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RICE UNIVERSITY

Combining Deterministic and Stochastic Velocity Fields in the Analysis of Deep Crustal Seismic Data

by

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Abstract

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Standard crustal seismic modeling obtains deterministic velocity models which ignore the effects of wavelength-scale heterogeneity, known to exist within the Earth’s crust. Stochastic velocity models are a means to include wavelength-scale heterogeneity in the modeling. These models are defined by statistical parameters obtained from geologic maps of exposed crystalline rock, and are thus tied to actual geologic structures. Combining both deterministic and stochastic velocity models into a single model allows a realistic full wavefield (2-D) to be computed. By comparing these simulations to recorded seismic data, the effects of wavelength-scale heterogeneity can be investigated.

Combined deterministic and stochastic velocity models are created for two datasets, the 1992 RISC seismic experiment in southeastern California and the 1986 PASSCAL seismic experiment in northern Nevada. The RISC experiment was located in the transition zone between the Salton Trough and the southern Basin and Range province. A high-velocity body previously identified beneath the Salton Trough is constrained to pinch out beneath the Chocolate Mountains to the northeast. The lateral extent of this body is evidence for the ephemeral nature of rifting loci as a continent is initially rifted. Stochastic modeling of wavelength-scale structures above this body indicate that little more than 5% mafic intrusion into a more felsic continental crust is responsible for the observed reflectivity. Modeling of the wide-angle RISC data indicates that coda waves following \( PmP \) are initially dominated by diffusion of energy out of the near-surface basin as the
wavefield reverberates within this low-velocity layer. At later times, this coda consists of scattered body waves and $P$ to $S$ conversions. Surface waves do not play a significant role in this coda.

Modeling of the PASSCAL dataset indicates that a high-gradient crust-mantle transition zone or a rough Moho interface is necessary to reduce precritical $PmP$ energy. Possibly related, inconsistencies in published velocity models are rectified by hypothesizing the existence of large, elongate, high-velocity bodies at the base of the crust oriented to and of similar scale as the basins and ranges at the surface. This structure would result in an anisotropic lower crust.
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To my Parents,
Bill and Sharon Larkin
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Chapter 1

Introduction

This thesis describes the first attempt to use combined deterministic and stochastic velocity models in the interpretation of near-offset and wide-angle deep crustal seismic data. The deterministic velocity structure of the crust has been the traditional target of refraction/wide-angle reflection seismology. The wavelength-scale velocity structure, however, has been a more elusive target, and the details of small-scale velocity heterogeneity within the Earth's crust, as well as its effects on the transmitted and reflected wavefield, are still being investigated. By combining both large and small-scale velocity structures in the analysis of observed deep crustal seismic data, a more complete and detailed interpretation of the data is possible.

The thesis is organized as follows. The remainder of the introductory chapter provides the historical context of the research with regard to the study of the continental crust, and explains what is meant by deterministic and stochastic velocity models. Following the introduction, two seismic datasets are analyzed, the 1992 RISC (RIce-uSC) experiment in southeastern California and the 1986 PASSCAL experiment in northern Nevada. Chapter 2 introduces the RISC experiment, and develops a deterministic crustal model using traditional methods of seismic data processing and analysis. Chapter 3 describes the application of stochastic velocity fields to the analysis of the near-offset RISC experiment, and Chapter 4 extends the analysis to the wide-angle experiment. Chapter 5 introduces the PASSCAL experiment and the modeling of both the near-offset and wide-angle data. Finally, Chapter 6 summarizes the thesis research and discusses the implications of the results. An appendix is added at the end of the thesis which describes a wave-equation datuming program written to remove the effects of near-surface basin structures on the recorded data.
1.1 Goals of controlled-source crustal seismic techniques

The beginnings of crustal seismology can be traced to the early 20th century work of Mohorovičić, who first identified a global seismic interface separating the Earth's crust from the mantle (Mohorovičić, 1909) [87]. However, the lack of regular spatial coverage of earthquake sources, as well as the sometimes complicated source signature, limits the applicability of the earthquake source to the study of crustal structure in most regions of the globe. The development of seismic techniques in the search for oil and gas introduced controlled-source seismology to the science of imaging the interior of the Earth. Controlled-source seismology, be it dynamite or Vibroseis, has the advantage of a source which can be placed in predetermined locations to provide consistent subsurface coverage of reflected and refracted energy. Although limitations do exist in the design of controlled-source experiments, mainly environmental and financial, these methods have provided the most detailed information on the structure of the Earth's crust to date.

Over the past 30-35 years, a proliferation of controlled-source deep crustal seismic experiments have been conducted throughout the world (Mooney and Brocher, 1987) [88]. These experiments have increased in size and scope, providing improved images of varied tectonic regions. Deep crustal seismic experiments are traditionally divided into two types: reflection and refraction seismology. The increase in source and receiver density has caused the distinction between these two methods to become somewhat blurred, but they still provide a useful description of the endmember modeling and processing which is commonly undertaken.

