The Complexity of Teleseismic P-Waves

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Abstract

Complex short-period teleseismic P-waveforms (consisting of the direct P wave and surface reflections pP and sP) are observed from many earthquake sources. It is often not possible to easily interpret these waveforms in terms of those three phases. This is necessary to obtain accurate earthquake depths and P and S wave radiation patterns. This thesis examines the contribution made by various factors to P-wave seismogram complexity using both synthetic and real data.

First, using a number of synthetic waveforms it is confirmed that long duration sources can contribute significantly to the complexity of short-period waveforms. However, it is highlighted that by using broadband recordings much of this complexity can be accounted for, and attributed to the limited passband of the short-period recording system. In addition, S-to-P mode conversions at near-source structure can also contribute significantly to the complexity of the short-period waveform.

Second, the causes of differences in the complexity of the short-period waveforms from the 1987 Whittier Narrows and the 1991 Sierra Madre earthquakes are examined. Originally these earthquakes were thought to be separated by a distance approximately the size of the first Fresnel Zone, and hence should, in theory, have indistinguishable near-source structure, when seen at teleseismic distances. Using relative amplitudes, the published CMT focal mechanism for these events is confirmed. In the case of Sierra Madre earthquake it was also possible to positively identify the one surface reflection, visible on the short-period seismogram, as pP. Even with complex waveforms the relative amplitude method can be used to place constraints on the focal mechanism of the Whittier Narrows earthquake. Using forward modelling, with a simple kinematic source
model, synthetic seismograms are matched to the observed broadband seismograms for both earthquakes. Using this simple source model, the variation in the source duration, caused by the difference in source rupture areas between the two earthquakes, is sufficient to account for the first-order variations in complexity seen. To second order, the near-source structure is sufficiently different even at the limit of resolution of the data, to contribute to some extent to the observed complexity variation, most likely due to the large thickness of sediment west of the Sierra Madre Fault.

Third, a suite of seismograms from the 29 October 1995 Caspian Sea earthquake is examined. Using relative amplitudes, the surface reflection on these seismograms is correctly identified and the actual depth estimated to be 48 km. From this it is shown that an arrival mis-identified by the Prototype International Data Center as a surface reflection is most likely to be a mode conversion at an interface 80 km beneath the source. Forward modelling of the broadband and short-period waveforms shows that these mode-conversions are enhanced by the downward propagating line rupture, and are best seen when the position of the stations are at a node in the P-wave radiation pattern. This produces an apparently complex waveform. Visible S waves from this earthquake at European stations show the very low attenuation in the mantle path and this may contribute to the greater than usual complexity observed for this event.

Finally several earthquakes that appear to show seismogram complexity that cannot be explained using a simple kinematic source model or path effects are examined. By modifying an existing finite-difference fault modelling code I present a possible dynamic source model that may provide one explanation for this additional complexity. This model includes real source physics (friction law, rupture criteria) and material heterogeneities. It produces complex far-field pulse shapes that vary with fault length, material heterogeneity, initial state of stress and attenuation.
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Chapter 1

Introduction

Our understanding of the internal structure of the Earth is predominantly based on the interpretation of earthquake seismograms. Similarly, the source of a seismic disturbance is interesting in its own right, both for the study of global tectonics, and the practical need to correctly distinguish between explosion and earthquake sources. The latter became increasingly important after the commencement of underground nuclear testing in the early 1960s, when seismology was identified as the only way that this could be done successfully.

Thirlaway (1963) was one of the first to investigate the problem of discrimination of earthquakes from nuclear explosions. He noted that the arrival of energy in the first minute of the seismogram after the first motion, i.e. the P-wave section, provides critical information about the nature of the seismic disturbance at the source. Figure 1.1 shows a cartoon of a typical short-period seismogram of the P-wave section, and the ray paths of three important phases that it may contain: direct P, and the surface reflections pP and sP. In principle, the best resolution within this P-wave section can be obtained at high frequencies, but in practice such short-period seismograms have often proved difficult to interpret. This thesis examines the general problem of interpreting the P-wave seismogram, using a variety of data types and techniques to resolve the factors that contribute to the observed complexity, particularly of short-period seismograms.

In the early history of nuclear discrimination it had been thought that it was necessary to have recordings close to the event. However, in 1962 seismic signals
CHAPTER 1. Introduction

Near-source structure above source
Near-source structure beneath the source
Seismograph

Figure 1.1.
Ray geometry of the phases comprising the P-wave section of the short-period teleseismic earthquake seismogram and the zones along the ray path that contribute to the shape of the short-period seismogram shown. The shallow earthquake depth controls the timing of the phases P, pP and sP shown. (after Pearce & Rogers (1994))

Figure 1.2.
Sketch of amplitude vs distance curve (Gutenberg and Richter) for P-phase and the almost flat amplitude response at teleseismic distances in the source window. (Bob Pearce, pers.comm.)
from fully-coupled underground nuclear explosions of a few kilotons, showed clear P at a distance of 10000 km (epicentral distance of \(\approx 90^\circ\)) from the source. These seismograms were recorded at the maximum epicentral distance where the path to the receiver is mainly through the lower mantle, with little crustal component. In fact the shape and amplitude of the seismic signal is similar from 30 to 90° (figure 1.2). This plot is a function of the Earth's velocity and attenuation structure, showing large fluctuations associated with the triplication of P at regional distances. The flat response within the 'source window' defined at teleseismic distance (from 30 to 90°), is also evidence for the low attenuation in the mantle for deep ray paths. As a consequence, seismograms are easier to interpret at teleseismic distances, where much of the early work in seismic source identification effort was concentrated. In fact, teleseismic data proved useful in source identification for sources as small as \(m_b=4.0\) (Evernden 1969).

Short-period recordings at frequencies of 1 Hz are necessary to resolve P, pP and sP from a seismic source of depths down to 1.5 km. This represents one quarter of a wavelength for a typical crustal velocity of 6 kms\(^{-1}\). At teleseismic distances, earthquake P-seismograms often appear relatively complex on the short period record, with many uninterpretable phases. In contrast, those from explosions initially appeared more simple, with only one phase of relatively large amplitude, followed by a coda of much lower amplitude. This is because explosion sources have a very short source duration, and produce little S energy. They are also at a shallow depth, so surface reflections are not identifiable. A United Kingdom Atomic Energy Authority report (UKAEA 1965) noted that 90% of earthquakes could be distinguished from explosions using this complexity criterion alone.

However, as examples of complex seismograms from underground nuclear tests, such as at Yellowknife (YKA), Canada from Novaya Zemlya in the Russian Arctic (see section 2.4 for more detail) became available, it was realised that P-wave complexity alone could not act as a reliable discriminant. Nevertheless, detailed information about the earthquake source process, depth, and focal mechanism is contained in the P-wave seismogram. As Douglas (1981) comments, it is only in understanding how the Earth and the source affect the short-period P signal to produce a complex seismogram, that an assessment can be made of how this will limit our ability to identify a source as an earthquake or explosion. This thesis attempts to identify and assess using a variety of case studies and forward
The causes of P-wave teleseismic seismogram complexity fall into two main categories: the Earth structure along the path from source to the receiver and the rupture process itself. Although to the first order the ray path, especially in the lower mantle, appears homogeneous, Douglas, Young & Hudson (1974) showed that for a simple earthquake source, complex near-source structure can generate complex teleseismic seismograms by multiple reflections and mode-converting of S wave energy to P energy. Douglas et al. (1973b) also noted that a zone of local attenuation above or below the source can reduce the size of one phase on a seismogram with respect to other phases and make a seismogram appear complex, although high attenuating paths also filter out high frequency energy thereby also reducing the complexity of the signal. However, the existence of simple seismograms from both explosion and earthquake sources confirm that for many regions of the world, the path effects do not vary over a wide area or are negligible (Husebye & Janson 1964, Pearce 1996).

In the source rupture process, both geometric and dynamic effects contribute to the complexity of the source-time function that can be resolved at teleseismic distances. Geometric effects include asperities and barriers on the fault plane and the finite size and shape of the rupture area. Dynamic effects are thought to be controlled by the velocity of the rupture, how this varies with time, and the initial state of stress on the fault. Thus the earthquake source is fundamentally a complex system. Due to the limit in computing time and the large number of variables in the inverse problem, in order to identify individual elements, simplifying assumptions have to be made.

For example, the kinematic model of Savage (1966) represents the source as an ellipse, with rupture initiating at a focus and propagating at a constant velocity. This is unlikely to be true for every earthquake but it produces more complexity than a point source. Das & Aki (1977a) and Das (1980) employed numerical techniques to generate source-time functions from a finite rupture area with material heterogeneity modelled by up to three asperities on the fault plane, and Nielsen & Tarantola (1992) generated near-field waveforms using numerical methods for a fully dynamic rupture in an arbitrary heterogeneity. This included realistic source physics in the rupture model such as a friction law and a rupture criterion. However as Brune (1991) states, “given the increase in the number
of papers one might have hoped that the uncertainties about the seismic source would have been greatly reduced, but this does not appear to be the case.” I believe this still applies today but these models of the earthquake source can provide an additional insight into causes of seismogram complexity.

In this thesis I examine the contributions of source duration, near-source structure, fault propagation direction, and attenuation, to complexity in a variety of seismograms. I use established techniques of phase identification and forward modelling to see how successfully these contributions can be resolved in a number of case-studies. These provide an insight into the problems associated with analysing complex seismograms. Also by using a new numerical model of fault rupture I show how variations in the factors that control fault propagation can improve our understanding of the source contribution to the complex seismogram.

Chapter 2 examines the issues surrounding complexity and simplicity of seismograms in detail, and outlines the previous work that forms the background to this study. I also consider how seismograms from known short-duration explosion sources have contributed to the identification of near-source and path complexity. Chapter 3 shows, using synthetic seismograms generated by the method of Douglas, Hudson & Blaney (1972), that altering the source duration, the near source structure, the propagation direction and path attenuation can all affect the apparent complexity of the seismogram.

Chapters 4 to 6 then investigate these factors further, using a number of case studies of individual earthquakes where seismogram complexity was observed. Chapter 4 compares the Whittier Narrows and Sierra Madre earthquakes in Southern California. I show that despite their geographical proximity, differences in near source structure could not be ruled out as a cause of the difference in the degree of complexity evident in their seismograms. I also show from forward modelling that differences in source duration can also have a dramatic effect on the complexity of the short-period seismogram from two supposedly ‘similar’ earthquakes. This chapter also shows that the narrow passband of the short period recording instrument emphasises the complexity of seismograms from the longer duration sources by filtering out the low-frequency part of the pulse. However, using broadband seismograms from apparently complex, intermediate-sized earthquakes (5 < m<sub>b</sub> < 6.5) at short-period, I show that longer sources can
produce interpretable seismograms at wider passbands. This highlights the im-
portance of using both broadband and short-period seismograms in conjunction.

Chapter 5 argues that near-source structure, together with the radiation pattern
from a uni-directional rupture, can explain the complexity observed at some
stations from the 25 October 1995 Caspian Sea earthquake. Unexpected arrivals
between P and the surface reflections had been misidentified as surface reflections
by the Prototype International Data Centre (PIDC) and an incorrect depth
assigned. The accurate determination of depth due to P-wave complexity may be
a problem in routine seismogram analysis. I show how a detailed treatment of a
suite of complex seismograms can resolve this. This shows that mode conversions
from S to P within a layered structure beneath the Caspian Sea region can
produce such complex phases, in particular they can become prominent in the
signal if the station is situated at a node in the P-wave radiation pattern. This
leads on to an examination of path effects outside the source region, which
was highlighted by an unusually simple S-wave signal from the Caspian Sea
earthquake. This simple S is unusual, it is more sensitive to attenuation than
P and is often complex on teleseismic records. I show that variations in the
upper mantle and crustal structure of Europe can affect the ability to identify
the S-wave phase. Thus important information about the simple nature of the
source may be hidden.

Having investigated path effects, chapters 6 and 7 consider how the nature of
the earthquake source can contribute to P-wave complexity. Chapter 6 examines
several earthquakes which show evidence for enhanced high frequency energy at
around 2 Hz which cannot be modelled using a simple elliptical source model
of Savage (1966). This simple kinematic source model only provides smooth
broadband synthetic seismograms as material properties and rupture velocity
are both constant in space and time. These model seismograms are inconsistent
with the observed small-scale variations in the high frequency component of the
broadband signal, which is dominant on the short-period signal. This suggests
that a more complex source model involving more realistic source physics, such
as friction, a rupture criteria, and material heterogeneity on the fault, may be
needed to explain high frequency fluctuations in the source time functions.

Chapter 7 describes how the dynamic source model of Nielsen & Tarantola (1992)
can produce dynamic seismic rupture on a fault. I have modified an existing
finite-difference modelling code to include a dynamic rupture element that produces complex source-time functions in the far-field. I show in various model scenarios that by allowing more extended rupture to evolve, and a greater distribution of material heterogeneity on the fault, that increasingly more complex but realistic rupture can be achieved. The resulting synthetic far-field pulses possess some of the higher frequency component seen in chapter 6. Finally, in chapter 8 I discuss how different sources of complexity can be identified or eliminated from a set of given seismograms, present a general strategy for dealing with observed P-wave seismogram complexity and summarise my conclusions.
Chapter 2

Complexity and Waveform Analysis

2.1 Introduction

Before analysing seismograms from individual earthquakes, it is important to show the reasons for the interest in complexity, the techniques used to study it, and the rationale for the methodology I have used in this thesis. In this chapter I introduce the concept of complexity and discuss the types of data used. By reviewing previous work on particular nuclear explosions and earthquakes, where the nature of the source is independently known, I describe the development of our understanding of the causes of complexity and outline the novel contribution this thesis makes. I then review the representation of the seismic source, showing that an improved understanding of source processes is useful in the study of seismogram complexity. I also discuss the various seismogram analysis techniques that are used in subsequent chapters for phase identification, focal mechanism determination, and the generation of synthetic seismograms.

This study of seismogram complexity, forms part of work on the verification of the Comprehensive Nuclear Test Ban Treaty (CTBT) and source discrimination. Much of the research on teleseismic P-wave complexity was carried out in the 1960s and 1970s by the research group at AWE Blacknest in the UK but also in the US. However after this time, the US moved to the monitoring of explosions
of the order of ‘a few kilotons fully decoupled’ (Douglas 1981). This meant that much of their recent research is focused at regional or local distances while the UK effort (including the work presented here) remained at teleseismic distances. The balance of references in this chapter reflects this split.

2.2 Complexity and Simplicity

Firstly I define the terms ‘simple’ and ‘complex’ in relation to the short-period P-wave teleseismic seismograms. A simple seismogram is defined here as one which can be interpreted uniquely in terms of some or all of the phases, P, pP and sP (figure 2.1a). One or more of the phases may be weak due to nodes in the radiation pattern, for example the sP in figure 2.1a. This definition is similar to that of Bowers (1996) who defines simplicity in terms of relative amplitudes and requires any of the three phases to have a large amplitude relative to the other two. This takes into account the possible misidentification of P. A complex seismogram in terms of this study is one where there is an initial P pulse followed by a coda (which may contain surface reflections) that is of a comparable size to, or larger than, the first arrival (figure 2.1b). This seismogram is not easily interpretable in terms of the three phases mentioned earlier. Seismograms may initially appear complex on first inspection but I will show that, with some investigation, many are in fact simple and ‘interpretable’. An important factor in this can be the type and quality of data used.

2.3 Seismic data

The quality of seismic data plays a key role in the amount of information that can be gained from a seismogram. The modern damped electromagnetic seismograph was developed in the early 1900s and long period recordings of events lead to the discovery of the gross structure of the Earth (Oldham 1906) following identification of P, S and surface waves. Work was also concentrated on routinely monitoring the locations of large events and on magnitude studies. Much of the seismology theory was in place at this time (Ben-Menahem 1995) but it was not until after the Second World War, with the first American (1945)
Figure 2.1.
(a) 'Simple' short-period seismogram recorded at Yellowknife Array, Canada (YKA) from an earthquake near Alma Ata, Kazakhstan showing P and a possible surface reflection (b) 'Complex' seismogram recorded at Eskdalemuir Array (EKA) from an earthquake in southern China in which individual phases cannot be distinguished easily (Douglas 1981)

and Soviet (1949) nuclear bomb tests, that significant funding was provided to improve instrumentation and fund the study of seismology. Large atmospheric tests produce considerable fallout and it was believed that these, along with underwater and space tests, could be easily monitored. A partial test-ban on these types of tests was signed in 1963. But monitoring of underground testing was more difficult. Seismology funding was then primarily aimed at gaining the ability to detect signals from underground nuclear explosions, estimate yields and discriminate between explosions and earthquakes.

In 1961 the Worldwide Standardised Seismograph Network (WWSSN) was established, consisting of well-calibrated short period (1 s) and long period (20 s) seismometers placed either side of the peak in background microseismic noise at 6 s. It was at this time that there were improvements in matching observations with theoretical predictions, in particular the finding that P and S radiation patterns are consistent with the double couple source model. Generally, short-period recordings have a greater signal-to-noise ratio in the P waveform, allowing individual phases to be seen separately and clearly compared to the long-period
record. At teleseismic distances, this P signal is free from crustal phase contamination caused by triplications and scattering. Hence, the study of teleseismic phases remains important, even when regional data, which might suffer from these problems, are available.

The nature of the short-period seismogram is more easily seen after signal-to-noise enhancement by beam forming of the individual signals from an array of seismometers. Beam-forming involves adding a systematic time-delay to each record, determined by source location and local crustal velocity, and then summing the individual signals across the array. The maximum amplitude in beam-formed record occurs when the direction to the source and phase velocity for the event-station pair are correct. Hence the beam-forming can give an approximate location, and distinguish events occurring at the same time, but arriving from different locations. The stacking also suppresses random noise, as only coherent arrivals are stacked in phase across the array. Four ‘T’ or ‘L’ shaped arrays were installed by the Blacknest Seismological Group in the 1960s at Eskdalemuir (EKA) (Figure 2.2) in Scotland, Yellowknife (YKA) in Canada, Gauribidanur (GBA) in India and Warramunga (WRA) in Australia. It is also possible to produce a ‘correlogram’, where the two beamforms from each arm of the array are multiplied together point by point to highlight coherent arrivals across the array. Coherent arrivals on both arms show up as positive amplitudes on the correlogram. Correlograms are very useful for picking out small-amplitude phases (Douglas 2000a).

Historically only vertical short-period seismometers had been installed at long range because these were the best recorded signals from underground explosions. Many of the seismogram analysis techniques used in this thesis were developed to utilise these data. However, for detailed modelling and analysis of complex waveforms, broadband data are required. This allows forward modelling of the full waveform and not just the high frequency components, at the expense of recording microseismic noise at periods around 6 s. The broadband instrument response function is shown, in comparison with the short-period instrument response, in figure 2.3. Broadband seismometers became more widely deployed in the 1980s. The combination of both broadband and short period records can provide a powerful tool for studying seismograms in general, and complexity in particular. In this thesis I will be using data gathered from several sources.
Chapter 2. Complexity and Waveform Analysis

Figure 2.2.
Layout of the seismic array station at EKA. Other arrays are in a similar cross shape due to original analogue computing of beam-forms. Black circles are approximately 1 km apart in this diagram. (Douglas 1997)

Figure 2.3.
Standard instrument response functions used in this thesis for real and synthetic data. Solid line: broadband velocity response. Dashed line: short period velocity response. These were generated from the standard responses available in the Seismic Analysis Code (SAC)
These are the short-period arrays installed by AWE Blacknest, who provided data directly, and from The Incorporated Research Institutions for Seismology (IRIS), who supply both broadband and some short-period data via their website (http://www.iris.washington.edu).

### 2.4 Seismogram Complexity

Much work has been carried out over the last thirty years on the factors which cause complex earthquake seismograms. This has looked at source, path and receiver effects. Such effects can reduce the size of the P compared to other phases, increase the size of the coda, or add extra phases other than those expected from a standard 1-D earth model. (Douglas \textit{et al.} (1973b) and Douglas (1981) give a comprehensive summary of previous work.)

In one of the first attempts to determine the causes of seismogram complexity, Thirlaway (1963) and Thirlaway (1966) assumed that the major cause was the nature of the seismic source. Natural earthquakes could produce complex seismograms if the source pulse was long in duration, or itself complex. In contrast, nuclear tests would be expected to have a small source volume and a short duration, due to the explosive nature of the source. Path effects were presumed to be negligible. This was consistent with the relatively simple seismograms produced by known nuclear explosions. For example Carpenter & Thirlaway (1966), Carpenter (1966) and Carpenter (1967) modelled several known nuclear explosions using only simple source and earth models.

A quantitative definition of complexity was suggested by Douglas (1967), who defined an energy ratio (ER) as the ratio of the energy in the first 5 s of the signal to that between 5 and 35 s. For consistency of comparison between different recording instruments, this definition was based on seismograms bandpass filtered between 1-2 Hz, then squared and smoothed by an exponential window with a time constant of 1.5 s. Complex signals tend to have an ER of less than 1.0, and simple records have a value of approximately unity or greater. With this definition, it soon became clear that not all explosions produce simple seismograms, and more earthquakes than might be expected also produce simple seismograms.
of papers one might have hoped that the uncertainties about the seismic source would have been greatly reduced, but this does not appear to be the case." I believe this still applies today but these models of the earthquake source can provide an additional insight into causes of seismogram complexity.

In this thesis I examine the contributions of source duration, near-source structure, fault propagation direction, and attenuation, to complexity in a variety of seismograms. I use established techniques of phase identification and forward modelling to see how successfully these contributions can be resolved in a number of case-studies. These provide an insight into the problems associated with analysing complex seismograms. Also by using a new numerical model of fault rupture I show how variations in the factors that control fault propagation can improve our understanding of the source contribution to the complex seismogram.

Chapter 2 examines the issues surrounding complexity and simplicity of seismograms in detail, and outlines the previous work that forms the background to this study. I also consider how seismograms from known short-duration explosion sources have contributed to the identification of near-source and path complexity. Chapter 3 shows, using synthetic seismograms generated by the method of Douglas, Hudson & Blamey (1972), that altering the source duration, the near source structure, the propagation direction and path attenuation can all affect the apparent complexity of the seismogram.

Chapters 4 to 6 then investigate these factors further, using a number of case studies of individual earthquakes where seismogram complexity was observed. Chapter 4 compares the Whittier Narrows and Sierra Madre earthquakes in Southern California. I show that despite their geographical proximity, differences in near source structure could not be ruled out as a cause of the difference in the degree of complexity evident in their seismograms. I also show from forward modelling that differences in source duration can also have a dramatic effect on the complexity of the short-period seismogram from two supposedly 'similar' earthquakes. This chapter also shows that the narrow passband of the short period recording instrument emphasises the complexity of seismograms from the longer duration sources by filtering out the low-frequency part of the pulse. However, using broadband seismograms from apparently complex, intermediate-sized earthquakes ($5 < m_b < 6.5$) at short-period, I show that longer sources can
produce interpretable seismograms at wider passbands. This highlights the importance of using both broadband and short-period seismograms in conjunction.

Chapter 5 argues that near-source structure, together with the radiation pattern from a uni-directional rupture, can explain the complexity observed at some stations from the 25 October 1995 Caspian Sea earthquake. Unexpected arrivals between P and the surface reflections had been misidentified as surface reflections by the Prototype International Data Centre (PIDC) and an incorrect depth assigned. The accurate determination of depth due to P-wave complexity may be a problem in routine seismogram analysis. I show how a detailed treatment of a suite of complex seismograms can resolve this. This shows that mode conversions from S to P within a layered structure beneath the Caspian Sea region can produce such complex phases, in particular they can become prominent in the signal if the station is situated at a node in the P-wave radiation pattern. This leads on to an examination of path effects outside the source region, which was highlighted by an unusually simple S-wave signal from the Caspian Sea earthquake. This simple S is unusual, it is more sensitive to attenuation than P and is often complex on teleseismic records. I show that variations in the upper mantle and crustal structure of Europe can affect the ability to identify the S-wave phase. Thus important information about the simple nature of the source may be hidden.

Having investigated path effects, chapters 6 and 7 consider how the nature of the earthquake source can contribute to P-wave complexity. Chapter 6 examines several earthquakes which show evidence for enhanced high frequency energy at around 2 Hz which cannot be modelled using a simple elliptical source model of Savage (1966). This simple kinematic source model only provides smooth broadband synthetic seismograms as material properties and rupture velocity are both constant in space and time. These model seismograms are inconsistent with the observed small-scale variations in the high frequency component of the broadband signal, which is dominant on the short-period signal. This suggests that a more complex source model involving more realistic source physics, such as friction, a rupture criteria, and material heterogeneity on the fault, may be needed to explain high frequency fluctuations in the source time functions.

