spectra; what is used, is described in the inversion steps below.

### 3.4 Point source inversion procedure

### 3.4.1 Green's function

In this application we use a Green's function database containing body-wave phases, modeled with the GEMINI program by Friederich and Dalkolmo [1995] for the IASP91 earth model approximation. The Green's functions are sampled at 2 Hz and the modeling has been done for frequencies of up to 0.3 Hz . The lateral spacing of the Green's function traces is 1 km in distance and depth. The Green's functions have been evaluated for an epicentral distance range of up to 11000 km and for source depths of up to 100 km .

### 3.4.2 Rough moment

For the following inversion steps, a very rough estimate on the moment of the earthquake is needed. If not available from the location procedure, I estimate it using a simple 1D grid search: For an "arbitrary" moment tensor point source


Figure 3.3: Rough moment estimation for the L'Aquila earthquake. The misfit as evaluated for different values of $\mathrm{M}_{\mathrm{W}}$ (magnitude) is shown. This has been done using amplitude spectrum $l^{1}$-norm, adaptive station weighting and after removal of traces which give extremely large misfits. The double-couple orientation has been fixed at (strike $=0^{\circ}, \operatorname{dip}=45^{\circ}$, slip-rake $=0^{\circ}$ ).
(strike $=0^{\circ}$, dip $=45^{\circ}$, slip-rake $=0^{\circ}$ works fine), synthetic seismograms are calculated for a range of moments. Amplitude spcectra of synthetic seismograms and observations are compared using the $l^{1}$-norm misfit setup described above. The moment, where the misfit is minimal should in general (with more than a few stations) already be on the order of the true moment of the event. For the L'Aquila earthquake, we find a rough moment corresponding to a magnitude of 5.9. The misfit curve leading to this result is shown in figure 3.3.

The rough moment estimated with this procedure, typically underestimates the moment of the event, so that the corresponding magnitudes differ by about 0.25 for earthquakes with magnitudes 5.5 to 7 (see figure 3.4 for statistics on this issue).

### 3.4.3 Orientation of the double-couple from frequency domain $l^{1}$-norm gradient searches

Using the misfit setup described above, a gradient search for moment, depth and orientation (strike,dip,rake) of the point source is done. Because it is likely that this gradient search gets stuck in local minima, the search is repeated for a number of starting models, differing in depth and the angles of orientation. An ambiguity remains in the solution for the point source since the polarity of the double-couple cannot be determined using amplitude spectra alone (the second solution corresponds to the case, where all seismograms have opposite sign.).


Figure 3.4: Histogram of rough versus final $\mathrm{M}_{\mathrm{W}}$ (magnitude) estimates for 170 earthquakes from 2008-2010 between $\mathrm{M}_{\mathrm{W}} 5.5$ and $\mathrm{M}_{\mathrm{W}} 7.5$. The size of the circles is proportional to the number of earthquakes in each bin.

### 3.4.4 Centroid time and double-couple polarity

Keeping the double-couple orientation fixed at the result of the previous step, a 2D grid search for moment and centroid time is repeated for both possible slip directions. This reveals the polarity of the double-couple, as well as the centroid time. In this step, the misfit is defined as an $l^{1}$-norm on the difference of the timedomain traces, while keeping tapering and filtering setup as in the previous step. If the previous step failed to reveal the correct orientation of the double-couple, this leads to a highly unstable result for the centroid time. If this problem arises, it can be identified by a large error estimate for the centroid time. The bootstrap tests histogram for centroid time typically exhibits a multi-modal probability if a wrong double-couple orientation or source depth is assumed (see figure 3.5).

### 3.4.5 Refinement of all point source parameters using time domain misfit

With the previous steps it was possible to get first estimates for moment, time, depth, strike, dip, and rake. To make the set of point source parameters complete, it is now necessary to refine, additionally, the centroid location (for which we have used a rough estimate from prior knowledge up to this point). We use a gradient search for all point source parameters using the time domain $l^{1}$-norm from the previous step, assuming that we are already close to the global mini-

(a) Results obtained when the correct orientation of the double-couple is used. In the upper plot the evaluated misfit values for various combinations of moment and the two possible slip directions are marked as blue points. The probability of finding the result at a specific time as estimated by the bootstrap procedure is given as a histogram in the lower figure.

(b) Results obtained when assuming an (arbitrarily chosen) incorrect orientation of the double-couple (strike $=190^{\circ}$, $\operatorname{dip}=20^{\circ}$, sliprake $=40^{\circ}$ ). Several minima develop and the bootstrap histogram becomes multi-modal, indicating that the result is unstable (i.e. what minimum is chosen depends on the station selection).

Figure 3.5: Grid search results for the estimation of centroid time by a grid search on the source parameters moment and centroid time, repeated for both possible slip directions.
mum in this multi-dimensional minimization problem. This gradient search is repeated for different starting depths, because it is not unlikely, that the prior estimate for the source depth was biased.

Success depends on the previous step to derive the centroid time because this gradient search likely descends into a local minimum if a wrong centroid time is used as a starting value.
Fitted and observed seismograms for the L'Aquila earthquake are given in figures 3.6 and 3.7.


Figure 3.6: Observed (dashed black) and synthetic seismograms (blue) for best fitting double couple solution for the L'Aquila 2009 earthquake. The traces have been filtered, tapered, and scaled to a common maximum after multiplication with the station weights as described in section 3.3.1. The azimuth to the station is indicated by a straight line in the small maps to the right and station distance is indicated by a circle. Continued in figure 3.7.


Figure 3.7: Continued from figure 3.6: Observed and synthetic seismograms for best fitting double couple solution for the L'Aquila 2009 earthquake.


Figure 3.8: The event's duration is estimated by running a grid search for the point source rise-time.

### 3.4.6 Estimation of the event duration

In the previous steps, point-source parameters were estimated from the lower frequency content of the seismograms (i.e. at periods longer than the event duration, which was assumed from scaling laws, so far). The next step now aims to directly measure the event duration. The event duration is of particular interest in our application, because it can give us as a crude hint on the size of the rupturing surface (in the next steps of the procedure, when we try to estimate the extended source parameters, we adjust the filtering and the parameters of the grid search with respect to the value we find for the event duration).

The double-couple orientation is fixed and a grid search for point source risetime, moment, and depth is performed. Tapering and misfit setup is kept from the previous steps (amplitude spectrum $l^{1}$-norm), but the filter is changed so that higher frequencies are included (up to 0.3 Hz ). The results from this grid search for the L'Aquila Earthquake give 8 s (see figure 3.8).

Due to the directivity of the event, each station observes an apparent rise-time depending on azimuth and take-off angle of the observed phase, thus allowing the results of this grid search to be used to analyze directivity by investigating the individual station misfits dependence on the rise-time. In [Cesca et al., 2010, submitted] we show that in some cases, the apparent durations extracted with this method, can be successfully interpreted in terms of simple line source models, for which approximate analytical solutions for the apparent durations can be calculated. This is an interesting alternative to the determination of the extension as described in the next steps, because no modeling for the extended source has to be done. With this method, directivity of the source can be estimated very quickly, once a point source model solution is available.

| Parameter | Unit | Value | $68 \%$ confidence interval |  |
| :--- | :--- | :--- | :---: | :---: |
| Time | $(\mathrm{s})$ | 5.7 | 5.2 | 6.2 |
| North-Shift | $(\mathrm{km})$ | -4.5 | -9.5 | 0.5 |
| East-Shift | $(\mathrm{km})$ | 12. | 7. | 17. |
| Latitude | $\left({ }^{\circ}\right)$ | 42.29 | 42.24 | 42.33 |
| Longitude | $\left({ }^{\circ}\right)$ | 13.5 | 13.4 | 13.5 |
| Depth | $(\mathrm{km})$ | 5. | 4.5 | 5.5 |
| $\mathrm{M}_{\mathrm{W}}$ |  | 6.2 | 6.2 | 6.2 |
| Moment | $(\mathrm{Nm})$ | $2.4 \mathrm{e}+18$ | $2.3 \mathrm{e}+18$ | $2.5 \mathrm{e}+18$ |
| Strike | $\left({ }^{\circ}\right)$ | 138. | 130. | 140. |
| Dip | $\left({ }^{\circ}\right)$ | 48. | 46. | 50. |
| Slip-Rake | $\left({ }^{\circ}\right)$ | -95. | -102. | -92. |
| Rise-Time | $(\mathrm{s})$ | 8. | 7.5 | 9.5 |

Table 3.2: Results and errors for point source parameters estimated for the $M_{W} 6.3$ L'Aquila earthquake. Only the angles for the true fault plane are given here.

### 3.4.7 Stability analysis of the point source inversion

The stability of the point source solution is analyzed by investigating the misfit function in a volume in parameter-space around the final solution found in the previous step. Three individual grid searches are run for the standard procedure: a 2D grid search on source depth and moment, a 4D grid search for orientation of the double-couple and moment and a 3D grid search for centroid location and time. For each of the grid searches the bootstrap test described in section 2.7.2 is applied to get an approximation of the probability density function for the result. Using this method, a full grid search on all parameters of the point source model would be required to gain a probability density function revealing all ambiguities between the source parameters, but unfortunately this is not feasible at the moment.

Projections into one dimension for the results are given in figure 3.9, 3.11, and 3.13. 2D-projections are given in figures 3.10, 3.12, and 3.14. We calculate the $68 \%$ confidence interval from the per-parameter histograms of the bootstrap test and report these as error estimates in our result summaries. The results are listed in table 3.2.


Figure 3.9: Stability analysis for the estimation of the orientation of the doublecouple. A grid search on strike, dip, slip-rake and $\mathrm{M}_{\mathrm{W}}$ (magnitude) in a 4D volume around the solution has been done. The minimal misfits when fixing one of the parameters to given values are shown on the left hand side. The histograms on the right hand side give the probability of finding a certain parameter result, as estimated by the bootstrap procedure. See figure 3.10 for 2D projections of these results.


Figure 3.10: Stability analysis for the estimation of the orientation of the doublecouple. A grid search on strike, dip, slip-rake and $\mathrm{M}_{\mathrm{W}}$ (magnitude) in a 4D volume around the solution has been done. The minimal misfits found in the hypercube projected into two dimensions are shown as contour lines and color (each point in the plots is the minimal misfit found when varying the two other free parameters). The red star marks model with the best fit. The black circles represent the 2D-histogram of the bootstrap. An slight ambiguity between strike and slip-rake is revealed. If the strike indicated by InSAR and geomorphology of 140$145^{\circ}$ is assumed true, the slip-rake is required to be at about $-90^{\circ}$, according to this results.


Figure 3.11: Stability analysis for centroid location and time. Plot style and symbols are as in figure 3.9. Per-parameter misfit and histograms from 3D grid search on the parameters north-shift, east-shift, and time are shown.


Figure 3.12: Stability analysis for centroid location and time. Plot style and symbols are as in figure 3.10. 2D projections of the minimal misfits of a 3D grid search on the parameters north-shift, east-shift, and time are shown.


Figure 3.13: Checking the stability of the results with respect to depth and $\mathrm{M}_{\mathrm{W}}$ (magnitude). Plot style and symbols are as in figure 3.9.


Figure 3.14: Checking the stability of the results with respect to depth and $\mathrm{M}_{\mathrm{W}}$ (magnitude). Plot style and symbols are as in figure 3.10. A minimum depth of 5 km has been considered.

### 3.5 Inversion for kinematic source parameters

In the previous steps of the inversion (section 3.4), I have identified a point source model for the earthquake. This point source model consists of centroid time, location, and depth, moment magnitude, orientations of the two possible fault planes, slip direction, and a point source duration (point source rise-time). The wavelengths considered so far do not allow for more resolution. In the next step of the inversion we will now include some shorter wavelengths into the analysis. This enables us to get a first-order picture of the extension of the earthquake. In terms of our minimal kinematic source model (section 2.2), this means that we will estimate its border radius, position of the nucleation point with respect to the centroid location, and rupture velocity. The method described here is suited as an automatic method for earthquakes in the magnitude range $M_{W} 6-7.5$. This generic procedure is exemplified by the L'Aquila 2009 earthquake. For a detailed study of a specific earthquake however, it is possible to adapt the results with some manual tuning (i.e. by introducing some additional constraints, manual quality control, and data selection).
To reduce the number of free parameters in this inversion step, some constraints are applied: rupture is restricted to a maximum depth of two times centroid depth ${ }^{2}$, the rise-time $\tau$ is assumed to be short compared to the rupture duration $T_{e}$ (it is set to $\tau=1 \mathrm{~s}$ ), and rupture velocity is assumed to be proportional to shear wave velocity. As the rupture velocity is especially difficult to determine, it is initially fixed at $80 \%$ of the shear wave velocity. The shear wave velocities given by the global crustal model CRUST 2.0 by Laske, G. and Masters, G. and Reif, C. [2009] are used in this application.
When rupture velocity is held constant, three free parameters of the eikonal source model remain: its border radius and the two relative coordinates which locate the nucleation center. From the point source parameters, all but the scalar moment are held constant. I estimate values for these four parameters by applying a grid search. The grid search is customized so that a maximum radius of $T_{e} \times 4000 \mathrm{~m} / \mathrm{s}$ is checked, where $T_{e}$ is the event duration. In order to observe effects of the extension of the fault, the filter is set so that its frequencies $f_{3}$ and $f_{4}$ are $1.5 \times T_{e}^{-1}$ and $2 \times T_{e}^{-1}$, respectively. $f_{1}$ and $f_{2}$ are kept at the values used for the point source inversion. For the L'Aquila event with a duration of 8 s , this yields 0.19 Hz and 0.25 Hz , respectively.

The higher the frequencies which are included, the more detail of the rupture can in principal be resolved. On the other hand, if too high frequencies are included, problems due to the rough modeling and of noise in the data become

[^6]

Figure 3.15: Eikonal source model, as derived for the L'Aquila earthquake of 2009. Isochrones of the rupture fronts are contoured with black lines. The red star indicates the nucleation point of the rupture. The dips of fault and auxilliary plane are $48^{\circ}$ and $42^{\circ}$, respectively.
more prominent. Thats why I conservatively set the filter to the point where effects from the extension of the fault just happen to become observable, which is approximately at $1 / T_{e}$ (see also figure 4.23). The tapering from the previous steps is applied unchanged. The grid search is repeated for either of the two possible fault planes.