I will begin by briefly discussing the design and targets of these two types of experiments, as well as hybrid experiments which consist of a combination of the two. I will then discuss a non-traditional method of modeling the crust using stochastic descriptions of velocity heterogeneity, and describe how such a technique might improve our understanding of deep crustal refraction and reflection data.
1.2 Traditional controlled-source seismic techniques for crustal seismic studies - What can they determine about the nature of the crust?

1.2.1 The refraction experiment

The refraction experiment, or wide-angle reflection experiment, is designed to obtain information on the bulk seismic velocity of the crust and upper mantle. This is accomplished by developing velocity models which predict the arrival times of identifiable crustal and upper mantle phases. This is usually a non-trivial exercise due to the fact that the relationship between traveltime and seismic velocity is nonlinear. This relationship can be expressed as

\[ t = \int_{0}^{s} \frac{ds}{v}, \] (1.1)

where \( t \) is the traveltime of a ray, \( s \) is the travel path of this ray, and \( v \) is the velocity. In this equation, traveltime is dependent on both the travel path and the velocity. However, changing the velocity changes the travel path of the ray, and results in a problem that cannot be analyzed using linear methods. To address this problem, modern interpretations of traveltime data usually involve iterative forward modeling, linearized inversion methods, or global nonlinear inversion methods.

The priority in the design of refraction experiments is sufficient source-receiver offset, at the expense of dense receiver spacing. Sufficient offset depends on target depths and target velocities, but a crustal refraction experiment should record a first-break mantle refraction (\( Pn \)) and a postcritical crust-mantle reflection (\( PmP \)), which usually means offsets > 150 km. Modern receiver spacing for land surveys can be as small as 300 m, and shot spacing is generally > 5 km. The bandwidth for a refraction survey is generally 2-12 Hz, resulting in wavelengths within the crystalline crust which range from ~0.5 to 4 km.

The resolution of velocities varies with depth in the crustal refraction experiment. The raypaths and traveltime curves of the main phases recorded in a typical refraction survey are displayed in Figure 1.1. The sedimentary basin structure is generally not well resolvable due to the coarse shot and receiver spacing. The upper 10-15 km is usually well resolved from first-break diving waves (usually termed refractions) within the upper crust (\( Pg \)). The velocity of the upper mantle, being significantly higher in velocity than the crust, is also observed as a first break
\((Pn)\) at far offsets, and is therefore also well constrained. The lower half of the continental crust is a more difficult target for velocity determination because of the lack of first-break refractions from these depths. These velocities are determined by analyzing wide-angle traveltimes and amplitudes of deep-crustal reflections, refractions and \(PmP\), which are more difficult to identify and pick precisely compared to first arrivals, and can therefore add to the uncertainty of lower crustal velocity estimates.

The seismic velocity structure determined from refraction experiments forms the foundation for deterministic velocity models. The individual structures are much larger than the source wavelength, and only those phases interpreted to be due to specific, determinable horizons or bodies are modeled.

1.2.2 The reflection experiment

Traditional deep crustal seismic reflection experiments consist of dense linear arrays of receivers which record energy emanating from a number of shotpoints along the line. Modern shot and receiver spacing ranges from 50 to 100 m. Maximum source-receiver offsets rarely exceed the maximum target depth, and are usually much smaller. The near-offset recording geometry makes the reflection experiment ideal for imaging the impedance structure directly beneath the recording spread. In addition, the shorter propagation distances, relative to the refraction experiment, enable higher frequencies to be recorded, and thus increases the resolution of the recorded data. The bandwidth for a reflection survey is generally 8-32 Hz, resulting in wavelengths within the crystalline crust which range from \(\sim 0.1-1.0\) km.

Basin structure, faults and shear zones are common targets of the reflection experiment. In addition, it is now common to comment on the "reflectivity" or "reflective character" of the crust and crust-mantle boundary. By this, the interpreter is describing zones of reflected energy within the crust which make some layers or zones distinct from others based upon seismic character. This reflectivity can be characterized by its arrival time, amplitudes, reflection lengths, spatial density of reflections and/or by structural dips. The depth and location of the different reflective zones form another part of the deterministic velocity model. These zones may or may not correspond to changes in the velocity model from refraction data, but they do delineate gross changes within the crustal column.
Figure 1.1 Ray diagram and traveltime curves for the main phases recorded in a typical deep-crustal refraction experiment. Velocities at depths sampled by first-break refractions from the upper crust $P_g$ and upper mantle $P_n$ are generally well-resolved.
Determining the details of the velocity structures which are responsible for the different types of reflectivity can enhance the knowledge of the geologic and tectonic history of a region. It is this wavelength-scale detail which the stochastic velocity model tries to reproduce.