Chapter 7 describes how the dynamic source model of Nielsen & Tarantola (1992) can produce dynamic seismic rupture on a fault. I have modified an existing
finite-difference modelling code to include a dynamic rupture element that produces complex source-time functions in the far-field. I show in various model scenarios that by allowing more extended rupture to evolve, and a greater distribution of material heterogeneity on the fault, that increasingly more complex but realistic rupture can be achieved. The resulting synthetic far-field pulses possess some of the higher frequency component seen in chapter 6. Finally, in chapter 8 I discuss how different sources of complexity can be identified or eliminated from a set of given seismograms, present a general strategy for dealing with observed P-wave seismogram complexity and summarise my conclusions.
Chapter 2

Complexity and Waveform Analysis

2.1 Introduction

Before analysing seismograms from individual earthquakes, it is important to show the reasons for the interest in complexity, the techniques used to study it, and the rationale for the methodology I have used in this thesis. In this chapter I introduce the concept of complexity and discuss the types of data used. By reviewing previous work on particular nuclear explosions and earthquakes, where the nature of the source is independently known, I describe the development of our understanding of the causes of complexity and outline the novel contribution this thesis makes. I then review the representation of the seismic source, showing that an improved understanding of source processes is useful in the study of seismogram complexity. I also discuss the various seismogram analysis techniques that are used in subsequent chapters for phase identification, focal mechanism determination, and the generation of synthetic seismograms.

This study of seismogram complexity, forms part of work on the verification of the Comprehensive Nuclear Test Ban Treaty (CTBT) and source discrimination. Much of the research on teleseismic P-wave complexity was carried out in the 1960s and 1970s by the research group at AWE Blacknest in the UK but also in the US. However after this time, the US moved to the monitoring of explosions
of the order of 'a few kilotons fully decoupled' (Douglas 1981). This meant that much of their recent research is focused at regional or local distances while the UK effort (including the work presented here) remained at teleseismic distances. The balance of references in this chapter reflects this split.

2.2 Complexity and Simplicity

Firstly I define the terms 'simple' and 'complex' in relation to the short-period P-wave teleseismic seismograms. A simple seismogram is defined here as one which can be interpreted uniquely in terms of some or all of the phases, P, pP and sP (figure 2.1a). One or more of the phases may be weak due to nodes in the radiation pattern, for example the sP in figure 2.1a. This definition is similar to that of Bowers (1996) who defines simplicity in terms of relative amplitudes and requires any of the three phases to have a large amplitude relative to the other two. This takes into account the possible misidentification of P. A complex seismogram in terms of this study is one where there is an initial P pulse followed by a coda (which may contain surface reflections) that is of a comparable size to, or larger than, the first arrival (figure 2.1b). This seismogram is not easily interpretable in terms of the three phases mentioned earlier. Seismograms may initially appear complex on first inspection but I will show that, with some investigation, many are in fact simple and 'interpretable'. An important factor in this can be the type and quality of data used.

2.3 Seismic data

The quality of seismic data plays a key role in the amount of information that can be gained from a seismogram. The modern damped electromagnetic seismograph was developed in the early 1900s and long period recordings of events lead to the discovery of the gross structure of the Earth (Oldham 1906) following identification of P, S and surface waves. Work was also concentrated on routinely monitoring the locations of large events and on magnitude studies. Much of the seismology theory was in place at this time (Ben-Menahem 1995) but it was not until after the Second World War, with the first American (1945)
and Soviet (1949) nuclear bomb tests, that significant funding was provided to improve instrumentation and fund the study of seismology. Large atmospheric tests produce considerable fallout and it was believed that these, along with underwater and space tests, could be easily monitored. A partial test-ban on these types of tests was signed in 1963. But monitoring of underground testing was more difficult. Seismology funding was then primarily aimed at gaining the ability to detect signals from underground nuclear explosions, estimate yields and discriminate between explosions and earthquakes.

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In 1964 a 'complex' signal by the energy ratio definition was recorded at YKA, and at other North American stations, from a known underground test at the nuclear test site at Novaya Zemlya in the Russian Arctic (Thirlaway 1966, Greenfield 1971). It was observed that if this level of complexity was used as a threshold for identification of earthquakes, then only a small percentage of earthquakes actually had signals more complex than explosion records from Novaya Zemlya. Figure 2.4a shows a recording at YKA from a subsequent explosion at Novaya Zemlya in 1967 which shows similar complexity to the signals from the original 1964 explosion. The signal consists of numerous large arrivals within the first 10s of the P waveform, with one arrival, 1.5s after P, being larger than the first arrival. This is not a typical explosive signal which would normally consist of a single initial P pulse of a few cycles on the short-period records shown. Independent work on this short-period complexity criterion produced poor results with only 20% of earthquakes positively identified (Kelly 1968, Lacoss 1968). These findings showed that complexity could not be used as a sole discriminant.

The existence of complex explosion records implies that complex waveforms are not due solely to source effects. This has subsequently been confirmed from the study of smaller earthquakes, which have simple spike-like source functions at periods of 1s or so, and still can produce complex seismograms due to path effects (Douglas et al. 1974). Source effects for intermediate-sized events can be
The main alternative considered next was the effect of scattering at near-source structure (Kennett 1973, Landers 1974). By scattering we imply the statistical effect of many heterogeneities that cannot be resolved by investigation of the near-source velocity structure. This may or may not be exacerbated by scattering at near-receiver structure (Douglas et al. 1973b).

Greenfield (1971) suggested that the complexity of explosive sources was due to scattering from rough topography near the source, where Rayleigh wave energy can be converted into P wave energy. If true, this also must happen to some extent near the receiver. However, most receiver arrays are placed in relatively flat topography with a simple, known velocity structure so this is unlikely to be a strong effect. Key (1968) pointed out that the converted Rayleigh wave to P near the source would have a low wave speed. However, the beam-forming can in principle remove such converted waves, which will have different apparent speeds at the receiver, and hence be suppressed compared to the main P, pP and sP energy, although a finite amount of scattered energy would still be present. Douglas, Hudson & Barley (1981) however did not explicitly find any evidence for strong near-field scattering due to topography, so this hypothesis was rejected as a significant contribution to short-period complexity.

Scattered energy in the near-source volume might be expected to be significant if P and pP leave the fault plane along a node in the radiation pattern, thus producing a small first arrival (Douglas 1967). In this 'weak signal' hypothesis, the coda from the scattering, although small in absolute amplitude, would then be large relative to P, and hence produce a complex emergent signal. A weak signal could also be produced if the P-wave is reduced in amplitude by a region of high anelastic attenuation, along a raypath not traversed by the coda waves (Douglas, Marshall & Corbishley 1971, Douglas et al. 1973b). Thus the coda is large compared to P and the seismogram again appears complex. There is evidence for such areas of anomalously low Q in the upper mantle, such as in Japan and South America (Sacks & Okada 1975). The weak signal hypothesis was challenged by Simpson & Cleary (1977), who preferred the near-source strong scattering hypothesis of Cleary, King & Haddon (1975). However at these depths (within the crust and upper mantle) the potential scatterers produce arrivals that are too small to support this hypothesis in the general case.
CHAPTER 2. Complexity and Waveform Analysis

(Key 1968, Greenfield 1971). The weak signal hypothesis is therefore a better explanation for complexity than that due to near-source scattering.

Observations of complexity in seismograms where P has been attenuated by a region of low Q are available for explosions at the Soviet Nuclear test-site at Bukhara in Central Asia (Douglas et al. 1971). Figure 2.4b shows a seismogram from an explosion in central Asia recorded at GBA, India. The raypath is under the Himalayas so we might expect significant attenuation of P and pP from a region of low Q beneath the isostatic root to the mountain chain. The large arrivals after pP are just small scattered waves from near the source, that have travelled a different path, and have not been attenuated compared to P and pP.

The weak signal hypothesis (due to anomalous attenuation of P) gained extra credence with explosions at Amchitka Island in the Aleutian subduction zone. These events produced a variety of both complex and simple signals at stations in the western U.S.A., depending on the ray path. As the map in figure 2.5a shows, radiation would not only travel through the subducting slab [which may defocus P from certain areas (Davies & Julian 1972)] but also through the Rockies orogenic zone (Key 1968). Spatial variations in upper mantle Q in the Rockies have attenuated the P wave considerably for some seismograms, leading to spatial variations in complexity in all directions from what should be a uniform explosive source amplitude in all directions. The large variation in attenuation is evident from a comparison of the estimated event magnitude at the different stations, which range from mb 4.74 (complex seismograms) to mb 6.13 (simple seismograms).

Two seismograms, recorded at the stations KC-MO and SI-BC on the map in Figure 2.5a, are shown in figure 2.5b. The KC-MO seismogram is simple and has travelled a high-Q path at greater depths. To SI-BC the ray-path is at much shallower depths and may travel through the low-Q mantle wedge. Whatever the cause, this higher attenuation has reduced the size of the P-signal and allowed scattered energy in the coda to make the signal appear complex. To some extent, the tectonic province in which the receiver lies may also change the nature of the signal. SI-BC is situated in the northern Rocky Mountains whereas KC-MO is on the stable Canadian Shield. It should be noted that these are all single stations, and that such scattering could in principle be removed by beam-forming.
Figure 2.5.
(a) Map showing the great circle paths from the LONGSHOT explosion to stations in North America from the Long Range Seismic Monitoring network (LRSM) 11, KC-MO 6, SI-BC. (b) Seismograms from the LONGSHOT explosion SI-BC signal is complex due to the small P signal. KC-MO is comparatively simple even though it has travelled a further distance. These differences must be due to path effects (Douglas et al. 1973b).
Attenuation along the raypath changes the frequency content as well as the amplitude. By comparing earthquake seismograms from the Aleutian Islands and those from the LONGSHOT explosion, Lambert et al. (1969) showed that in both cases the more low frequency content the signal had, the more complex the signal would be. This is consistent with attenuation of the higher frequency phases in P and pP, leaving a small amplitude P, and relatively high-amplitude scatterers. The P, pP and sP phases have similar raypaths, whereas the scattered energy arrives over a range of raypaths. Thus, even if the attenuation is higher than average along some, it will be lower on others, always leaving a finite amount of scattered energy on the seismogram. In this case, the high frequencies transmitted along high-Q paths, from scattering and multipathing of radiation off heterogeneities on the scale of one wavelength (8-14 km in the mantle and 6 km in the crust for periods of 1 s) (Douglas et al. 1972, Aki, Christoffersson & Husebye 1977, Douglas et al. 1981) and from the source region, can swamp the signal, making it appear complex. Studies of crustal heterogeneity from reflection, refraction and regional tomography have shown that such scattering is very likely at periods of 1 s (Mereu 1981, Dainty 1981), implying that the Earth is spatially heterogeneous down to 10 km or less. As a consequence, Douglas (1981) stated that 'we should not be asking why the coda is so much larger in amplitude than P, but why P is so much smaller in amplitude than the coda'. This question was posed again in Douglas (2000a) in a comment on Weber et al. (1998) where they had also proposed a mechanism for why the coda of certain underground nuclear tests were as large as the initial P wave.

Due to the existence of such path effects, complexity is now no longer formally used on its own as a discriminant between explosion and earthquake sources. However, an understanding of the factors contributing to it will enable further information on the earthquake source and earth structure to be obtained from the seismogram, as well as contributing to the development of a more reliable discriminant.

### 2.5 Source models

Understanding the earthquake rupture process has been the focus of extensive research, from pioneering work in the 1960s on the kinematic representation of
the earthquake source (Haskell 1964, Savage 1966), to recent work on dynamic friction laws and fault geometry (Nielsen, Carlson & Olsen 2000, Madariaga, Olsen & Archuleta 1998). Brune (1991) and Beroza (1995) give a lengthy review of work on source dynamics. Even with numerous stations and the ability to perform routine inversions after important earthquakes, the data and the analysis do not often seem to be able to distinguish between various source models. This is a generic problem in geophysical inverse theory, due to the presence of limited sampling and noisy data [Shearer (1999) p.74]. In this branch of seismology the theory was often developed long before the observation was at sufficient detail to resolve between competing source models. Even now approximations to the source are still used due to the limited availability of computer time and the undetermined nature of the inversion due to multiple parameters in the problem.

Given incomplete data and finite computational resources, any attempt to forward model seismograms must assume an earthquake source that is conceptually simpler than it is in nature. This involves assumptions as to the kinematics of the source, its size, geometry, rupture velocity, and orientation. A variety of source models have been used, the simplest of which is a point shear dislocation time (Langston & Helmberger 1975), extended to have a finite duration triangular or trapezoidal source time function by Kanamori & Stewart (1976). This approach continues to the present day, with the point dislocation model in space, with a complex source time function being used in the moment tensor inversion. This simplification is valid when the wavelength is much longer than the source size, appropriate for most events in the Centroid Moment Tensor catalogue. At higher frequencies such as those studied here, the point dislocation is not valid.

A simple finite-sized source rupture geometry was proposed by Savage (1966). The finite source geometry is idealised as an ellipse, which may be long and thin (high aspect ratio), or circular (low aspect ratio) in the extreme cases. Figure 2.6a shows a schematic diagram of the Savage model at a time, \(t\), after rupture has begun. The slip initiates at one of the foci of the ellipse, and propagates outwards on the fault plane until it reaches the pre-defined boundary of the stressed region, when the slip stops. It assumes that there are no stress discontinuities across the fault, that the rupture propagates at a pre-defined constant rupture velocity, and that the slip is accumulated instantaneously, resulting in
a Heaviside function for the cumulative slip (displacement) at a given point, and a delta function for the slip velocity. The latter assumption (figure 2.6b) is not physical, since it requires infinite acceleration, but the data often do not justify the additional parameters that a more complicated model would require. The amplitude of the displacement is evaluated at a particular point in space on the fault surface and hence is 2D rather than a 3D model. Complexity arises predominantly due to the extended source time function, the stopping phases when the rupture front reaches the elliptical boundary of the stressed area, and the finite aspect ratio.

![Diagram](image)

**Figure 2.6.**

(a) The Savage Model. Shaded area represents area over which the fault has ruptured at time \( t \). Larger ellipse represents the boundary of the stressed region where the slip stops. (b) The slip function at a point, \( P \), on the rupture plane at angle \( \theta \) from the vertical, begins and ends with a step (Savage 1966).

The Savage model is kinematic: it does not take into account complex rupturing on the fault plane due to material heterogeneities. For example Das & Aki (1977a) examined the effect of barriers to rupture propagation in generating far-field pulses, from a unilateral rupture. By adding areas of high stress (barriers) they show that progressively more complex far-field pulses can be generated that vary with azimuth. A few barriers and asperities can account for a few sub-events appearing on the source time function. In nature geological heterogeneity
is normally present on all scales (Turcotte & Newman 1996), but seismometers filter out the higher frequencies produced from the smaller-wavelength heterogeneities. Thus seismograms showing a few sub-events may be considered a low-frequency filtered version of the actual source function.

Other authors have shown from scaling arguments and simple massless cellular automata how faults can produce complex seismic rupture that is scale invariant (Rundle 1988, Ito & Matsuzaki 1990), but such quasi-static models have not generally been utilised in forward modelling of seismograms. Nielsen & Tarantola (1992) took this one step further by generating near-field synthetic seismograms from a fully dynamic rupture model. They demonstrated that as a fault ruptures in an arbitrary heterogeneous medium, such as a rock, seismic energy will radiate in a complex fashion from all points along the fault and interact with each other. Large amounts of high-frequency seismic energy are also generated at the stopping phase when the fault stops rupturing. Nielsen, Knopoff & Tarantola (1995) also note that seismic energy from one part of the fault, which is travelling at the P-wave speed, may momentarily increase the stress on a separate part of the fault causing the rupture to propagate at a velocity greater than the inertial limit (the Rayleigh wave speed in the medium.) For a uniform medium the Rayleigh wave speed is a fixed upper limit [Aki & Richards (1980) vol. 2 p.865.]. The common observation of source rupture speeds greater than this is direct evidence for material heterogeneity, and hence source-generated complexity.

In the studies of Nielsen & Tarantola (1992) and Nielsen et al. (1995) the fault is embedded in a finite-difference grid (Virieux 1986) which is used to solve numerically the elastic-wave equation (see chapter 7 for more details) for a spatially variable pre-stress or strength. This model has been used in other recent work, for example to model fault population growth and reservoir-induced seismicity (Madariaga et al. 1998, Nielsen et al. 2000, Maillot, Nielsen & Main 1999). Such dynamic fault models contain source physics such as a friction law and a rupture criteria, as well as an arbitrary material heterogeneity. They provide a realistic model of the fault rupture which is superior to simply using a point source which produces a triangular source pulse or a sum of triangles which has no physical meaning. Although the Savage model provides a realistic kinematic model of a finite-sized rupturing fault, it produces intrinsically smooth source functions,
in contrast to the observation of sub-events other than starting and stopping phases. In this thesis I will use the simplest model consistent with the data. Sometimes the Savage model is sufficient (chapter 4 and 5), and sometimes a full dynamic solution is needed (chapter 6 and 7).

2.6 Forward Modelling

In this thesis I frequently use the method of Douglas et al. (1972) to generate synthetic seismograms. This has been used extensively in the literature to forward model real data, for example to validate the method of determining focal mechanisms using relative amplitudes to be discussed in section 2.7. Hudson (1969a) and Hudson (1969b) developed the analytical theory for modelling radiation from point and extended sources. A point source needs only a time function, that may be a triangle, a trapezoid or some combination of the two. Douglas et al. (1972) developed a numerical code from this work to create a method for generating teleseismic synthetic seismograms. It uses the Savage (1966) model of the fault rupture to generate spherical seismic waves leaving the source and then propagates these through a horizontal plane-layered elastic structure near the source, representing the crust and uppermost mantle. The source also has a finite rise-time component to remove the problem in the Savage model of infinite slip velocity.

The receiver is at the free surface of another horizontal plane-layered structure at an angular distance from the source. Only two path effects in the mantle are accounted for: geometrical spreading and anelastic attenuation. Thus most of the small arrivals forming the P-wave coda are assumed to be multiples or mode conversions from near-source or near-receiver Earth structure (see section 3.3). The signal is then convolved with the response of the instrument (figure 2.3) to obtain the synthetic seismogram. Crustal structure at receiving stations are generally well known from seismic refraction surveys carried out when the stations were initially set up. The method of generating synthetic seismograms is carried out in the frequency domain with the vertical P wave component of the seismogram, $U_P(\omega)$ being evaluated over the range of frequencies 0 to $\omega_{max}$ by the convolution model:
\[ U^P(\omega) = M(\omega).S(i, \nu, \omega).C(\omega).R(\omega).I(\omega) \]  

where \( R(\omega) \) is the response of the receiver structure, \( I(\omega) \) the response of the seismograph, \( M(\omega) \) is the path effect in the mantle, \( S(i, \nu, \omega) \) is the P wave amplitude radiated into the mantle at azimuth \( \nu \) and at take off angle \( i \) (from the vertical), derived from the source orientation and type, and \( C(\omega) \) is the response of the plane layered structure near the source.

Bowers (1994) gives a good summary of the application of this type of source modelling using absolute amplitudes. Forward modelling can give an estimation of the average stress drop, ray-path attenuation, fault size and orientation. From this, the seismic moment, proportional to the area under the radiated P-wave displacement function, can be calculated using standard scaling rules (Kanamori & Anderson 1975). For example, Kirsty, Burdick & Simpson (1980) obtained estimates of the scalar moment and the orientation of the source by visually fitting synthetic and observed long period waveforms by trial and error. However, the use of absolute amplitudes at short periods is very sensitive to the choice of attenuation parameters along the path which reduces the high frequency component of the waveform and thus the modelled amplitudes. The quality of the solution and its uniqueness are dependent upon the focal sphere coverage and the signal-to-noise ratio of the recorded seismograms. At certain positions on the focal sphere large variations in modelled parameters may produce acceptable matches in a poorly recorded waveform. The absolute amplitude technique as currently utilised involves trial and error fitting of the waveforms so a quantitative estimation of errors and uniqueness is difficult.

However, these problems can be overcome using relative amplitudes, because the P and surface reflections travel a similar path in the mantle (figure 1.1), and so the problems of attenuation can be removed. Also, the P and surface reflections sample different parts of the focal sphere so more information about the source can be gained from the relative amplitudes of the phases P, pP and sP. I will be making extensive use of the relative amplitude method in this thesis, so the next section gives a more extended summary of this technique.
2.7 The Relative Amplitude Method

The relative amplitude method uses the fact that P and the surface reflections pP and sP have travelled similar ray paths between source and receiver (apart from the distance travelled up to the surface and back to the source depth). The different phases sample different parts of the focal sphere (figure 2.7). Thus, except for structure above the source, the path effect of mantle, near-receiver structure, and attenuation can effectively be cancelled out by considering relative amplitudes. Thus a greater coverage, than using P alone, of the focal sphere (upper and lower) can be obtained by using fewer stations.

![Diagram showing how direct phases and surface reflections travel a similar path except for the above source structure and sample different parts of the focal sphere.](free_surface_CRUST.png)

**Figure 2.7.**

Diagram showing how direct phases and surface reflections travel a similar path except for the above source structure and sample different parts of the focal sphere.

The correct identification of surface reflections pP and sP can provide a reliable estimate of source depth. Once this is known the focal mechanism can be obtained more accurately. Common techniques for this, such as the Nabelek (1984) method and the CMT method (Dziewonski, Chou & Woodhouse 1981), rely on using the whole observed waveform at a dominant period of greater than 20s. At these frequencies the source is effectively a point, thereby simplifying the analysis. However, the details of the source rupture are not taken into account in a point source, and it is difficult to differentiate between the numerous
factors that affect the complexity of the waveform at short period. Hence there is a significant known trade-off between optimal depth and focal mechanism in this method. Because of this the focal mechanisms for the CMT solutions are quoted as a range of possible solutions. However, no uncertainties in the primary data of the arrival time and amplitude of the different phases are given, making it difficult to evaluate the underlying cause of complexity in the short-period seismograms.

The depth determination for these methods is good for deep earthquakes but at shallow depths resolution is poor and introduces considerable errors into the modelling which are difficult to quantify (Buchanan 1998). For example, at a dominant period of greater than 20s the shallowest depth at which surface reflections can be separated from the P-wave is for an earthquake at about 30 km, compared to 1.5 km at 1s period. Most earthquakes I have studied in this thesis are shallower than 30 km. Thus, the use of good quality short-period recordings can provide critical additional information when used in conjunction with longer period data. In particular short-period data can be used to identify and discriminate between P, pP and sP phases and hence obtain a more reliable depth.

The relative amplitude method (Pearce 1977, Pearce 1980) uses short-period data to test any interpretation of p, pP and sP, and hence estimate source-type and orientation. The method uses P and the surface reflections, pP and sP, where they can be identified, their amplitude with appropriate confidence limits, and polarity, where they can be measured (figure 2.8). If any of the phases is small or missing compared to the background noise, especially on complex seismograms, then an undefined polarity and a wide confidence limit can be assigned to allow formally for the fact that a phase may be ‘small’. Information is still contained in the fact that a phase is ‘small’ or even absent from the seismogram. This is another of the advantages of the relative amplitude method: a null observation can be significant. The amplitude bounds placed on each phase also take into account preferential attenuation of any phase and the degree of confidence in the phase identification.

Confidence limits are assigned by assuming a uniform probability function between the amplitude bounds, i.e. a ‘box-car’ probability function. The program searches for all possible source mechanisms compatible with these observations,
Figure 2.8.
Amplitude bounds and polarity as needed for relative amplitude studies. Small arrow is the minimum and large arrow the maximum amplitude bounds. Note the larger bounds placed on pP and sP due to various factors effecting these phases as described in this section.
using a grid of strike, dip and slip angle (2.9b). The method does not find the optimal solution, but rather works by eliminating all possible solutions that are incompatible with the relative amplitudes calculated from the individual confidence limits. The method is developed in detail in Pearce (1980), Pearce & Rogers (1989) and Hudson, Pearce & Rogers (1989). The Relative Amplitude Moment Tensor Program (RAMP) then outputs a list of compatible solutions which consist of the strike ($\sigma$), dip ($\delta$) and slip angle ($\psi$) (see figure 2.9). The range of possible focal mechanism solutions can then be plotted on a vectorplot (figure 2.9). RAMP can also look at a range of sources, from a positive or negative compensated linear vector dipole (CLVD), to an explosion or implosion. The double couple is the simplest acceptable earthquake model (Aki & Richards 1980) and forms a suitable null hypothesis that would have to be formally rejected before adopting a more complicated source model. In all cases studied here, the double couple proved to be compatible with the data, so more complex source models were not examined.