The misfit and bootstrap results from these grid searches are given in figures 3.17 and 3.18 and in table 3.3. The result for the rupture plane is visualized in figures 3.15 and 3.16. Differential station misfits (relative to the best point source solution) are shown in figure 3.19.

For both fault planes asymmetric bilateral rupture is revealed. According to the result, rupture propagated about $10-15 \mathrm{~km}$ to the northwest and $25-30 \mathrm{~km}$ the southeast. Total length of the rupture was about 40 km . The bootstrap results reveal a second possible solution which would correspond to asymmetric bilateral rupture with opposite proportions. The presence of this second minimum causes large uncertainties for the horizontal position of the nucleation center. For this event, both fault plane results give comparable minimal misfit values (fault plane: 0.5228 , auxiliary plane: 0.5224 ), making it impossible to determine which is the true fault plane.


Figure 3.16: Eikonal source model derived for the L'Aquila earthquake of 2009, projected on an epicentral map. The red star indicates the nucleation point of the rupture. Deeper parts of the fault plane are shaded dark in this type of plot.

Fault plane (strike $=138^{\circ}$ )




Auxiliary plane (strike $=325^{\circ}$ )



Figure 3.17: Histograms from bootstrap of grid search for extended source model. A grid search has been used to search the parameter border-radius and the coordinates of the nucleation point. The histograms shown here represent the probability of the finding specific values for the parameters within the space searched.

Fault plane (strike=138 ${ }^{\circ}$ )


Figure 3.18: Determination of the extended rupture model. The minimal misfits found in a 3D grid search on the parameter border-radius and the coordinates of the nucleation point are projected into two dimensions and visualized with contour lines and color (each point in the plots is the minimal misfit found when varying the third free parameter). The red star marks the model with the best fit. The black circles represent a 2D-histogram of the bootstrap results.


Figure 3.19: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depicts an improvement of fit with respect to the best fitting point source model, red colors indicate decline.

## Fault plane (strike=138 ${ }^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :--- | :--- | :---: | :---: |
| Border radius | $(\mathrm{km})$ | 19. | 18. | 21. |
| Nucleation along strike | $(\mathrm{km})$ | -9.6 | -11. | 11. |
| Nucleation down dip | $(\mathrm{km})$ | 3.2 | -1.6 | 4.8 |

Auxiliary plane (strike $=325^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :---: | :---: | :---: | :---: |
| Border radius | $(\mathrm{km})$ | 19. | 18. | 24. |
| Nucleation along strike | $(\mathrm{km})$ | 6.4 | -14. | 11. |
| Nucleation down dip | $(\mathrm{km})$ | 0. | -1.6 | 1.6 |

Table 3.3: Fault and auxiliary plane results and errors for kinematic source parameters estimated for the $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake under the assumption of a fixed rupture velocity of $80 \%$ of shear wave velocity.

### 3.6 Estimation of the rupture velocity

Up to this point, the rupture velocity has been fixed to reduce the computational cost and because inverting for this parameter likely causes the inversion to become unstable. The rupture velocity has been set to $80 \%$ of the S-wave velocity. I will now try to estimate this problematic parameter with an optional additional inversion step.

I use a grid search around the previously found solution for the extended rupture model, inverting for position of the nucleation point, border radius, and additionally, rupture velocity. The point source parameters are kept fixed.
Border radius and nucleation point ranges are set offset to the results from the previous inversion step (table 3.3). The rupture velocity is still assumed to be proportional to the shear wave velocity, but is allowed to vary in ratio from 0.4 to 1.4 of the latter. Again, a grid search is used to find the best fitting model. It is repeated for either of the two possible fault planes.
The results are given in table 3.4 and are visualized in figures 3.20. The fitted seismograms are given in figure 3.21 and 3.22 .

The joint parameter probabilities between rupture velocity and border radius, (figure 3.3) reveal an ambiguity between these two parameters. their ratio can be estimated as $R /\left(v_{r} / v_{s}\right) \approx 25 \mathrm{~km}$.

Allowing variations to the rupture velocity leads to a somewhat smaller rupture size and lower velocities, which would better fit aftershock locations. But also the errors to the estimates become large making this result questionable.

## Fault plane (strike=138 ${ }^{\circ}$ )

| Parameter | Unit | Value | $68 \%$ |  |
| :--- | :---: | :---: | :---: | :---: |
| confidence interval |  |  |  |  |
| Rel. rupture velocity |  | 0.7 | 0.45 | 0.75 |
| Border radius | $(\mathrm{km})$ | 15. | 12. | 18. |
| Nucleation along strike | $(\mathrm{km})$ | -8. | -9. | -5. |
| Nucleation down dip | $(\mathrm{km})$ | 1. | 0. | 2. |

## Auxiliary plane (strike $=325^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :---: | :---: | :---: | :---: |
| Rel. rupture velocity |  | 0.8 | 0.55 | 0.95 |
| Border radius | $(\mathrm{km})$ | 17. | 14. | 20. |
| Nucleation along strike | $(\mathrm{km})$ | 8. | 5. | 9. |
| Nucleation down dip | $(\mathrm{km})$ | 0. | -3. | 1. |

Table 3.4: Fault and auxiliary plane results and errors for kinematic source parameters estimated for the $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake, when rupture velocity is not fixed.

Fault plane (strike $=138^{\circ}$ )


Figure 3.20: Misfit and histogram plots from grid search to estimate the rupture velocity. A four-dimensional grid search has been used to estimate rupture velocity and to refine border radius and nucleation center of the event. Plot style and symbols are explained in the caption of figure 3.18.


Figure 3.21: Observed (dashed black), synthetic seismograms for best fitting point source (blue), and extended source (green) for the L'Aquila 2009 earthquake. The seismograms for the actual fault plane are shown. The azimuth to the station is indicated by a straight line in the small maps to the right and station distance is indicated by a circle. Continued in figure 3.22.


Figure 3.22: Continued from figure 3.21:

## Chapter 4

## Synthetic tests

### 4.1 Inversion without noise

To demonstrate the functional operability of our method, I present in this section an inversion of noise-free synthetic data. Synthetic data are calculated by forward modeling for a virtual earthquake, resembling the 2009 L'Aquila earthquake. Virtual station locations are the same as those used in the real application to the L'Aquila earthquake (section 3.1). The source parameters given in table 4.1 have been used to setup the eikonal source model for the virtual earthquake. The rupture is visualized in figure 4.15 (b).

The inversion is carried out exactly as in the application to real data (section 3.1). We can expect to get very low misfits after the point source inversion and zero misfit after the inversion for the extended model for the true fault plane if the above specified model can be retrieved correctly. The expected results show up as clear minima in in the misfit function as can be seen in figures 4.1-4.6. This result verifies the general ability of the tool to run correctly with respect to numerics and computation.
It is worthwhile to have a closer look at the 2D misfit cross sections from the stability analysis (done as in section 3.4.7) of the point source inversion given in figures 4.1, 4.2, and 4.3. With a few exceptions, discussed below, all 2D misfit surfaces have a single, clear, and symmetric minimum. From these we may conclude, that the whole setup is very well suited for an analysis of the point source parameters.

Slight ambiguities are revealed between strike and slip-rake and between northshift and time (by the diagonal elongated shape of the misfit contours in their respective plots in 4.2 and 4.3). The presence of these ambiguities in this synthetic test shows that they are inherent to the station configuration and data selection

## Source parameters of virtual earthquake for synthetic test

| Parameter | Unit | Input | Output |
| :--- | :--- | :--- | :--- |
| Time | $(\mathrm{s})$ | 0. | 0. |
| North-Shift | $(\mathrm{km})$ | 0. | -0.3 |
| East-Shift | $(\mathrm{km})$ | 0. | -1.5 |
| Depth | $(\mathrm{km})$ | 5.0 | 6.0 |
| Latitude | $\left({ }^{\circ}\right)$ | 42.33 | 42.32 |
| Longitude | $\left({ }^{\circ}\right)$ | 13.33 | 13.31 |
| Moment $^{\mathrm{M}_{\mathrm{W}}}$ | $(\mathrm{Nm})$ | $3.2 \mathrm{e}+18$ | $3.1 \mathrm{e}+18$ |
| Strike |  | 6.3 | 6.3 |
| Dip | $\left({ }^{\circ}\right)$ | 135. | 136. |
| Slip-Rake | $\left({ }^{\circ}\right)$ | 45. | 45. |
| Border radius | $\left({ }^{\circ}\right)$ | -90. | -88. |
| Nucleation along strike | $(\mathrm{km})$ | 16. | 14. |
| Nucleation down dip | $(\mathrm{km})$ | -8. | -8.4 |
| Rel. rupture velocity |  | -4. | 0. |
| Rise-Time | $(\mathrm{s})$ | 0.9 | 0.8 (fixed) |
|  |  | 1. | $2 .(f i x e d)$ |

Table 4.1: Source parameters of virtual earthquake for synthetic test (Input) and inversion results (Output). The reference model is chosen to resemble the 2009 L'Aquila earthquake (see tables 3.2 and 3.3). The results from the inversion with fixed rupture velocity are shown.
scheme used in this particular application. Both are also revealed in the real data example (see figure 3.10 and 3.12). The variation of misfit with depth shows a more complex behavior (figure 4.1). Discontinuities in misfit appear at depths of layer interfaces in the earth model which was used to generate the Green's functions. Additionally, an ambiguity between moment and depth exists.

The inversion for the extended source model with fixed rupture velocity (done as described in section 3.5) results in some deviations compared to the inserted source parameters (figure 4.4). This is because here, the rupture velocity is fixed at $80 \%$ of the shear wave velocity, whereas in the reference model it was arbitrarily set to $90 \%$. The result is that the size of the fault is underestimated by about $10 \%$, no perfect fit can be achieved, but the retrieved proportions of the earthquake match the reference model. The misfits of the best models obtained in this step are still 0.094 for the fault plane and 0.10 for the auxiliary plane. It would not be possible, from this result, to decide which is the true fault plane.

The improvement in fit, compared to a point source, is mainly attributed to the fit


Figure 4.1: Stability analysis of point source inversion of noise-free synthetic data for an extended source. The variation of misfit with respect to depth and moment (magnitude $\mathrm{M}_{\mathrm{W}}$ ) is shown as contour lines and color. The star indicates the best solution. The circle is a representation of the 2D histogram from the bootstrap results (all bootstrap results coincide with the best solution here). All other point source parameters have been fixed at their values for the best fitting point source.
of P-waves, especially at stations in south-easterly direction in this example (see figure 4.5).

When rupture velocity is also inverted for (the method is explained in section 3.6), we find a major ambiguity between rupture velocity and the border-radius parameter of the source model (figure 4.6). Despite the use of noise-free data in this example, the bootstrap test gives some scatter. Also, due to the rough gridspacing of the grid search, it is not possible to reproduce the exact input values.


Figure 4.2: Stability analysis of point source inversion of noise-free synthetic data for an extended source. Plot style and symbols are as in the preceding figure. Six 2D-projections of the minimal misfits of a 4D grid search in $\mathrm{M}_{\mathrm{W}}$ (magnitude), strike, dip, and slip-rake are shown.


Figure 4.3: Stability analysis of of point source inversion of noise-free synthetic data for an extended source. Plot style and symbols are as in the preceding figure. Three 2D-projections of the minimal misfits of a 3D grid search in the parameters north-shift, east-shift and time are shown.


Figure 4.4: Kinematic source inversion of noise-free synthetic data. Plot style and symbols are as in the preceding figures. These are 2D projections from a 3D grid search on the border-radius parameter and the coordinates of the nucleation point.


Figure 4.5: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depict an improvement of fit with respect to the best fitting point source model, red colors indicate decline.

Fault plane (strike $=135^{\circ}$ )


Figure 4.6: Kinematic source inversion of noise-free synthetic data. Attempt to additionally invert for rupture velocity. Plot style and symbols are as in figure 4.1. These are 2 D projections from a 4 D grid search within the vicinity of the previously found solution (figure 4.5).

### 4.2 Inversion with noise

In the previous section the general features of our method have been shown using an inversion applied to noise-free data. Even with the undistorted data some sensitivity effects have already occurred. Here, I want to further investigate potential sources of error and ambiguities as an intermediate step between the optimal case and the real application. Due to uncontrollable external influences, the presence of noise can't be neglected at any time, so it is an important test to verify that its presence does not affect the outcome of the inversion routine. Although not taking into account local specialities, an analysis of synthetic data with added white noise already shows the main points. Here, I choose to add gaussian white noise with station-dependent amplitudes. The amplitudes are scaled according to the maximum amplitudes observed over the whole trace at each station. This type of noise is not typical for reality, but is a simple way to partly account for imprecise forward modeling.

The same course of action as in the preceding section is taken, but this time with noise added to the synthetic observations.

In figures 4.7-4.9 the point source inversion results are shown. In comparison with the results from the inversion without noise changes are small. Very few ( $<1 \%$ ) of the bootstrap results are assigned to the neighboring grid nodes of the grid searches. The amount of noise has been choosen so that misfit is now on the order of 0.5 (comparable with real observations).

Figures 4.10 and 4.11 summarize the results of the inversion for the kinematic model, with the rupture velocity fixed to $80 \%$ of the shear wave velocity. The fit of seismograms is shown in figures 4.12 and 4.13. Considerable scatter is indicated by the bootstrap test in the inversion for the extended source. The borderradius of the model is estimated to 17 km with a $68 \%$ confidence interval of ( 12.6 , 18.2) km and the offset of the nucleation point along strike is found to be -5 km with a confidence interval of $(-10,-4) \mathrm{km}$. Additionally to the less stable result, a local minimum is present in the misfit subspace between border-radius and horizontal shift of the nucleation point. It corresponds to a bilateral rupture having approximatly the same proportions as the true result, but instead of having the nucleation point north-west of the centroid, it has it in the south-east. A similar observation is found in the inversion of real data of the L'Aquila earthquake 3.18.

Trying to invert additionally for rupture velocity causes the inversion to become unstable. The corresponding grid search results are depicted in figure 4.14. Only a ratio between border-radius and rupture velocity can reliably be obtained with the presented method.


Figure 4.7: Stability analysis of point source inversion of noisy synthetic data for an extended source. The variation of misfit with respect to depth and moment (magnitude $\mathrm{M}_{\mathrm{W}}$ ) is shown as contour lines and color. The star indicates the best solution. The circle is a representation of the 2D histogram from the bootstrap results (all bootstrap results coincide with the best solution here). All other point source parameters have been fixed at their values for the best fitting point source.