1.2.3 Hybrid experiments

It is now common to combine both the refraction and reflection experiments when collecting deep crustal seismic data in a particular region. The reflection survey is sometimes termed a "piggyback" survey upon the larger refraction survey. These experiments generally consist of a seismic line with a denser shot and receiver spacing in the center, though the shot spacing rarely approaches the receiver spacing. Both the RISC 1992 data and the PASSCAL 1986 northern Nevada data are from hybrid experiments. The benefits of these experiments are obvious: both a reflection image and a bulk velocity model can be obtained at the same location.

In summary, the refraction and reflection data provide complimentary information on the velocity structure of the continental crust. The former is controlled by the large-scale bulk velocity structure, and the latter is dominated by the wavelength-scale impedance structure. This combination of velocity fields can be written as

\[ v(x, z) = v_0(x, z) + \delta v(x, z). \]  

(1.2)

The ultimate goal of crustal seismology is to obtain \( v(x, z) \). Refraction seismic methods attempt to define \( v_0(x, z) \). The reflection experiment is dominated by \( \delta v(x, z) \), and therefore provides information on the smaller-scale velocity structures within the crust. As will be argued below, defining this variable deterministically is not realizable. The following section explains why stochastic velocity models can be used to investigate \( \delta v(x, z) \).

1.3 Non-traditional technique - Wavelength-scale velocity fields based upon statistics of geologic outcrop maps

Geologic outcrops of the crystalline basement often reveal a complicated juxtaposition of rock types and structures at many scales. Examples of such exposures of upper, middle and lower crustal rocks are found in the Strona-Ceneri Zone of
Italy (Holliger and Levander, 1994) [62], the Franciscan terrane of California, the Hafait gneiss complex of Egypt (Goff and Levander, 1996) [43], the Lewisian gneiss complex of Scotland (Levander et al., 1994) [76], and the Ivrea Zone of northern Italy (Holliger and others, 1993) [63]. Due to the complexity of these exposed rock outcrops, it might be suggested that a deterministic model, that which attempts to identify specific bodies within the crust, may be difficult to achieve on the wavelength scale (100s of meters). However, if the distribution of rock velocities obeys a unique set of simple statistical parameters, then it might be possible to obtain these statistics from the recorded data and therefore obtain quantitative measures of the rock distributions at depth. This last statement defines the inverse problem which is the driving force behind investigations of stochastic models of velocity heterogeneity. While this thesis does not attempt to solve the inverse problem, it does attempt to understand the forward propagation of waves through a stochastic medium, and thus define the effects of specific statistical parameters on the wavefield. Thus, in one sense I am seeking to define tractable inverse problems by investigating and understanding the forward problem. In particular, I limit myself to experimental geometries which are similar to those used in the deep crustal seismic experiment, and to targets of interest defined by observations of recorded data.

1.3.1 Stochastic velocity fields

Recently, it has been shown that stochastic models for small-scale velocity heterogeneity (\(\delta v\)) based upon statistics derived from maps of exposed mid and lower crust can qualitatively reproduce the crustal wavefield at small and large offsets (Levander and Holliger, 1992b; Holliger et al., 1993; Holliger and Levander, 1994a,b; Levander et al., 1994a,b) [80, 63, 60, 61, 76, 79]. Statistical analysis of geologic maps indicate that the spatial distribution of rock velocities expected in the crustal column can be described by a 2 or 3 dimensional von Kármán auto-covariance function describing crustal fabric and a probability density function (pdf) describing seismic velocity population (Holliger and Levander, 1992b; Levander et al., 1994a) [59, 76]. The auto-covariance function is defined by the horizontal and vertical characteristic scales (\(a_x\) and \(a_z\), respectively) and the Hurst number (\(\nu\)), which defines the roughness of the media. The Hurst number is related to the fractal dimension by the equation \(D = N + 1 - \nu\), where \(N\) is the Euclidean
dimension of the velocity field. \( a_x, a_z \) and \( \nu \) are obtained from statistical analysis of the spatial fabric of different rock types in fine-scale geologic maps. The pdf is obtained by determining the appropriate petrophysical properties from laboratory measurements of the exposed rock types.

Stochastic velocity models are a convenient means to construct a complex seismic velocity model which is a statistically valid representation of both geologic outcrop and the velocity field traversed by seismic waves. The models are developed directly from observable small-scale velocity distributions which actually exist in the Earth.

1.3.2 Combining the deterministic and stochastic velocity fields

To investigate the effects of velocity heterogeneity on the seismograms, a stochastic distribution of velocities is superimposed upon the deterministic velocity models obtained from analysis of wide-angle travel times. Stochastic models are chosen which from experience will reproduce qualitatively the recorded wavefield, but are also tied to the in situ seismic velocities and geologic maps from exposed basement rocks. To create the full-wavefield simulation from the velocity models, I utilize a viscoelastic finite-difference algorithm which simulates the propagation and attenuation of compressional and shear waves through an arbitrarily heterogeneous velocity field (Robertsson et al., 1994) [98].