The RAMP method works best for deeper sources at about 120 km (Stimpson 1987) where P, pP and sP are always distinct. For deep events only three good quality, well positioned stations are needed to produce a reliable focal mechanism (Pearce 1980). The RAMP method can also be useful when dealing with a multiple explosions with relatively few stations (Clark & Pearce 1988). In this case the second explosion could be confused with a surface reflection. This hypothesis could, though, be rejected with good azimuthal coverage of the event. For shallow sources generally, the phases are close together and the seismograms relatively complex, so less information can be extracted from each seismogram in the form of polarity and amplitude of each phase. Specific problems can occur for predominantly reverse and normal faults, where teleseismic data (near a P-wave antinode) do not sample the radiation pattern sufficiently to make a reliable estimation of the focal mechanism. For these (the majority of global earthquakes), the relative amplitudes pP/P and sP/P are insensitive to azimuth and to a lesser extent the take-off angle. Again this highlights the utility of null observations near the nodes of the radiation pattern in constraining the earthquake source.

In summary longer period data does not provide a suitable means for examining P-wave complexity because it estimates the focal mechanism over the whole
CHAPTER 2. Complexity and Waveform Analysis

Figure 2.9.
(a) Vectorplot for displaying focal mechanisms. Lower hemisphere stereographic projections indicate the type of fault plane orientation. Italicised words and lower numbers indicate Aki and Richards convention. Other numbers are the convention used in the relative amplitude method which is illustrated (b) (Pearce 1977).
waveform and does not utilise the information in the P-wave section of the waveform. The relative amplitude method works well at short period and for a small number of stations because it extracts the maximum amount of information from the amplitude and polarity of both the P and surface reflection arrivals. However the relative amplitude method must be applied with caution as it has its limitations in certain circumstances.

2.8 Application to source discrimination

Understanding earthquake seismogram complexity forms part of the necessary ongoing work in the discrimination between earthquake and explosion sources. In 1958 a Committee of Experts met in Geneva to design a monitoring system for underground nuclear tests and it realised that the only way to do this effectively was by their seismic effect. Much work has been carried out over the last forty years into this problem and it gained impetus with the signing in 1963 of a partial test ban prohibiting tests under water, in outer space, or in the atmosphere and the CTBT in 1996 (since when the CTBT has not yet come into force and ratification is continuing).

This chapter highlights how the understanding of the complexity of explosion sources is still developing. In the meantime, one of the strategies in deliberately cheating on a test-ban treaty could be to make a test look deliberately more complex by setting off charges in a complex source area, or by setting off a series of explosions. Similarly explosions can be ‘muffled’ by setting off charges in a cavity. As a consequence Pearce (1996) has suggested that work on looking for deliberate detection evasion will require detection at a lower magnitude threshold.

At low magnitude levels it has always been known that regional (up to 3000 km) seismic monitoring would be essential for signal detection. However, from the complexity of regional seismograms due to crust and upper mantle structures, and the need for many networks near possible test sites, a conclusive discriminant has remained elusive for forty years. Here I use only teleseismic data, which has proved to be useful in source identification down to as low as $m_b = 3.8$.
(Douglas et al. 1999, Ringdahl 1976), equivalent to an explosion in certain circumstances of 0.1 kton TNT equivalent. Interest in teleseismic data had waned since it was thought that the detectable magnitude threshold would be too high for useful monitoring, but with modern digital seismometers and good signal to noise ratios (especially with the improved processing of array data and high quality broadband recording) teleseismic data can provide useful information about seismic sources even at lower magnitudes. It has the advantage that a relatively sparse global network of stations is required for teleseismic monitoring (Pearce 1996), whereas many regional networks are needed to monitor all possible test sites at a regional scale.

Very recent work has highlighted the fact that some earthquakes fail conventional source discrimination techniques ($m_b$:$M_s$, first motion) due to the complexity of the seismograms (Pearce 1996). For theoretical and practical reasons, the phenomenon of seismogram complexity at high frequencies clearly requires further research. This thesis attempts to use modern data to unravel the causes of such seismogram complexity for some specific case studies, and suggest strategies for resolving the individual contributions to complexity.

### 2.9 Summary

In this chapter I have introduced the background to the study of seismogram complexity and the underlying rationale for this thesis. The study of both explosion and earthquake complexity has developed with the use of short-period, long-period, and broadband data. These data tend to provide complementary constraints on the source model and wave propagation effects, and so are best used in conjunction. Using a simple kinematic source model, forward modelling and relative amplitude techniques can identify surface reflections, obtain depth and focal mechanism of events and resolve various individual factors that contribute to complexity. However, a fully dynamic source model may be necessary to explain some forms of complexity, particularly at the highest frequencies. It is clear that seismogram complexity can be due to a variety of factors, and these can be hard to unravel.

In the following chapters I present a series of case studies that illustrate how
the origin of complexity can be identified uniquely under certain conditions of data quality and source type. Before we examine actual seismograms, various synthetic seismograms are presented in the next chapter, to demonstrate quantitatively how the source duration (controlled by the rupture velocity and fault radius), fault propagation direction, source model, near source structure and path effects can all contribute to seismogram complexity.
Chapter 3

Factors affecting complexity

3.1 Introduction

This chapter describes in more detail the most important individual factors which contribute to P-wave complexity. The method used will be to calculate synthetic seismograms for different source durations and with varying near-source and path structure, using the convolution model as implemented by Douglas et al. (1972), already outlined in section 2.6. Synthetic seismograms will be calculated for source and path characteristics similar to those that will be examined in the next two chapters. The aim of the exercise is to break down the synthetic seismogram into individual elements that can cause or mask true complexity.

First, I examine the effect of source duration (section 3.2). This is affected both by the geometry of the fault rupture and rupture speed. Although fault ruptures may have an arbitrary geometry, for simplicity I assume a general elliptical fault of varying aspect ratio. Even such simple rupture geometries can produce complexity due to starting and stopping phases which increases the complexity of the seismogram. I then examine the effect of the average rupture speed, which controls the duration of the source-time function. In particular I show that slow rupture speeds can introduce complexity, also by affecting the prominence of starting and stopping phases. Second, I show how mode-conversions caused by near-source velocity structure add to the complexity of the P arrival (section 3.3). Third, I examine the effect of the direction of rupture propagation on
source complexity (section 3.4). Propagation direction on its own can generate complexity on narrow-band instruments by producing systematically different frequencies of the P and surface reflection phases. However, here I show that propagation direction can also amplify the effect of mode conversions on the resulting seismogram. Finally I illustrate how the effect of anelastic attenuation along the ray-path contributes significantly to the apparent P-wave complexity (section 3.5). Attenuation does not cause complexity on its own, but, by preferentially removing high-frequency energy, it can filter out arrivals such as mode conversions and multiple reflections, that mask the true complexity.

The seismograms to be calculated in this chapter are all based on a single velocity structure at the receiver. This control allows the effect of the different elements listed above to be isolated more easily, since receiver structures are usually independently known (in fact are used to choose the locations of the arrays, and hence are selected to be relatively simple). In addition near-receiver multiples and mode conversions affect the P, pP and sP phases only to a minor extent. Here I use the velocity structure at the receiver at EKA (Eskdalemuir, Scotland) (table 3.1b). It is preferable to use an actual structure, rather than an oversimplified half-space assumption for example. This control is also justified because the path in the near receiver structure can be assumed to be identical for teleseismic P, pP and sP, and any other phases created near the source.

In contrast, near source structure does have a first-order effect on the seismograms. Here I first choose a near-source structure, and calculate ray-paths, appropriate for the location and focal mechanism of the 1st October 1989 Whittier Narrows earthquake (table 3.1a) (Linde & Johnston 1989). These will be used for comparison with seismograms to be examined in chapter 4. For this earthquake’s focal mechanism, EKA is not nodal for P, enabling the effects of source geometry and rupture speed to be highlighted clearly.

Path effects, such as S-to-P mode conversions near the source, are exacerbated when the path between the source and receiver is nodal for P. This occurs because S is anti-nodal, so there is a large potential source of energy for mode-conversions near the source that might affect the P-wave section. To highlight this effect I choose the location and focal mechanism of the 25 October 1995 Caspian Sea earthquake (table 3.1a). This event has a steeply dipping fault plane (table 3.1a), and as will be demonstrated in chapter 5, with a rupture direction towards the
receiver. Thus mode conversions from near-source structure will be significant. Similarly, when examining the effect of rupture direction, I use the actual focal mechanism of this event. This choice was made mainly to highlight the effect of the mode conversions and aids comparison with seismograms to be presented in chapter 5.

The source itself may be arbitrarily complex. In this chapter, and in the three to follow, I use the simplest finite-sized source model (Savage 1966). This kinematic model assumes uniform material properties near the source, and a constant rupture velocity (section 2.5). These simplifications allow the individual contributions from other factors described above to be seen more clearly. The general case of fully dynamic rupture is discussed in (chapter 7). The rationale is to examine simpler models first, and resort to more complex ones only where the factors discussed in this chapter can be eliminated as a cause.

<table>
<thead>
<tr>
<th>Source</th>
<th>Whittier Narrows type fault</th>
<th>Caspian Sea type fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dip of fault plane</td>
<td>30°</td>
<td>89°</td>
</tr>
<tr>
<td>Strike of the fault plane</td>
<td>330°</td>
<td>0°</td>
</tr>
<tr>
<td>Angle between strike and direction of slip</td>
<td>98°</td>
<td>0°</td>
</tr>
<tr>
<td>Depth of focus (initial rupture point)</td>
<td>14 km</td>
<td>14 km</td>
</tr>
<tr>
<td>attenuation (t*)</td>
<td>0.2</td>
<td>0.2</td>
</tr>
</tbody>
</table>

(b) Depth to base of layer (km) | $V_p$ kms$^{-1}$ | $V_s$ kms$^{-1}$ | density kgm$^{-3}$ |
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>5.3</td>
<td>6.14</td>
<td>3.54</td>
<td>2.8</td>
</tr>
<tr>
<td>19.7</td>
<td>7.28</td>
<td>4.20</td>
<td>3.2</td>
</tr>
<tr>
<td>half space</td>
<td>8.09</td>
<td>4.67</td>
<td>3.38</td>
</tr>
</tbody>
</table>
3.2 Source Duration

The source geometry and average rupture speed both control the total duration of the seismic source. In the simplest case of an ellipse with identical major and minor axes (a circular fault) we have:

\[
\text{source duration} = \frac{\text{radius}}{\text{average rupture speed}}.
\]

In the Savage (1966) model the rupture speed is assumed constant. The rupture geometry changes the shape of the source pulse as well as the duration, but the rupture speed controls only the duration. For a given fixed geometry, the size and rupture speed have an equivalent effect. Here I calculate the effect of how changes in the average rupture speed and the size of the fault generate complexity on short-period records. Since no near-source complexity is assumed, the models retain a simple broadband record.

3.2.1 Fault Size and Geometry

Here I show that a constant fault rupture speed and varying fault size can produce a variety of short-period seismogram complexities. The rupture speed is assumed constant at 3.5 km/s, and a relatively simple near-source structure corresponding to the Southern California model was used (table 3.2). The rupture speed is only slightly lower than the shear-wave velocity in the source layer (layer 3 for its depth of 14 km - see tables 3.1 and 3.2), producing a relatively sharp source pulse for a given geometry, while remaining in the observed range of rupture speeds. Although strictly the inertial limit is the Rayleigh wave speed, higher velocities can be expected due to diffractions (section 2.5).

Figure 3.1 shows (a) short-period and (b) broadband synthetic seismograms for a circular fault for different source rupture diameters up to a realistic maximum of 12 km for intermediate-sized earthquakes. On this and other figures the predicted P and surface reflection times are independently known and are marked on the diagrams. The smallest fault size, of radius 1 km, produces a simple seismogram on both the short-period and broadband record (figure 3.1). On
Table 3.2. Southern California velocity model used in this section (Hauksson & Jones 1989)

<table>
<thead>
<tr>
<th>Layer</th>
<th>P-Wave Velocity</th>
<th>S-Wave Velocity</th>
<th>Density</th>
<th>Depth to Top of Layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>4.10</td>
<td>2.37</td>
<td>2.7</td>
<td>0.0</td>
</tr>
<tr>
<td>2</td>
<td>5.16</td>
<td>2.80</td>
<td>2.8</td>
<td>2.0</td>
</tr>
<tr>
<td>3</td>
<td>6.35</td>
<td>3.67</td>
<td>2.8</td>
<td>5.5</td>
</tr>
<tr>
<td>4</td>
<td>6.54</td>
<td>3.78</td>
<td>2.8</td>
<td>16.0</td>
</tr>
<tr>
<td>5</td>
<td>7.87</td>
<td>4.54</td>
<td>3.0</td>
<td>32.0</td>
</tr>
</tbody>
</table>

both, the short duration P pulse is followed 6 s later by pP, separated in turn by 1 s from sP. The amplitude of P is approximately 45 nm, much larger than the small multiple reflections and mode conversions seen between P and pP. The seismograms have similar properties up to a rupture radius of 4 km where the amplitude of P has increased to 300 nm because of the effect of increasing source moment from a larger fault. At 5 km radius the source duration, as seen on the broadband, is now 1 s, and the P pulse on the short-period record begins to separate into distinct phases corresponding to the starting and stopping of the rupture. These separate phases may lead to the mis-identification of the stopping phase as a surface reflection if a very shallow depth is assumed. If the correct depth is assumed, pP and sP can still be seen clearly. However on the broadband record the seismogram remains simple, with a prominent positive excursion, followed by a negative, corresponding to the onset and termination of the P pulse.

As the source size increases, the seismogram becomes increasingly complex. For the largest ruptures, up to 12 km in source radius, the P pulse on the short-period has several distinct phases, and there are several visible phases at the time predicted for pP and sP. This produces a truly complex seismogram on the short-period record, where only the onset of P can be uniquely identified (figure 3.1a). In contrast, the seismogram retains simplicity on the broadband record (figure 3.1b) where the full source duration can be seen although pP and sP would be difficult to identify given finite background noise not present in the ideal synthetics. The amplitude of the band-limited short-period signal saturates at a maximum of around 600 nm for fault ruptures larger than 8 km, whereas the broadband signal reflects the true amplitude of P at 3000 nm. The surface reflections on the broadband record are small compared to this but two small negative pulses can still be seen for a 12 km fault. This example illustrates the
need to record both short-period and broadband to get accurate phase arrival times, source duration, and amplitude for both small and large fault sizes, and the complementary nature of the two records in identifying these.

![Figure 3.1](image)

**Figure 3.1.** Synthetic (a) short-period and (b) broadband seismograms for a circular source of increasing radius: 1 km at top to 12 km bottom in increments of 1 km. (c) Maximum peak to peak amplitude against fault radius

Figures 3.2a and 3.2b show the equivalent set of seismograms for an elliptical fault, where the major axis is twice the length of the minor axis of the ellipse. This fault rupture has an aspect ratio of two, an average of the data compiled by Pegler & Das (1996). Here the fault 'size' is defined as the length of the major axis. For this definition the overall fault area is therefore smaller for an ellipse than for the equivalent circular fault. Hence the seismograms are simpler for a given 'size', but they show the same broad characteristics as for the
circular fault. The elliptical rupture also produces a much smaller amplitude for a given size, only 1300 nm compared to 3000 nm for the circular rupture. The smaller the major axis of the fault, the shorter the source duration and the simpler the short-period seismogram. As the fault size increases on the short period seismograms, the P pulse separates into two distinct starting and stopping phases, similar to the results for the circular fault. However even at 12 km in size the elliptical fault produces an interpretable seismogram although it would be difficult to identify surface reflections uniquely. The broadband records clearly show the source duration is always less than 1 s which is why the short-period is still interpretable. In summary an elliptical fault of similar rupture length, but greater aspect ratio, produces a simpler seismogram.

3.2.2 Rupture Speed

Here I show how variations in the speed of the fault rupture at a fixed fault size cause complexity to emerge due to changes in the source duration. For simplicity I use a circular fault with three different radii, and six different rupture speeds in each case. In the last section I assumed a relatively fast rupture speed only slightly lower than the shear-wave velocity in the source layer. However, rupture speeds are often as slow as \( \frac{1}{\sqrt{3}} \) of \( V_s \) (Madariaga 1976). In this section I calculate synthetic seismograms within the range \( \frac{1}{\sqrt{3}} \) of \( V_s \) to 0.95\( V_s \). Figure 3.3 shows synthetic seismograms for source sizes of 1 km, 5 km and 10 km for the short-period instrument.

For a small fault radius (1 km) the seismograms for all rupture speeds shown are simple (figure 3.3a). In this case, even for slower rupture velocities a single P pulse and identifiable surface reflections can be seen at the correct predicted times. At a radius of 5 km the seismograms are still interpretable, for P and surface reflections, at the fastest rupture speeds, but become more difficult to interpret correctly for lower velocities, e.g at 2.2 km/s where the P pulse consists of several arrivals. This is because the starting and stopping of the rupture now produce individual cycles on the short-period record because the pulse duration is now longer than the natural period of the instrument. As you approach \( V_s \) for the crust in this area (3.78 km/s), the seismograms are more simple. The P pulse consists of a single cycle and possible surface reflections. Finally,
Figure 3.2.
Synthetic (a) short-period and (b) broadband seismograms for an elliptical source of increasing major axis size and an aspect ratio of 2:- 1 km at top to 12 km bottom in increments of 1 km.
for the largest of 10 km, the P pulse itself produces multiple peaks due to the starting and stopping phases, so that pP and sP overlap, especially for the slowest rupture velocities where the source duration is longer than the sP-pP travel time difference. This source effect could lead to the mis-identification of the surface reflections.

I now examine the effect of rupture duration on the broadband record. Figure 3.4 shows a single example for a 5 km radius source. The broadband seismogram has one simple arrival on the P pulse that shortens in duration with increasing rupture velocity. The surface reflections are interpretable, although the resolution is poor, and diminishes with decreasing rupture velocity.

In conclusion, rupture velocity can greatly effect the complexity of the short-period seismogram, especially for larger rupture areas. In particular, a fast rupture velocity shortens the duration of the source-time function and produces a simpler seismogram on the short-period record. A longer rupture duration produces additional stopping phases, and a pulse duration longer than the separation between pP and sP, making sP hard to identify uniquely. In addition the stopping phase may be misidentified as pP. However, the combined use of short-period and broadband records can help identify surface reflections uniquely, and place bounds on the rupture duration in many cases.
CHAPTER 3. Factors affecting complexity

2.1 kms

3.7 kms

Figure 3.3.
Synthetic short-period seismograms for a fixed fault radius with increasing rupture velocity: 2.1 kms$^{-1}$ top to 3.7 kms$^{-1}$ bottom in increments of 0.3 kms$^{-1}$. (a) circular fault radius 1 km, (b) radius 5 km and (c) 10 km.
Figure 3.4.
Synthetic broadband seismograms for a fixed fault radius with increasing rupture velocity: 2.2 km/s top to 3.7 at the bottom in increments of 0.3 km/s. 5 km radius fault
3.3 Near-Source Structure

The previous section used only a simple layered source structure with a velocity increasing gradually with depth (table 3.2). If however this structure is changed to include a layer of anomalously different velocity, then the resulting waveform has extra arrivals generated by multiple reflections and mode conversions. In particular conversions from S to P waves can arrive between P and pP and can be mistaken for surface reflections. These arrivals consist of downgoing S-wave energy that is incident on a boundary beneath the source with a high acoustic impedance contrast. Energy will be partitioned at the interface and a small proportion of the S-wave energy will be converted to a transmitted P-wave, arriving after the P phase by an amount that depends on the depth of the interface below the source. Conversions to P can also occur when upgoing S energy is reflected at boundaries other than the free surface, but these tend to be small due to the smaller reflection coefficients between near-surface layers. In this section I examine the effect of adding a low-velocity or a high velocity layer below the source depth.

Figure 3.5 compares synthetic seismograms for the example of a 2 km radius source, with three different near-source structures corresponding to (a) a 5 km thick high-velocity layer inserted between layers 4 and 5 of table 3.2, (b) the standard Southern California velocity model, and (c) a 5 km thick low-velocity layer at the same position. The additional layers have their top at 32 km (fig. 3.5). The P, S velocities and densities are given in the figure caption. The small source radius produces a pulse less than 1 s in duration, so that any complexity is not source-generated. For the standard crustal model we have a simple P phase with small surface reflections. The overall size of the seismogram is 5 nm as the P-wave arrival is near nodal for this earthquake-station pair (section 3.1). The mode conversions between P and pP are relatively small in amplitude compared to the other phases. With a high velocity layer (figure 3.5a) the amplitude of P, and the S to P mode conversions, decreases, but the mode conversions are now more visible. In inverse mode, this would make the identification of pP and sP difficult for this near-nodal arrival. In the lower seismogram, figure 3.5c, the low velocity zone produces several even larger mode conversions, although the overall size of the surface reflections has increased as well.
Synthetic short-period seismograms generated from a 2 km radius steeply dipping fault plane with the station and rupture direction in the direction of strike of the fault. (a) includes a 5 km thick high velocity zone ($V_p = 8.0 \text{ kms}^{-1}$, $V_s = 4.6 \text{ kms}^{-1}$, $\rho = 3.1 \text{ kgm}^{-3}$ (d)), (b) is for the standard Southern California earth model, and bottom (c) a 5 km thick low velocity zone ($V_p = 5.0 \text{ kms}^{-1}$, $V_s = 2.9 \text{ kms}^{-1}$, $\rho = 2.8 \text{ kgm}^{-3}$ (e)). The difference in arrival times of P on (a) and (c) is due to the high and low velocity zones.
In summary, mode conversions tend to be small in amplitude on most seismograms and are generally not noticeable unless the P-wave arrival is nodal. A low velocity zone beneath the source produces a more complex seismogram overall in which it is difficult to identify surface reflections. Although a high velocity zone also produces complex seismograms, it is easier to interpret in the example shown here. In the next section, I show that this effect is enhanced by source ruptures with higher aspect ratio.

3.4 Propagation Direction

The simplest fault rupture model is a circular expanding rupture which starts from a point and propagates in all directions. The Savage model introduces a sudden stopping phase when the circular rupture hits an elliptical boundary, which typically has aspect ratio of two for most earthquake ruptures (section 2.5). However it has been noted in some cases that fault ruptures can have much higher aspect ratio than this. Such ruptures may be better approximated by a line source e.g. as in (Barley, Hudson & Douglas 1982) where a seismogram for an example from Chile can be better modelled with a fault propagating from a point downwards along the ray-path. This can be regarded as a special case of the Savage model with an infinite aspect ratio. A downward-propagating unilateral fracture on a steeply dipping (>45°) plane can also focus S wave energy downwards, and cause the amplitude of S-to-P conversions to be large compared to other phases. This is a Caspian Sea-type model which will be investigated in chapter 5.

In general, a moving rupture front will cause the amplitude of the signal and the duration of the pulse to change, depending on the propagation direction relative to the station. For example, if the direction of propagation is towards the station then the duration of the pulse is shortened (the Doppler effect, figure 3.6) and the amplitude of the signal increased. This will increase the observed corner frequency of the source (on a log amplitude versus log frequency plot) in that direction, so a larger signal is seen in the short-period passband as the amplitude does not fall off until higher frequencies. In the opposite direction from the direction of rupture propagation the signal is of lower amplitude but longer duration. This decreases the corner frequency observed at a receiver and so
the signal amplitude falls from a lower frequency. The signal seen on the short-period seismogram is therefore reduced in amplitude. If the rupture propagation direction has a vertical component, then the amplitude and duration of the P and pP, sP phases will be systematically different.

![Diagram](image)

**Figure 3.6.**
Idealised version of a down-going propagating line rupture showing the far-field pulse radiated in and away from the direction of propagation.

If the station azimuth is also close to a P wave node (and hence close to S wave antinode) then the relative amplitude of S-to-P conversion with respect to P will be increased, and the seismogram will appear complex (Douglas, Richardson & Hutchins 1990) solely due to this directional effect. This model corresponds to the origins of P-wave complexity suggested by Douglas (1967) and Douglas *et al.* (1973b) when P is suppressed relative to the coda. Figure 3.7 shows synthetic seismograms for a down-going line rupture of the same source parameters as those used to generate figure 3.5, including the three different near source structures in order to see the effect of possible mode conversions.