Figure 4.8: Stability analysis of point source inversion of noisy synthetic data for an extended source. Plot style and symbols are as in the preceding figure. Six 2Dprojections of the minimal misfits of a 4 D grid search in $\mathrm{M}_{\mathrm{W}}$ (magnitude), strike, dip, and slip-rake are shown.


Figure 4.9: Stability analysis of point source inversion of noisy synthetic data for an extended source. Plot style and symbols are as in the preceding figure. Three 2D-projections of the minimal misfits of a 3D grid search in the parameters northshift, east-shift and time are shown.


Figure 4.10: Kinematic source inversion of noisy synthetic data. Plot style and symbols are as in the preceding figures. These are 2D projections from a 3D grid search on the border-radius parameter and the coordinates of the nucleation point.


Figure 4.11: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depict an improvement of fit with respect to the best fitting point source model, red colors indicate decline.


Figure 4.12: Synthetic observations with noise added (dashed black), seismograms for best fitting point source (blue), and extended source (green) for inversion with noise. The seismograms for the actual fault plane are shown. The azimuth to the station is indicated by a straight line in the small maps to the right and station distance is indicated by a circle. Continued in figure 4.13.


Figure 4.13: Continued from figure 4.12:

Fault plane (strike $=135^{\circ}$ )


Figure 4.14: Kinematic source inversion of noisy synthetic data. Attempt to additionally invert for rupture velocity. Plot style and symbols are as in figure 4.7. These are 2D projections from a 4D grid search within the vicinity of the previously found solution (figure 4.11).

### 4.3 Sensitivity study of source directivity effects

In order to understand what kind of effects the extension and direction of rupture produce on seismograms, I present in this section a forward modeling and sensitivity study of source directivity effects [e.g. Aki and Richards, 2002]. Rather than showing how idealized directivity effects look like my focus here is towards the application to real data. I will show where and how strong directivity effects show up in the seismograms, how they depend on distance and azimuth with respect to the source and what frequencies should be included into an analysis of the source characteristics.

My approach to this topic will be somehow reversed, starting off at synthetic observations: I will first setup a synthetic reference earthquake which is only slightly idealized and investigate what errors are made when using a point source model to describe it. This way is chosen, because this is also the situation we are facing when looking at real data. However, with real data we would not be completely sure about the true input and we would not have the possibility to look at noise-free data.

Synthetic seismograms are calculated for a rupture process similar to $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake of 2009 (3.1), using a model within the parameter-space provided by the eikonal source model. Normal faulting with a dip of $45^{\circ}$ is assumed. The rupture itself is made partly bilateral, breaking more to the south-east than to the north-west. The complete setup is depicted in figure 4.15 and the chosen source parameters are given in table 4.1.
The Green's functions as defined in 2.8 used to calculate the seismograms were generated using the GEMINI program by Friederich and Dalkolmo [1995]. Full waveform seismograms for frequencies of up to 0.3 Hz were evaluated for the IASP91 earth model [Kennet, 1991].
Next, I try to fit the seismograms produced for the extended fault model using seismograms for a point source model. I assume here, that the exact location and time origin of the centroid, as well as the orientation of the fault plane can be correctly derived. However, I will determine magnitude, depth, and rise-time of the event, because these are the parameters which are more likely biased when a point source is used as a model for an extended rupture. The parameters are fitted using a grid search on all combinations of the three free parameters. Amplitude spectra of the P-phase pulses are fitted first, in order to isolate the effect of the extension of the fault on the P-phase (which I want to study first). The spectra are taken from a time window of 60 s length. An $l^{1}$-norm is used to compare the two sets of seismograms. The filtering is set broad enough, so that we have resolution in the full range of rise-times considered. Finally distance dependent weighting has to be introduced, so that the influence of stations near to the source

(a) A grid of receivers (black triangles) is distributed logarithmically in distance and evenly in azimuth around the source. The source shows normal faulting behavior, strikes at $135^{\circ}$ and has a dip angle of $45^{\circ}$. The actual fault plane is indicated as a black line in this focal mechanism representation.

(b) Shape of the rupture surface and isochrones of the rupture front (black contours). The rupture has a size of approximately $30 \mathrm{~km} \times 12.7 \mathrm{~km}$. The upper boundary of the rupture surface is at 1 km , the lower boundary is at 10 km depth. The location of the nucleation center (red star) is chosen so that the rupture propagates mainly from northwest to south-east. The black dots indicate the sub-fault centroids used during the calculation of synthetic seismograms.

Figure 4.15: Experiment setup used for synthetic study of directivity effects from a normal faulting event. Size and source parameters have been chosen to resemble the 6.3 L'Aquila earthquake of 2009.
does not inappropriately dominate. Adaptive station weighting, as described in section 2.6.3, is applied.

The results of the grid search are summarized in figure 4.16. As a first (obvious) result, one can see that the magnitude does not deviate much from the true magnitude of the event. A point source depth of 6 km best fits the extended rupture stretching from 1 to 10 km depth, but the bootstrap errors indicate some scatter. An overall rise-time of 8 seconds is the preferred point source duration using this data and misfit setup.
The largest disagreement between the point source and the reference seismograms can be expected to be found along the strike direction of the event, while the smallest should be found perpendicular to this direction. Two sections of seismograms in these directions are plotted in figure 4.17.
To further inspect the angular and distantial dependence of the disagreement of the waveforms one has to look at the misfit obtained at each node of the grid of receivers. Because the misfit values are dominated by the influence of distance

(a) Minimal misfits found in the grid search. The misfit was calculated using $l^{1}$-norm on P wave amplitude spectra. Distance dependent station weighting was employed to compensate for the amplitude decay with distance. Each blue dot marks the smallest misfit value of all tested combinations of the other two search parameters.

(b) Histograms approximating the probability of finding the grid search's misfit minimum at any specific choice of the search parameters, with respect to data selection. The probabilities have been calculated using 1000 bootstrap iterations with randomly chosen receiver configurations, as described in section 2.7.2.

Figure 4.16: Results from 3D grid search on the source parameters $M_{W}$ (magnitude), depth, and rise-time, which was used to find the point source model which best approximates the synthetic test event defined in fig. 4.15.

(a) Section aligning with direction of strike

(b) Section perpendicular to strike direction

Figure 4.17: P-waveform comparison for extended source (red) and point-source (blue) for the synthetic normal faulting test event defined in fig. 4.15. The vertical component of displacement is shown with flipped sign. The effect of directivity can be clearly seen on the seismograms of the section aligning with the strike direction. It is much smaller on the seismogram section perpendicular to that.


Figure 4.18: Per-receiver P-wave misfits for the point source model which best approximates the reference data of the test event defined in fig. 4.15. The misfits have been calculated using the vertical component of P -wave amplitude spectra. The misfits are locally normalized per receiver by the trace misfit normalization factors as defined in (2.13). This normalization effectively removes the radiation pattern and emphasizes the differences in the waveforms. See fig. 4.19 for the arguments to the division used for normalization. The function has been evaluated only at the black dots and was interpolated elsewhere in the image.
to the receiver and the radiation pattern of the event, I show locally normalized misfit values in figure 4.18. The arguments of this normalization are shown in figure 4.19. Adaptive station weights have been applied to the data in the latter figure. The station weights enter both in numerator and denominator and are thus here only used to improve figure 4.19. The station weights dependence on distance is shown in figure 4.20.

The effects of directivity are largest in the directions aligning with the strike direction of the event (which is also the main direction of rupture). Due to the radiation pattern, amplitudes directly in these direction are small compared to all other directions. The two effects superpose in a way that there is a wide range of azimuths $\left(90^{\circ}-180^{\circ}\right)$ yielding comparable values for the absolute value of misfit (figure 4.19 b ).

(a) Distance dependent weighted trace misfits $w_{i} m_{i}$ globally normalized by $N \sum w_{i} n_{i}$.

(b) Distance dependent weighted trace misfit normalization factors $w_{i} n_{i}$ globally normalized in the same manner as (a), revealing the radiation pattern of the synthetic test event.

Figure 4.19: Numerator (a) and denominator (b) used to calculate the locally normalized misfits shown in fig. 4.18.


Figure 4.20: Example P-wave station weights as a function of distance obtained by forward modeling (explained in section 2.6.2). These station weights compensate for an averaged radiation pattern for double-couple point sources. Different colors indicate different source depth: red: 1 km , blue: 2.7 km , green: 7 km , orange: 18 km , purple: 50 km .

As a comparison, the same procedure is applied to SH phases on the transversal component of the seismograms. The results are given in figures 4.21 and 4.22.

(a) Locally normalized misfit as explained in the caption of fig. 4.18.

(b) Distance dependent weighted misfit as in fig. 4.19(a).

(c) Distance dependent weighted misfit normalization factor as in fig. 4.19(b).

Figure 4.21: Point source approximation error as a function of distance and azimuth, as displayed in 4.18 and 4.19, but using SH waves on the transvesal component of the seismograms. The point source used here for comparison had a duration of 5 s and a depth of 6 km .


Figure 4.22: SH waveform comparison for extended source (red) and pointsource (blue) for the synthetic normal faulting test event defined in fig. 4.15. The transversal component of displacement is shown.

### 4.3.1 Frequency dependence

At long wavelengths, a point source model is a valid representation for earthquake rupture. If we look at higher frequencies, this is no longer the case. I will here give, as an example, the frequency dependence of the approximation error made when using a point source to model an extended earthquake. The same setup, as in the previous section is used. A point source with a duration of 8 s is used to match the rupture model from figure 4.15. Normalized misfit is calculated as a function of the corner frequency of the low-pass. The result is shown in figure 4.23 and compared to the cases, where rupture is symmetrically bilateral and purely unilateral. Each of these other cases has been fitted against point sources of durations matching their mean apparent duration. The approximation error is always smallest for the bilateral model and largest for purely unilateral rupture.


Figure 4.23: Frequency-dependence of point source approximation error. Solid red line: global misfit obtained when fitting a point source approximation with 8 s duration against the extended test source defined in figure 4.15 , as a function of the corner period of the low-pass applied to both sets of spectra. This calculation has been repeated, replacing the asymmetrical bilateral rupture of the test event with purely unilateral rupture (dashed blue line) and symmetrical bilateral rupture (dashed green line), respectively. Because the different source models yield different apparent rise-times, each one is compared to its own specific point source approximation: The purely unilateral source is compared to a point source of 10 s duration, the symmetrical bilateral source is compared to a point source of 5 s duration.

(a) Locally normalized misfit as explained in the caption of fig. 4.18.

(b) Distance dependent weighted misfit as in fig. 4.19(a).

(c) Distance dependent weighted misfit normalization factor as in fig. 4.19(b).

Figure 4.24: Point source approximation error as a function of distance and azimuth, as displayed in 4.18 and 4.19 , but using $l^{1}$-norm on the difference between the time-domain traces instead of spectra. The point source used here had a duration of 6 s and a depth of 5 km .

### 4.3.2 Time domain norm

In the preceding discussion of the azimuthal and distantial dependence of source directivity effects, an $l^{1}$-norm misfit was used on the differences between point source and extended source seismograms amplitude spectra. As a comparison I replace here the amplitude spectra-based comparison with a comparison of the time domain traces. The results are shown in figure 4.24. The normalized misfit values reach slightly higher values, but the qualitative behaviour is similar.

## Chapter 5

## Examples and evaluation

### 5.1 Strike-slip: $M_{W}$ 6.9 Gulf of California, 2009

As an example for a strike-slip event I present in this section inversion results for the $\mathrm{M}_{\mathrm{W}} 6.9$ earthquake which occured on August $3,2009,17: 59: 56$ UTC in the Gulf of California. USGS located the epicenter to ( $29.066^{\circ} \mathrm{N}, 112.871^{\circ} \mathrm{W}$ ). The centroid location given in the Global CMT (gCMT) catalog [Ekström and Nettles, 1982] is $\left(29.27^{\circ} \mathrm{N}, 113.50^{\circ} \mathrm{W}\right)$.

The earthquake was in the transform fault region of the plate boundary between the North America and Pacific plate. The Pacific plate moves northward with a speed of about $45 \mathrm{~mm} / \mathrm{yr}$ with respect to the North America plate [USGS, U.S. Geological Survey and NEIC, National Earthquake Information Center, 2009].

Hayes, G. [2009] derive a kinematic source model for the $\mathrm{M}_{\mathrm{W}}$ 6.9 Gulf of California earthquake using the finite fault slip inversion method by Ji et al. [2002]. They derive unilateral rupture propagating about 30 km northwestward.

A comparison of my point source results with those provided by gCMT is given in table 5.1. Except for the moment estimation which yields a slightly smaller value in my case, all other parameters are in good agreement.
My results of the inversion for the extended fault model are given in table 5.2 and are visualized in figures 5.1 and 5.2. Histograms of the source parameter probabilities are shown in figure 5.3. The misfits yielded by the grid searches together with joint probability histograms are plotted in figure 5.4. The best model for the fault plane striking at $312^{\circ}$ (which alignes with the orientation of the fault system), gives a smaller misfit (0.436) than the misfit of the best model obtained assuming the other plane (0.479). According to my results, rupture propagated almost unilaterally about $35-40 \mathrm{~km}$ northwestward.

## Point source parameters

|  | $g C M T$ | My study | My 68\% confidence interval |  |
| :---: | :---: | :---: | :---: | :---: |
| Time | 18:00:05.8 | 18:00:05.5 | -0.5 s (offset) | +0.5 s (offset) |
| Latitude | $29.27^{\circ}$ | $29.22^{\circ}$ | -15. km (offset) | +5. km (offset) |
| Longitude | -113.50 ${ }^{\circ}$ | -113.43 ${ }^{\circ}$ | -15. km (offset) | +5. km (offset) |
| Depth | 15.4 km | 5. km (limit) | 4.5 km | 5.5 km |
| M ${ }_{\text {W }}$ | 6.9 | 6.8 | 6.7 | 6.8 |
| Moment | $2.49 \mathrm{e}+19 \mathrm{Nm}$ | $1.7 \mathrm{e}+19 \mathrm{Nm}$ | $1.4 \mathrm{e}+19 \mathrm{Nm}$ | $1.8 \mathrm{e}+19 \mathrm{Nm}$ |
| Strike | $312 .{ }^{\circ}$ | $312 .{ }^{\circ}$ | $310 .{ }^{\circ}$ | $315 .{ }^{\circ}$ |
| Dip | $87 .{ }^{\circ}$ | $86 .{ }^{\circ}$ | $84 .{ }^{\circ}$ | 89. ${ }^{\circ}$ |
| Slip-Rake | -174. ${ }^{\circ}$ | $178 .{ }^{\circ}$ | $171 .{ }^{\circ}$ | $186 .{ }^{\circ}$ |
| Duration | 6.6 s | $12 . \mathrm{s}$ | $12 . \mathrm{s}$ | 14.5 |

Table 5.1: Results and errors for point source parameters estimated for the $M_{W} 6.9$ Gulf of California earthquake of 2009.