If the distribution of rock properties can be defined as a self-affine field, then analysis of the full wavefield using heterogeneity which mimics this distribution is the most appropriate way to analyze the seismic wavefield. Therefore, investigations into the nature of this heterogeneity and its effect on the recorded wavefield can be a useful form of analysis of certain aspects of deep crustal data.
Chapter 2

The RISC Seismic Experiment: A Deterministic Model for the Salton Trough/Basin and Range Transition Zone

2.1 Introduction

The Salton Trough is situated at the southern termination of the San Andreas transform system into the transtensional Gulf of California (Figure 2.1). It is believed to be one of the few places on Earth where oceanic rifting is actively impinging upon continental crust (Lonsdale, 1989) [81], with consequent hydrothermal activity, normal and right-lateral faulting, and active seismicity and volcanism. Despite the unique tectonic setting, the crustal structure along the flanks of the Salton Trough has not previously been explored.

In 1992, Rice University and the University of Southern California (RISC) collected a seismic reflection profile across the transition from the Salton Trough to the southern Basin and Range Province in southeastern California. This was a piggyback experiment on the final leg of the much larger Pacific to Arizona Crustal Experiment (PACE) refraction profile. The two experiments were designed to determine the crustal structure of this region at both large and small scales, from the gross velocity structure provided by the refraction data (10-30 km structures) down to the wavelength-scale impedance structure of the reflection data (100s of meters). To accomplish this, both deterministic and statistical velocity information are combined to create velocity models for both the large and small structures present in the crust. In this chapter, I present the deterministic model of the crust determined from the RISC seismic experiment data. In Chapter 4 I will describe the stochastic model and its implications.
Figure 2.1 Regional plate tectonic map centered around the Salton Trough showing the interaction between the Pacific and North American plates (Lachenbruch et al., 1985).
2.2 Tectonic framework of southeastern California from geologic and geophysical observations

Southeastern California has experienced a complicated tectonic history, and a knowledge of this history and its impact on the seismic velocity structure is necessary before a proper interpretation of the seismic data can be made. I begin with an overview of the complex geologic and tectonic history of southeastern California, paying special attention to the rocks exposed within the Chocolate Mountains and the events which may contribute to the seismic response of the present-day crust.

The oldest rocks exposed along the RISC profile (Figure 2.2) are the Early Proterozoic gneisses and granites of the Chocolate Mountains (Crowell, 1981) [19]. The Proterozoic gneisses range from hornblende meta-gabbro to diorite and minor meta-ultramafic rocks to banded granitic orthogneisses. These rocks are part of the Proterozoic orogenic belts which accreted to Archean North America between 1.8 and 1.6 Ga (Crowell, 1981) [19].

In western North America, the Mesozoic was dominated by orogens and plutonism related to oblique subduction of the Kula and/or Farallon plates. The Mesozoic plutonic rocks exposed today in the Chocolate Mountains are granitic orthogneisses (Dillon, 1976) [23]. Together with the Proterozoic gneisses which they intrude, these rocks presumably form a large part of the basement terrane for southeastern California and southern Arizona (Barth and Ehlig, 1988; May and Walker, 1989; Tosdal et al., 1989) [4, 84, 115]; however, they do not form the entire basement beneath the Chocolate Mountains.

Forming the core of the Chocolate Mountains is the Orocopia Schist, a quartzofeldspathic schist thought to be related to the Rand and Pelona schists exposed to the northwest in the Transverse Ranges (Dillon, 1976; Haxel, 1977; Haxel and Dillon, 1978; Haxel et al., 1987) [23, 51, 53, 52]. The protolith of the schist is believed to be marine sediments and volcanics deposited in an offshore basin during the mid to late Jurassic (Haxel and Tosdal, 1986; Tosdal et al., 1989) [54, 115]. The sediments were then metamorphosed to amphibolite facies at depths of 25-35 km (Jacobson et al., 1988) [68] during a thrusting/subduction event in the Late Cretaceous or earliest Tertiary (May and Walker, 1989; Jacobson, 1990) [84, 67]. It is believed that the Orocopia Schist was subducted beneath an offshore continental block along the northeast verging Chocolate Mountains thrust, placing the Proterozoic and Mesozoic continental crust described above onto the Orocopia
Figure 2.2  Map of southeastern California showing RISC experimental geometry. The recording was accomplished during two separate deployments resulting in a total line length of 45 km extending from the edge of the Salton Trough across the Chocolate Mountains into the southern Basin and Range. Off-end shots were recorded from both the Salton Trough and the southern Basin and Range providing offsets as great as 180 km. Currently active transform faults (solid bold lines) are confined to the center of the Trough, but fossil transforms (dashed bold lines) exist to the edges of the basin both to the northeast and southwest, suggesting that spreading loci may have moved over time.
Schist (Dillon et al., 1990) [24]. The lateral extent and depth of the Orocopia Schist are currently unknown. Industry reflection data suggest that a set of upper crustal reflections may mark the base of the schist at 5 to 10 km depth (Morris et al., 1986) [90]. However, because the basal contact of the schist does not outcrop, this conclusion is hypothetical, and the rock types which lie beneath the schist are still unknown. It is worth noting that Dillon et al. [1990] [24] comment on the local and regional heterogeneity of the Proterozoic and Mesozoic igneous and metamorphic terranes compared to the relative homogeneity of the Orocopia Schist.