The line source produces a smaller absolute amplitude (see scale) than the equivalent circular fault as the seismic moment is less for a given fault 'size' (length in this case). Only the onset of P is clearly visible in each seismogram with numerous phases in the coda of comparative size to P. All three seismograms in figure 3.7 show large S-to-P conversions between P and the predicted pP arrival time. These are of higher frequency and a larger relative amplitude compared to other phases, than from a circular source (compare figure 3.7 with figure 3.5).
Figure 3.7.
Synthetic short-period seismograms from a steeply dipping line source length 2 km propagating down dip. (a) includes a 5 km thick high velocity zone, (b) is for a standard Southern California earth model and (c) a 5 km thick low velocity zone (see figure 3.5 caption for parameters). Again the delays are due to the presence of the anomalous zones at 32 km
In particular, the low velocity zone (c) does produce a slightly larger S-to-P conversion than the other two seismograms in figure 3.7.

All three seismograms are overall very complex so the additional effect of the low and high velocity zones are less apparent. In conclusion, the combined effect of propagation direction and high aspect ratio source geometry can introduce a very high degree of complexity.

### 3.5 Attenuation

Attenuation is the energy lost by the seismic wave along the length of the ray path and is measured by a value, \( t^* \), where

\[
t^* = \int_0^T \frac{1}{Q(t)} dt = \frac{T}{\bar{Q}}
\]

\( Q \) is the local quality factor measurement of attenuation with units of \( s^{-1} \) at a point \( x \) corresponding to time \( T \), \( \bar{Q} \) is the average quality factor along the path and \( T \) is the source-station travel time. At low values of \( t^* \), attenuation is low, and very sharp clear arrivals can be seen on both the short-period and broadband seismograms. As attenuation increases to a maximum plausible value of \( t^* = 1 \) (Abercrombie et al. 1995, Sharrock, Main & Douglas 1995), more energy is removed and the waveform reduces in amplitude. Due to the fact that the higher frequencies are preferentially removed the shape of the waveform changes, and the signal appears less complex and lower in frequency. If the surface reflections, mode conversions, or starting and stopping phases are at a higher frequency, then attenuation could make an apparently complex short-period waveform (figure 3.8a, top trace) appear deceptively simple.

Figure 3.8 shows (a) short-period and (b) broadband seismograms for a Whittier Narrows-type event (simple source geometry, anti-nodal P), for different values of \( t^* \). For a very small \( t^* \) the amplitude of many of the phases are of similar size, with one large phase at 18s on the short period record. Phase pP and sP are clearly visible, and their arrival times can be measured very accurately. Some multiples or mode conversions can also be seen between P and pP. As
(a) Short-period synthetic seismograms and (b) broadband for a 2 km radius circular fault with Whittier Narrows focal mechanism. Top trace $t^* = 0.01$, second 0.2 and then increments of 0.2 to 1.0. Time in seconds. Note the amplitude scale is reduced from top to bottom in (a) and (b). The delay is due to the removal of high frequencies which means that as you increase attenuation the first arrival is of lower frequency and arrives slightly later.
CHAPTER 3. Factors affecting complexity

attenuation is increased, the larger amplitude but higher frequency phases are preferentially filtered out leaving only two interpretable phases, P and pP, and thus a simple seismogram. For the broadband seismogram the three discrete phases P, pP and sP are seen at all levels of attenuation. This highlights the utility of broadband records for source mechanism studies, where the overall amplitude of the seismogram is reduced but the relative size of individual arrivals remains similar. As the frequency of the signal is reduced, the surface reflections in figure 3.8b become less discrete. It could therefore be argued that the broadband record becomes less interpretable at very large values of t*. In summary on the short period record, large, sharp arrivals can be obscured in the coda of earlier arrivals by high attenuation along the ray path, whereas on the broadband record they remain, but at greatly reduced resolution.

3.6 Summary

In this chapter I have shown that the passband of the recording instrument (whether it is short-period or broadband), the duration of the source (controlled by the source area, geometry and the rupture velocity), the near-source structure, the direction of propagation, and path attenuation all have a systematic bearing on the complexity of the seismogram in forward modelling. This in turn affects the ability to correctly identify individual phases and obtain information about the source from actual recorded seismograms. I have shown how some of these effects have characteristic features that can be identified, and I have shown qualitatively the contribution each makes to complexity in certain circumstances. The models presented here are realistic scenarios and in the following chapters I present how certain complex seismograms in individual case studies can be explained using the different underlying causes set out in this chapter.
Chapter 4

Complexity due to Source Duration

4.1 Introduction

The aim of this chapter is to examine complexity due to a long source duration for two intermediate-sized events. This is illustrated by comparing two events in the same source area, with long and short duration source pulses respectively. The motivation for this comparative study was to use the short source pulse event as a control, and hence to isolate the contribution of the source to the observed seismogram complexity. Here I examine in detail the 28 June 1991 Sierra Madre and the 1 October 1987 Whittier Narrows earthquakes in Southern California. Both earthquakes caused significant damage, and represent a significant hazard for the city of Los Angeles and environs. They are thus important events to study in their own right. The source complexity studied here in the far-field can also affect to some extent the spectrum of the radiated energy in the near-field, and hence affect seismic design criteria for buildings and infrastructure. Their source parameters as reported by the ISC are listed in table 4.1. These parameters provide a starting point for the modelling to be undertaken here.

Much work has been carried out on these earthquakes over the last few years. The focal mechanisms have been determined seismically from the CMT method.
(Hauksson & Jones 1989, Bolt, Lomax & Uhrhammer 1989, Wald 1992, Hauksson 1994) and from geodetic data (Linde & Johnston 1989). Forward modelling has also been used to model the short-period data from the Whittier Narrows earthquake, with varying degrees of success. Bent & Helmberger (1989) approximately matched the short-period seismogram at one station by using a source pulse with two-subevent model. However, the model was not validated on three other stations for which they presented real data, possibly due to difficulties in obtaining a unique solution consistent with all azimuths. However, Douglas (1997) did succeed in accurately matching broadband seismograms for two independent stations for this event, and approximately matching some of the complexity seen in two short-period seismograms, but did not attempt to match the short-period data in this short communication. For the Sierra Madre, Wald (1992) matched six broadband stations and obtained a good match for one short-period station. In this chapter I aim to explain all of the observations (broadband and short-period) at several stations with a single model.

Table 4.1. Whittier Narrows and Sierra Madre earthquake parameters given by the International Seismological Centre (Aki & Richards (1980) notation)

<table>
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<tr>
<th>Parameter</th>
<th>Sierra Madre</th>
<th>Whittier Narrows</th>
</tr>
</thead>
<tbody>
<tr>
<td>time</td>
<td>14:43:54.1</td>
<td>14:42:20.0</td>
</tr>
<tr>
<td>location</td>
<td>34.24°N 118.03°W</td>
<td>34.06°N 118.08°W</td>
</tr>
<tr>
<td>depth</td>
<td>11.0±3.1 km</td>
<td>13.4±1.2 km</td>
</tr>
<tr>
<td>strike</td>
<td>93°</td>
<td>248°</td>
</tr>
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<td>dip</td>
<td>53°</td>
<td>16°</td>
</tr>
<tr>
<td>rake</td>
<td>130°</td>
<td>72°</td>
</tr>
</tbody>
</table>

Figure 4.1 shows the seismicity in the region, including the locations of the two events studied here. This area of Southern California includes the location of the 'big bend' in the San Andreas fault, where its strike rotates nearer an east-west direction. This rotation is associated with the development of thrust faulting with an oblique component. This thrusting produces a fold belt consisting of the San Gabriel and Santa Monica Mountains. Many of the thrusts are moderately dipping, east-west striking faults, such as the Hollywood and Raymond faults, but there are a variety of strikes in other orientations (Hauksson & Jones 1989). Some strike-slip faults following the regional NNW trend of the San Andreas fault have also been mapped. However, the intermediate-sized thrust events
often do not break the surface, and so cannot be mapped directly. As a consequence many earthquakes in the greater Los Angeles area occur on previously unknown blind thrusts such as the 1995 Northridge earthquake (Hauksson, Jones & Hutton 1995, Song, Jones & Helmerberger 1995). The earthquakes in table 4.1 are predominantly thrust events, with neither breaking the surface.

Figure 4.1.
Map of southern California with seismicity for the period 1964-1993. Earthquake magnitude is represented by size of circle. Epicentres of Whittier Narrows (1987) and Sierra Madre (1991) earthquakes are highlighted by dense seismicity due to aftershocks. (Wald 1992)

For both earthquakes I obtain a focal mechanism using the relative amplitude method. I also show that Wald (1992) misidentified the surface reflection for the Sierra Madre earthquake as sP rather than pP. I produce forward models for both broadband and short period data, for both earthquakes. The results improve on the number of seismograms fitted at different azimuths, and on the quantitative closeness of fit compared to the studies listed above. The results imply that source duration is a major factor in the differences in seismogram complexity, and that it is possible to obtain a first-order division of seismograms
into simple and complex categories, depending on their source area and rupture speed.

One of the main motivations for choosing these events is that they were both thought to be within the same Fresnel Zone. Thus at teleseismic distances they should appear to come from the same source region within the spatial resolution of the seismic waves that can be transmitted through the Earth. With this control, path and source-structure effects could in principle be eliminated as a potential cause of a difference in complexity between two seismograms from the same station.

The Fresnel Zone (or volume in 3D) determines the the spatial resolution of a seismic wave at a given wavelength, as illustrated schematically in figure 4.2, and can be defined in several ways. Here we use the simple definition proposed by Kravtsov & Orlov (1980). The shortest wavelengths investigated here correspond to frequencies up to 2 Hz, i.e. the upper cut-off on the short-period record.

\[ |T(F, S) + T(F, R) - T(R, S)| < \frac{1}{2} f^{-1} \]

where \( T(F, S) \) is the travel time from \( F \) to \( S \), \( T(F, R) \) from \( F \) to \( R \), \( T(R, S) \) from \( R \) to \( S \) and \( f \) is the dominant frequency of the signal. At teleseismic distances \( T(F, R) \approx T(R, S) \). For a separation of 23 km and layer velocity of 6 km/s, \( T(F, S) \approx 4 \). For \( f = 2 \) Hz, the right hand side is less than the left. This would imply that the Sierra Madre earthquake source is not within the Fresnel volume of the Whittier Narrows event. Here I investigate whether the assumption that
CHAPTER 4. Complexity due to Source Duration

path affects are negligible and do not contribute to the variation in complexity, is in fact correct, given that this result.

4.2 Focal Mechanisms

Before looking at the seismogram complexity it is necessary to obtain a reliable focal mechanism solution for the two earthquakes. This is needed to fit the amplitude and shape of synthetic seismograms to the recorded data, and in particular correctly identify the surface reflections when the relative amplitude moment tensor method is used. The rationale for using the RAMP method is that alternative methods, such as the CMT method of determining focal mechanisms by modelling long period body waves, are known not to be reliable in certain situations (Buchanan 1998), and do not quote uncertainties about the best-fitting solution. In particular CMT solutions may not be reliable for shallow-depth events such as those investigated in this chapter (section 2.7). The RAMP method includes both broadband and short-period information where available, and in principle has a higher resolution than the 50-70 s period bandwidth used by the CMT. The RAMP solutions for the two events are presented below.

4.2.1 Sierra Madre

Short-period and broadband seismograms were selected by quality as listed in table 4.2. Seismograms at particular azimuth and epicentral distances were selected where the signal to noise ratio for surface reflections was adequate. This data set still represents a larger suite of seismograms than previous studies. Following this selection procedure, data were available for up to 9 GDSN stations. The relevant seismograms are shown in figure 4.3a.

The short-period seismograms are comparatively simple at a variety of azimuths, and show that it is possible to identify both P and one surface reflection (that turns out to be pP, as demonstrated later). To demonstrate this the relative amplitudes must be tested against the competing hypotheses of a pP or sP arrival. The shallow depth of the earthquake in table 4.1 is confirmed by the short P-pP time of around 5 s on figure 4.3. Earthquakes at this depth generally
do not require a significant volumetric component (Dziewonski & Woodhouse 1983, Pearce & Rogers 1989). It is therefore reasonable to assume as a null hypothesis that the displacement field can be modelled using the double couple equivalent body force representation. The seismograms examined in this chapter are all consistent with this null hypothesis, so alternative radiation patterns were not examined. The five broadband seismograms presented in figure 4.3 are the only ones available for the short-period data presented.

First I test the hypothesis that the visible surface reflection is sP. This implies that pP is small, and that the earthquake would be at a shallower depth. A quantitative test of this hypothesis needs measurements of the relative amplitudes of the different phases. Amplitudes (in arbitrary units) and polarities measured on the individual seismograms are listed in table 4.2. After testing all possible double couple focal mechanisms for the relative amplitudes calculated from the measurements listed in table 4.2a no possible solutions are found to be compatible with these observations if the phase is sP. Thus we can formally reject the hypothesis that the phase is sP.

If instead we assume that the surface reflection is pP (table 4.2b), then we obtain a variety of potential solutions compatible with the observations, as illustrated in figure 4.4. First consider the short-period solution. This is consistent with a reverse fault of dip between 20° and 60°, slip direction between 40° and 110°. The strike is poorly constrained, indicated by the range of possible orientations of the vector at each point, but the focal mechanisms are restricted solely to the thrust faulting regime. The compatible range of focal mechanisms are moderately dipping with a possible oblique slip component. The CMT solution is compatible with this range of possible solutions.

An additional constraint on the best-fitting double couple focal mechanism can be obtained by considering the area under the broadband pulse for P, pP and sP. This is more consistent with the moment tensor determination, since the scalar moment is proportional to the area under the pulse (figure 3.6). In the above it is assumed that the amplitude alone provides a reasonable estimate of the radiation pattern for the short-period record. However, when the corner frequency of the radiated far-field pulse is greater than the passband of the short-period instrument, measuring a simple peak-to-peak amplitude does not give an accurate measurement of the phase amplitude at the source. This is
Figure 4.3.

(a) Short-period and (b) displacement broadband seismograms for the Sierra Madre earthquake, used in the relative amplitude method.
### Table 4.2. Relative amplitude measurements for the 28 June 1991 Sierra Madre earthquake

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<td>45.0°</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>
Figure 4.4.
Focal mechanisms compatible with the polarities and relative amplitudes shown in (a) Table 4.2a for short-period Sierra Madre seismograms and (b) 4.2b for broadband seismograms.
because on the seismogram the energy is spread over several cycles and thus the relative amplitudes can be in error if a phase amplitude is measured on a single cycle. The broadband record has no such distortion.

Several broadband recordings are available for the Sierra Madre earthquake (table 4.2c). These have been converted to a standard broadband response prior to plotting on figure 4.3, as described in section 2.3. On this record, the phases overlap more than the short-period record, hence requiring larger error bounds for the reflected phases, thereby offsetting the potential gain from using the area under the pulse. The main advantage of the broadband record is therefore the more accurate determination of the P-wave radiation pattern. The broadband recordings (using the area under the pulses) produced a smaller range of focal mechanisms than was obtained using short-period, containing a sub-set of the focal mechanisms compatible with the short-period records, and again the CMT solution.

In summary, the Sierra Madre earthquake can be fitted with source parameters that agree with the quoted ISC/CMT solution quoted in table 4.1. It is not possible to fit the data if the surface reflection phase is sP, so this hypothesis is rejected. The broadband data provides the strongest constraint on the focal mechanism, primarily due to the more accurate determination of the P-wave pulse area. The short-period record provides stronger constraints on depth because the pP phase can be uniquely identified. Solutions compatible with both the short-period and broadband records have strike=93 ± 10°, dip=43±15° and rake=130 ± 20° (figure 4.4b).

4.2.2 Whittier Narrows

The short-period Whittier Narrows seismograms are clearly more complex than those of the Sierra Madre earthquake (figure 4.5a). This event was larger than the previous one, so there are one or two additional stations on both the short-period and the broadband record for this earthquake. Nevertheless, it is not possible to identify a surface reflection in the coda of the P wave at any of the stations. The broadband record shows a clear large amplitude, long duration P pulse (figure 4.5b). Both are greater than the values for the Sierra Madre earthquake. In particular the Harvard scalar moment is larger \(0.84 \times 10^{18} \text{Nm}\).
CHAPTER 4. Complexity due to Source Duration

compared to $2.8 \times 10^{17} \text{Nm}$ (Wald 1992). The duration for this earthquake is considerably longer than the 1s duration needed for the relative amplitude measurements to be made on the short-period records. Therefore no analysis of the short-period data could be carried out as in the previous section for a shorter P-wave pulse.

![Figure 4.5.](image)

(a) Complex short-period records and (b) displacement broadband recorded seismograms for the Whittier Narrows earthquake.

Using the broadband data alone, and measuring the area under the pulse for the P-phase and possible pP, as above, I obtained a large range of focal mechanisms compatible with the observed amplitudes (figure 4.6). The data used to constrain the vectorplot are listed in table 4.3. The surface reflections are not easily identifiable on the broadband record as they interfere with P, so large amplitudes bounds, based on the maximum amplitude of the coda, are listed in the table.
In summary, the longer source duration makes it harder to resolve the phases \( P, pP \) and \( sP \), even on the broadband record, for an event of this depth. This represents an upper bound to the resolution of the RAMP method, and shows how an extended source can generate very complex seismograms, even for the relatively simple kinematic Savage model considered here, as proposed in the last chapter (section 3.2). This complexity degrades the confidence limits in the focal mechanism.

**Table 4.3.** Relative amplitude measurements, measured from broadband recordings for the Whittier Narrows earthquake.

<table>
<thead>
<tr>
<th>Observation</th>
<th>( \Delta )</th>
<th>Azimuth</th>
<th>( \mathbf{P} )</th>
<th>( \mathbf{pP} )</th>
<th>( \mathbf{sP} )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>( \mathbf{P} )</td>
<td>( \mathbf{pP} )</td>
<td>( \mathbf{sP} )</td>
</tr>
<tr>
<td>GRFO 85.3°</td>
<td>30.2°</td>
<td>+</td>
<td>8.0</td>
<td>10.0</td>
<td>U</td>
</tr>
<tr>
<td>SCP 32.4°</td>
<td>66.1°</td>
<td>+</td>
<td>20.0</td>
<td>30.0</td>
<td>U</td>
</tr>
<tr>
<td>ZOBO 69.1°</td>
<td>128.0°</td>
<td>+</td>
<td>18.0</td>
<td>22.0</td>
<td>U</td>
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<tr>
<td>BDF 83.1°</td>
<td>114.0°</td>
<td>+</td>
<td>26.0</td>
<td>34.0</td>
<td>U</td>
</tr>
<tr>
<td>MAJO 80.0°</td>
<td>307.4°</td>
<td>+</td>
<td>7.0</td>
<td>10.0</td>
<td>U</td>
</tr>
<tr>
<td>BOCO 90.3°</td>
<td>115.6°</td>
<td>+</td>
<td>6.0</td>
<td>10.0</td>
<td>U</td>
</tr>
<tr>
<td>BJI 90.8°</td>
<td>321.4°</td>
<td>+</td>
<td>5.0</td>
<td>9.0</td>
<td>U</td>
</tr>
</tbody>
</table>

The focal mechanism for the Whittier Narrows event is not as well constrained as the Sierra Madre focal mechanism, but it is nevertheless consistent with the thrust nature of faulting as reported by Harvard and NEIC in the ISC bulletin (figure 4.6). For these intermediate-sized earthquakes we might expect the CMT method to give a good result because it utilises longer periods for an assumed point source. The solutions I obtain for the Whittier Narrows event at the higher frequencies considered here have considerable variation in strike and rake, but the dip is constrained to a thrust or strike slip event. The poor constraint is due to the lack of identifiable polarities in the surface reflection readings, and the resultant wide amplitude bounds that were placed on most readings due to the complexity of the seismograms.

For both events, the thrust nature of the event means that most teleseismic stations tend to lie within the same (anti-nodal) region of the focal sphere for \( P, pP \) and \( sP \), and the radiation pattern is not adequately sampled. The focal mechanism could in principle be constrained better using regional data, but the interpretation may not be improved significantly since the paths of the \( P, pP \) and \( sP \) cannot be assumed to be indistinguishable. In summary, the combination of moderately dipping thrust focal mechanisms of source pulse duration longer
than 1s can lead to a very complex seismograms on the short-period record, and very poorly constrained focal mechanism on the broadband record. The solution is given in table 4.1.

![Figure 4.6.](image)

Focal mechanisms compatible with the polarities and relative amplitudes shown in table 4.3 for broadband Whittier Narrows seismograms

### 4.3 Forward Modelling

Having established a range of potential focal mechanisms for the two earthquakes, I now examine whether or not the variations in complexity seen in the seismograms can be reproduced using the simple source model of Savage (1966). I will do this by comparing observed short-period and broadband seismograms with synthetics as described in the previous chapter. The displacement broadband seismograms for both earthquakes appear simple, so that in principle a measure of fault size can be obtained by matching the length of the pulse. I begin with the Whittier Narrows earthquake because it is the more complex. Since
there is no published evidence that this earthquake had a fast rupture velocity, I thus assume a typical rupture velocity of 2.2 kms$^{-1}$ (0.6Vs at the source) and a typical stress drop of 1 MPa (Douglas 1997).

4.3.1 Whittier Narrows

Synthetic broadband seismograms for the Whittier Narrows broadband seismograms in figure 4.7 show a good match of amplitude and duration for a circular fault of source radius 12 km. The synthetics were computed for both velocity structures listed in table 4.4 for the regional Southern California average, and for the local Los Angeles basin. In fact, the broadband recordings are insensitive to the variation in the near-source velocity structure between a Southern California or Los Angeles Basin Models. However I used the Los Angeles basin model as the Whittier Narrows event lies in the region. The good match of the synthetics to the data on figure 4.7 for the broadband record indicates that it is not necessary to have an elliptical rupture area. Having looked at several values of $t^*$, I chose $t^* = 0.3$ which reproduces the frequency content of the seismograms adequately, and is within the usual range (Douglas 1981).

| Table 4.4. Velocity models used in this chapter (Hauksson & Jones 1989) |
|-----------------------------|-----------------------------|-----------------------------|-----------------------------|
| P-Wave Velocity kms$^{-1}$  | S-Wave Velocity kms$^{-1}$  | Density kgm$^{-3}$          | Depth to Top of Layer km    |
| Southern California Model   |                             |                             |                             |
| 4.10                        | 2.37                        | 2.7                         | 0.0                         |
| 5.16                        | 2.80                        | 2.8                         | 2.0                         |
| 6.35                        | 3.67                        | 2.8                         | 5.5                         |
| 6.54                        | 3.78                        | 2.8                         | 16.0                        |
| 7.87                        | 4.54                        | 3.0                         | 32.0                        |
| Los Angeles Basin Model     |                             |                             |                             |
| 3.31                        | 1.91                        | 2.7                         | 0.0                         |
| 4.08                        | 2.35                        | 2.8                         | 2.0                         |
| 6.02                        | 3.47                        | 2.8                         | 5.5                         |
| 6.37                        | 3.68                        | 2.8                         | 13.0                        |
| 6.68                        | 3.86                        | 2.8                         | 16.0                        |
| 7.87                        | 4.54                        | 3.0                         | 30.0                        |
Figure 4.7. (a) Displacement Broadband and (b) short-period comparisons of real (upper trace) and synthetic (lower trace) seismograms for the Whittier Narrows earthquake.
For the short period seismograms a fault size of 12 km produces a very complex seismogram using the Los Angeles Basin source velocity model. The data cannot be fitted by the regional Southern California velocity structure in this case. We obtain a good match to the data with a depth of 14 km, i.e. indistinguishable from the ISC quoted depth. It is not possible to match every cycle, but the beam-formed EKA seismogram in figure 4.7b seismograms is better fitted by the model presented here than that of Douglas (1997) (figure 4.8), who used a depth of 16 km.