## Fault plane (strike=312 ${ }^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :--- | ---: | ---: | ---: |
| Border radius | $(\mathrm{km})$ | 21. | 18. | 23. |
| Nucleation along strike | $(\mathrm{km})$ | -16. | -18. | -13. |
| Nucleation down dip | $(\mathrm{km})$ | 0. | -3. | 3. |

## Auxiliary plane (strike $=42^{\circ}$ )

| Parameter | Unit | Value | $68 \%$ confidence interval |  |
| :--- | :--- | :---: | ---: | :---: |
| Border radius | $(\mathrm{km})$ | 36. | 18. | 44. |
| Nucleation along strike | $(\mathrm{km})$ | 0. | -13. | 13. |
| Nucleation down dip | $(\mathrm{km})$ | 0. | -3. | 3. |

Table 5.2: Fault and auxiliary plane results and errors for kinematic source parameters estimated for the $\mathrm{M}_{\mathrm{W}}$ 6.8 Gulf of California earthquake under the assumption of a fixed rupture velocity of $80 \%$ of shear wave velocity.

Allowing for a variable rupture velocity in the inversion leads to unstable results (see figure 5.6). The result is ambiguous in a way that larger velocities require larger rupture length. The ratio between border-radius $R$ and the relative rupture velocity $v_{r} / v_{s}$ can be given as $R /\left(v_{r} / v_{s}\right) \approx 28 \mathrm{~km}$ for this event.
My results are roughly in accord with the slip inversion results by Hayes, G. [2009]. The pattern of directivity can be nicely seen in the station-wise breakdown of misfit improvements, given in figure 5.5.


Figure 5.1: Eikonal source model, as derived for the $\mathrm{M}_{\mathrm{W}}$ 6.8 Gulf of California earthquake of 2009. Isochrones of the rupture fronts are contoured with black lines. The red star indicates the nucleation point of the rupture.


Figure 5.2: Eikonal source model derived for the $\mathrm{M}_{\mathrm{W}}$ 6.8 Gulf of California earthquake of 2009, projected on an epicentral map. The red star indicates the nucleation point of the rupture.

Fault plane (strike $=312^{\circ}$ )


Figure 5.3: Histograms from bootstrap of grid search for extended source model. A grid search has been used to search the parameter border-radius and the coordinates of the nucleation point. The histograms shown here represent the probability of the finding specific values for the parameters within the space searched.


Figure 5.4: Determination of the extended rupture model. The minimal misfits found in a 3D grid search on the parameter border-radius and the coordinates of the nucleation point are projected into two dimensions and visualized with contour lines and color (each point in the plots is the minimal misfit found when varying the third free parameter). The red star marks the model with the best fit. The black circles represent a 2D-histogram of the bootstrap results.


Figure 5.5: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depict an improvement of fit with respect to the best fitting point source model, red colors indicate decline.


Figure 5.6: Misfit and histogram plots from grid search to estimate the rupture velocity. A four-dimensional grid search has been done in order to estimate rupture velocity and to refine border radius and nucleation center of the event.

### 5.2 Oblique-slip: M $_{W}$ 7.0 Haiti, 2010

As a further example, I show in this section the automatic solutions derived for the recent $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti earthquake of 2010, which was with 222.570 casualties one of the deadliest earthquakes on record.

The earthquake occured at 21:53 UTC on January 12, 2010. The centroid location given by gCMT is $\left(18.62^{\circ} \mathrm{N}, 72.59^{\circ} \mathrm{W}\right)$, which points us to the Enriquillo-Plantain Garden fault system in the southwestern tip of Hispaniola. Hispaniola is frequently hit by disastrous earthquakes [e.g. Scherer, 1912]. An overview of the tectonic settings is given by Mann et al. [1991]: Seismicity is driven by westward shift ( $1.1 \mathrm{~cm} / \mathrm{yr}$ ) of the North American plate with respect to the Caribbean plate. In the eastern part of the island, motion between the American and Caribbean plate is partitioned into two distinct oblique-slip fault-systems, one in the north (Septentrional) and one in the south (Enriquillo-Plantain Garden). The westcentral part of the island is furthermore under a east-west oriented compressional stress regime, yielding oblique-slip and reverse faulting.

Hayes, G. [2010a] derive a kinematic source model for the $\mathrm{M}_{\mathrm{W}}$ 7.0 Haiti earthquake using the finite fault slip inversion method by Ji et al. [2002]. They derive bilateral rupture propagating about 8 km eastward and about 25 km westward.

A comparison of my point source results with those provided by gCMT is given in table 5.3. My results of the inversion for the extended fault model are given in table 5.4 and are visualized in figures 5.7 and 5.8. Histograms of the source parameter probabilities are shown in figure 5.9. The misfits yielded by the grid searches together with joint probability histograms are plotted in figure 5.10. The best model for the fault plane striking at $253^{\circ}$ (which alignes with the EnriquilloPlantain Garden fault system), gives a slightly smaller misfit (0.486) than the misfit of the best model obtained assuming the auxiliary plane (0.495). Bilateral rupture propagating about 10 km to the east and 34 km to the west is indicated by my best model. The bootstrap results reveal a second possible but slightly less probable model with a slightly larger size and opposite proportions.

Allowing for a variable rupture velocity in the inversion leads to unstable results (see figure 5.12). The result is ambiguous in a way that larger velocities require larger rupture length. The ratio between border-radius $R$ and the relative rupture velocity $v_{r} / v_{s}$ can be given as $R /\left(v_{r} / v_{s}\right) \approx 30 \mathrm{~km}$ for this event.

## Point source parameters

|  | gCMT | My study | My 68\% confidence interval |  |
| :---: | :---: | :---: | :---: | :---: |
| Time | 21:53:17.7 | 21:53:14.7 | -0.5 s (offset) | +0.5 s (offset) |
| Latitude | $18.62^{\circ}$ | $18.47^{\circ}$ | -5. km (offset) | +5. km (offset) |
| Longitude | -72.59 ${ }^{\circ}$ | $-72.64{ }^{\circ}$ | -5. km (offset) | $+5 . \mathrm{km}$ (offset) |
| Depth | 12. km (fixed) | 6. km | 4. km | 8. km |
| M ${ }_{\text {W }}$ | 7.1 | 7.0 | 7.0 | 7.0 |
| Moment | $4.74 \mathrm{e}+19 \mathrm{Nm}$ | $3.6 \mathrm{e}+19 \mathrm{Nm}$ | $3.1 \mathrm{e}+19 \mathrm{Nm}$ | $4.2 \mathrm{e}+19 \mathrm{Nm}$ |
| Strike | $151 .{ }^{\circ}$ | $150 .{ }^{\circ}$ | 147. ${ }^{\circ}$ | 152. ${ }^{\circ}$ |
| Dip | $64 .{ }^{\circ}$ | $58 .{ }^{\circ}$ | $56 .{ }^{\circ}$ | $66 .{ }^{\circ}$ |
| Slip-Rake | $158 .{ }^{\circ}$ | $156 .{ }^{\circ}$ | 148. ${ }^{\circ}$ | 163. ${ }^{\circ}$ |
| Rise-Time | 8.2 s | 9.5 | 8.5 s | $10 . \mathrm{s}$ |

Table 5.3: Results and errors for point source parameters estimated for the $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti earthquake of 2010. The angles for the auxilliary plane are shown.

## Fault plane (strike=253 ${ }^{\circ}$ )

| Parameter | Unit | Value | $68 \%$ confidence interval |  |
| :--- | :--- | ---: | ---: | ---: |
| Border radius | $(\mathrm{km})$ | 22. |  | 20. |
| Nucleation along strike | $(\mathrm{km})$ | -11. | -13. | 13. |
| Nucleation down dip | $(\mathrm{km})$ | 4. | 2. | 6. |

## Auxiliary plane (strike $=150^{\circ}$ )

| Parameter | Unit | Value | $68 \%$ |  |
| :--- | :--- | :--- | :--- | :--- |
| confidence interval |  |  |  |  |
| Border radius | $(\mathrm{km})$ | 25. | 20. | 27. |
| Nucleation along strike | $(\mathrm{km})$ | -4. | -9. | 9. |
| Nucleation down dip | $(\mathrm{km})$ | 4. | -6. | 6. |

Table 5.4: Fault and auxiliary plane results and errors for kinematic source parameters estimated for the $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti earthquake of 2010 under the assumption of a fixed rupture velocity of $80 \%$ of shear wave velocity.


Figure 5.7: Eikonal source model, as derived for the $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti earthquake of 2010. Isochrones of the rupture fronts are contoured with black lines. The red star indicates the nucleation point of the rupture.


Figure 5.8: Eikonal source model derived for the $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti earthquake of 2010, projected on an epicentral map. The red star indicates the nucleation point of the rupture.


Figure 5.9: Histograms from bootstrap of grid search for extended source model. A grid search has been used to search the parameter border-radius and the coordinates of the nucleation point. The histograms shown here represent the probability of the finding specific values for the parameters within the space searched.


Figure 5.10: Determination of the extended rupture model. The minimal misfits found in a 3D grid search on the parameter border-radius and the coordinates of the nucleation point are projected into two dimensions and visualized with contour lines and color (each point in the plots is the minimal misfit found when varying the third free parameter). The red star marks the model with the best fit. The black circles represent a 2D-histogram of the bootstrap results.


Figure 5.11: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depict an improvement of fit with respect to the best fitting point source model, red colors indicate decline.


Figure 5.12: Misfit and histogram plots from grid search to estimate the rupture velocity. A four-dimensional grid search has been done in order to estimate rupture velocity and to refine border radius and nucleation center of the event. Plot style and symbols are explained in the caption of figure 5.10

### 5.3 Thrust faulting: $M_{W}$ 7.2 Sumatra, 2010

As an example for a thrust faulting event, I show in this section the automatic solutions derived for the $\mathrm{M}_{\mathrm{W}} 7.2$ Sumatra earthquake of May 9, 2010.

The earthquake occured at 05:59 UTC on May 9, 2010. It occured in the subduction zone at the between the Australia-India and Sunda plate.
Hayes, G. [2010b] derive a kinematic source model for this event, using the finite fault slip inversion method by Ji et al. [2002]. They derive two slip patches, a stronger one to the northwest of the epicenter and weaker one to the southeast. They find an extension of about 40 km .

A comparison of my point source results with those provided by gCMT is given in table 5.5. My results of the inversion for the extended fault model are given in table 5.6 and are visualized in figures 5.13 and 5.14. Histograms of the source parameter probabilities are shown in figure 5.15. The misfits yielded by the grid searches together with joint probability histograms are plotted in figure 5.16.

I assume that the shallow dipping plane is the fault plane although the misfit values obtained for the steeper dipping plane are slightly smaller. Unilateral rupture is proposed by my result. This is approximately consistant with the dominating stronger patch in the result by Hayes, G. [2010b]. Our estimate for the size of the rupture is in agreement with the size given there.

## Point source parameters

|  | $g C M T$ | My study | My $68 \%$ confidence interval |  |
| :--- | :---: | :---: | :---: | :---: |
| Time | $05: 59: 51.4$ | $05: 59: 50.4$ | -0.5 s (offset) | +0.5 s (offset) |
| Latitude | $3.36^{\circ}$ | $3.7^{\circ}$ | $-15 . \mathrm{km}$ (offset) | $+5 . \mathrm{km}$ (offset) |
| Longitude | $95.78^{\circ}$ | $95.96^{\circ}$ | $-5 . \mathrm{km}$ (offset) | $+5 . \mathrm{km}$ (offset) |
| Depth | 37.2 km (fixed) | $43 . \mathrm{km}$ | $42 . \mathrm{km}$ | $46 . \mathrm{km}$ |
| M | 7.2 | 7.2 | 7.2 | 7.3 |
| Moment | $9.41 \mathrm{e}+19 \mathrm{Nm}$ | $8.1 \mathrm{e}+19 \mathrm{Nm}$ | $7.7 \mathrm{e}+19 \mathrm{Nm}$ | $8.5 \mathrm{e}+19 \mathrm{Nm}$ |
| Strike | $130 . \circ^{\circ}$ | $126 .^{\circ}$ | $124 . .^{\circ}$ | $129 .^{\circ}$ |
| Dip | $75 .^{\circ}$ | $70 . .^{\circ}$ | $67 .^{\circ}$ | $72 .^{\circ}$ |
| Slip-Rake | $91 .^{\circ}$ | $94 .^{\circ}$ | $91 .^{\circ}$ | $96 . .^{\circ}$ |
| Rise-Time | 10.3 s | $12 . \mathrm{s}$ | $10 . \mathrm{s}$ | $12 . \mathrm{s}$ |

Table 5.5: Results and errors for point source parameters estimated for the $\mathrm{M}_{\mathrm{W}} 7.0$ Sumatra earthquake of 2010. The angles for the auxilliary plane are shown.

## Fault plane $\left(\right.$ strike $=294{ }^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :--- | :---: | ---: | :---: |
| Border radius | $(\mathrm{km})$ | 22. | 20. | 23. |
| Nucleation along strike | $(\mathrm{km})$ | -18. | -20. | -13. |
| Nucleation down dip | $(\mathrm{km})$ | 11. | 9. | 16. |

## Auxiliary plane (strike $=126^{\circ}$ )

| Parameter | Unit | Value | 68\% confidence interval |  |
| :--- | :---: | ---: | :--- | :--- |
| Border radius | $(\mathrm{km})$ | 22. | 20. |  |
| Nucleation along strike | $(\mathrm{km})$ | 14. | 13. | 16. |
| Nucleation down dip | $(\mathrm{km})$ | 0. | -2. | 2. |

Table 5.6: Fault and auxiliary plane results and errors for kinematic source parameters estimated for the $M_{W} 7.2$ Sumatra earthquake of 2010 under the assumption of a fixed rupture velocity of $60 \%$ of shear wave velocity.