The present surface exposure of the Orocopia Schist is the result of tectonic denudation during Tertiary Basin and Range extension. This extension is responsible for the present-day basin and range topography observed northeast of the Salton Trough. In the Chocolate Mountains, mafic lavas associated with this extension are exposed in the ranges bordering the Salton Trough and have age dates within Oligocene-Miocene time (Hawkins, 1970; Crowe, 1973) [49, 18]. The youngest of these lavas are 13 Ma basalts. Attenuation of the crust may have been quite extensive during this time, with a proto-Salton Trough well-developed by the end of the period (Herzig and Jacobs, 1994) [56].

The most recent tectonic event affecting this region is the formation of the Salton Trough during the last 5 Ma. The Salton Trough is the landward extent of the transtensional Gulf of California, and forms the transition from a divergent plate boundary in the Gulf to a transform plate boundary along the San Andreas fault system. (Lonsdale, 1989) [81] (Figure 2.1). A number of geophysical and geological studies within the Salton Trough indicate that it is at an advanced stage of rifting. Abnormally high heat flow values suggest asthenospheric upwelling and active magmatism beneath the Trough (Lachenbruch et al., 1985) [73]. Quaternary volcanic activity in the Salton Buttes rhyolite domes contain mafic xenoliths of tholeiitic composition similar to those along the East Pacific Rise, supporting the argument that the asthenosphere is at shallow levels (Elders et al., 1972; Robinson et al., 1976; Herzig and Jacobs, 1994) [30, 100, 56]. Teleseismic studies using the extensive southern California seismic array have mapped the upper mantle seismic structure beneath the Trough (Humphreys and Clayton, 1990; Humphreys and Hager, 1990) [65, 66]. These workers detect a 3-4% decrease in upper mantle P-wave velocity beneath the Salton Trough and the surrounding region to the east, including the Chocolate Mountains, and interpret this anomaly as evidence for
1-4% partial melt in the upper mantle consistent with upwelling of the underlying asthenosphere.

Seismic refraction data collected in the late 1970s (Fuig et al., 1984) [33] first identified a high-velocity lower crust (estimated velocity 7.4 km/s) beneath the Salton Trough, although its existence had been hypothesized by previous workers from the analysis of gravity and surficial geology (Biehler, 1964) [11]. Fuig et al. [1984] [33] also suggested that as much as 15 km of sediments now occupy parts of the Salton Trough and that the original continental crust has been completely replaced by sediments and intrusion. This is supported by recent geochemical analysis of the Salton Buttes xenoliths by Herzig and Jacobs [1994] [56], who conclude that these basalts have had no interaction with continental material.

Thus, the area around the RISC survey has experienced dramatically different and extensive tectonic events since its original formation in the Proterozoic. Basin and Range and Salton Trough rifting may have overprinted much of the pre-Cenezoic history, but most of the basement exposure still consists of Mesozoic and Proterozoic gneisses.

2.3 Data acquisition and analysis

The RISC profile was designed as a high-resolution piggyback on the final phase of the United States Geological Survey’s Pacific to Arizona Crustal Experiment (PACE). The experimental geometry is shown in Figure 2.2. Twenty-seven explosive shots were fired at 22 shotpoints and were recorded by 435 three-component recorders deployed twice, end-to-end, at 50 m group spacing. The resulting 45 km long profile has source-receiver offsets ranging from 0 km to greater than 180 km. Sixteen shots from 11 shotpoints were within the recording spread, and 11 shots were located off-end of the recording spread for wide-angle coverage. Average shot spacing within the receiver array was 5 km.

The goal of the RISC experiment was to identify and model not only the large-scale deterministic structure of the continental crust, but also the wavelength-scale impedance structure. For this reason, in addition to traditional processing of the reflection data, a significant amount of research involved the modeling of the reflection response from stochastic velocity fields using finite difference techniques. I begin with the traditional processing to determine the large-scale crustal model.
and then, using this model, I develop an integrated deterministic and stochastic model in Chapter 3 which predicts the reflectivity in the observed data.

2.3.1 Reflection data

To produce a low-fold vertical incidence CMP reflection image of the crust, shots and receivers were selected which produced and recorded the highest quality signal, with the offset range limited to less than 15 km. The fold ranged from 1 to 4. Figure 2.3 displays the resulting stacked section, with the processing sequence listed in Table 2.1. Amplitudes are squared for display.