4.3.2 Sierra Madre

The Sierra Madre earthquake has a much shorter duration broadband seismogram and it is possible to see two distinct pulses corresponding to P and pP. This suggests a smaller source area than the Whittier Narrows earthquake, consistent with the smaller observed scalar moment. The best fit obtained here from forward modelling confirms this time that an elliptical source of semi-major axes 9 km by 5.5 km and a depth of 11 km (figure 4.9a) fits the data better than the circular fault of similar moment. This is consistent with an elliptical-shaped source area from local strong-motion records and the earthquake aftershock zone (Wald 1992). The inferred fault area is comparable to the area of aftershocks if a fault rupture velocity approaching $V_2$ is used, otherwise the observed pulse durations cannot be matched. The elliptical nature of the fault is independently confirmed by the steep rise of the P pulse at some stations, which cannot be
modelled by the Doppler effect on a circular fault rupture, as previously shown (Douglas, Hudson & Pearce 1988) (also chapter 3).

In this case the simpler Southern California velocity model (table 4.4) contains enough information to fit the short-period seismograms closely. Polarities, amplitude and time of P and pP are reproduced, at most stations. The more detailed Los Angeles basin model would also fit the data adequately, but it is not necessary to add this level of near-source path effect as an underlying cause of the seismogram complexity. Again, the simpler hypothesis is retained until the data require a more complicated model.

4.4 Comparison

The two earthquake studied here have very different seismogram complexities. The most significant difference is due to different rupture areas and rupture velocities. This can affect the complexity seen on short-period seismograms, mainly due to the duration, ellipticity and directivity of the rupture (chapter 3). For a Whittier Narrows-type source, figure 4.10 summarises how seismograms change from being simple to complex, depending on the rupture velocity and radius.

Using the synthetic seismograms generated for section 3.2 it is possible to create this predictive tool for seismogram complexity. For an earthquake source of known focal mechanism these graphs can define what we would expect the complexity of an earthquake to be for different source durations. These graphs are based primarily on whether or not it is possible to identify surface reflections. If seismograms had clear surface reflections then they were classed as simple. If surface reflections were not readily identifiable then the seismogram was classed as complex. Using the method of analysis in chapter 3 similar graphs could be developed for any source orientation. As I have done here with the elliptical source, other factors such as attenuation or directivity could be included to see how this moves the transition line. This gives a good representation of the transition from simple to complex seismograms and could be used to predict the complexity of seismograms for a similar earthquake with known source duration and rupture velocity.
(a) Displacement Broadband and (b) short-period comparisons of real (upper trace) and synthetic (lower trace) seismograms for the Sierra Madre earthquake.
The best fitting solution determined here for the Whittier Narrows earthquake lies to the mid-right of (a) in the complex region, whereas the fault area and velocity of the Sierra Madre earthquake lies in the simple portion of (b). For these particular earthquakes the effect of the source duration on the short-period instrument is the controlling factor in the difference in complexity of the seismograms.

The transition zone from simple to complex seismograms, at low rupture velocities, is at approximately 5 km. At a source layer velocity of 4.5 kms$^{-1}$ and a dominant frequency of 1 Hz, this is slightly longer than one wavelength. This would appear to indicate that on a short-period seismogram the limit of the seismologist’s ability to clearly identify P and the surface reflections is at the wavelength of the dominant arrival. At faster rupture velocities the dominant frequency of the signal is higher due to the short duration source. At lower frequencies we would expect the transition from a simple to complex seismogram to be at a larger radius and this is what is seen in figure 4.10.

The ISC epicentres for the Whittier Narrows and Sierra Madre earthquakes are only 23 km apart. There is also a depth difference of at least 3 km. The smallest structures resolvable on the short period record will be approximately $V/4f$ (Kravtsov & Orlov 1980) which is a maximum of 1.3 km at 1 Hz. Given the large known variation in near-source velocities on either side of the Sierra Madre Fault (between the Southern California and Los Angeles Basin models) it is likely that this will have an impact on the difference in the seismograms. This distinction apparent after completing the modelling exercise in the last section. Thus we cannot say that the travel paths over teleseismic distances are the same, as postulated in the introduction to this chapter. The primary evidence that this assumption is incorrect is that, even accounting for the differences in duration, differences in structure either side of the Sierra Madre fault are required to provide the best fit to the data. Figure 4.11 shows predicted ray paths for the direct and surface reflected phases in the two source region velocity structures. In particular the low average velocity of the Los Angeles Basin (Whittier Narrows epicentre) produces multiple reflections, and larger mode conversions of S to P, when compared to the Southern California structure (Sierra Madre epicentre). Although these effects are second order here, they broadly confirm the predictions of the forward modelling exercise carried out in chapter 3.
Diagram presenting how the complexity of a seismogram is dependent on rupture velocity and fault dimensions for (a) a circular fault and (b) an elliptical fault.
Figure 4.11.
Diagram of phase arrivals from the left Sierra Madre and right Whittier Narrows earthquakes. This shows the extra arrivals from the more complex Los Angeles Basin Model.
4.5 Conclusions

The two events examined here produce radically different seismograms, mainly due to differences in source duration, but also due to differences in local velocity structure near the source. For both earthquakes, it is possible to place significant constraints on the focal mechanism, even when surface reflections are difficult to identify. The ISC/CMT solution for depth and focal mechanism is compatible with range of the source mechanisms found here for both events. The surface reflection on the Sierra Madre short-period seismograms can be identified uniquely as pP, and not sP, as incorrectly identified by Wald (1992).

The Whittier Narrows earthquake has a source area three times the size of the Sierra Madre earthquake. This is the primary cause of the variation in complexity seen on short-period records, as the longer duration source of the Whittier Narrows earthquake is greater than the period of the short-period instrument. This produces many starting and stopping phases and creates a very complex short-period signal.

As the two sources are 23 km apart, compared to the Fresnel zone of 1.3 km, the differences between the Southern California and Los Angeles Basin near-source velocity structure may be significant at teleseismic distances, and contribute to the additional complexity seen from the Whittier Narrows earthquake. The main reason for this difference in structure is the basin bounding Sierra Madre fault that lies in between the two earthquake locations.
Chapter 5

Complexity due to Near-Source Structure

5.1 Introduction

The aim of this chapter is to examine complexity due to the effect of the near-source structure on the seismic signal. This is done by examining a suite of seismograms from the 25 October 1995 Caspian Sea earthquake. The earthquake occurred in a band of seismicity striking WNW-ENE across the central section of the Caspian Sea (figure 5.1), in the area of the Apsceron-Balkhan sill where earthquakes with both normal and reverse fault mechanisms occur near to the relatively aseismic southern Caspian Sea (Priestley, Baker & Jackson 1994). The motivation for this case study was the very simple and large amplitude S-wave signal observed at EKA from this event. A large simple S is comparatively rare at teleseismic distances (Marshall et al. 1975) and its presence indicates a simple earthquake source and low attenuation in the source-receiver path. By having this as a control it can be inferred that the complexity seen in other seismograms must be due to path effects such as mode-conversions, multiple reflections, attenuation or the direction of the ray-path. The Caspian Sea area is also near the former Soviet Union Central Asian Nuclear test sites, and hence data is available for all of the Blacknest array sites which were sited at teleseismic distance from this region. The results are of potential significance for the discrimination of earthquakes and explosions using depth, although this is not
the main focus here.

The conversion of seismic energy from one mode to another occurs when a seismic wave is incident upon a boundary. Depending on the angle of incidence and the acoustic impedance contrast across the interface, energy is partitioned into both reflected and refracted compressional (P) and transverse (S) waves. Exact analytical solutions for this partition are given by the Zoeppritz's equations (Richter 1941). If the incident and converted wave velocities, and the depth of the converting boundary, are of suitable values, then the converted waves can arrive within the P-wave portion of the seismogram. This can lead to complexity if these additional arrivals are between P and the surface reflections (see section 3.3). Here I concentrate mainly on complexity caused by the conversion of an S-wave to a P-wave at a boundary beneath the source (S-to-P). Such conversions have been attributed by various authors (Thirlaway 1963, Barley 1977, Barley et al. 1982, Douglas, Stewart & Richardson 1984, Iidaka et al. 1990, Douglas, Sheehan & Stewart 1992) to the presence of structure at depth, for example subduction slab boundaries at depths within the crust up to the 660 km discontinuity itself. The presence of such discontinuities in the Caspian Sea area is consistent with the geological evidence for a remnant thrust zone caused by subduction in the late stages of closure of the Tethyan Ocean (Jackson & McKenzie 1988).

5.2 Data Analysis

The locations of this event from the different agencies are listed in Table 5.1. For example the PIDC assigned a depth of 30 km for this event. Figure 5.2 shows a selection of data available from the PIDC, normalised to have the same P-wave time, along with their interpretation of the near-surface reflections marked T1 and T2. Some stations show a single large P arrival only (DBIC, SPITS), some a small P with larger phases in the first 30 s of the coda (stations labelled HFS, NORES, NPO) while others consist of two identifiable phases within the first 30 s (stations ULM, WALA). These phases are in addition to the T1 and T2 arrivals picked by the PIDC, from which their 30 km depth estimate was obtained.
Figure 5.1. ISC seismicity 1964-1995 for the Caspian Sea area. The ISC location for the 25 October 1995 event is plotted with the focal mechanism determined by relative amplitudes and used for forward modelling in this study.
The USGS National Earthquake Information Center (NEIC) also gives a relatively shallow depth of 33 km but this is the default depth of the location inversion, and hence implies an essentially unconstrained depth. The ISC depth of 50 km utilises 24 pP phases, but the ISC is unable to verify depth phase identifications because it does not have access to waveform data. The ISC Bulletin of Events information is also not available until approximately two years after an earthquake. These inconsistencies in the earthquake depth due to the different identification of pP due to the complexity of the seismogram is a cause for concern. In terms of discriminating between earthquakes and nuclear explosions, the depth is critical, since deep events must be natural earthquakes. Hence, inaccuracies in depth determination can lead to the earthquakes being labelled incorrectly as potentially 'suspicious' in an area near a nuclear site.

First, using the relative amplitude method, I aim to positively identify the surface reflections and obtain a reliable focal mechanism and depth. Then by forward modelling I show how mode conversions in the near-source structure, the direction of source rupture propagation, and the position of the station on the radiation pattern, are all necessary to explain the observed complexity.

In this investigation I examine seismograms from the four short period, medium aperture seismometer arrays reporting to AWE Blacknest (table 5.2). These arrays produce seismograms with excellent signal-to-noise ratio and provide good azimuthal coverage for earthquakes in the Caspian Sea region. In addition, I use three stations from the primary network of the PIDC, to improve azimuthal
Figure 5.2.
Seismograms from PIDC with times of analyst-picked surface reflections. T1 represents pP. T2 represents sP.
coverage. Many of the seismograms show an anomalous phase after P which causes many of the seismograms to have apparent complexity. This type of phase has been defined previously as ‘bP’ and assigned to an S-to-P mode conversion beneath the source (Kind & Seidl 1982). The seismograms will be presented later in this chapter in comparison with the forward modelling results, but here figure 5.3 shows schematically the possible origin of this phase. Here I will attempt to determine the depth of the earthquake, the depth of the refractive boundary within the mantle below the source, and the velocity contrast across this interface. Other properties of the near-source structure are taken from the results of Clark & Graham (1989).

On the short-period seismograms (presented later in figure 5.8) the phase bP is seen clearly, approximately 10 s after P on both GBA and YKA seismograms, and can also be identified as a small arrival on EKA, PDY and WRA. The phase is not visible at BGCA or VAF. The correlograms, as defined in section 2.3, which identify energy arriving with the correct vector slowness, show a possible arrival on all four array seismograms (figure 5.4a), although it is very small at WRA. Any interpretation of the bP phase must explain its range of amplitudes, the constant arrival time after P of 10 s, and why it is not seen on seismograms from other earthquakes in the same region.

Here I examine six possible hypotheses for the origin of the bP phase.

1. P wave arrival from a second earthquake in the same region
2. Near-receiver P-to-S conversion
Figure 5.3. Diagram showing the origin of phases mentioned in this chapter. Above-source structure is simplified for clarity.

3. Above-source S-to-P conversion

4. Multiple reflections of P at both receiver and source

5. Surface reflection pP or sP

6. S-to-P conversion at an interface below the source (Figure 5.3)

If the arrival bP is from a second earthquake (1), it must also have an identical hypocentre within the resolution of the Fresnel zone, because as it beamforms constructively with the same slowness and azimuth as the first arrival. If the second event was in the same source region then a second S arrival should be visible at EKA. Only one very simple S is seen at EKA (figure 5.8) so this hypothesis can be rejected. If phase bP is due to a near-receiver mode conversion (2), then it would exist only at one station, since otherwise a similar interface would be required at exactly the same depth under many stations. The structure beneath the four arrays is independently known (Arora 1969, Underwood 1967, Hasegawa 1971, Parks 1967), and does not predict such conversions. In addition, the simple S-wave seen at EKA without preceding mode conversions of P to S under the station, but a clear bP phase, independently confirms that hypothesis (2) can be rejected.
Figure 5.4.
(a) Correlograms showing arrivals of phased energy across the array, and the identified arrival bP (b) Synthetic ground displacement function at EKA for a steeply dipping fault with focal mechanism as determined in this chapter
If the arrival bP was a reflection or mode conversion at an interface above the source (3), this could be consistent with the arrival times of bP. However, for this earthquake, we shall see later in this chapter that the best fitting source model has a downward propagating rupture, enabling the upgoing and downgoing waves to be distinguished by their frequency content. Figure 5.4b shows the displacement function (that is the far-field pulse radiated into the mantle) generated using the method of Douglas et al. (1972) at station EKA for the best fitting source function to be determined later in the chapter. The downward propagating rupture produces a short duration P and bP, and a longer duration pP phase. This is needed to explain the lower amplitude of short-period pP energy in the seismograms, because such a rupture would produce a pP phase of duration greater than 1 s and a corner frequency outside the pass-band of a short period instrument (Douglas et al. 1988). The large amplitude S also indicates that the path to EKA has low attenuation and hence rules out the possibility that the higher frequency components have simply been attenuated. We can therefore reject hypothesis (3).

Finally, hypothesis (4) can be rejected because (a) bP is seen at only some stations and (b) it has the same amplitude as P at YKA. Multiple reflections of P near the source or receiver would be unlikely to have such a large amplitude at any station, and would not have such a large azimuthal variation.

The two remaining hypotheses are (5) that bP is a surface reflection (either pP or sP) or (6) an S-to-P conversion at an interface below the source. These two possibilities are now examined in more detail.

### 5.3 Is bP a Surface Reflection?

The interpretation of bP as a surface reflection can be tested using the relative amplitude method (section 2.7). Here I test three possible interpretations to explain the observed phases bP, P, and sP, at least one of which can be seen on each seismogram. In the RAMP method null observations (the absence of a phase) can be as important as clear arrivals (see section 2.7). In fact vital information about the source mechanism is contained in the seismogram from both the presence and absence of a phase. If a phase is not visible then maximum
amplitude bounds can be placed on the individual phases in the P-wave portion of the seismogram at the time it is predicted to arrive.

If (a) we postulate that the phase labelled bP on figure 5.4 is instead the surface reflection pP, then the large arrival labelled pP on the seismogram occurs too late after bP to be the expected sP phase for a Poisson's ratio in the normal range. If (b) the large arrival labelled pP on figure 5.4 is correctly identified, then bP is not a surface reflection. If (c) the labelled pP phase is sP, with a small pP before it, then it is hard to explain bP. These three possibilities are now tested quantitatively using RAMP, using the phase polarities and amplitudes, in arbitrary units in table 5.3, where rows (a)-(c) correspond to the three alternatives listed above.

For model (a), when phase bP labelled on figure 5.4 is interpreted instead as pP, we obtain no compatible fault orientations from a 5° grid search of strike, dip and slip angle, so bP is unlikely to be pP. For model (c), when phase pP labelled on figure 5.4 is instead assumed to be sP, we also obtain no compatible focal mechanisms. For model (b), when the large arrival is assumed to be pP, we obtain two compatible focal mechanisms (figure 5.5). Moreover, the focal mechanisms for this interpretation are very tightly constrained due to the use of several stations with high-quality data, including some with nodal arrivals for P, pP and sP. Thus we can have a high confidence in the interpretation of these phases as presented in figure 5.4, and hence reject hypothesis (5) that the phase labelled bP is a surface reflection.

The compatible focal mechanisms shown in figure 5.5 both involve steeply dipping fault planes, and are very similar to each other. One of these is used to model the velocity broadband data, because it produced a better match to P, pP and sP. Synthetic velocity broadband seismograms were generated using the source velocity model determined in the next section from the short-period records (fig. 5.6). This model includes the S-to-P mode conversions under the source, but the broad-band seismograms are relatively insensitive to the mode conversion S-to-P, and we do not consider this here.

The best fitting source model is most tightly constrained by the S-wave train at EKA. Figure 5.7 shows two models (b) and (c) for the observed seismogram (a). These are (b) a downward-propagating line source and (c) a circular fault. The
Table 5.3. Relative amplitude measurements for (a) P large, pP large shallow source, sP small; (b) P large, pP large (deeper source); (c) P large, pP small, sP large. U is uncertain polarity. Each also has 2 S-wave amplitude readings at EKA S, min 15.0 max 50.0 U; Sh min 0.0 max 100.0 U.

<table>
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<td>U</td>
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<td>U</td>
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<tr>
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<td>U</td>
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<td>U</td>
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<td>15.0</td>
<td>U</td>
<td>10.0</td>
<td>20.0</td>
</tr>
<tr>
<td>BGCA</td>
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<td>18.0</td>
<td>U</td>
<td>0.0</td>
<td>19.0</td>
<td>U</td>
<td>0.0</td>
<td>20.0</td>
</tr>
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</table>
Two compatible focal mechanisms

Figure 5.5.
Fault-plane solutions from relative amplitudes of P, pP, sP obtained assuming a P-wavespeed in the source layer of 8.1 km s$^{-1}$. Each compatible solution is shown as a short vector parallel to the strike of the fault plane and radiating from a point defined by the slip angle and dip of the fault plane.
Figure 5.6.
Broadband data for the 29 October 1995 earthquake (upper traces) and synthetics (lower traces) as generated using the fault plane solution derived from relative amplitudes.
line source clearly matches the S-wave arrival better, since it produces no sS, as observed. The best fit source has a length of 4 km and depth of 48 km. This depth is more similar to the ISC and CMT depths listed in table 5.1. Additional information from the S-wave train is not central to the argument for this chapter, and is described in Appendix A for completeness.

Figure 5.7.
(a) Observed seismogram at EKA; (b) Modelled Sv at EKA from a line rupture; (c) S and sS from circular source rupture

The line source model fits the P-wave portion of the velocity broadband seismogram at a range of azimuths very well, if the more steeply-dipping of the two nodal planes are identified as the true fault plane (figure 5.6). This highlights the ability of the Doppler effect, where it is present, to distinguish the true fault plane from the auxiliary plane, since it introduces asymmetry to the radiation pattern (section 2.7). Many of the stations (including EKA and YKA) are near
nodes in the best-fitting P-wave radiation pattern. The best fitting focal mechanism and depth is listed in table 5.4. It is incompatible with the hypothesis that the phase labelled bP on figure 5.4 is a surface reflection, implying that the earthquake has a depth greater than that reported by PIDC and NEIC.

Table 5.4. Source parameters used to generate synthetic seismograms in this chapter

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dip of fault plane</td>
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</tr>
<tr>
<td>Strike of the fault plane</td>
<td>300°</td>
</tr>
<tr>
<td>Angle between strike and direction of slip</td>
<td>130°</td>
</tr>
<tr>
<td>Depth of focus (initial rupture point)</td>
<td>48 km</td>
</tr>
<tr>
<td>Length of line rupture</td>
<td>4 km</td>
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</table>

5.4 S-to-P conversions at an interface below the source?

Having rejected all of the alternative explanations for the bP phase and obtained a reliable focal mechanism solution and source rupture model, I now show that bP can plausibly be modelled as an S-to-P conversion at an interface below the source with the same source model. The downward propagating rupture required to explain the absence of sS at EKA on figure 5.7 predicts a long duration, small amplitude upgoing S pulse and a short-duration large amplitude downgoing S-wave. This in turn would produce larger than normal S-to-P conversions at any sharp boundary below the source. In this section I will demonstrate that the combination of a rapid, downward propagating rupture and a sharp velocity contrast below the source can produce significant complexity in the seismogram between P and the surface reflections, and hence explain the bP phase.

Another contributing factor to the large bP amplitude on the short period record is the fact that many stations are at a node in the radiation pattern for P, and an antinode for S. Thus any mode conversions below the source have a strong input energy from S, and a weak P-wave signal, consistent with the weak signal hypothesis of Douglas (1967) and Douglas et al. (1973b). Figure 5.8
shows the short-period seismograms for this event, compared to the best-fitting forward model for a downward propagating rupture for the same source model and velocity structure as used in section 5.3. Figure 5.8 shows the large S-wave arrival at EKA that is a critical minimum condition for strong S-to-P conversions below the source.

The remaining parts of Figure 5.8 show the P-wave portion of the seismogram, including bP and the surface reflections. The phase bP is more clearly seen at GBA and YKA, where the amplitude of P (20 and 5 nm respectively) was relatively small. Hence both stations lie close to the nodal planes. At EKA we might expect P to be small if S is large, but in fact P has a relatively large amplitude (800 nm) compared to YKA, implying an unusually high Q averaged along this particular ray path. Nevertheless, both EKA and YKA are near-nodal, as confirmed by the best fitting solution of figure 5.6. The first-order observations of prominent and absent phases, and frequency content, of both the short period (figure 5.8) and broadband records (figure 5.6) are well matched by the same source model and near-source velocity structure. Some second-order residual additional sources of 'noise' are not accounted for by this model.

The synthetic seismograms of Figures 5.6 and 5.8 were generated assuming the velocity structure for the Caspian Sea region of Clark & Graham (1989), modified to include the large velocity contrast between 8.1 and 8.5 kms\(^{-1}\), at a depth of 60 km below the source (108 km below the surface). This combined model is listed in table 5.5. A good match to the observed pP-P and bP-P times was achieved using a depth of 48 km, consistent with the ISC and CMT solutions (table 5.1).

In summary the short period record confirms the hypothesis that bP is an S to P conversion below the source, and validates the source model proposed from the velocity broadband record. In this case the broadband record constrains the source more closely, and the short-period record constrains the near-source velocity structure, indicating the complimentary nature of the different seismograms.
Figure 5.8.
The short period waveforms (upper seismograms) recorded at the stations used in this chapter for the 29th October 1995 Caspian Sea earthquake. Lower traces are the synthetic data generated using the method described in the text. The position of each station is shown on a lower hemisphere focal projection in figure 5.6.

Table 5.5. Velocity model used to fit the broadband and short-period data in this chapter

<table>
<thead>
<tr>
<th>P-Wave Velocity (kms$^{-1}$)</th>
<th>S-Wave Velocity (kms$^{-1}$)</th>
<th>Density (kgm$^{-3}$)</th>
<th>Layer Thickness (km)</th>
</tr>
</thead>
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5.5 Conclusions

The complexity in seismograms for the 29 October 1995 Caspian Sea earthquake can be modelled using S-to-P conversions at an interface at a depth of 108 km below the surface. All other alternative hypotheses can be rejected by different aspects of the available data. In particular, the hypothesis that phase bP is a surface reflection cannot produce double-couple S and P radiation patterns that fit the available data. After formally rejecting this hypothesis, the focal mechanism is well constrained using relative amplitudes, and has a nodal plane which corresponds to a steeply-dipping normal fault with a strike-slip component, consistent with the range of focal mechanisms reported for the area.

The absence of an sS phase at EKA implies a downward propagating line rupture that introduces a strong Doppler effect. This enables the more steeply-dipping nodal plane to be identified as the true fault rupture plane. The downward propagating rupture produces a larger S than usual at P wave nodes, and hence amplifies the amplitude of S-to-P conversions relative to P and pP. The S to P mode conversions below the source are also enhanced relative to P when the overall amplitude of the seismogram is small, such as at YKA and GBA.

The best fitting source model has a depth of 48 km. The 30 km depth given by the PIDC apparently resulted from the misidentification of bP as pP on some seismograms, and also the misidentification of other phases.