Figure 5.13: Eikonal source model, as derived for the $\mathrm{M}_{\mathrm{W}} 7.2$ Sumatra earthquake of 2010. Isochrones of the rupture fronts are contoured with black lines. The red star indicates the nucleation point of the rupture.

Fault plane $\left(\right.$ strike $\left.=294^{\circ}\right)$


Auxiliary plane (strike $=126^{\circ}$ )


Figure 5.14: Eikonal source model derived for the $\mathrm{M}_{\mathrm{W}} 7.2$ Sumatra earthquake of 2010, projected on an epicentral map. The red star indicates the nucleation point of the rupture.


Figure 5.15: Histograms from bootstrap of grid search for extended source model. A grid search has been used to search the parameter border-radius and the coordinates of the nucleation point. The histograms shown here represent the probability of the finding specific values for the parameters within the space searched.


Figure 5.16: Determination of the extended rupture model. The minimal misfits found in a 3D grid search on the parameter border-radius and the coordinates of the nucleation point are projected into two dimensions and visualized with contour lines and color (each point in the plots is the minimal misfit found when varying the third free parameter). The red star marks the model with the best fit. The black circles represent a 2D-histogram of the bootstrap results.


Figure 5.17: Misfit breakdown by station. Two flags are drawn at each station. The flag pointing to the left is for the P-phase window, the flag pointing to the right is for the S-phase window. The sizes of the flags are proportional to the misfit contributed by the represented phase. Blue colors depict an improvement of fit with respect to the best fitting point source model, red colors indicate decline.


Figure 5.18: Misfit and histogram plots from grid search to estimate the rupture velocity. A four-dimensional grid search has been done in order to estimate rupture velocity and to refine border radius and nucleation center of the event.

### 5.4 Comparison of points source solution catalogs

In the former sections I gave detailed comparisons of the inversion results for extended sources for a few selected earthquakes. In contrast to such isolated case studies for extended sources, in the case of point source inversion, there exist several catalogs of inverted source parameters. Arguably the most important of these is the Global CMT (gCMT) catalog [Ekström and Nettles, 1982]. At this point a comparison of our point source inversion results with the gCMT catalog is appropriate. Events with magnitude larger than $\mathrm{M}_{\mathrm{W}} 6$ and depth of less than 100 km have been analyzed for a time frame from August, 2008 to June, 2010. The list of events is given in appendix D .

Focal sphere diagrams for all processed events are shown in figure 5.20, together with the respective solutions from the reference catalog. Projections on a global map are shown in figure 5.19. Here it is to mention that gCMT, of course, does not only invert for double-couple, but the full moment tensor information, For a direct comparison, I show the double couple compontents of these. For few events, no solution was found in the gCMT catalog.

At first glance nearly all solutions are in good agreement. Only for a few events distinct deviations in the orientation can be found. In even less cases, a disagreement in polarity shows up.

Our magnitude estimates are systematically a bit smaller than the corresponding estimates given by gCMT (figure 5.21). This might be due to the use of a slightly different earth model and Green's functions. Also our depth estimates tend to be slightly shallower (figure 5.22).

The high degree of correlation with a well tested catalog containing results gained with an independent method, give reason to trust in the results of our point source inversion.

A more detailed comparison is planned, once a larger dataset has been processed with the most recent version of our tools.


Figure 5.19: Focal solutions and locations of 240 earthquakes with $\mathrm{M}_{\mathrm{W}}>6$ and depth $<100 \mathrm{~km}$ for a time frame from August, 2008 to June, 2010, as estimated with our automatic point source inversion. The symbols have been scaled by magnitude and drawn ordered by centroid depth.


Figure 5.20: Comparing focal solutions of the gCMT catalog (grey symbols) with our results (black symbols) for 240 earthquakes with $\mathrm{M}_{\mathrm{W}}>6$ and depth $<100 \mathrm{~km}$. The double-couple component of the gCMT solutions is shown. For few events, no solution was found in the gCMT catalog. The list of events can be found in D


Figure 5.21: Comparison of our estimates for the moment magnitude $\mathrm{M}_{\mathrm{W}}$ of 240 test events with the corresponding solutions published in the gCMT catalog.


Figure 5.22: Comparison of our estimates for the centroid depth of 240 test events with the corresponding solutions published in the gCMT catalog. The clipping at 12 km for the centroid depths given by gCMT appears because this method uses a fixed minimum depth of 12 km .

## Chapter 6

## Conclusions

In this work, I described a rupture model and the tools required to solve the forward and inverse problems associated with it.

A flexible kinematic model of earthquake rupture, the eikonal source, has been introduced. This model has been devoloped in order to prevent problems due to overparameterization, as they are present in other methods to image rupture [Beresnev, 2003], right from the start. It is based on 13 inversion parameters of which five are needed to specify extension and propagation of rupture. The low number of parameters has been achieved by the use of geometrical and physical constraints and a coarse level of detail.

The forward modeling is performed on the basis of pre-calculated Green's functions. The benefit is two-fold: the inversion is independent of the Green's function generation in a way that external tools can be used, and the forward modeling is very fast, at the price of an increased but manageable demand on disk storage and computer memory. Rules on how to safely discretize the parameterized source model into point sources have been given and implemented in the codes to make the handling of the source models as easy as possible. A storage scheme for Green's functions in a well defined and platform independent format has been established to simplify reuse and sharing of Green's functions databases. The possibility of using higher order interpolation methods to reduce the number of stored Green's functions has been considered and implemented. Such an interpolation allows for a decrease in the demand on disk space at the price of more computation power, which was not apropriate for my applications, so far.

How seismograms and simulation results are compared is subject to the definition of misfit, data weighting, and data selection using tapering and filtering. I have chosen to use misfits based on an $l^{1}$-norm in order to improve the robustness of the inversions. Unweighted misfit contributions gained in time windows
of differing lengths at different epicentral distances for different phase arrivals vary strongly, so an unbiased adaptive method to determine equalizing weighting factors for the different datasets entering into an inversion has been developed.

To find solutions to the inverse problem, it is necessary to search misfit space for a global minimum, i.e. to find the choice of model parameters which produces the best fit in the data. This is a delicate task because of the presence of local minima and sometimes ambiguities in the treated misfit functions. Robust solutions are found using combinations of grid and gradient searches. To quantify errors or probabilities on the retrieved results a bootstrap technique has been adapted and applied exhaustively. It has turned out to be a very useful tool to uncover not only uncertainties but also ambiguities in the results.

Finally, some tools for the automatic processing of events have been presented. A problem which sometimes arises here is that too much data is available to be handled efficiently, so I proposed a procedure to evaluate a priori station qualities which can be used in combination with a special station selection algorithm which additionally takes into account azimuthal and distantial coverage.

A multi-step inversion strategy, based on the tools described above, suitable to operate without human supervision, has been developed and implemented. Its application has been exemplified, using the $\mathrm{M}_{\mathrm{W}} 6.3$ L'Aquila earthquake of 2009 as a test case. The overall operational functionality has been demonstrated in a test with noise free synthetic data as a proof of concept. In a further test, it has been shown which source parameters can be stably estimated also under nonoptimal conditions, when the data is distorted with noise. Even the synthetic tests already indicate some possible sources of error inherent to the distinct geometrical setup for this event: A slight ambiguity between strike and slip-rake, and a separated local minimum in the grid search for the extended source, corresponding to a spurious rupture pattern. Both of these features consistently appear in the detailed error analysis done for the L'Aquila test case study with real data. The distribution of strike and slip-rake from the centroid moment tensor solutions given by others (Global CMT [Ekström and Nettles, 1982], USGS CMT [U.S. Geological Survey, National Earthquake Information Center, 2009], [Walters et al., 2009]) for this event, roughly align with the detected ambiguity, hinting that this feature is probably inherent to the event geometry. The derived point source parameters are in good agreement with the findings of other analyses. The estimate for the extension of the rupture is larger ( $36-42 \mathrm{~km}$ length) than what would be expected from aftershock distributions and the estimates given by other groups (12-19 km) [Chiarabba et al., 2009, Walters et al., 2009, Cirella et al., 2009], but our results consistantly show that rupture nucleated northwest of the mean centroid and propageted more to the southeast than to the northwest. Inverting additionally for rupture velocity leads to unstable results in the sense that rup-
ture velocity and rupture size cannot be determined independently. According to this result, a rupture velocity of about $45 \%$ of shear wave velocity would have to be assumed for the model to fit a rupture length of 20 km . This would yield an average rupture velocity of about $2 \mathrm{~km} / \mathrm{s}$ which would be consistent with the results of Cirella et al. [2009]. A remarkable result is that even in the optimal case of noise free synthetic data, it seems impossible to distinguish between fault and auxiliary plane for this type of event, using this method.

Three additional earthquakes (one strike-slip, one oblique-slip, one thrust-fault) have been analyzed using the same automatic method which was used in the L'Aquila case study in order to demonstrate its applicability to other types of events. The point source models derived for the three events are in good agreement with the results given in the Global CMT catalog, which can be seen as further indication that the method is reliable and yields stable results.

For the strike slip and the oblique-slip examples ( $\mathrm{M}_{\mathrm{W}}$ 6.9 Gulf of California, August 3, 2009 and $\mathrm{M}_{\mathrm{W}} 7.0$ Haiti Earthquake January 12, 2010) the estimated parameters are in good agreement with the results of Hayes, G. [2009] and Hayes, G. [2010a], though both of my estimates for rupture size seem to be slightly larger. In these two cases it is well possible to distinguish between fault and auxiliary plane.
The reason for the often observed ambiguity between rupture velocity and rupture size is that longer rupture length, as well as slower rupture velocity, change the observed pulse widths in a very similar way. Whether it is still possible to resolve this ambiguity is case-dependent. The chances are raised, when near-field data is available, when station coverage is very good or when the waveforms radiated from different positions on the rupture surface change significantly (e.g. change is stronger with depth than laterally). Independent constraints on rupture size, e.g from aftershock distributions, GPS measurements, or InSAR results may be helpful here.
Unlike as is the case for extended source parameters, where comparison studies are not routinely available, for point source parameters there exist established catalogs. To statistically check the robustness of the part of our method providing point source parameters, a comparison with a well tested method (Global CMT catalog) has been done. The results are generally in good agreement except our magnitude estimates are systematically smaller, possibly due to slightly different earth models beeing used for the generation of Green's functions.
The method is now in a state which works stable and automatic. A catalog with extended source parameters, as is already existing for point source parameters, can now be created based on this work. The automatic routines to set up such a catalog exist and have been tested.

As with any scientific achievement, there is a lot of room for improvements. A
limiting factor to our method currently is the rough forward modeling. To improve that, the Green's functions could be tuned station-wise. The speed of our method could possibly be improved by using more sophisticated search algorithms to solve the minimization problem. When using only body-wave phases, a simple improvement would be to use bilinear interpolation on time-shifted traces using phase-specific reduction velocities, which would reduce the storage requirements by at least $95 \%$. A mixed time/spectral domain norm could possibly be used to avoid the switching between different norms, as is currently done and which in some cases leads to inconsistencies. An exciting problem, difficult to tackle is the ambiguity between rupture velocity and rupture size.
The presented work covers the complete range from the model of earthquake rupture to automatic event processing for catalog generation. It was a satisfactory experience to see that it is possible to find solutions and to create valuable new tools to handle each encountered link in this elaborate chain of challenges.

## Chapter 7

## Acknowledgments

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## Appendix A

## Software

## A. 1 The Kiwi Tools

A modular framework to analyze earthquake source parameters has been developed in the scope of the KINHERD project: The Kiwi Tools. It is an open source toolkit which can be used to implement various kinds of waveform inversion setups and automatic processing schemes.

The source code of the Kiwi Tools is available on the KINHERD project web page, http://kinherd.org/.
The Kiwi Tools consist of two parts, the Kiwi Core Tools and the Kiwi Extensions. The Kiwi Core Tools, written in Fortran 95, are modules and tools to efficiently calculate synthetic seismograms for extended earthquake sources by using pre-calculated Green's functions, and a framework to taper, filter, and calculate misfits between seismograms. It also contains tools and modules to create and modify Greens function databases for use with the Kiwi Tools and to interpolate Green's functions.

The Kiwi Core Tools are used internally by the Kiwi Extensions, an open set of programs and modules written using the Python programming language to perform higher level tasks, flow control, data handling, and plotting.

## A.1.1 Command line tools

- minimizer

Synthetic seismogram generator for extended earthquake sources and misfit calculation engine. This is the main front-end to the Kiwi Core Tools.

- source_info

Utility to display information about the built-in earthquake source models

- gfdb_build

Create and fill a Green's function database

- gfdb_extract

Extract traces from a Green's function database

- gfdb_redeploy

Copy traces from one Green's function database to another

- gfdb_info

Display information about a Green's function database

- gfdb_build_ahfull

Create a Green's function database with analytical solutions for a homogeneous full-space

- gfdb_phaser

A script to create subsets of Green's function databases containing specific arrival phases

- gfdb_scale

A script to create a scaled copy of a Green's function database

- gfdb_downsample

A script to create a downsampled copy of a Green's function database

- kinherd_sourceview

Visualization of the source models

- autokiwi

Automatic acquisition, preprocessing, and processing framework

## A.1.2 Core modules

The Kiwi Core Tools are built from a few Fortran 95 modules, the following list is an overview on the most important of them:

- comparator Misfit calculation framework
- crust $2 \times 2$ Read CRUST 2.0 model
- eikonal Eikonal equation solver
- elseis Calculate elementary seismograms for a homogeneous full-space
- geometry Geometrical utility functions
- $\mathrm{g} f \mathrm{db}$ Green's function storage caching functionality
- interpolation Implementation of Gülünay's 3D interpolation method
- orthodrome Distance and azimuth calculation
- receiver Receiver data structure and associated methods
- seismogram Seismogram generator for discretized sources
- seismogram_io Seismogram input and output
- source_bilat Bilateral rupture model
- source_circular Circular rupture model
- source_eikonal Eikonal source model
- source_moment_tensor Moment tensor point source
- sparse_trace Stacking of sparse traces


## A.1.3 The Minimizer tool

## A.1.3.1 Usage

```
> minimizer <<EOF
[minimizer commands]
EOF
```


## A.1.3.2 Description

The Minimizer tool is a text interface driven application to calculate synthetic seismograms and misfits for different point source and extended rupture models. It is intended to be run by other applications or scripts, which feed commands to this tool and retrieve their results. The application is completely configured by running series of commands, so no other configuration is needed.