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<th>Process</th>
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<tr>
<td>Trace edit</td>
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<td>Bandpass filter</td>
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The stack section in Figure 2.3 shows three main levels of reflectivity, one at 3 s beneath the Chocolate Mountains [Zone A], another at 5.5–7.0 s [Zone B], and a third between 8 and 9 s [Zone C]. The upper crustal reflections beneath the Chocolate Mountains [Zone A] correspond to the reflections observed in nearby industry data by Morris et al. [1986] [90] mentioned above, which have been interpreted to be the base of the Oroopia Schist. The RISC data shed no new light on this problem, and this interpretation still remains valid, though untested.

The midcrustal reflective zone [Zone B] can be up to 1.5 s in thickness. It is undulatory in character, especially to the northeast where it is also much lower in amplitude. Individual reflection lengths at vertical incidence are on the order of 100-150 meters in the shot gathers, and increase to at least twice that in the stack, while the band of energy can be followed for tens of kilometers. The low amplitude undulatory band of reflections in the northeastern half of the line have been observed in industry profiles recorded in the Milpitas Wash valley, and have
Figure 2.3  Low-fold stack of the RISC data. Note the midcrustal reflectivity at 5.5 s, the Moho beneath at 8.5 s, and the northeast dipping energy connecting the midcrust with the Moho. Amplitudes are squared for display.
been termed the "Milpitas Wash Reflections" (MWR) (Morris et al., 1986) [90]. In the southwestern half of the profile beneath the Chocolate Mountains, the mid-crustal reflectivity has higher amplitudes, suggesting stronger impedance contrasts within the zone. These higher amplitude reflections will be the main focus of the stochastic modeling described in Chapter 3.

The reflection band between 8 and 9 s [Zone C] is interpreted to be the Moho. It contains higher amplitudes within the northeastern half of the profile, though it is observed throughout the entire spread. An explanation for this amplitude change is found when the RISC stack is combined with the velocity model determined from traveltime modeling of the PACE refraction data (Figure 2.4) (Parsons and McCarthy, 1996) [94]. The most outstanding feature of this model is the high-velocity (6.9 km/s) body or layer beneath the Salton Trough, which pinches out northeast of the Chocolate Mountains. Due to the inherent low lateral resolution in refraction data, the northwestern extent of this body is difficult to constrain by the PACE refraction data alone. However, a northeast dipping reflection sequence [Zone D] connecting the midcrustal reflections beneath the Salton Trough with the Moho reflection beneath the Chocolate Mountains is observed in the stack in Figure 2.3. This is also in the region where the Moho reflections become higher in amplitude. We have therefore interpreted the dipping reflectors to be the northeastward lateral extent of the high-velocity body. Using this interpretation, the travel-time error for the PACE refraction modeling was reduced (Parsons and McCarthy, 1996) [94].

2.3.2 Wide-angle data

A number of off-end shots were recorded into the reflection spread, providing offset ranges of up to 180 km for wide-angle reflections from both the Salton Trough and the southern Basin and Range. The northern shots were recorded by the southern half of the reflection spread, and the southern shots were recorded by the northern half in a separate deployment. This geometry was chosen so that the reflection energy from the off-end shots would have bottoming points beneath the recording spread.

A particularly interesting phase is observed at later times than \( PmP \) from a number of northern shots. Figure 2.5 displays this phase as observed from Shotpoint 111, and Figure 2.6 displays this phase as observed from Shotpoint
Figure 2.4  Deterministic velocity model for the RISC experiment based upon analysis and interpretation of the PACE 1992 refraction data and the RISC 1992 reflection data (modified from Parsons and McCarthy, 1996). A line drawing from the near-offset reflection stack is superimposed upon the bulk velocity model. Numbers along top of figure in black ovals are shotpoint numbers. The line connecting the shotpoint numbers is topography. Numbers within the velocity models are P-wave velocities in km/s.
20. The phase is spatially stationary with respect to the receivers from shot to shot, indicating that it is a diffraction (Claerbout, 1985) [15]. The top of the diffraction hyperbola is located near where the high-velocity body is interpreted to pinch out in the stacked image (beneath Shotpoint 107), and provides further evidence for the termination of this body. The southern half of the diffraction hyperbola merges with the $PmP$ phase, indicating that this diffraction originates from the crust-mantle boundary.

2.4 Discussion and implications

The near-offset and wide-angle seismic data presented here have illuminated the large-scale crustal structure of the edges of the Salton Trough. This deterministic velocity model is displayed in Figure 2.4. Though not the most obvious feature in the reflection data, the most interesting feature imaged is the top of a high-velocity layer at the base of the crust. The top of this layer is identified by a dipping zone of weak reflections beneath the southern deployment of the RISC data, and intersects the Moho approximately beneath Shotpoint 107 (Figure 2.3 and 2.4). In addition to the stacked image, wide-angle shots to the north record a high-amplitude diffraction centered beneath Shotpoint 107. Though the velocity within this layer is not defined by the RISC data, the presence of high velocities at the base of the crust beneath the Chocolate Mountains has been identified by traveltime modeling of the PACE refraction data (Parsons and McCarthy, 1996) [94]. This layer is centered beneath the Salton Trough, but its northeastward lateral extent was not resolved by the PACE data alone. The combined evidence of a thinning high-velocity body, a dipping zone of reflectors intersecting the Moho beneath Shotpoint 107, and a diffraction originating from that same point as well, suggests that combining the reflection and refraction data has successfully constrained the northeastward lateral extent of this body.