The velocity interface responsible for S-to-P mode conversions below the source is in the upper mantle, at a depth below the base of normal continental lithosphere in this area. The interface may plausibly be the top of an old subducting slab under the central Caspian Sea, but this hypothesis cannot be tested in detail using the methods applied here. Further work on deep seismic reflections or seismic tomography would have to be carried out to determine the nature of the interface required to fit the data here. The results of this chapter highlight how a detailed study of all phases is essential in correctly identifying surface reflections and explaining complex phases not predicted by standard travel-times. They also confirm the critical role of near-nodal stations in identifying the causes for seismogram complexity.
Chapter 6

Evidence of Complex Source Rupture

6.1 Introduction

Having looked, in the previous chapters, at various factors effecting complexity using a simple source model (Savage 1966), I now move on to look at the possible contribution of dynamic source complexity in the following two chapters. In this short chapter I will present results from two earthquakes (table 6.1) that appear to show evidence for a complex dynamic source rupture. For each of these earthquakes I will highlight how complex seismograms may indicate that complex dynamic rupture may be occurring and that a standard representation of the seismic source cannot explain the high frequency energy seen on broadband seismograms. This will lead on to a study in the following chapter of a method of generating complex source functions with additional high frequencies which are similar to those which would be needed to reproduce the complexity shown in this chapter.

6.2 Analysis of the Seismic Source

It is easy in some seismograms to observe additional high frequency energy especially if the overall size of the signal is low and perhaps the high frequency energy
Table 6.1. Parameters for the two earthquakes discussed in this chapter given by the ISC

<table>
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<th>Egypt</th>
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</tr>
<tr>
<td>Location</td>
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</tr>
<tr>
<td>Depth (km)</td>
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<td>11.1</td>
</tr>
<tr>
<td>Body-wave magnitude</td>
<td>5.7</td>
<td>6.1</td>
</tr>
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</table>

is riding on top of a lower frequency signal. This can be seen quantitatively however in the earthquake source spectrum. This is the Fourier transform of the earthquake time-series and plots the amplitude of the signal at each frequency contained within the signal. This frequency spectrum of the P wave radiation is simply the Fourier transform of the P wave component of the earthquake seismogram. By using a teleseismic broadband seismogram this approximates the far-field pulse and the source spectra can be obtained from approximately 0.1 to 4 Hz. This is the range over which the broadband instrument response is the same for all frequencies.

A standard seismic source approximated to a point source in the far-field has a moment-rate that has a constant value over a duration known as the rise time, \( \tau_r \) (figure 6.1). This is the Haskell source model. A moving source is the superposition of several of these point sources and the actual slip at a point on the fault can be described by a ramp function in the displacement as the rupture passes. In the far-field the displacement pulse will be the convolution of two boxcar functions of width \( \tau_r \) and, the apparent rupture duration, \( \tau_d \). This produces a Haskell fault model which is a trapezoid shaped pulse. Due to directivity effects the apparent duration will change depending on the source-receiver azimuth and this in turn will change the pulse length and amplitude (Douglas et al. 1988).

In the frequency domain this Haskell fault model is a sinc function for which the far-field amplitude spectrum can be expressed as

\[
|A(\omega)| = gM_0|\text{sinc}(\omega\tau_r/2)||\text{sinc}(\omega\tau_d/2) |
\]

Plotted on a log-log plot this gives the standard amplitude versus frequency plots
which will be used in this chapter (figure 6.2). In this standard example three slopes can be identified corresponding to no amplitude fall off at low frequencies, then $\omega^{-1}$ at intermediate frequencies and $\omega^{-2}$ at higher frequencies. In practice only 2 portions can usually be identified corresponding to constant amplitude and then the fall-off of $\omega^{-2}$. Any variations in the source model away from the standard Haskell Source should be evident in the spectra curves. If there are larger amplitudes over a range of higher frequencies then this would be seen as a shallower slope to the $\omega^{-2}$ portion. If there were higher amplitudes at specific frequencies than a upward curve might be observed in the spectra.

Figure 6.2.
Amplitude spectrum for the $\omega^{-2}$ source model (Shearer 1999), where $\tau_d$ is the apparent rupture duration, $\tau_r$ is the rise-time of the source and $\omega$ is the frequency.
6.3 Egypt

The 22 November 1995 Gulf of Aqaba earthquake was one of the largest in the Middle East to be instrumentally recorded (Rabinowitz & Steinberg 1998). The focal mechanism determined by Foster & Jackson (1998) corresponds to a strike-slip faulting event with a moderate normal component. The regional tectonics suggests that the fault plane lies in the nodal plane that strikes in a north-south direction parallel to the faults forming the Gulf of Suez. The ISC bulletin states that the event had an extended fault rupture which may have consisted of 2 periods of moment release and this is what lead to the earthquake being of interest as this source complexity may be evident in the seismogram. As figure 6.3 shows the short-period waveform for this event is very complex. Part of this is due to the fact that the source is longer than 1 s and so is outside the pass-band of the recording instrument. But, as can be seen in figure 6.4, the broadband seismogram consists of a number of discrete arrivals and there is evidence for a higher frequency signal riding on top of the longer period arrivals. This may not be visible on the other seismograms because the higher frequency signal is of a much smaller amplitude compared to the longer period component. The frequency content of the P-waveform also shows evidence for higher frequency arrivals at 1 Hz. As was mentioned previously the amplitude of a normal Haskell fault model will fall off with frequency by $\omega^{-2}$. However it can be seen that the fall off with frequency in the range 1 to 10 Hz only has a gradient of 1. This may indicate that there is more higher frequency energy than would normally be expected for a simple source model.

The other three seismograms and amplitude versus frequency curves presented in figures 6.5 to 6.7 also contain additional energy at around 1 Hz which is not present in the noise. As this energy is visible at a number of stations then it might be possible to assume that it is coming from the source region and may show complex rupture on the fault due to material heterogeneities. I observed similar results from an earthquake in the Chiapas region of Mexico but this was at a depth of 175 km, which is in the region of deep earthquakes, whose source processes are still uncertain. Thus it is difficult to link complexities in the seismic signal directly to fault rupture processes.
Figure 6.3.
Short Period P-wave seismogram at YKA, Canada. Axes are amplitude (nm) against time (sec)
Figure 6.4.
Broadband seismogram with the instrument response removed for station ALE, Canada. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
CHAPTER 6. Evidence of Complex Source Rupture

Figure 6.5.
Broadband seismogram with the instrument response removed for station ERM, Japan. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
CHAPTER 6. Evidence of Complex Source Rupture

**Figure 6.6.**

Broadband seismogram with the instrument response removed for station WMQ, China. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
Figure 6.7.
Broadband seismogram with the instrument response removed for station LSZ, Zambia. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
6.4 Gansu

The 21 July 1995 earthquake was a small seismic event in western China in an area of strong seismicity. This event would not have been unusual if it was not for the strong high frequency signal which was observed on several seismograms. As the magnitude of this event is only $m_b = 5.7$ the overall size of the broadband seismogram is small enough for these high frequency arrivals to be visible. They can be clearly seen on the three seismograms presented here in figures 6.8 to 6.10. In the seismic traces in figures 6.8 and 6.10 the high frequencies are only visible in the first few seconds of the initial arrival but they may also be visible after the larger arrivals (which may be surface reflections). However on figure 6.9 the higher frequency pulse can be seen contributing to each of the main arrivals. Badly placed ()'s

For this earthquake the frequency plots seem less reliable as the noise level is only slightly less than the signal, so the amplitude response of the P-wave may contain considerable noise. However a flattening of the P-wave amplitude response curve can be observed on figures 6.9 and 6.10. At this point the P-wave is at a higher amplitude than the noise. This flattening of the curve is away from the normal $\omega^{-2}$ fall off that would be expected from a Haskell Model and may indicate the presence of high amplitude arrivals at about 1-2 Hz. This may be due to the breaking of heterogeneities on the fault plane which is investigated in the following chapter.

6.5 Summary

One possible reason for the complexity seen here is for the source to be made up of two separate fault ruptures. These ruptures are distinguishable at the resolution of the local data but when viewed at teleseismic distance they may not be resolvable and would appear to give a different focal mechanism. This would be because the second earthquake, which may happen only a few seconds after the first, occurs within its Fresnel Zone. These two events would be much closer than the Whittier Narrows and Sierra Madre earthquakes which were resolvable at teleseismic distances.
Figure 6.8.
Broadband seismogram with the instrument response removed for station ARU, Russia. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
Figure 6.9.
Broadband seismogram with the instrument response removed for station OBN, Russia. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
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Figure 6.10.
Broadband seismogram with the instrument response removed for station LID, Russia. Frequency response for the background noise in the seismogram before the first arrival, and frequency response of the P-waveform itself.
A situation like this occurred in the Mammoth Lake area of California and it provided some clues as to the possible mechanism of these type of earthquakes. In May 1980, $m_b > 6$, earthquakes occurred in the space of a few days. By August there had been 600 moderate-sized aftershocks. The fault system bounds the eastern side of the Sierra Nevada where it intersects with the Long Valley Caldera. It is in a similar extensional regime similar to the Basin and Range province (Given, Wallace & Kanamori 1982) and has a well documented volcanic history (Bolt 1993). For the initial large event the teleseismic $P$ is incompatible with the strike-slip determined from the local and regional stations. This could be due to complex multiple sources, artifacts or near-source structure. By using relative amplitudes it was possible to show that there was a significant volumetric component to the focal mechanism derived from the teleseismic data (Stimpson 1986) which would begin to explain the discrepancy in terms of a complex source, as RAMP is insensitive to local velocity structure. Given et al. (1982) propose that the differences in local and body wave data may be due to structural complexity under the caldera but this seems ruled out by Stimpson (1986). They also proposed that the event started off as a strike slip event and then moved to more oblique-slip corresponding to the two types of focal mechanisms obtained from the local data.

From trial and error fitting of the complex $P$-wave data, which is similar to the seismograms presented in this chapter, they obtained a good fit for two sources, 4s apart, in a multiple rupture. This would bring together the local and teleseismic data. In this case it has been proposed that volcanic intrusion produced the multiple events (Archuleta et al. 1982) and the volumetric component in the focal mechanism would seem to support this. Knopoff (1983) showed that many of the first motions were not consistent with a double couple source and that the epicentres of the swarm rose from 9 to 3.5 km, suggesting rising magmatic intrusion into pre-existing structures.

In this short chapter I have presented some interesting observations, from two earthquakes, which may indicate that the source rupture process itself is contributing to the complexity of the observed teleseismic $P$-wave seismogram. Observations of higher-than expected amplitudes, at frequencies of between one and 2 Hz may come from the source region. They are seen in both the time-series and the amplitude-frequency plot and may indicate the complex nature of these
particular seismic rupture processes. This could be modelled in a similar way to the main Mammoth Lake event which consisted of two pulses but the choice of source-time function is arbitrary in this case. In the following chapter I develop a different approach where the complexity is generated from a new numerical model for the seismic rupture process. Rather than fitting each wiggle of the seismogram I am seeking to generate source-time functions which contain the high frequencies seen in earthquakes presented here, but which develop from real sources and contain no assumptions about the number of source pulses.
Chapter 7

Source Modelling

7.1 Introduction

Having looked at a number of factors effecting P-wave complexity and shown using new case studies how these have contributed to shape of the P-wave short-period seismogram, I will now show how the complexity of the source rupture process itself can cause some of the complexity examined in chapter 6. This is not readily explainable by other processes.

For this, a number of assumptions about the nature of the earthquake source need to be made and these are described in the appropriate section. Even with these broad assumptions some understanding as to the nature of the source and its contribution to the teleseismic seismogram can be gained. Less attention is often given to source processes as a point or line approximation is generally used, especially at teleseismic distances, when the moment tensor of the earthquake or the regional Earth structure are being investigated. Here I offer the beginnings of a dynamic approach to the seismic source which constrains those parameters which are well known and allows others to evolve freely.

In the previous chapters the method used for generating synthetic seismograms (Douglas et al. 1972) employs a kinematic model for the earthquake source. It treats the earthquake source as a sum of radiating point sources with fixed rupture velocity and stress drop and pre-defines the fault plane as a circular or elliptical limit to which the point sources may propagate. I now introduce the
basis of the model where the source is allowed to evolve freely depending only on the strength of the medium and an imposed tectonic strain.

From the seismograms that were investigated in chapter 6 it can be seen that additional high frequency energy in the range 0.5 - 2 Hz is present in some seismograms. This is not readily apparent on broadband seismograms because for large earthquakes the small amplitude of this signal is masked by the low frequency component. An additional high frequency component to a seismogram has been explained by the presence of asperities on the fault plane (Das & Aki 1977a, Das & Aki 1977b). Das and Aki generated far-field seismic pulse shapes for a variety of simple asperity geometries and receiver azimuths which show that for a unilateral in-plane shear crack a range of source-time functions can be generated. These range from a simple triangle to complex pulses and those which include some type of harmonic ringing and is seen as an additional high frequency component which is seen on the spectra (fig. 7.1).

![Figure 7.1.](image)

(a) Far field seismic P-wave pulse shapes and (b) frequency content, generated by an in-plane shear crack. After Das & Aki (1977b).

As the number of asperities on the fault plane increases the complexity of the far-field pulse also increases. Theoretically any number of asperities could produce more and more complex pulses but Das and Aki limited their study to two so as to be able to resolve the effects of azimuth. They also noted that a low stress
drop can increase high frequencies in the seismogram. Although prescribing a certain number of asperities is useful I use random heterogeneity to get a better understanding of how real faults rupture during earthquakes. Other work which has shown increased high frequencies, but in the near field, is that of Nielsen & Tarantola (1992). In this they present near-field seismograms which show an additional higher frequency component (fig. 7.2) and they show an effect similar to that seen in broadband seismograms in the previous chapter with a higher frequency signal riding on top of a lower one.

![Figure 7.2.](image)

Near-field synthetic seismogram generated using a finite-difference model with dynamic rupture (Nielsen & Tarantola 1992).

Most of the complexity in these seismograms is caused by the interactions between the rupture at heterogeneities in the medium but their locations are not defined explicitly. Instead a random distribution of shear strength (or cohesion) was assigned in a given area with a percentage spread around a given mean strength. As I am not looking at the role played by each individual asperity, but at the observed seismograms in the far-field, this model of faulting is suitable for the purposes of this investigation. Unfortunately the code for the 1992 work was no longer available (Stefan Nielsen pers. comm.) so here I use a different numerical implementation of the same model. The numerical modelling technique, using finite differences to solve for real physical parameters, is simple to
implement on the limited computing resources available and a dynamic model can be achieved by building on existing numerical models that were available to me. I will outline the current implementation of the model first and then describe how this has been extended to produce a dynamic model.

7.2 The Numerical Model

This model is based on a numerical implementation of Nielsen & Tarantola (1992) by Fernando Nino, Bertrand Maillot and Patience Cowie called BEWE3D (Brittle-Elastic-Wave equation in 3 Dimensions). In this they simulate the birth and growth of fractures or faults in a brittle-elastic medium. Tectonic loading cycles can be applied to the edges of the grid. A Mohr-Coulomb fracture criteria is applied to each node to see which has exceeded its shear strength due to the tectonic loading. At this node shear stress is dropped due to a pre-defined friction law. Seismic radiation is attenuated away and then the fracture criteria is re-applied to see if the new static stress conditions cause other nodes to fail. When no more nodes are beyond the fracture criteria then that is the end of that 'earthquake' and another tectonic loading cycle is applied. Thus a model of earthquake recurrence and the growth of large scale faults over time can be mapped (Cowie, Sornete & Vanneste 1995). This had been done in 2D with loading applied either in-plane or anti-plane and was used to map faults on the surface in their lateral extent (Nino, Maillot & Cowie 1997). This software however did not take into account the possibility that the dynamic stress due to the seismic radiation may trigger further ruptures, as Nielsen & Tarantola (1992) did, rather than just the static stress. This would be an important addition to the seismic source in computing teleseismic seismograms. I want to focus on the evolution of a fault during one seismic rupture after loading (one 'earthquake'), so I have amended the numerical model to take account of the possibility of dynamic rupture (rupture moving at shear or possibly super-shear velocities) as well as allowing a rupture to evolve due to the static stress.
7.3 Simplifications

With such a complex phenomenon as seismic rupture simplifications and assumptions need to be made to allow for the modelling to be carried out in a reasonable time frame. These are all similar to those used in Nielsen & Tarantola (1992).

- The effects of gravity and pore fluids are ignored. This is reasonable as these are not known to have a great effect within the earthquake rupture.

- As I am dealing with time scales of an individual earthquake the study is restricted to the brittle-elastic regime and the plastic regime is ignored.

- Displacements at a broken node are considered to be infinitesimally small, which is realistic as they are one or two orders of magnitude smaller than the grid spacing of approximately 100 m.

- On failure, energy is released as a shear stress drop on the fault and the faults remain closed.

- Medium heterogeneity is defined as variation in cohesion at each node of the lattice. This makes it computationally easier and will be discussed later.

- Once a node brakes within one earthquake it remains broken.

- The boundary conditions are periodic.

7.4 Theoretical Model

Any model which attempts to simulate fault growth and elastic wave propagation needs to include several essential elements:

1. The equation of motion governing the evolution of the system

2. The constitutive relations that describe how the system behaves mechanically to stresses imposed. These include
• Fracture criteria which governs the fracturing of nodes in response to both static and dynamic stress
• Friction Law which controls the stress drop at broken nodes
• Specification of the heterogeneity in the model
• How the medium responds elastically due to the relationship between stress and strain.

**Equation of motion**

I will not derive the equation of motion for the displacement acceleration in the absence of external forces here. This can be found in Shearer (1999), p. 25. As we are assuming infinitesimal displacements then the acceleration of the displacements are proportional to the stress tensor and are related by Cauchy’s equation of motion

\[
\rho \frac{\partial^2 u_i}{\partial t^2} = \nabla_j \sigma_{ij}, \tag{7.1}
\]

where \(\sigma_{ij}\) are the stress elements in the displacement field \(u\) and \(\rho\) is the rock density. Also throughout this work \(\nabla_j \sigma_{ij} = \nabla_1 \sigma_{i1} + \nabla_2 \sigma_{i2} + \nabla_3 \sigma_{i3}\). This is the fundamental equation in this model and is solved using a finite difference approximation which shall be described later. In order to do this we need a constitutive relation linking stress to the strain \(\epsilon_{ij}(x, t)\).

**Constitutive relations**

**Stresses**

In the model we have an original tectonic loading and lithostatic pressure which provides a component of stress from before rupture, \(\sigma_{ij}^P(t)\), and a component from the time of rupture which is due to the stress drops at the broken nodes in the medium, \(\sigma_{ij}^F(x, t)\), where

\[
\sigma_{ij}(x, t) = \sigma_{ij}^P(t) + \sigma_{ij}^F(x, t), \tag{7.2}
\]
where the stress before rupture

\[ \sigma_{ij}^P(t) = \sigma_{ij}^P(0) + tf(i, j)c_{ijkl}\varepsilon_{kl}, \]

(7.3)
is the sum of stress due to the lithostatic pressure \( \sigma_{ij}^P(0) \) which is pre-defined and that due to the tectonic loading strain-rate \( \dot{\varepsilon}_{ij} \) applied during \( t \). \( c_{ijkl} \) are the elastic stiffnesses and \( f(i, j) \) is a function with values 1 or 0 specifying whether the component \( \dot{\varepsilon}_{ij} \) of the loading strain rates is applied or not.

Fracture Criteria

Previous work (Andrews 1976, Das & Aki 1977b) has used a Griffith fracture criterion (Griffith 1924) to determine if failure occurs at a node. This approximates the failure envelope on a shear stress against normal stress plot to a parabola (fig. 7.3) and works for tensile and compressive cracks. For compressive regimes this was modified to include the resistance of sliding by friction after the crack closes. This approximates to the Mohr-Coloumb fracture criterion in the compressive regime which I am using here. As a straight line this is easy to implement computationally and predicts where and when fracture will occur (Scholz 1990). This is an empirical fracture criterion and has been used successful in the previous model on which I am basing this work (Nielsen & Tarantola 1992, Nielsen et al. 1995). The criterion states that fracture will occur across a plane where the shear stress first reaches a value that depends on both the normal stress to the plane and the material dependent parameters.

\[ |\tau| = C + \nu\sigma \]

(7.4)

where \( C \) is the cohesion and \( \nu \) is a friction coefficient (both are material-dependent parameters). \( \tau \) is the shear stress and \( \sigma \) is the normal stress. Figure 7.3 shows how either an increase in the shear stress (and hence the radius of the Mohr circle) or a reduction in the normal stress (but not the differential stress \( \sigma_1 - \sigma_3 \)) would bring the circle nearer the failure envelope and cause rupture.

This can also be used to find the orientation of the fracture planes. The fracture plane is the plane where the difference between the shear stress and \( \nu \) times the
Mohr circles and the Coulomb and Griffith failure criteria. The friction coefficient $\nu$ where $\nu = \tan \phi$ and the cohesion $C$ ("$S_{\text{max}}$" in the software) are medium properties that define the failure envelope (bold diagonal line). $(S_1, S_2, S_3)$ are the principal components of the total stress $\sigma_{ij}$; $(S'_{1}, S'_{2}, S'_{3})$ represent a stress required for rupture.

Figure 7.3.

normal stress is maximum. Figure 7.4 shows that the vectors normal to these planes form an angle $\beta$ with the direction of maximum stress, where $\tan(2\beta) = -\frac{1}{\nu}$. $\beta$ can be either positive or negative and the actual fault plane is chosen as the positive angle. As I am looking at the radiated energy from the fault the negative would have been as good.

This fracture criterion is applied first after the initial loading sequence to determine the initial point of rupture, and then at each time step during the stress drop and seismic energy radiation to check for further rupture. This may be due to either the seismic energy or static stress.

**Stress Drop**

In the original BEWE3D code the friction on the fault was not velocity dependent and the stress drop increased smoothly from 0 to a pre-assigned value over a period of time which was controlled by the strength of the seismic radiation required. As seismic radiations were not allowed to cause ruptures then this time period was not important. After each rupture the medium was allowed to stabilise to a static state before other ruptures were tested for. Each node was in effect a separate earthquake on a separate section of fault.

Having added a dynamic rupture component seismic radiations are allowed to
cause rupture and the fault propagates dynamically over the period of one earthquake. Each node is now simply the location of some stress drop on the fault plane during one earthquake. The stress drop at each node still follows the same cosine curve (figure 7.5), but as the rupture propagates, the timing of stress release is effectively dependent on the rupture velocity even though we have a constant stress drop at each node. This is consistent with seismological observations.

The evolution of shear stress on the fracture plane can then be written as:

$$\tilde{\sigma}_{ij}(x^b, t > t_0) = g(t)\tilde{\sigma}_{ij}(x^b, t_0) . \quad (7.5)$$

where $x^b$ is a location where failure is occurring, $t_0$ is the onset of failure, $\tilde{\sigma}_{ij}$ is
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\[ g(t) \]

Figure 7.5.

Time dependence of the deviatoric stress decay during failure. \( t_0 \): time of onset of rupture; \( t_r \): relaxation time. \( t_r \) and \( \Delta \sigma \) are the two free parameters of the friction law.

The deviatoric stress defined as

\[ \bar{\sigma}_{ij} = \sigma_{ij} - \frac{1}{3} \delta_{ij} \sigma_{kk}, \quad (7.6) \]

and \( g(t) \) is a cosine function going from 0 to the maximum stress drop \( \Delta \sigma \) over a relaxation time \( t_r \). The relaxation time \( t_r \) and the fixed amount of stress drop \( \Delta \sigma \) are the free parameters of the friction law. Obviously the stress cannot drop more than is present at the node before the stress drop occurred.

It is the deviatoric components of the stress tensor that evolve due to the prescribed friction law and the homogeneous parts evolve freely due to the equations of motion. Within the deviatoric component it is the shear component that is prescribed (by the friction law) and the normal stress that evolves freely. In the global co-ordinate system however the whole stress tensor effectively evolves freely as each component contains both a shear and normal sub-component. The orientation of the fracture depends on both the state of stress before rupture due to the imposed strains and on the cohesion.

Unlike friction laws which use a characteristic length scale (Nielsen et al. 2000, Knopoff, Landoni & Abinante 1992), here this is replaced by a characteristic time of decay \( t_r \), which controls the duration of the rupture and how much seismic energy is generated. This must be carefully chosen so as to allow radiations to
cause rupture but without unrealistic numerical dispersion. Within this model this is quantified by a relaxation time scale $t^*_r$ where 1 is a quick stress drop and strong radiation and 10 produces a much weaker seismic radiation and

$$t^*_r = t_r \frac{C_s}{L} \quad (7.7)$$

where $C_s$ is the speed of shear waves and $L$ a characteristic length of fracture. $L$ is set as the length of spatial sampling, $\Delta x$, because one broken node represents the smallest amount of break on the fault which is resolvable.