The internal state of the application consists of a source model, which is discretized into a number of points sources, a Green's function database, a setup how Green's functions should be interpolated, a number of receivers (stations)
each containing several components, each of which having attached reference seismograms (observations), and a setup how misfits should be produced. The misfit configuration consists of tapers, which can be given per receiver, a filter, which is applied in the frequency-domain, and the kind of misfit to be used. The misfit norms can be applied to either the differences of the time domain traces or the differences of their amplitude spectra.

Traces from the Green's function database are cached in memory to speed up successive calculations. Lazy evaluation is used, where possible, so that if for example the kind of misfit norm is changed, only the misfit calculation is repeated and neither the source model discretization nor the seismogram stacking has to be done again. Some additional shortcuts are built into the code, such that if for example only the moment of a source model is changed, the already calculated seismograms are scaled according to the new moment, overleaping the discretization and stacking.
The Minimizer tool is run by feeding commands to its standard input and by retrieving results from its standard output using a very simple ascii protocol. After each executed command, Minimizer prints 'ok' or 'nok', depending on whether the command was executed successfully or not. 'ok' and 'nok' may be followed by ' $>$ ', in which case a further line of output is to be expected. In the case of failure, this line contains an error message, in the case of success it contains the results of the command. Additional error messages and warnings may be printed via stderr.

## A.1.3.3 Minimizer Commands

```
set_database dbpath [ nipx nipz ]
```

Select Green's function database to use for the calculation of synthetic seismograms.
dbpath is the path to a Greens function database created with $g f d b \_b u i l d$. This is the path without the filename extensions.index or .chunk.
nipx nipz turn on Gülünay's interpolation in the Greens function database if set to values other than one. A Greens function database opened this way will pretend to have nipx times the number of traces in the horizontal direction, inserting interpolated traces as needed. Same applies with nipz to the vertical. Gülünay's generalized FK interpolation is used to fill the interpolated traces. If either of nipx or nipz is set to one, a 2D interpolation (time-distance or time-depth) is performed. If both nipx and nipz are set to the same value, a 3D (time-distance-depth) interpolation is performed. If nipx and nipz are not set to the same value, first horizontal 2D interpolation is applied followed by a vertical 2D interpolation.

Note: Gül"unay's interpolation works in the spectral domain and uses FFTs and thus has cyclic properties. To prevent wrap-around artifacts, the interpolation is done block-wise with some overlap. At the boundaries of the database, repeating end points are used to gain a margin and enough traces for the interpolation. Nevertheless, this introduces errors near to the surface and at the maximum depth, as well as at the ends of the distance range of the database.

```
set_local_interpolation ( nearest_neighbor | bilinear )
```

Set local interpolation method used during calculation of synthetic seismograms.

```
set_spacial_undersampling nxunder nzunder
```

Tell Minimizer to use only a subset of the databases Green's functions. nxunder: use every nxunder'th horizontal Green's function distance. nzunder: use every nzunder'th vertical Green's function depth.

```
set_receivers filename
```

Read a list of receiver coordinates from three column (lat lon components) ascii file filename.

The file format is as follows:

- first column: latitude in degrees
- second column: longitude in degrees
- third column: selected components of the station; for every component needed, add one of the following component characters:
radial component:
- $\mathrm{a}=$ positive is displacement away from source
- $\mathrm{c}=$ positive is displacement coming towards source
- transversal component:
- $\mathrm{r}=$ positive is rightwards seen from source
- l = positive is leftwards seen from source
- vertical component:
- $\mathrm{d}=$ positive is downwards
- $\mathrm{u}=$ positive is upwards
- horizontal component (north-south)
- $\mathrm{n}=$ positive is north
- $\mathrm{s}=$ positive is south
- horizontal component (east-west)
- $\mathrm{e}=$ positive is east
- $\mathrm{w}=$ positive is west

Adding the same component more than once is not allowed, so at most 5 components may be given. Lines starting with a ' $\#$ ' are considered to be comment lines.

An example receivers file might look like this:

```
# ok:
42.35 13.4 ard
49.78 17.54 ard
# north component broken:
45.49 25.95 ed
# only vertical component available
47.92 19.89 d
35.87 14.52 d
```

switch_receiver ireceiver ( on | off )

Turn receiver number ireceiver on or off.

```
set_ref_seismograms filenamebase format
```

Read a set of reference seismograms.
For every component at every of the receivers which have been set with set_receivers one file must be povided.
Currently the following formats are available:

- table: ASCII tables with two columns: time [s] and displacement [m].
- mseed: Single trace Data Only SEED Volume (Mini-SEED, http:// www.iris.edu/manuals/SEED_appG.htm.
- sac: SAC binary file. Please note, that this file format is platform dependant.

The files are expected to be named using the following scheme:
\$filenamebase-\$ReceiverNumber-\$ComponentCharacter. \$format where

- \$ReceiverNumber is the number of the receiver, as defined by the ordering of receivers in the receiver file (see set_receivers).
- \$ComponentCharacter is one of the characters defining receiver components as described in set_receivers.
shift_ref_seismogram ireceiver shift
Timeshift reference seismogram.
Shift reference seismogram at receiver number ireceiver by shift seconds.
autoshift_ref_seismogram ireceiver min-shift max-shift
Automatically timeshift reference seismogram.
Shift reference seismogram at receiver number ireceiver to where the cross-correlation has a maximum in the interval [min-shift,max-shift]. If ireceiver is set to zero, all seismograms are auto-shifted.

```
set_source_location latitude longitude reference-time
```

Sets the source location and reference time.

- latitude, longitude: Geographical coordinates of source reference point in [degrees]. All locations given in the source model description are measured relative to this reference point.
- reference-time: source reference time in seconds. All times given in the source model description are measured relative to this reference time.
set_source_constraints px1 pyl pz1 nx1 nyl nzl...
Set constraining planes which affect source geometry for certain source models.

Each constraining plane is defined by a point and a normal vector. They are specified in the local carthesian coordinate system at the source, which has its principal axes pointing north, east, and downward, and whose origin is at the surface at the coordinates given with set_source_location.

- px1 py1 pz1: coordinates of point for plane number 1 in [m]
- nx1 ny1 nz1: components of normal vector of plane number 1
set_source_crustal_thickness_limit thickness-limit
Limit crustal thickness at the source.
thickness-limit: Maximal thickness of crust in [m].
Default values for the thickness are retrieved from the crust $2 \times 2$ model.

```
get_source_crustal_thickness
```

Returns crustal thickness at the source in [m].

```
set_source_params source-type source-params ...
```

Sets the source type and parameters.
The available source types and a complete description of their parameters are given in the source type documentation. Short descriptions can be queried using the source_info tool.

This function detects if the same source parameters have already been set, so that seismograms are not recalculated when the same source is set several times.

```
set_source_params_mask mask ...
```

Select inversion parameters for the next minimization with minimize_lm. mask is built by giving a ' T ' or ' F ' for every source parameter of the source type that is currently in use. ' T ' makes the corresponding parameter an actual inversion parameter, ' $\mathrm{F}^{\prime}$ fixes the corresponding parameter to its current value. The 'T's and 'F's must be separated by whitespace.
The values of the selected parameters can be set using set_subparams and queried using get_subparams.
set_source_subparams subparams ...
Assignes values to the currently selected inversion parameters.
This command expects one value for each parameter selected with the command set_source_params_mask.
set_effective_dt effective_dt
Sets the effective dt controlling the source parameterization.

```
set_misfit_method ( l2norm | l1norm | ampspec_l2norm |
    ampspec_llnorm | scalar_product | peak )
```

Set the misfit calculation method.
Available methods are:

- 12 norm: L 2 norm is done on difference of time traces
- l1norm: L1 norm is done on difference of time traces
- ampspec_12norm: L2 norm is done on difference of amplitude spectra
- ampspec_11norm: L2 norm is done on difference of amplitude spectra
- scalar_product: instead of a norm, the scalar product is calculated
- peak: instead of a norm, the peak amplitudes are returned
set misfit_filter x0 y0 x1 y1 ...
Defines a piecewise linear function which is multiplied to the spectra before calculating misfits in the frequency domain.
- x0 y0 x1 y1 .... Control points with xi: frequency [Hz] and yi: multiplicator amplitude.

The amplitude drops to zero before the first and after the last control point. Example: set_misfit_filter 0.210 .51 defines a rectangular window between 0.2 and 0.5 Hz .

```
set_misfit_taper ireceiver x0 y0 x1 y1 ...
```

Defines a piecewise linear function which is multiplied to seismogram traces before calculating spectra or misfits.

- ireceiver: Number of the receiver to which the taper shall be applied.
- x 0 y 0 x 1 y 1 .... Control points with xi: time [s] and yi: multiplicator amplitude.

The amplitude drops to zero before the first and after the last control point.
Example: set_misfit_taper 12011501 defines a rectangular window between 120 and 150 s

```
set_synthecics_factor factor
```

Scale amplitude of synthecic seismograms by this factor during misfit calculation.

```
minimize_lm
```

Runs Levenberg-Marquardt minimization.
This tries to invert for the source parameters selected with the command set_source_params_mask by searching for a minimum in the currently selected misfit function, starting from the current source model parameterization.

This function makes use of $\operatorname{lmdif}()$ from MINPACK from Netlib.
Returns: info iterations misfit

- info: Information on convergence, returned by lmdif(). See MINPACK documentation for details.
- iterations: Number of function evaluations $=$ number of source models tested.
- misfit: Final global misfit.

```
output_source_model filenamebase
```

Output information about the current source model.

```
output_seismogram_spectra filenamebase
    (synthetics|references) (plain|filtered)
```

Output the seismogram spectra which are used during misfit calculation.

- If the first argument is references, the spectra of the reference seismograms are outputted, if it is synthetics, those of the synthetic seismograms for the current source model are outputted.
- If a filter has been set using set_misfit_filter, the filtered spectra are written.
- For every selected component at every defined receiver one file is written.

The files are named using the following scheme:
\$filenamebase-\$ReceiverNumber-\$ComponentCharacter.table where

- \$ReceiverNumber is the number of the receiver, as defined by the ordering of receivers in the receiver file (see set_receivers).
- \$ComponentCharacter is one of the characters defining receiver components as described in set_receivers.

```
output_seismograms filenamebase fileformat
    (synthetics|references) (plain|tapered|filtered)
```

Output current synthetic or reference seismograms.

- filenamebase: Stem for the creation of filenames, see below for details.
- fileformat: Format of the ouputted files.
- [ tapered ]: If this argument is present and a taper has been set using set misfit_taper, the tapered seismograms are written.

The files are named using the following scheme:

```
$filenamebase-$ReceiverNumber-$Component.$Format
```

where

- \$ReceiverNumber is the number of the receiver, as defined by the ordering of receivers in the receiver file (see set_receivers).
- \$Component is one of the characters defining receiver components as described in set_receivers.

```
get_source_subparams
```

Returns the current values of the source parameters selected with set_source_params_mask.
get_global_misfit
Returns the global misfit between the synthetic seismograms for the current source model and the reference seismograms.

```
get_misfits
```

Returns the misfit and normalization factors between the synthetic seismograms for the current source model and the reference seismograms.
Disabled stations are omitted in output list.
Returns: misfit-receiver1-component1
normfactor-receiver1-component1 ...
get_peak_amplitudes
Get the horizonal and vertical peak amplitudes of the synthetic traces.
Disabled stations are omitted in output list.
Returns: maxabs_receiver_1_horizontal
maxabs_receiver_1_vertical ...
get_principal_axes
Get the orientation of the principal axes P and T of the current source model.
Returns: p-axis-phi p-axis-theta t-axis-phi t-axis-theta

- p-axis-phi, p-axis-theta: Spherical coordinates of the direction of $P$.
- t-axis-phi, t-axis-theta: Spherical coordinates of the direction of T .

These are ordinary spherical coordinates based on the local carthesian north-east-down coordinate system at the source.

```
output_distances filename
```

Dump epicentral distances and azumiths to ascii file filename.

```
output_cross_correlations filenamebase shift-min shift-max
```

Output cross-correlations between synthetics and references

- filenamebase: Stem for the creation of filenames, see below for details.
- shift-min shift-max: Range of shifts for which cross-correlations are evaluated. (in [s]).

The files are named using the following scheme:
\$filenamebase-\$ReceiverNumber-\$Component. \$format
where

- \$ReceiverNumber is the number of the receiver, as defined by the ordering of receivers in the receiver file (see set_receivers).
- \$Component is one of the characters defining receiver components as described in set_receivers.


## get_cached_traces_memory

Get memory usage by Green's function database cache
Returns number of bytes allocated for traces in the Greens function database cache. This number does not contain the overhead of header data in the traces, and index tables. It is the plain number of bytes used to hold the seismogram traces.
set_cached_traces_memory_limit nbytes
Set maximum memory usage by Green's function database cache
Sets the approximate maximum of memory used by the Greens function database cache. This limit does not include the overhead of header data in the traces, and index tables. It is the plain number of bytes which the Green's function database is allowed to use to cache seismogram traces.

```
set_verbose (T|F)
```

Toggle verbose operation.
T turns on verbose operation. F turns it off.

## A.1.3.4 Example

To calculate synthetic seismograms at 11 receivers for the Izmit event:

```
# setup receivers, indicating that for each receiver
# north, east and down components should be calculated.
> cat >izmit-receivers.table <<EOF
42.350 13.400 ned
49.780 17.540 ned
45.490 25.950 ned
47.920 19.890 ned
35.870 14.520 ned
34.960 33.330 ned
35.280 24.890 ned
35.180 25.500 ned
49.630 22.710 ned
36.370 25.460 ned
42.620 23.240 ned
EOF
> minimizer <<EOF
set_database /gfdb/gemini-prem/db
set_effective_dt 0.5
set_local_interpolation bilinear
set_receivers izmit-receivers.table
set_source_location 40.75 29.86 0
set_source_params bilateral 0 0 0 10000 2e20 91 87 164 0 40000 20000 18000 3500 2
output_seismograms izmit-seismogram mseed synthetics plain
EOF
```


## A.1.4 The Autokiwi tool

This command line tool is a frontend for aquisition of event-data, preprocessing, processing, and posting of results. For acquisition and preprocessing the Pyrocko library is used. For processing, currently, a user provided program is called. The basic functionality of Autokiwi can be extended with plugins.