A 6.9 km/s $P$-wave velocity is consistent with laboratory measurements of gabbros corrected for pressure and high temperature (Holbrook et al., 1992) [58]. For this reason, Parsons and McCarthy [1996] [94] interpret the high velocities at the base of the crust beneath the Salton Trough as evidence for emplacement of mafic magmas during Salton Trough rifting, in agreement with other workers (Fuis et al., 1984; Lachenbruch et al., 1985) [33, 73]. The presence of such a body beneath an active rift is not surprising considering the present knowledge of
Figure 2.5  a) Vertical component seismogram from Shotpoint 111. Precritical PmP is clearly visible. Note the high-amplitude diffraction below PmP. b) Synthetic seismogram from deterministic velocity model showing location of diffraction from the edge of high-velocity body.
Figure 2.6  Vertical component seismogram from Shotpoint 20. Precritical $PmP$ is clearly visible. Note the high-amplitude diffraction below $PmP$. 
oceanic rift systems and the creation of new crust by magmatic accretion. What is surprising is the lateral extent over which this body is observed.

1-D heat flow modeling indicates that the base of the crust beneath the Salton Trough is likely at the basalt solidus, and that magmatic intrusion or emplacement at the base of the crust has kept pace with sedimentation at the surface (Lachenbruch et al, 1985) [73]. They calculated magmatism to cover a width of 150 km parallel to the spreading direction, indicating that the active locus of spreading and seismicity is ephemeral, at least in the NW-SE direction. 2-D models of extension also predict a changing locus of extension at early stages of rifting, parallel to the spreading direction (eg. Hopper and Buck [in press] [64]). Therefore, a gabbroic layer beneath the Salton Trough can be expected over a large area to the northwest and southeast of the current active zone of spreading. But what is expected in the SW-NE direction?

Along oceanic ridges, magmatism is generally observed to decrease approaching the bounding transform (Detrick et al., 1993) [21]. However, in the Salton Trough, the closest active transform to the edge of the high-velocity body is more than 35 km to the southeast. This implies that either the oceanic ridge analogy is not appropriate for the Salton Trough, or that the high-velocity layer beneath the Chocolate Mountains is not due to Salton Trough rifting. There are a number of reasons why the oceanic ridge analogy may be inappropriate. The most obvious reason is that the Salton Trough is a continental rift, not an oceanic rift. Another reason is that the Salton Trough is a transtensional rift system, and is dominated by transform faults rather than spreading ridges. A number of continental rift systems have been studied using modern geophysical techniques (e.g., the East African rift, the Rhinegraben, and the Rio Grande rift). However, like oceanic ridges, these rifts are also dominated by normal faults rather than transforms, and are at a relatively limited stage of rifting.

One area often compared to the Salton Trough is the Red Sea rift, which is part of the Afro-Arabian rift system between the Arabian peninsula and the African continent. Like the Salton Trough, the northern Red Sea is an active continental rift believed to be at the initial stages of seafloor spreading. However, again the Salton Trough is substantially different. The borders of the Red Sea are characterized by normal faults (Cochran and Martinez, 1988; Martinez and Cochran, 1988) [16, 83], whereas the borders of the Salton Trough are defined by transform faults. The present-day transforms within the trough and the Gulf of
California are > 50 km in length, as opposed to the much shorter, < 10 km, rift segments (Lonsdale, 1989) [81], and is thus much more of a transtensional system. Furthermore, the Red Sea is sediment starved (Martinez and Cochran, 1988) [83] relative to the Salton Trough, which may contain up to 15 km of sediments (Fuis et al., 1984) [33].

Extending north of the Red Sea rift is the Jordan-Dead Sea Rift, the northernmost extension of the East African rift system, and the landward extent of the Red Sea rift. Like the Salton Trough, the rift is currently dominated by strike-slip motion (Ben Menahem et al. 1976) [86], and its tectonic development prior to this present stage is still a subject of inquiry (Ginzburg et al. 1979b) [39]. Deep crustal refraction data along the axis of the rift reveal a crustal discontinuity at 19 km depth based upon identification of an intracrustal wide-angle reflection (Ginzburg et al. 1979a) [38]. The velocity below this interface is estimated to be 6.52 km/s, significantly less than that observed beneath the Salton Trough (Ginzburg et al. 1979a) [38]. However, reliable velocity estimates from lower crustal phases using refraction data of this vintage can be difficult, and the possibility remains that this layer and the lower crust beneath the Salton Trough are the result of similar processes. Unfortunately, seismic data perpendicular to the rift does exist, and hence the lateral extent of this higher velocity lower crust is unknown.