**Linear elasticity**

The friction law is applied at broken nodes but at all other points in the medium, during tectonic loading and wave propagation, stresses and strains are linearly related through Hooke's law.

$$\sigma_{ij}(x,t) = c_{ijkl}(x)\epsilon_{kl}(x,t). \quad (7.8)$$

We will consider homogeneous isotropic media, for which

$$c_{ijkl} = \lambda \delta_{ij} \delta_{kl} + \mu (\delta_{ij} \delta_{kl} + \delta_{il} \delta_{jk}) \quad (7.9)$$

does not depend on position $x$ and the only heterogeneity is the cohesion $C$. $\lambda$ and $\mu$ are the Lamé parameters. Incorporating equation 7.8 and 7.9 into the equation of motion (equation 7.1) yields an equation that describes seismic waves. This model hence explicitly calculates the seismic energy associated with failure at a collection of nodes which represent a fault plane. Having outlined this theoretical model I now discuss the numerical implementation of this model building on the work of Nino et al. (1997) in using BEWE3D for establishing a static state.
7.5 Numerical Implementation

The numerical implementation of the model falls into six subsections which I will deal with in the order that they are carried out within the code. This order represents a complete earthquake cycle; loading, rupture and wave propagation.

7.5.1 Initialisation of Parameters

All displacements are set to zero. So are the deviatoric components of stress, with the overall stress, $\sigma_{ij}$, set to the background state of stress $\sigma_{ij}^0$, which is pre-defined.

7.5.2 Loading Cycle

The strains are applied at a pre-defined strain-rate in the medium and the stresses updated. These strains can be at any angle to the edge of the medium either in plane or antiplane to the grid. Although the software can calculate in full 3D, computing limitations meant that loading was done on a 2D slice. This provides sufficient information as to the formation of in-plane fractures (fig. 7.6) and the associated seismic radiations and so a full 3D implementation is not necessary. The algorithm checks at each time step if any node has reached failure (according to the Mohr-Coulomb fracture criteria). When one node is detected as having reached failure then it moves onto the rupture cycle.

7.5.3 Finite-Difference approximation towards the static state

The rupture cycle will stop when a static state has been reached. Before checking for this a finite difference approximation to various fields (stress, displacement, strain-rate) is calculated using the equation of motion (equation 7.1) and the constitutive relations. This is done on a discrete staggered grid drawing on the results of Nielsen & Tarantola (1992) and Nielsen et al. (1995).
It is not necessary to calculate all the components of velocity, stress and strain at every node on the grid. Figure 7.7 shows where the components of these fields are necessary, using a conceptual grid of half the size of the actual computational grid. For example, $\epsilon_{11} = \frac{\partial u_1}{\partial x_1}$ is known between the nodes where $u_1$ is known, and vice versa.

Instead of solving the equation of motion (7.1) as it is, it solves

$$\rho \frac{\partial v_i}{\partial t} = \nabla_j \sigma_{ij} - \alpha |\nabla_j \sigma_{ij}| \text{sign}(v_i),$$

(7.10)

where $v(x, t) = \frac{\partial u_i}{\partial t}(x, t)$ is the velocity of displacements, because it is more practical to deal with the velocities than the displacements. The second member on the right hand side is a force proportional to $\nabla_j \sigma_{ij}$ acting against the velocity of displacement: it attenuates the elastic waves by dissipating the kinetic energy in the medium, hence allowing it to reach the static state. $\alpha$ is an attenuation coefficient whose best value, after testing for accuracy against efficiency, appears to be 0.47 (Nino et al. 1997). The space and time steps of discretisation will be respectively noted $\Delta x_i$ and $\Delta t$. We start with $v(x, t)$ known everywhere on the lattice at the time $t$. 

**Figure 7.6.**
Sketch of in-plane loading used in this study. Thick arrows represent the applied loading.
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Local strain rates

The model then computes the local strain rates (not the tectonic strain rates) through a finite-difference approximation:

\[
\dot{\epsilon}_{ij}(\vec{x}, t) \approx \frac{1}{2} \left[ \frac{v_i(\vec{x} + \vec{e}_j, t) - v_i(\vec{x} - \vec{e}_j, t)}{\Delta x_j} \right. \\
\left. + \frac{v_j(\vec{x} + \vec{e}_i, t) - v_j(\vec{x} - \vec{e}_i, t)}{\Delta x_j} \right]
\]  

(7.11)

where vectors \( \vec{e}_i, (i = 1, \ldots, d) \) are defined on figure 7.7
Constitutive relations

- Hooke’s law: At unbroken (i.e., elastic) points, the model uses the linear elasticity equations (7.8) and (7.9) which can be combined to give:

\[
\sigma_{ij}(\bar{x}, t + \Delta t) = \sigma_{ij}(\bar{x}, t) + \lambda \delta_{ij} \Delta t \dot{\varepsilon}_{kk}^f(\bar{x}, t) + 2\mu \Delta t \dot{\varepsilon}_{ij}^f(\bar{x}, t) \quad (7.12)
\]

- Stress Drop: First the plane of failure has to be calculated at the broken node. This is at an angle $\beta$ to the direction of maximum stress (fig. 7.4). The size and direction of the principal axes of stress are simply the eigenvalues and eigenvectors of the stress tensor. At a failed node the deviatoric stress is then dropped by removing the homogeneous component from the stress tensor to calculate the deviatoric stress. This is then rotated onto the plane of failure. The amount of stress drop is calculated by using the curve from figure 7.5 depending on the pre-defined total stress drop and relaxation time. The deviatoric stress is then rotated back to the global coordinate system, and the homogeneous component added as I am working with the total stress tensor. The resulting static stress field is then calculated using finite-differences (Virieux 1986).

Stress Divergence

Through a finite-difference approximation, the model computes the stress divergence appearing in the equation of motion (7.1),

\[
\frac{\partial \sigma_{ij}(\bar{x}, t)}{\partial x_j} \approx \sum_{j=1}^{3} \frac{\sigma_{ij}(\bar{x} + \mathbf{e}_j, t) - \sigma_{ij}(\bar{x} - \mathbf{e}_j, t)}{\Delta x_j} \quad (7.13)
\]

7.5.4 Further Rupture

The original BEWE3D code allowed the grid to reach a static state after the rupture of each point. There were two time scales. One was the time of the loading which stops when the rupture starts and the other is the time taken to go through the finite-difference routine towards the static state. In the new dynamic stress drop code proposed here the time that each rupture takes within the finite-difference routine is also important as each point will start and stop radiating
energy at different times. Thus at each time within the finite-difference routine, as the seismic energy is propagating, the Mohr-Coulomb fracture criterion is applied at each node to see if any further ruptures occur due to the propagating P and S-waves or by the static build up of stress. These ruptures then start radiating energy at this time. This is an improvement on the original BEWE3D code as these nodes would not be able to radiate seismic energy until after the static state had been reached. Each point is allowed to rupture once during a single earthquake and a time limit has to be placed on rupture. As Nielsen & Tarantola (1992) found it is not possible to stop ruptures in a homogeneous or weakly heterogeneous medium, so a time-limit needs to be imposed. This allows the rupture to grow to realistic length and then for all the seismic radiation to propagate out of the grid. The rupture can be stopped by unrealistically high areas of strength but this was not used in the results presented here.

7.5.5 Time extrapolation

Eventually, the model extrapolates the velocities \( \mathbf{v} \) to the next time step \( t + \Delta t \) by the finite-difference approximation

\[
v_i(\mathbf{x}, t + \Delta t) \approx v_i(\mathbf{x}, t) + \frac{\Delta t}{\rho(\mathbf{x})} \left[ \frac{\partial \sigma_{ij}(\mathbf{x}, t)}{\partial x_j} \right] \left( \frac{\partial \sigma_{ij}(\mathbf{x}, t)}{\partial x_j} \right) \text{sign}(v_i) \]

and the displacements

\[
u_i(\mathbf{x}, t + \Delta t) \approx u_i(\mathbf{x}, t) + \Delta t \mathbf{v}_i(\mathbf{x}, t + \Delta t). \]

Notice that, even in an anisotropic medium, both the stress and strain tensor are symmetrical (i.e., \( \sigma_{ij} = \sigma_{ji}; \epsilon_{ij} = \epsilon_{ji} \)). Therefore, only six components have to be declared in the program (in three dimensions).

7.5.6 Static state

Once the rupture has been terminated the stresses continue to propagate through the medium until a static state is reached. The static state is defined by the
equilibrium of forces. At each time step $t$ we calculate the maximum of the absolute value of the components of forces throughout the lattice:

$$f(t) = \max \left( \left| \frac{\partial \sigma_{ij}}{\partial x_j}(\bar{x}, t) \right| \right) \text{ over all nodes } \bar{x} \text{ and all components } i$$  \hspace{1cm} (7.16)

We define a reference force $f_{\text{ref}}$ against which this maximum unbalanced force (7.16) should be compared; it is simply the minimum applied force on the system, as deduced from the tectonic loading (i.e. the forces that can be deduced from the tectonic strain rates).

The static state is then numerically defined as

$$\frac{f(t)}{f_{\text{ref}}(t)} \leq \gamma$$  \hspace{1cm} (7.17)

where $\gamma$ is a small number which determines the accuracy of the static state (in practice, $\gamma$ or stat.acc is set to a value between $10^{-5}$ to $10^{-7}$). If it is too high, accuracy will not be satisfying, if it is too low, the static state calculations become too long.

The following section will present results from the static displacement in the near field to show that this correctly predicts the static field as shown in Nielsen & Tarantola (1992). It will then investigate increasingly more complex fault models showing the effect this has on the far-field pulse. I will then show how combining seismograms from various take-off angles a representation of the teleseismic seismogram can be obtained.

### 7.6 Testing the Model

Firstly to test that the code is running correctly and that it is giving realistic results, I compare it with the results of Nielsen & Tarantola (1992). The parameters used in this modified BEWE3D test are given in table 7.1. The fault developed slightly longer than the Nielsen model and produced a reverse fault of approximately 500 m in length, that ruptured in approximately 0.04 s. This is 100 m longer than the Nielsen fault rupture and may be due to a difference in
initial strain-rates, which Nielsen & Tarantola (1992) do not specify. Figure 7.8 shows the final vertical displacement induced by the rupture in the medium, using the Nielsen and modified BEWE3D models. The two lobes of upward and downward movement are divided by the visible fault trace. Both are similar shape and indicate that I can have confidence in the new dynamic model presented here, given that it produces realistic fault displacements. From this static test on a small scale I now present the results of extending the model to far-field pulses. This simply involves scaling-up the finite difference grid dimensions to produce fault lengths of several kilometres rather than the 200 m shown in figure 7.8.

![Figure 7.8.](image)

(a) Vertical component of the static displacement field using the Nielsen Model. Purple is downward - Green upward (Nielsen & Tarantola 1992) (b) Vertical component of the static displacement from my model. Blue upward - red downward.

### 7.7 Dynamic Model

Nielsen & Tarantola (1992) looked at the small scale faulting in the near-field, I now scale this up to examine faults that would produce detectable seismic radiations at teleseismic distances. These would be of the order of $m_b \approx 5.0-6.0$, with a fault size up to approximately 4 km. I present here P-wave far-field source pulses which, due to the differences in the physical parameters of the medium in
Table 7.1. Input parameters for new dynamic rupture model to simulate the static displacement results of Nielsen & Tarantola (1992)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grid size</td>
<td>2048 m</td>
</tr>
<tr>
<td>node size</td>
<td>8 m</td>
</tr>
<tr>
<td>strain angle w.r.t. x3</td>
<td>45°</td>
</tr>
<tr>
<td>strain rate</td>
<td>1x10^{-7} Nm^{-1}</td>
</tr>
<tr>
<td>initial stress</td>
<td>1x10^6 Pa</td>
</tr>
<tr>
<td>angle of internal friction</td>
<td>45°</td>
</tr>
<tr>
<td>max stress drop</td>
<td>1x10^6 Pa</td>
</tr>
<tr>
<td>cohesion</td>
<td>180000 Pa</td>
</tr>
</tbody>
</table>

which the faults occur, present a range of complexity in the far-field. In each of the uses of the dynamic model I have maintained some common finite-difference and source parameters and these are shown in table 7.2. The lattice dimensions are chosen so as to give an overall grid size of approximately 25 km and the time dependent parameters are model derived so that the S-wave does not propagate more than one grid node per time step.

Table 7.2. Input parameters common to each of the results presented in the results section

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grid size</td>
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</tr>
<tr>
<td>node size</td>
<td>100 m</td>
</tr>
<tr>
<td>strain angle w.r.t. x3</td>
<td>45°</td>
</tr>
<tr>
<td>strain rate</td>
<td>1x10^{-7} Nm^{-1}</td>
</tr>
<tr>
<td>initial stress</td>
<td>1x10^6 Pa</td>
</tr>
<tr>
<td>angle of internal friction</td>
<td>45°</td>
</tr>
<tr>
<td>max stress drop</td>
<td>1x10^6 Pa</td>
</tr>
<tr>
<td>cohesion</td>
<td>180000 Pa</td>
</tr>
</tbody>
</table>

The four sets of results presented here build up the complexity of the fault from the length of a single node in a homogeneous medium to an extended fault approximately 9 km in length in a heterogeneous medium. Each of them thus represents the P-wave displacement, recorded over time at the same position, in the far-field region of the grid. Figure 7.9 shows how the near-field displacement is proportional to the seismic moment whereas the moment-rate function is proportional to the far-field displacement. The far-field displacement presented here is thus proportional to the source-time function which gives a record of
the moment-rate at the source. The displacement function shown here is the component of the displacement in the direction of the propagation of the seismic wave. To obtain this the vertical and horizontal components of the displacement obtained from the model are rotated to the direction of, and tangential to, the wavefront, with the initial arrival positive in this quadrant of the radiation pattern. This is why in each of these examples the dominant energy is the initial P-wave arrival because the S-wave energy is in the tangential component.

![Graph showing near-field displacement, far-field displacement, and moment over time.](image)

**Figure 7.9.**
The relationship between near-field displacement, far-field displacement and moment (Shearer 1999).

In the first result I examine the failure of one grid node which represents a fault, 100m in length within a homogeneous grid. This single node is specified by allowing one weak node at the centre of the grid to rupture over a period of 19 time steps. The main feature of figure 7.10 is the single simple positive displacement which corresponds to the seismic energy released over a short period of time from the single node. In this example the displacement does not return to 0 after the passing of the seismic wave and may indicate that there may be some S energy in this component or that the receiving position is not wholly within the far-field. However this does not mask the single simple short initial arrival.

Figure 7.11 shows how if the fault is allowed to propagate into the surrounding medium to approximately 4 km in length then the initial arrival is longer as one would expect from an elongated fault. But after the initial arrival, which is dominated by the displacement from the time of maximum stress drop at the
Figure 7.10.
Radial displacement measured in the far-field from the rupture of a single node in a homogeneous medium.
node, there is seismic energy which is lower in amplitude and of slightly longer
duration. This displacement comes from the lower amplitude seismic energy
generated as the stress drop at the node decreases due to the friction law. This
energy is the sum from various points on the fault which have broken at different
times and thus it has an irregular shape compared to the first arrival. Thus even
for a simple line rupture some complexity in the source-time function has been
introduced.

\begin{figure}
\centering
\includegraphics[width=\textwidth]{figure711}
\caption{Radial displacement measured in the far-field from the rupture of
a 4 km fault in a homogeneous medium.}
\end{figure}

Having looked at a homogeneous medium, heterogeneity is now introduced in
the central region of the grid in a square of side length 10 km centred on the
middle of the grid. The central node is defined to be much weaker than the other
nodes so rupture begins at this point and propagates to other nodes depending
on the strengths of adjacent nodes. The heterogeneity is a randomly generated
grid in which the strength (cohesion) of the nodes can vary by up to 50% of
the maximum value. As can be seen in figure 7.12 this produces a seismogram with a shorter initial period of displacement and a complex tail to the pulse. The shorter initial displacement may be due to the fault being shorter than the corresponding fault in the homogeneous medium, as stronger nodes may have prevented it from propagating as far. The complex tail to the pulse may be generated by the seismic radiation from various nodes, which are at different stages in their stress drop, interacting with each other. Again with this source-time function there is a static displacement, which is because the observer is not fully into the far-field. This however does not affect the overall shape of the complexity seen.

![Figure 7.12](image)

*Figure 7.12.*

Radial displacement measured in the far-field from the rupture of a 4 km fault in a heterogeneous medium.

In the final example a zone of hardness five nodes by five nodes in size was placed adjacent to the weak central node. The cohesion of this node was large compared to the other nodes and could not be broken by the propagating fault.
This is similar to the single asperity model of Das & Aki (1977b). Figure 7.13 shows that this zone of hardness produces a much more complex seismogram with the initial displacement having higher frequency arrivals on top of the longer duration pulses. This is due to the irregular contour of the fault which propagates around the zone of hardness. The fault also does not propagate smoothly in time so the initial displacement is of a longer duration than previous examples. The heterogeneous medium also adds complexity to this source-time function.

Figure 7.13.
Radial displacement measured in the far-field from the rupture of an irregular fault propagating around a zone of hardness adjacent to the central node.
7.8 Summary

Here I have presented some basic results which show that with just a slight variation in some of the medium parameters very complex source-time functions can be produced using a dynamic source rupture model. This has been developed from previous numerical codes which dealt only with the static displacement. The medium parameters are just one of the near limitless number of variables that can be introduced into this dynamic rupture model. Others would be the angle of the initial tectonic loading on the grid, the time for the stress drop to occur at each node and the angle of internal friction of the fracture criterion at each node. Here I have shown that by fixing these parameters to produce a realistic earthquake rupture process the complexity in the seismograms, such as those shown in chapter 6 could be due to non-uniform fault rupture, controlled by the irregular structure in the source region.

7.9 Limitations and Problems

With this work I was not able to generate synthetic seismograms from the source time functions produced, which had been my original plan. I was unable to merge the new source-time functions with existing synthetic seismogram techniques mainly due to the differences in source code and problems with computer architecture. For instance the original code for generating synthetic seismograms, used in this thesis, was written in Fortran4 in 1972 and was originally designed for punch-cards. The dynamic rupture code was written in Fortran77 but used a complex message passing interface to ideally work on a parallel computer which was not available to me. I was unable to merge these two structures. With more time and close support from the original code authors further progress could be made.
Chapter 8

Discussion and Conclusions

8.1 Discussion

This thesis has investigated the various factors that can contribute to the complexity of the P-wave seismogram, using a number of different techniques. The aim of this chapter is to draw these various analyses together and hence form a strategy for resolving the factors that can contribute to observed seismogram complexity. First I present a discussion of the most pertinent findings.

Synthetic seismograms were presented (chapter 3) that demonstrated, under certain conditions, the extent to which a number of source and path effects contribute to the complexity. Although similar to the work presented in Douglas, Young & Hudson (1973) here I extended the work using more realistic earthquake source orientations. This shows that even for complex sources it is possible to separate out individual factors effecting the seismogram. In particular (section 3.2) the duration of the earthquake source (controlled by both the rupture area and the rupture velocity) is the most important factor in the complexity of the short-period seismogram. This has implications for looking at large magnitude events, where the large source duration will swamp the seismogram and mask any other features. However, even when the short-period seismogram is complex, it is usually possible to identify individual phases within the P waveform on broadband data. Even where phases are small or indistinguishable any constraint which forward modelling can place on these phases is useful when
CHAPTER 8. Discussion and Conclusions

applying the Relative Amplitude Method. This was noted in a series of papers by Pearce (1977, 1980) using short-period but it applies here to broadband, as the study of the Whittier Narrows earthquake showed.

When there is a long duration source, the forward modelling identified the cause of complexity in the short-period P waveform to be the starting and stopping phases produced by the kinematic source model used. These phases are more noticeable on synthetic seismograms than in real data and may indicate that real earthquakes do not have the same abrupt start and end points as the Haskell Fault Model. When P is nodal, S-to-P conversions at interfaces beneath the source can produce complexity in the short-period waveform. For a circular fault these conversions are small (section 3.3). If however the rupture has a high aspect ratio and behaves as a line source then the size of these arrivals can be as large as direct P and the surface reflections (section 3.4) making the short-period seismograms uninterpretable. I would agree with Douglas et al. (1992) that these S-to-P conversions are rare in real data due to the number of factors which must come together for them to be seen but as they produce arrivals of appreciable size they are a significant factor in test-ban monitoring.

The comparison of the Whittier Narrows and Sierra Madre earthquakes (chapter 4) was originally proposed to identify differences in source effects, since both events were believed to occur within the same source region, as seen from teleseismic distances. But although the two earthquakes have been shown to be outside each others' Fresnel Zones the longer duration of the Whittier Narrows source was identified as the primary reason for the variation in complexity seen in the short-period seismograms. It was not necessary to postulate a more complex source model for this event. However, the more complex near-source structure within the Los Angeles sedimentary basin is likely to have added significantly to the complexity seen in a number of the Whittier Narrows seismograms, although it was not possible to resolve this component uniquely. These two earthquakes show that care must be taken when comparing seismograms from different earthquakes whilst using one as a control. Here the simple Sierra Madre seismogram had originally been taken as an indication of simple earth structure for this particular region and that another earthquake up to 25 km away could be compared to it. This has been shown not to be the case and it would seem that looking at the Whittier Narrows compared to its aftershocks, which occur within 3 km,
CHAPTER 8. Discussion and Conclusions

could tell us more about the Whittier Narrows source. Hauksson & Jones (1989) had started this by comparing moment tensors from the whole earthquake sequence but did not compare seismograms.

For the Caspian Sea earthquake (chapter 5) it was possible to determine, using a suite of seismograms, a number of factors contributing to seismogram complexity. Critically, a simple S-wave seen at a single station demonstrated that the source must have been simple. Thus any factors contributing to seismogram complexity must have been along the source-receiver path. This technique of obtaining a simple control from S is not one that is generally used, except by Marshall et al. (1975). If S is more widely looked for at teleseismic distances then this could be a powerful tool for finding simple earthquake sources. Explosion sources should not generate S-waves.

Again this earthquake showed that the relative amplitude method is a powerful technique to correctly identify the surface reflections on the seismogram. Although the Relative Amplitude method has not gained much usage outside the Edinburgh and Blacknest group it should be used for analysing suspicious events and to get a positive identification of P, pP and sP to obtain a reliable depth. By using this method here, followed by forward modelling, the complexity observed on some seismograms was shown to be due to S-to-P conversions from an interface beneath the source. At the P-wave node these arrivals are relatively large and occur between P and the surface reflections. I have shown that this caused the mis-identification of P by the PIDC. Use of relative amplitudes would have prevented this.

These observations also confirm the 'weak signal hypothesis' (Douglas et al. 1971, Douglas et al. 1973b) where more complex signals occur at stations at a node in the P-wave radiation pattern. In contrast, those at an anti-node for P were relatively simple. The best fitting source model required a down-going line rupture to explain directional variations in amplitude of P and S waves. This Doppler effect, due to a downward-propagating rupture near a P-wave node, also contributed to the large S-to-P mode conversions seen.

For the two earthquakes presented in chapter 6 it was not possible to explain all the complexity seen in both the short-period and broadband records using a simple kinematic model. This is often the case with larger shallow events
where, in general, a dynamic source model is needed. By modifying an existing numerical code, it was possible to derive very complex source functions for a dynamic source model (chapter 7). Such source functions could in principle be compared to the broadband record. However, it was not possible to derive synthetic seismograms at short period within the time frame of this thesis, due to a fundamental incompatibility encountered between the two source codes needed. Even if this was possible the complexity of the sources used would prevent a wiggle-for-wiggle comparison with real data. What would be more useful would be to see if these new sources can produce the overall frequency content of real data and from this an idea of the asperities on an particular fault could be obtained.

By including real source physics in the source model, very complex far-field source pulses can be achieved by including material heterogeneity along the fault. The complexity of the pulse increases considerably as the heterogeneity is increased. Such complex source pulses could produce very complex seismograms, with similar additional high-frequency energy to that seen in chapter 6. Such modelling would be especially significant where a high-Q path allows higher frequency arrivals to be recorded at the receiver. Many of the world's seismograms are located specifically to utilise high-Q paths from the world's principal earthquake regions so more complex source models would be required.

In this thesis I have employed the principal of Occam's Razor which states "one should not increase, beyond what is necessary, the number of entities required to explain everything." This is very important concept in examining complexity as there is inherent non-uniqueness with many of the results. By only using a minimum number of factors required to explain complexity, it has been possible to resolve the contribution made by each one in certain circumstances. Any more complicated explanation, using extra degrees of freedom in the model, would have to be shown to outperform this null hypothesis before being adopted. This principle underpins the general strategy for examining complexity suggested in section 8.3.
8.2 Usefulness of Data and Analysis Techniques

The results of this thesis confirm the usefulness of relative amplitudes and forward modelling of seismograms (Douglas et al. 1972).