## A.1.4.1 Usage

```
> autokiwi --help
Usage: autokiwi [options] command[,command2[,...]] [args]
autokiwi [options] pull [ first | all | eventname ]
autokiwi [options] prepare ( all | eventnames ... )
autokiwi [options] process ( all | eventnames ... )
autokiwi [options] report ( all | eventnames ... )
autokiwi [options] post ( all | eventnames ... )
autokiwi [options] list
Options:
    -h, --help show this help message and exit
    --loglevel=LOGLEVEL set logger level to "error", "warning", "info", or
        "debug". Default is "warning".
    --config=CONFIG_FILENAME
        set name of config file to use. Default is
        "__search_parent_dirs_for_autokiwi.conf__"
    --force force pulling of events that have already been
        downloaded
```


## A.1.4.2 Autokiwi subcommands

Autokiwi has several subcommands which can be run one-by-one or grouped. The built-in subcommands are:

```
pull
```

Retrieve event data from configured source.

```
prepare
```

Preprocess data for use with the Kiwi Tools.

## process

Run user-supplied program to process the event data.

```
report
```

Run user-supplied program to generate a report on the processing output.

```
post
```

Post results somewhere.

## list

List the events Autokiwi is aware of.

## A.1.4.3 Autokiwi configuration file

If no explicit configuration filename is given using the --config option, Autokiwi looks for a file named autokiwi.conf, first in the current directory, then in its parent directories.

The following is an example configuration for Autokiwi .

```
#!/usr/bin/env python
This is a configuration file for autokiwi.
It is a Python script, so that autokiwi can be configured in a very flexible
# way. In this example it contains mainly simple assignments, so that no
# deeper knowledge of Python is required to modify it.
All values are to be given in SI units, unless stated explicitely.
This configuration is built from several sub-configurations, each containing
key-value pairs. A sub-configuration can be made to extend another one, by
using a "base" entry, having as value the sub-configuration to be extended.
```

```
# All intermediate directories in pathnames given in this file are created as
# needed. Any pathname given in this configuration can be made to extend
# other pathnames, by using python string interpolation ("%(KEY)s") like in the
# following example:
base_dir = '/kinherd'
plugins_dir = '%(base_dir)s/plugins', # evaluates to: '/kinherd/plugins'
All string interpolation is postponed to when the path is actually needed, it
# is not done when the configuration file is processed (so ordering, for
# example, does not matter).
from tunguska.configurator import Config
from tunguska.phase import Timing, Taper
# Some unit multiplicators, to make this file more readable:
km = 1000.
m=60.
h = 3600.
days}=60*60*2
base_config = Config(
    # autokiwi puts all its stuff into subdirectories of base_dir
    base_dir = '/kinherd/fullwave_geofon_100',
    # for each event a direcory is created
    event_dir = '%(base_dir)s/events/%(event_name)s',
    # this tells 'edump' where to put data and 'prepare', where to look for it
    edump_data_dir = '%(event__dir)s/data',
    # where to look for plugins (not needed here)
    #plugins_dir = '%(base_dir)s/plugins',
)
# This part of the configuration tells autokiwi to get Geofon data through
# SeedLink/ArcLink through SeisComP3.
edump_config = Config(
    # Extends 'base_config' configuration
    base = base_config,
    # Get event descriptions through GEOFON online catalog
    use_geofon_catalog = True,
    # Tells autokiwi where to look for stream badnesses. Stream badnesses are
    # quality indicators for all/some streams in the network. A stream is
    # the data from a specific comonent of a specific station.
    # Autokiwi expects badnesses in a simple text file with two columns
    # separated by white-space. The first column is the name of the stream, as
    # dot-separated string of the form NETWORK.STATION.LOCATION.CHANNEL, the
    # second column contains the corresponding badness values.
    # The filenames in 'streams_badness_dir' must be of the form
    # 'badness_START_END', where START and END are of the form
    # YEAR-MONTH-DAY-HOUR-MIN-SEC.
    # Example File: badness_2010-03-05_11-47-09_2010-04-13_23-49-37
    # GE.RGN..BHE 0.981431
    # GE.KBS.OO.BHE 99
    # IA.PCI..BHE }9
    # ...
    streams_badness_dir = '%(base_dir)s/badness',
```

```
    # Only download data from streams with a badness smaller or equal to this
    streams_badness_limit = 1.0
    # Time range in which to look for matching events
    time_range = (('now', -31*days), ('now', -1*h)),
    # A filter defining what kind of events should be downloaded
    event_filter = lambda ev: 6.5 <= ev.magnitude,
    # A filter defining which stations should be included
    station_filter = lambda sta: 3.0 <= sta.dist_deg and sta.dist_deg <= 120.,
    # List of channel names to look for
    channels = ['BHZ', 'BHN', 'BHE'],
    # Data time window to download, relative to event time
    time_window = (-20*m, 120*m),
    # Edump arguments (SeisComP3 Host to connect to)
    argv = [ '--host', 'geofon-cluster', '--debug' ],
    # Up to this many stations are selected; stations are selected
    # automatically, based on azimuth, distance, and stream badness
    nwanted_stations = 100
)
# This part of the configuration tells autokiwi how to prepare data
prepare_config = Config(
    base = base_config,
    # If the following two variables are defined, station selection based on
    # badness is applied in the 'prepare' step of autokiwi. This selection can
    # also be made during acquisition
    #streams_badness_dir = '%(base_dir)s/badness',
    #streams_badness_limit = 1.0,
    # Frequency band for restitution. This defines a cosine flanked taper, which
    # is applied to the seismogram spectra
    restitution_frequencyband = (0.001, 0.002, 0.05, 0.1),
    # Cosine shaped fadein/fadout time applied to the traces before restitution
    restitution_fade_time = 2./0.002,
    # List of allowed restitution methods, in the order they are tried.
    # Possible values for data downloaded through SeisComP3 are: 'polezero' and
    # 'integration'.
    restitution_methods = [ 'polezero' ],
    # Cut seismograms to given span. The times are offsets to given phases,
    # the phases are defined in $KIWI_HOME/aux/phases.
    cut_span = [Timing('begin', -2.*m), Timing('end', +20.*m)],
    # Exclude traces with aplitudes larger than this value.
    displacement_limit = 1.,
    # If present, prepare data to be used with this database. This mainly sets
    # the sampling rate to which the traces are downsampled.
    gfdb_path = '/kinherd/gfdb/gemini-iasp91-20000km-0.2hz/db',
    # Exclude traces from stations which are outside of this margin from the
    # ends of the database. This is done so that stations near to the ends of
    # the database do not cause problems for extended sources.
    gfdb_margin = 150*km,
```

```
    # Defines component projections. Using the following setup tells autokiwi to
    # project raw traces to east-north-up orientation. The channel names given
    # are the input and output channel names to be used.
    projection_functions = [ lambda station:
    station.projection_to_enu(('BHE','BHN','BHZ'), ('E','N','Z')) ],
    # Defines rotations of horizontal components. Using the following setup
    # tells autokiwi to rotate to radial(away)-transversal(right) components.
    rotation_functions = [ lambda station:
        station.rotation_ne_to_rt(('N', 'E'), (' R', 'T')) ],
)
# Preprocessing configuration specific to setup an event directory for use with
# Kiwi inversion scripts.
kiwi_config = Config(
    base=prepare_config
    # Shift traces so that a time of zero is at the preliminary time origin of
    # the event. (The other possibility is 'system').
    trace_time_zero = 'event',
    # Scale traces by this factor
    trace_factor = 1.0e9,
    # List of channels to be selected for the Kiwi inversion
    wanted_channels = ['Z', 'R', 'T'],
    # How to rename the channels for Kiwi internal
    kiwi_component_map = {
        'Z' : 'u',
        'N': 'n'
        'E': 'e'
        'T': 'r'
        'R': 'a'
    },
    # Each station can be made to appear multiple times, so that it is
    # possible to treat different phases as individual datasets.
    # This value defines by how many stations should be made out of each.
    nsets = 1,
    station_splitting = ['ura'],
    # Files from this directory are copied into the Kiwi main directory
    skeleton_dir = '%(base_dir)s/skeleton/kiwi',
    # Main directory for the Kiwi inversion
    main_dir = '%(event_dir)s/kiwi',
    # Where to put the data for the Kiwi inversion
    data_dir = '%(main_dir)s/data',
    # Filenames to be used
    stations_path = '%(data__dir)s/stations.table',
    receivers_path = '%(data_dir)s/receivers.table',
    source_origin_path = '%(data_dir)s/source-origin.table',
    event_info_path = '%(data_dir)s/event.txt',
    reference_time_path = '%(data_dir)s/reference-time.txt',
    displacement_trace_path = \
                            \prime%(data_dir)s/reference-%(ireceiver)s-%(component)s.mseed',
    # Command executed in main_dir to run the processing
    processing_command = ['./kiwi', 'work', ' -', 'durationfinder2'],
```

```
    # command executed in main_dir to generate the report
    report_command = ['./kiwi', 'report'],
)
# Configuration of where to put the processing results
post_config = Config(
    base = kiwi_config,
    # Directory of results to be posted
    source_dir = '%(main_dir)s/report',
    # Where to put the results
    target_host = '',
    target_dir = '/kinherd/web/reports-import/%(event_name)s',
    # This command is executed after copying the files with rsync
    trigger_command = [
        '/kinherd/web/code/kinherd/import.py',
        '%(target_dir)s'
    ],
)
```


## Appendix B

## The DFG project KINHERD

This work has been funded by DFG project KINHERD (DA478/14-1/2), partly in cooperations with the BMBF/DFG (BMBF07/343) "Geotechnologien" project RAPID.

A project web page is available at http://kinherd.org/.
The following participants collaborated within KINHERD

- Simone Cesca (University of Hamburg) - development and testing of inversion schemes, application to regional events [Cesca et al., 2010], rapid directivity detection [Cesca et al., 2010, submitted]
- Torsten Dahm (University of Hamburg) - concept, supervision, project applicant, principal investigator
- Frank Krüger (University of Potsdam) - concept, supervision, project applicant
- Francesco Pacchiani - Green's function interpolation
- Klaus Stammler (BGR) - technical infrastructure, supervision, project applicant
- Thomas Plenefisch (BGR) - supervision
- Rainer Kind (GFZ Potsdam) - project applicant


## Appendix C

## Derived work

The software tools developed in the scope of this work, have been successfully applied and presented in several cases:

- Published article: Automated procedure for point and kinematic source inversion at regional distances [Cesca et al., 2010]
My contribution to this work has been the development of the core tools used for the handling of Green's function databases, synthetic seismogram generation, misfit calculation and inversion.
- Published article: A seismological study of shallow weak earthquakes in the urban area of Hamburg city, Germany, and its possible relation to salt dissolution [Dahm et al., 2010]

My contribution to this work has been the development of all software tools needed for the inversion for a partial moment tensor which is presented in the article, and the tools to calculate shakemaps by forward modeling based on pre-calculated Green's functions.

- Submitted article: Rapid directivity detection by azimuthal amplitude spectra inversion [Cesca et al., 2010, submitted],
My contribution to this work has been the implementation of the core functionality to forward model synthetic seismograms based on pre-calculated Green's functions.
- Bachelor thesis: Source modelling of the 2001 Ekofisk, North Sea, induced earthquake, [Juretzek, 2009]

The inversions were done using the Kiwi Tools.

- International workshop for applying the Kiwi Tools for moment tensor and kinematic invesion, has been held at Hamburg University from 19.10.2009 20.10.2009. Two groups, one from Tessaloniki, Greece and another from Lissabon, Portugal plan to integrate the Kiwi Tools in their routine processing of regional earthquakes.
- The Kiwi Tools have been integrated for routine processing at the BGR, Hannover (SZGRF) and GFZ, Potsdam (GEOFON) and is currently in a testing stage. I have coded the interfaces and implemented the methods.