The relative amplitude method was used several times in this thesis, first to determine the focal mechanisms of the Whittier Narrows earthquake. For this earthquake the short-period data was complex, but with just a few broadband stations with polarity observations, and large error bounds on the readings, it was possible to place some constraint on the focal mechanism solution. Specifically, compatible solutions were in the thrust regime, consistent with regional tectonics and the published centroid moment tensor solution. For the Sierra Madre earthquake it was possible to test two interpretations of the two prominent arrivals on the short-period seismogram and positively identify P and the surface reflection, pP, thereby fixing the depth of the event at 11 km.

For the Caspian Sea earthquake the relative amplitude method was used to test the interpretation of the phases P, pP and sP made by the PIDC. No focal mechanisms were compatible with these observations, so this interpretation could be rejected. I was then able to confirm my interpretation of a larger arrival, after the phase identified by the PIDC, as pP. This interpretation produced a very well constrained focal mechanism solution in the form of a normal fault with a significant strike-slip component, consistent with the local tectonics. The surface reflections gave a depth estimate of 48 km, consistent with the ISC and CMT result.

Using the Douglas et al. (1972) method for generating synthetic seismograms I was able to show again that, with a simple kinematic source model, the principle features of short-period and broadband seismograms can be reproduced. By using a range of source rupture areas and rupture velocities it was possible to classify earthquakes as simple or complex (section 4.4), and identify where the transition occurs as the rupture duration is varied. This transition ranged from 5 km source radius for slow rupture velocities to 12 km radius for very fast rupture velocities, roughly corresponding to body-wave magnitudes between 5.5 and 6.0.
CHAPTER 8. Discussion and Conclusions

Even for very complex short-period seismograms, such as those from the Whittier Narrows earthquake, it was possible to obtain a reasonable fit to the data using the forward modelling technique developed by Douglas et al. (1972), which employs the Savage source model. This highlights the continued power of this method 30 years after its development. However the seismograms in chapter 6 could not be explained by the simple kinematic source model used in the numerical code. In future, it should be possible to generate synthetic seismograms at teleseismic distances using a fully dynamic source model.

8.3 Strategy for Examining Complexity

Having investigated the main findings in the previous section, it is now possible to present a general strategy for dealing with complex seismograms. First the quality of the data is important in identifying complexity above background noise. As has been demonstrated many times, array stations are vital in looking at short-period complexity. Beam-formed recordings from these stations suppress the random scattered noise, making picking of arrivals and forward modelling much easier. By using teleseismic data the seismograms are generally free of crustal phase contamination from the source and receiver areas.

Second it is important to have both short-period and broadband records. Often the short-period signal is complex when the broadband signal is simple, for example in the seismograms from the Whittier Narrows earthquake. The broadband record gives the best measure of the source duration, and the short-period the best indication of the arrival time and amplitude of the P, pP and sP phases, and hence event depth. If the source duration is longer than 1 s, then the short period seismogram will be complex. Figure 4.10 in section 4.4 gives an indication of the thresholds where source rupture velocities and fault radius begin to produce complexity on the short-period seismogram. Forward modelling of the broadband seismogram will give a first-order estimate of the rupture size, shape, scalar seismic moment and focal mechanism. However, these parameters are all constrained better to second order when the short-period record is used in conjunction. Source directivity effects can be assessed using a suite of seismograms from a range of azimuths. This will further constrain the shape of the rupture
area but add a determination of the rupture direction. By looking at seismograms from different stations in the same receiver area for the same earthquake, and from different earthquakes recorded at the same station, it is also possible to assess the relative contribution of source and mantle path effects.

Third, if the complexity cannot be explained by source duration effects, then a major potential cause of additional phases may be S-to-P conversions below the source. Recently Geographical Information Systems data bases have become available which quote the local crustal structure around source and receivers (Bowers, Pearce & Douglas 2000). This can indicate independently if S-to-P conversions are likely, and hence remove one of the uncertainties before the seismograms are forward modelled. The key seismograms to investigate for mode conversions are the nodal recordings, where both P and the surface reflections are small relative to the size of the mode conversions.

Fourth, it is possible, even with complex seismograms, to use the relative amplitude method (section 2.7) to test various interpretations for P, pP and sP, and hence obtain some constraint on the focal mechanism solution, when the correct interpretation is made. From this interpretation an accurate measure of the earthquake depth is possible. In terms of discrimination between earthquakes and nuclear explosions, depth is a key parameter, because all explosions are shallow. Explosions are typically set off in boreholes or cavities limited to depths of less than 1 km, i.e. below the limit of resolution of pP and sP for periods of 1 s. In contrast, intermediate-sized earthquakes all nucleate at depths greater than 2 km or so, for a typical geotherm where the rheology of quartz-feldspar materials changes from velocity strengthening to velocity weakening [(Scholz 1990) figure 3.19]. That is, unstable seismogenic slip nucleates below 2 km, and hence produces resolvable pP and sP at 1 s period.

Finally, if none of these hypotheses can explain all the complexity on a seismogram, then it may be necessary to consider dynamic complexity in the rupture itself. Critically, if only one observation of a simple P or S waveform seismogram is seen at any station, then a dynamically complex source model is not required. Otherwise, a dynamically complex rupture may be needed. Again I would stress that the importance of first attempting to find a simple explanation for the available data, before adding additional complexity to the model only where necessary.
8.4 Conclusions

The passband of the recording instrument (whether it is short-period or broad-band), the duration of the source (controlled by the source area, geometry and the rupture velocity), the near-source structure, the direction of propagation, and path attenuation all have a systematic bearing on the complexity of the seismogram in forward modelling. This in turn affects the ability to correctly identify individual phases and obtain information about the source from actual recorded seismograms. I have shown how some of these effects have characteristic features that can be identified, and I have shown qualitatively the contribution each makes to complexity in certain circumstances.

Despite the geographic proximity, the Whittier Narrows and Sierra Madre earthquakes produce radically different seismograms, mainly due to differences in source duration, but also due to differences in local velocity structure near the source. For both earthquakes, it is possible to place significant constraints on the focal mechanism using relative amplitudes, even when surface reflections are difficult to identify. The ISC/CMT solution for depth and focal mechanism is compatible with the range of the source mechanisms determined for both events. The surface reflection on the Sierra Madre short-period seismograms can be identified uniquely as pP, and not sP, as incorrectly identified by Wald (1992).

The Whittier Narrows earthquake has a source area three times the size of the Sierra Madre earthquake. This is the primary cause of the variation in complexity seen on short-period records, as the longer duration source of the Whittier Narrows earthquake is greater than the period of the short-period instrument. This produces many starting and stopping phases and creates a very complex short-period signal. Although the two sources are only 23 km apart, the differences between the Southern California and Los Angeles Basin near-source velocity structure may be significant at teleseismic distances, and contribute to the additional complexity seen from the Whittier Narrows earthquake. The main reason for this difference in structure is the basin bounding Sierra Madre fault that lies in between the two earthquake locations.

The complexity in seismograms for the 29 October 1995 Caspian Sea earthquake can be modelled using S-to-P conversions at an interface at a depth of 108 km
below the surface. All other alternative hypotheses can be rejected by different aspects of the available data. In particular, the hypothesis that an additional phase between P and pP is a surface reflection, cannot produce double-couple S and P radiation patterns that fit the available data. After formally rejecting this hypothesis, the focal mechanism was well constrained using relative amplitudes, and has a nodal plane which corresponds to a steeply-dipping normal fault with a strike-slip component, consistent with the range of focal mechanisms reported for the area.

The absence of an sS phase on the EKA record for the Caspian Sea earthquake implies a downward-propagating line rupture that introduces a strong Doppler effect. This enables the more steeply-dipping nodal plane to be identified as the true fault rupture plane. The downward propagating rupture produces a larger S than usual at P wave nodes, and hence enhances the amplitude of S-to-P conversions relative to P and pP. The S to P mode conversions below the source are also enhanced relative to P when the overall amplitude of the seismogram is small. The best fitting source model for this event has a depth of 48 km. The 30 km depth given by the PIDC apparently resulted from the misidentification of the surface reflection.

The velocity interface responsible for S-to-P mode conversions below the source is in the mantle at depths corresponding roughly to the base of normal continental lithosphere. The interface may plausibly be the top of an old subducting slab under the central Caspian Sea, but this hypothesis cannot be tested in detail using the methods applied here. Further work on deep seismic reflections or seismic tomography would have to be carried out to determine the nature of the interface required to fit the data here. This highlights how a detailed study of all phases is essential in correctly identifying surface reflections and explaining complex phases not predicted by standard travel-times. They also confirm the critical role of near-nodal stations in identifying the causes for seismogram complexity.

Finally I have shown that for those earthquakes which fail conventional forward modelling techniques it is possible to explain some of the observed complexity by using a dynamic source model. This shows that barriers on a fault plane can produce the unusually high frequencies seen with some earthquakes.
8.5 Application to Test-Ban Monitoring

The work in this thesis comes within the scope of general research into monitoring the Comprehensive Test-Ban Treaty. Although much of the current research involves using large travel-time databases of regional P to calibrate the structure of the earth for regional monitoring (Megnin & Romanowicz 2000, Bowers et al. 2000), this work on complexity at teleseismic distances has shown that long-range monitoring still has a role to play. The strategy for examining complexity presented in this chapter, developed from the various case-studies and modelling in this thesis, is the kind of systematic study that should go on when a suspicious event, of suitable magnitude to be recorded teleseismically, is detected using normal discrimination criteria. As the title of Douglas (2000b) states 'Recognising underground tests from long range is simplicity itself'. By using forward modelling of events and the relative amplitude method, and eliminating possibilities, it would be relatively straight-forward to identify earthquakes. The analysis also provides information into the individual factors that contribute to the complexity of the seismograms. If routine discrimination methods fail to correctly identify crucial parameters, such as the depth (which according to seismologists from the PIDC is a common occurrence), then this thesis provides examples of the way this can be tackled using relative amplitudes and forward modelling.

The results from the previous chapter showed that it is possible to obtain realistic source-time functions from numerical models of the earthquake source. Part of the research into test-ban monitoring is the better understanding of the earthquake source and this work, although limited at present, is a first step to using realistic earthquake sources in forward modelling studies. This adds significantly to the forward modelling of events, especially those which are known to have a complex source and where there are barriers on the fault plane. I have shown these to significantly increase the complexity of the far-field pulse and hence the teleseismic seismogram.
8.6 Suggestions for Future Work

It was hoped that from this work it would be possible to generate synthetic seismograms for an arbitrary source rupture model at teleseismic distances. This turned out not to be an achievable goal within the available time period, given the breadth of the work undertaken. The current method for generating synthetic seismograms is tied at a fundamental level within the numerical code to the kinematic Savage model, and needs a fundamental re-write of sections of the code to include a more complicated model where the data justify the additional complexity. The work presented here forms a very small part of what would be possible using the dynamic rupture models. A more complete treatment using a finer source grid and a full 3D model would require more powerful computers than were available to me. This could possibly be done in conjunction with the Edinburgh parallel computing centre using the supercomputers they have available.

Finally, it has been possible to examine only a few case studies here. The next stage should involve a more systematic study of complexity perhaps for a fixed window in time and source region. By studying numerous earthquakes from the same source area, a greater understanding of the factors that contribute to complexity could be gained. For instance, I have shown that a comparative study of the Sierra Madre and Whittier Narrows earthquakes provides more definitive conclusions than studying one earthquake alone. Both of these earthquakes produce large aftershock sequences from the same source area. Including these in a future study may provide additional information about the contribution of the near-source structure and source duration to seismogram complexity. One way of doing this may be to use the seismogram from the small earthquake as the Earth impulse response function for the main event.


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Appendix A

S-wave observations

A.1 Introduction to Path Complexity

Having considered how near source effects contribute to the complexity of the P-wave seismogram I now look at one factor that can affect the ability to see this complexity. Due to the low attenuation on the ray path to EKA it was possible to see high frequency S-to-P conversions even though P was large. Here I examine how the mantle path may effect the seismogram complexity using evidence from S-wave arrivals which are more sensitive to attenuation.

The mantle path is usually assumed to be structurally homogeneous with only attenuation a significant factor. This is true for recordings at teleseismic distances as the take off angle for P is steep. This is also true for surface reflections which after the reflection at the free surface enter the mantle at a steep angle of incidence. Also it is generally assumed that the P and surface reflections travel the same mantle path to teleseismic distances. Attenuation and scattering should therefore effect both equally and can be assumed constant when dealing with relative amplitudes.

But when looking at individual seismograms and considering the origins of complex phases it is important to establish how attenuation, or lack of it, may enhance certain features or prevent others from contributing to the teleseismic seismogram. In this section I examine how upper mantle structure varies across
Europe using S-wave attenuation and how these structures may contribute to the complexity of the teleseismic P and S-wave seismograms.

A.2 Attenuation

Attenuation causes the loss in amplitude of the waves due to anelastic processes within the bonds in the rock and due to frictional losses. This type of attenuation is dependent on rock properties even in homogeneous rock units and may be distinguished from scattering energy at small scale heterogeneities. In this case however the overall energy content of the wavefield remains the same. We will be considering only the actual energy loss due to the intrinsic attenuation. This is measured by the dimensionless quantity $Q$, the fractional energy loss per cycle, and is inversely related to the attenuation.

When dealing with real earth situations where $Q \gg 1$ it is more convenient to model attenuation as an integral value along the whole wavepath including the section of crust beneath the source and receiver and the mantle path. This is the value $t^*$ (see section 3.5), which generally takes a value between 0 and 1 in real earth models. The attenuation is dependent on frequency with high frequencies being preferentially reduced in amplitude due to there being more cycles, in a given distance.

A.3 Teleseismic S-waves

Before examining attenuation, shown by S wave variations, it is important to note how the amplitude of S at teleseismic distances is effected by the position of the receiver in the radiation pattern and nature of the seismic rupture. Figure A.1 shows the S-wave radiation pattern for a double couple point source with the fault and auxiliary planes marked. In the direction of the fault or auxiliary planes the S wave is at its maximum amplitude. This is also an approximate minimum in the P-wave radiation pattern.
The equation for the radiation of S-waves from a point source is

\[ u^S(x, t) = \frac{1}{4\pi \rho \beta^3} \left( \cos 2\theta \cos \phi \hat{\theta} - \cos \theta \sin \phi \hat{\phi} \right) \frac{1}{r} \frac{\dot{M}_0}{\beta} \left( t - \frac{r}{\beta} \right) \]

where \( u^S(x, t) \) is the amplitude of the far-field S pulse at position vector \( x \), \( \hat{\theta} \) and \( \hat{\phi} \) are unit Cartesian vectors in the \( \theta \) and \( \phi \) directions, \( \beta \) and \( r \) is the distance from source to receiver at time \( t \) (Shearer 1999). The amplitude of this S-wave can be plotted from this for a point source. If a station lies along the strike of the fault or the auxiliary plane then you would expect a large S. The amplitude of the radiated energy for S is also 5 times that of the P-wave. However S-wave energy is generally not seen at teleseismic distances. Shear attenuation is observed to be much stronger in the Earth than bulk attenuation and so shear waves are preferentially attenuated

As was discussed previously, the complexity of the 29 October 1995 Caspian Sea earthquake is controlled by S-to-P conversions in the uppermost mantle. As I have shown reason that these S-to-P conversions are large at some stations is that the rupture propagates downwards and the location of the stations are in the
strike of the fault. This direction is a P-wave node, and as can be seen in figure A.2, it is also the direction of the S-wave anti-node and increased S amplitude due to the propagating rupture. This large S is radiated at an azimuth towards northern Europe, which is in the strike of the fault, and is smaller towards the southern Europe. I will show is borne out by the observations.

\begin{figure}
\centering
\includegraphics[width=0.8\textwidth]{s-wave-radiation-pattern}
\caption{Amplitude of the far-field radiation pattern for a unilateral propagating rupture. Azimuth scale in degrees around edge and for take-off angles from 0 to 90° from centre to edge.}
\end{figure}

A.4 Identification of S

At teleseismic distances S-waves are hard to identify as they are often small compared to the background noise. Figure A.3 shows S stations from a range
of azimuths with increasing epicentral distance aligned on the P-wave arrival time.

S is clearly visible on many of the stations and the amplitudes vary in size even over the epicentral distance range of 30-40° which covers those stations within Europe. The S arrival times correspond to the predicted times using the IASPEI91 earth model so I can be confident that this phase I have identified is S. The large amplitude at ESK can be clearly seen on this figure even above the noise.

Using the method of (Douglas et al. 1972), figure 5.7a and b show the observed and modelled seismograms for $S_v$. I have used the focal mechanism obtained in this chapter, with a propagating downward line rupture. Figure 5.7c shows that for a circular rupture $SS'$ would be expected 15s after S and with similar amplitude and this is not seen. This adds weight to the interpretation of the Caspian Sea earthquake as a downward-propagating line rupture which enhanced the amplitude of downgoing S and caused the S-to-P conversions see in the teleseismic P seismogram at many stations.

### A.5 European attenuation variation

In this section I will now examine how the amplitude of the S-wave recorded across Europe varies and what could be the reasons behind. Figure A.4 shows the stations which have S-wave measurements and these are listed in table A.1.

They are grouped according to the amplitude and fall into 3 distinct groups. Southern European readings lie between 800 and 1000 nm/s, Central European between 200 and 800 nm/s and Northern European are greater than 1200 nm/s. Variations is S-wave amplitude likely to be due to either the structure in the upper mantle or the position in the S-wave amplitude pattern.

There is the possibility that the pattern seen may be due to more local effects in the near-source and near-receiver structure. Many of the stations studied are only within the teleseismic source window and so the take off angles of the S-waves are shallow. This means that it spends more time within the higher attenuating crust, near the receiver. The structural complexity of the crust in
Figure A.3.
Broadband seismograms showing picked S wave arrival with increasing epicentral distance from left to right. IPUO - P arrival. ISUO-S arrival
Figure A.4.
European map showing stations with earthquake focal mechanism at earthquake epicentre
Table A.1. Locations of stations with S-wave observations showing there azimuth and epicentral distance from the 25 October Caspian Sea earthquake

<table>
<thead>
<tr>
<th>Station</th>
<th>Location</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Distance (°)</th>
<th>Azimuth (°)</th>
<th>Amplitude (nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>KIEV</td>
<td>Ukraine</td>
<td>50.69°N</td>
<td>29.21°E</td>
<td>19.39</td>
<td>312.4</td>
<td>10000</td>
</tr>
<tr>
<td>KONO</td>
<td>Norway</td>
<td>59.65°N</td>
<td>9.60°E</td>
<td>33.16</td>
<td>321.3</td>
<td>1500</td>
</tr>
<tr>
<td>ESK</td>
<td>Scotland</td>
<td>55.33°N</td>
<td>3.15°W</td>
<td>39.32</td>
<td>312.3</td>
<td>2500</td>
</tr>
<tr>
<td>DSB</td>
<td>Ireland</td>
<td>53.25°N</td>
<td>6.38°W</td>
<td>41.27</td>
<td>309.2</td>
<td>40</td>
</tr>
<tr>
<td>KBS</td>
<td>Spitsbergen</td>
<td>78.92°N</td>
<td>11.92°E</td>
<td>42.49</td>
<td>349.4</td>
<td>1200</td>
</tr>
<tr>
<td>HFS</td>
<td>Norway</td>
<td>60.13°N</td>
<td>13.70°E</td>
<td>31.41</td>
<td>323.6</td>
<td>1100</td>
</tr>
<tr>
<td>ECH</td>
<td>Switzerland</td>
<td>48.22°N</td>
<td>7.16°E</td>
<td>32.96</td>
<td>300.1</td>
<td>200</td>
</tr>
<tr>
<td>BRG</td>
<td>Switzerland</td>
<td>50.87°N</td>
<td>13.94°E</td>
<td>28.70</td>
<td>305.8</td>
<td>400</td>
</tr>
<tr>
<td>MOX</td>
<td>Germany</td>
<td>50.65°N</td>
<td>11.61°E</td>
<td>30.14</td>
<td>304.9</td>
<td>800</td>
</tr>
<tr>
<td>WET</td>
<td>Germany</td>
<td>49.15°N</td>
<td>12.88°E</td>
<td>29.19</td>
<td>302.0</td>
<td>400</td>
</tr>
<tr>
<td>DFC</td>
<td>Czech Republic</td>
<td>50.36°N</td>
<td>16.41°E</td>
<td>27.07</td>
<td>305.2</td>
<td>600</td>
</tr>
<tr>
<td>MORC</td>
<td>Poland</td>
<td>49.78°N</td>
<td>17.54°E</td>
<td>26.25</td>
<td>304.2</td>
<td>800</td>
</tr>
<tr>
<td>HGN</td>
<td>Netherlands</td>
<td>50.76°N</td>
<td>5.93°E</td>
<td>33.75</td>
<td>304.7</td>
<td>&lt;600</td>
</tr>
<tr>
<td>GRFO</td>
<td>Germany</td>
<td>49.69°N</td>
<td>11.22°E</td>
<td>30.32</td>
<td>303.0</td>
<td>&lt;600</td>
</tr>
<tr>
<td>AQU</td>
<td>Italy</td>
<td>42.35°N</td>
<td>13.51°E</td>
<td>29.03</td>
<td>288.0</td>
<td>1000</td>
</tr>
<tr>
<td>CLL</td>
<td>Italy</td>
<td>51.31°N</td>
<td>13.00°E</td>
<td>29.35</td>
<td>306.5</td>
<td>900</td>
</tr>
<tr>
<td>PAB</td>
<td>Spain</td>
<td>39.54°N</td>
<td>4.35°W</td>
<td>42.73</td>
<td>288.7</td>
<td>800</td>
</tr>
</tbody>
</table>

central Europe may also add to this. Crustal structure is not a factor near the receiver as the source is in the upper mantle but there may be some effect from azimuthal variation in near-source structure. Also anomalous readings such as that at DSB must have a strong local effect. There is also the possibility that some of the S phases may have been misidentified but I am confident of the picks given the large relative amplitude of the signal compared to the surrounding coda.

Figure A.5 shows that ESK approximately lies on the azimuth of the fault plane, with stations to the north having a high amplitude and to the south a low amplitude. This indicates that the anomaly at ESK could be due to its position at an S-wave antinode but that other may be dominated by the structure of the upper mantle through which the wave passes. The reading at ESK is also polarised with most of the energy in $S_v$. This indicates that the line rupture was propagating towards ESK. One would expect the amplitude of $S$ to fall off to the north and south if it was controlled by the S-wave radiation pattern, as shown in figure A.2, but this is not the case. Also if the amplitude differences are due to the difference in epicentral distance then those in central Europe should have a higher amplitude and this is not seen either.
Figure A.5.
Map centred on earthquake showing focal mechanism with tectonic regions TTZ = Tornquist-Teisseyre Zone, NC = Norwegian Caledonides, SC = Scottish Caledonides, EH = Eiffel Hotspot, NS = North Sea Basin, BS = Baltic Shield, PB = Pannonian Basin (Marquering & Sneider 1996).
The variations in amplitude approximately correspond to the change in lithosphere between northern European shield and central Alpine region and southern European convergent margin. The shield area of Russian platform and Baltic shield represents low attenuation and high $Q$. This is seen in higher S-wave amplitudes and is separated from the lower amplitude region by the Tornquist-Teisseyre Zone. These lower amplitudes are seen from stations in central Europe where attenuation is higher in more heavily deformed rock of the continental collision zone. This is born out by tomographic studies at 600 km. Figure A.6 shows the percentage velocity change.

Figure A.6.
Tomography of the earth at mantle at 600 km. Red high velocity; Blue low velocity.
A.6 Relevance to P-wave complexity

In this section I have aimed to show how the variation in S-wave amplitude is affected by the structure through which the wave travels and if attenuation is low then information regarding the nature of the rupture process and what causes P-wave complexity can be gained. In this example a large S-wave seen at teleseismic distances due to a low attenuating path indicated that strong S-wave radiation was present at this azimuth. Forward modelling of the seismogram showed that the rupture process was a unilateral rupture which radiated S energy downwards. This produced S-to-P conversions mentioned in the previous chapter and contributed to the complexity of the P seismogram. The station EKA also lay at a P-wave node (approximate S-wave antinode) so the S-to-P conversions were more easily identifiable.