## Appendix D

## List of events (point source inversion)

On the following pages, the list of events used for the point source catalog comparison (section 5.4 ) is given.

| Time | Latitude | Longitude | Depth/km | $M_{W}$ | Moment/Nm | Strike | Dip | Slip-Rake | Duration | Misfit | Stations | Region |
| :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: | :---: |
| 2010-06-12 19:26:56 | 7.670 | 91.816 | 43 | 7.49 | $1.89 \mathrm{e}+20$ | 114 | 74 | 155 | 16 | 0.6301 | 112 | Nicobar Islands, India Region |
| 2010-06-05 05:22:01 | 43.513 | 147.017 | 59 | 5.44 | $1.60 \mathrm{e}+17$ | 207 | 14 | 69 | 4 | 0.9716 | 78 | Kuril Islands |
| 2010-06-03 09:23:56 | 5.844 | 97.050 | 73 | 5.71 | $4.08 \mathrm{e}+17$ | 314 | 87 | 28 | 30 | 0.9658 | 83 | Northern Sumatra, Indonesia |
| 2010-06-02 01:49:41 | -57.679 | -26.467 | 57 | 6.06 | $1.40 \mathrm{e}+18$ | 352 | 84 | -74 | 12 | 0.9398 | 60 | South Sandwich Islands Region |
| 2010-05-31 10:15:59 | 7.014 | 124.172 | 19 | 5.79 | $5.41 \mathrm{e}+17$ | 342 | 62 | -30 | 10 | 0.8393 | 75 | Mindanao, Philippines |
| 2010-05-28 04:25:55 | -14.766 | 166.908 | 87 | 6.18 | $2.09 \mathrm{e}+18$ | 248 | 43 | 78 | 30 | 0.9484 | 44 | Vanuatu Islands |
| 2010-05-27 20:48:05 | -13.507 | 166.159 | 98 | 6.22 | $2.42 \mathrm{e}+18$ | 252 | 44 | 98 | 30 | 0.9644 | 40 | Vanuatu Islands |
| 2010-05-27 17:14:49 | -13.899 | 166.853 | 34 | 7.09 | $4.80 \mathrm{e}+19$ | 187 | 42 | 89 | 14 | 0.6639 | 39 | Vanuatu Islands |
| 2010-05-26 08:53:10 | 25.840 | 129.813 | 5 | 6.44 | $5.09 \mathrm{e}+18$ | 75 | 69 | -58 | 9 | 0.6754 | 82 | Southeast of Ryukyu Islands |
| 2010-05-25 10:09:07 | 35.125 | -35.825 | 8 | 6.20 | $2.25 \mathrm{e}+18$ | 99 | 86 | 4 | 13 | 0.8089 | 75 | Northern Mid Atlantic Ridge |
| 2010-05-23 22:46:51 | -13.936 | -74.438 | 95 | 6.04 | $1.31 \mathrm{e}+18$ | 140 | 62 | -94 | 9 | 0.8222 | 74 | Central Peru |
| 2010-05-21 18:52:05 | -35.043 | -71.365 | 99 | 5.80 | $5.57 \mathrm{e}+17$ | 229 | 62 | -60 | 28 | 0.9326 | 45 | Near Coast of Central Chile |
| 2010-05-09 05:59:49 | 3.894 | 95.981 | 34 | 7.12 | $5.45 \mathrm{e}+19$ | 127 | 69 | 79 | 13 | 0.5666 | 107 | Northern Sumatra, Indonesia |
| 2010-04-26 02:59:52 | 22.206 | 123.784 | 3 | 6.45 | $5.24 \mathrm{e}+18$ | 62 | 77 | -147 | 10 | 0.5827 | 215 |  |
| 2010-04-23 10:03:06 | -37.646 | -72.928 | 21 | 5.95 | $9.28 \mathrm{e}+17$ | 135 | 76 | 76 | 5 | 0.4569 | 17 |  |
| 2010-03-16 02:21:58 | -36.229 | -73.321 | 18 | 6.47 | $5.63 \mathrm{e}+18$ | 172 | 75 | 86 | 1 | 0.6725 | 106 | near-coast-of-central-chile |
| 2010-03-15 11:08:30 | -35.800 | -73.341 | 4 | 6.36 | $3.91 \mathrm{e}+18$ | 189 | 84 | 91 | 5 | 0.7466 | 108 | off-coast-of-central-chile |
| 2010-03-14 20:33:14 | -2.613 | 83.723 | 32 | 5.57 | $2.58 \mathrm{e}+17$ | 244 | 34 | 88 | 7 | 0.7685 | 116 | south-indian-ocean |
| 2010-03-14 08:08:04 | 37.708 | 141.911 | 33 | 6.42 | $4.85 \mathrm{e}+18$ | 196 | 22 | 83 | 6 | 0.6072 | 155 | near-east-coast-of-honshu-japan |
| 2010-03-14 00:57:45 | -1.752 | 128.205 | 53 | 6.40 | $4.52 \mathrm{e}+18$ | 136 | 62 | 19 | 7 | 0.6407 | 124 | halmahera-indonesia |
| 2010-03-11 15:06:31 | -34.200 | -72.605 | 49 | 6.50 | $6.25 \mathrm{e}+18$ | 118 | 53 | 127 | 9 | 0.9971 | 108 | near-coast-of-central-chile |
| 2010-03-11 14:55:30 | -34.291 | -71.727 | 6 | 6.96 | $3.12 \mathrm{e}+19$ | 156 | 87 | -91 | 8 | 0.8928 | 109 | near-coast-of-central-chile |
| 2010-03-11 14:39:48 | -34.388 | -71.735 | 27 | 6.64 | $1.04 \mathrm{e}+19$ | 283 | 38 | -92 | 5 | 0.6335 | 115 | near-coast-of-central-chile |
| 2010-03-05 16:07:03 | -3.871 | 100.978 | 31 | 6.52 | $6.77 \mathrm{e}+18$ | 127 | 76 | 88 | 11 | 0.6114 | 66 | southwest-of-sumatera-indonesia |
| 2010-03-05 11:47:09 | -36.694 | -73.133 | 21 | 6.49 | $6.18 \mathrm{e}+18$ | 20 | 19 | 108 | 3 | 0.6214 | 45 | near-coast-of-central-chile |
| 2010-03-05 09:19:33 | -36.692 | -73.486 | 19 | 5.83 | $6.22 \mathrm{e}+17$ | 319 | 13 | 67 | 1 | 0.6782 | 47 | near-coast-of-central-chile |
| 2010-03-04 01:59:48 | -33.156 | -72.138 | 12 | 5.89 | $7.72 \mathrm{e}+17$ | 199 | 75 | 37 | 1 | 0.8811 | 51 | off-coast-of-central-chile |
| 2010-03-04 00:18:52 | 22.951 | 120.872 | 23 | 6.16 | $1.97 \mathrm{e}+18$ | 312 | 35 | 47 | 8 | 0.7229 | 70 | taiwan |
| 2010-02-28 11:25:35 | -34.927 | -71.700 | 45 | 6.18 | $2.09 \mathrm{e}+18$ | 1 | 29 | 106 | 6 | 0.6993 | 101 | chile-argentina-border-region |
| 2010-02-27 18:59:40 | -33.423 | -71.921 | 98 | 6.04 | $1.29 \mathrm{e}+18$ | 255 | 42 | -156 | 1 | 0.9712 | 99 | near-coast-of-central-chile |
| 2010-02-27 15:45:00 | -25.935 | -64.632 | 56 | 6.27 | $2.89 \mathrm{e}+18$ | 326 | 35 | -73 | 29 | 0.9832 | 97 | salta-province-argentina |
| 2010-02-27 08:24:56 | -35.277 | -72.510 | 66 | 7.18 | $6.63 \mathrm{e}+19$ | 160 | 44 | 116 | 19 | 0.9831 | 97 | near-coast-of-central-chile |
| 2010-02-27 08:01:49 | -38.514 | -75.413 | 4 | 7.12 | $5.45 \mathrm{e}+19$ | 33 | 72 | -177 | 27 | 0.9928 | 98 | off-coast-of-central-chile |
| 2010-02-27 07:36:49 | -37.077 | -72.003 | 65 | 7.11 | $5.21 \mathrm{e}+19$ | 330 | 59 | -103 | 30 | 0.9886 | 98 | near-coast-of-central-chile |
| 2010-02-27 07:12:55 | -33.719 | -70.843 | 19 | 7.18 | $6.52 \mathrm{e}+19$ | 166 | 45 | -96 | 30 | 0.9888 | 97 | near-coast-of-central-chile |
| 2010-02-27 06:53:04 | -35.930 | -74.142 | 95 | 7.66 | $3.46 \mathrm{e}+20$ | 159 | 52 | 85 | 16 | 0.9931 | 91 | near-coast-of-central-chile |
| 2010-02-27 06:35:28 | -35.744 | -72.550 | 13 | 8.84 | $2.06 \mathrm{e}+22$ | 11 | 12 | 98 | 90 | 0.7236 | 53 | near-coast-of-central-chile |
| 2010-02-07 06:09:59 | 23.161 | 123.764 | 27 | 6.23 | $2.47 \mathrm{e}+18$ | 141 | 67 | 21 | 7 | 0.7480 | 75 | southwestern-ryukyu-islands |
| 2010-02-06 04:45:00 | 46.772 | 152.768 | 31 | 5.90 | $8.06 \mathrm{e}+17$ | 92 | 68 | 106 | 14 | 0.7154 | 80 | kuril-islands |
| 2010-02-05 06:59:05 | -47.880 | 99.591 | 1 | 6.13 | $1.74 \mathrm{e}+18$ | 306 | 78 | 21 | 1 | 0.7281 | 42 | southeast-indian-ridge |
| 2010-02-04 20:20:21 | 40.503 | -124.931 | 7 | 5.71 | $4.15 \mathrm{e}+17$ | 224 | 61 | 20 | 1 | 0.8874 | 77 | near-coast-of-northern-calif |
| 2010-02-01 22:28:17 | -6.186 | 154.439 | 33 | 6.08 | $1.47 \mathrm{e}+18$ | 134 | 47 | 89 | 6 | 0.7282 | 68 | solomon-islands |
| 2010-01-29 09:19:31 | -18.840 | 169.713 | 15 | 5.56 | $2.49 \mathrm{e}+17$ | 303 | 33 | -100 | 8 | 0.9506 | 48 | vanuatu-islands |


| philippine－islands－region |
| :--- |
| haiti－region |
| drake－passage |
| south－of－mariana－islands |
| haiti－region |
| near－coast－of－northern－calif |
| east－of－south－sandwich－islands |
| solomon－islands |
| solomon－islands |
| solomon－islands |
| south－of－mariana－islands |
| banda－sea |
| tanzania |
| taiwan |
| loyalty－islands－region |
| mindanao－philippine－islands |
| kermadec－islands－new－zealand |
| south－of－sumbawa－indonesia |
| tonga－islands |
| queen－charlotte－islands－region |
| near－coast－of－northern－chile |
| nicobar－islands－india |
| sumbawa－region－indonesia |
| azores－islands－region |
| south－of－tonga－islands |
| ryukyu－islands |
| drake－passage |
| santa－cruz－islands |
| irian－jaya－region－indonesia |
| south－of－panama |
| samoa－islands－region |
| sunda－strait |
| sulawesi－indonesia |
| samoa－islands－region |
| fox－islands－aleutian－islands |
| halmahera－indonesia |
| fox－islands－aleutian－islands |
| santa－cruz－islands |
| mauritius－reunion－region |
| santa－cruz－islands |
| santa－cruz－islands |
| santa－cruz－islands |
| vanuatu－islands |
| santa－cruz－islands |
















高家家家首


vanuatu-islands
taiwan
tonga-islands
southern-sumatera-indonesia
 southern-sumatera-indonesia
samoa-islands-region
ff-coast-of-jalisco-mexico off-coast-of-jalisco-mexico
macquarie-islands-region bhutan
mindoro mindoro-philippine-islands
easter-island-region

 | $n$ |
| :---: |
| 3 |
| 0 |
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| 1 |
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| 0 |
| 0 |
| 3 |
| 3 |
| 0 | south-of-jawa-indonesia

jawa-indonesia

年
 southern-sumatera-indonesia south-of-honshu-japan
near-s-coast-of-honshu-japan near-s-coast-of-honshu-japan
andaman-islands-india santa-cruz-islands
southwestern-ryukyu-islands southwestern-ryukyu-islands
gulf-of-california
 irian-jaya-region-indonesia
southern-east-pacific-rise southern-east-pacific-rise
southern-sumatera-indonesia
 new-ireland-region-png
off-w-coast-of-s-island-nz
 sumba-region-indonesia
baffin-bay rat-islands-aleutian-islands rat-islands-aleutian-islands
panama
gulf-of-california苋
new-ireland-region-png
bouvet-island-region














mindanao-philippine-islands
vanuatu-islands
northern-mid-atlantic-ridge
hokkaido-japan-region
prince-edward-islands-region
vanuatu-islands
north-of-honduras
kermadec-islands-region
kermadec-islands-region
new-britain-region-png
off-coast-of-ecuador
guerrero-mexico
kuril-islands
near-coast-of-northern-chile
south-sandwich-islands-region
solomon-islands
southern-sumatera-indonesia
kuril-islands
central-italy
kyushu-japan
philippine-islands-region
near-n-coast-of-new-guinea-png
near-north-coast-of-irian-jaya
tonga-islands-region
talaud-islands-indonesia
pacific-antarctic-ridge
south-of-panama
nort-of-svalbard
south-sandwich-islands-region
south-of-mariana-islands
talaud-islands-indonesia
kermadec-islands-region
kermadec-islands-new-zealand
near-coast-of-northern-peru
kermadec-islands-region
talaud-islands-indonesia
talaud-islands-indonesia
talaud-islands-indonesia
talaud-islands-indonesia
talaud-islands-indonesia
near-coast-of-northern-peru
near-coast-of-peru
near-east-coast-of-honshu-japan
southern-sumatera-indonesia
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tajikistan-xinjiang-border-reg
kermadec-islands-new-zealand
loyalty-islands-region
kermadec-islands-new-zealand
mindoro-philippine-islands
xizang
off-coast-of-jalisco-mexico
central-east-pacific-rise
hokkaido-japan-region
halmahera-indonesia
central-mid-atlantic-ridge
southern-iran
solomon-islands
vanuatu-islands
eastern-new-guinea-reg-png
north-of-ascension-island
vancouver-island-region
lake-baykal-region-russia
xizang
mauritius-reunion-region
myanmar-china-border-region
tonga-islands
santa-cruz-islands
central-mid-atlantic-ridge









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[^0]:    ${ }^{1}$ Attributing to their derivation from kinematic models, based on the observed motion, as opposed to dynamic models, which try to integrate the causes of the motion, the physics of the rupture process.

[^1]:    ${ }^{1}$ An algorithm to compose seismograms by superposition of Green's function traces has been implemented in the Kiwi core module seismogram. (See appendix A.1)

[^2]:    ${ }^{2}$ Discretization schemes for different source models into sub-fault centroids has been implemented in the the Kiwi core modules source_eikonal, source_bilateral, source_circular, amongst others (see A.1).

[^3]:    ${ }^{3}$ The API to read and write Green's function databases and the Green's function cache is implemented in the Kiwi core module $\mathrm{g} f \mathrm{db}$ (see A.1).

[^4]:    ${ }^{4}$ In collaboration with Francesco Pacchiani, we have implemented Gülünay's interpolation method for spatially aliased data in the Kiwi tools. The implementation can be found in the Kiwi core module interpolation. The module for accessing Green's function databases $\mathrm{g} f \mathrm{db}$ can transparently make use of this feature, providing on-the-fly interpolation of Green's function traces.
    ${ }^{5}$ The Kiwi tools provide a customizable, modular, and efficient misfit calculation engine, supporting tapers, filtering and different norms. The Kiwi core module comparator provides an API to compare seismograms and calculate misfits (see A.1)

[^5]:    ${ }^{1}$ The taper $W$ was chosen to be flat between 0.01 and 1 Hz , i.e. optimized for events with a rupture duration of less than 50 s .

[^6]:    ${ }^{2}$ Another simple possibility for crustal earthquakes would be to allow rupture to happen only within the crust, as given by e.g. CRUST 2.0