It appears, then, that a simple analogy with other rift systems cannot be used to explain the presence of the high-velocity layer outside the Salton Trough proper. The most plausible explanation of this feature comes from the fact that southeastern California has for some time been the locus of active extension and magmatism. In particular, geochemical analysis of basalt xenoliths in the Salton Buttes suggests that continental crust is not present beneath the active rift zones (Herzig and Jacobs, 1994) [56]. This is interpreted as evidence for a two-stage rift evolution of the Salton Trough in which Oligocene-Miocene Basin and Range extension initiates crustal rupture, followed by 4-0 Ma Salton Trough rifting. In addition, fossil transforms and ridges have been identified as far northeast as Shotpoint 106 (Figure 2.2), within 10 km of the edge of the high-velocity layer (Olmsted et al., 1973; Lucchitta, 1978; Lonsdale, 1989; Parsons and McCarthy, 1996) [93, 82, 81, 94]. This indicates that the locus of spreading has moved over a broad area, not just within the current active transforms. While these spreading centers were active, mafic material was likely added to the base of the crust. This model explains the presence of the high-velocity body beneath the Chocolate Mountains, the current lack of seis-
micity outside the active spreading zones (Hearn and Clayton, 1986) [55], the low
$Pn$ velocities beneath the Chocolate Mountains (Parsons and McCarthy, in press)
[94], and the broad low velocity upper mantle observed from traveltime residuals
(Humphreys and Clayton, 1990) [65], and is thus the preferred interpretation of
high-velocity body at the base of the crust.

2.5 Summary and conclusions

By combining both vertical incidence and refraction/wide-angle reflection data, a
high-velocity body has been identified beneath the Salton Trough and Chocolate
Mountains. Seismic velocities and geochemical analyses indicate that this body
is predominantly of gabbroic composition, added to the base of the crust during
riifting. This body extends more than 35 km north of the closest active trans-
form fault within the Trough, which, based on current knowledge of oceanic ridge-
transform systems, should define the northeastward lateral extent of magmatic
intrusion. However, fossil transforms indicate that the locus of rifting within the
Salton Trough has moved over a broad area, to within 10 km of the edge of the
high-velocity body identified in the RISC data, and that a proto-Salton Trough
may have existed as early as the Miocene. Thus, the gabbroic body was probably
emplaced within the last 15 Ma, and its wide lateral extent is indicative of the
ephemeral and mobile nature of spreading loci during the initial stages of conti-
nental rifting.
Chapter 3

Small-Scale, Stochastic Crustal Structures of the Salton Trough-Southern Basin and Range Transition Zone from the RISC 1992 Seismic Experiment

3.1 Introduction

The velocity models determined from the PACE refraction data and RISC reflection data delineate the gross structural features of the crust, namely crustal thickness, average crustal velocity, and the presence of the high-velocity lower crust beneath the Salton Trough. The near-offset reflection spread was designed to analyze the wavelength-scale impedance structure within the crustal column, so that smaller structures which might indicate shearing or intrusion could be identified. To investigate the effects of velocity heterogeneity on the seismograms, stochastic distributions of velocities have been superimposed upon the deterministic velocity model obtained from analysis of wide-angle traveltimes (Parsons and McCarthy, 1996) [94] (Figure 2.4) and interpretation of the RISC stacked image (Figure 2.3). The stochastic models are designed to reproduce qualitatively the recorded wavefield, but are also tied to the in situ seismic velocities and geologic maps from the exposed basement rocks in the Chocolate Mountains. The computed synthetic seismograms from these models must not only resemble the recorded data qualitatively, but also contain similar amplitude relations between the reflective and nonreflective zones.

3.2 Justification of stochastic velocity modeling of RISC data

Before proceeding further with the stochastic analysis of the RISC data, the applicability of stochastic velocity models to the RISC data must be justified. There are
a number of ways of testing whether a stochastic velocity model is appropriate for simulating the recorded wavefield. The most direct evidence comes from geologic maps of upper, mid, and lower crustal rocks exposed at the surface. As discussed in the introduction, the velocity distributions derived from such outcrops can indeed be described stochastically. Therefore many geologic structures, especially within the crystalline crust, can be modeled stochastically. Given this observation, it needs to be shown that the recorded wavefield displays characteristics consistent with reflectivity from a stochastic medium.

One such test uses the measurements of the coefficient of coherence (Foster and Guinzy, 1968) [31]. The target zone of interest in this study is the midcrustal band of reflections (at 5.5-7.0 s TWT) observed above the thinning 6.9 km/s high velocity body. Figure 3.1 displays an example of this reflectivity at near offsets from Shotpoint 106 in the Chocolate Mountains. Gibson and Levander [1991] [37] observed that the coherence of reflections increases with offset in a characteristic manner for a wavefield scattered from a heterogeneous target zone of velocity perturbations having a smooth 2-D fractal fabric and a Gaussian pdf. Figure 3.2 shows the coefficient of coherence for the midcrustal reflectors at a frequency of 10 Hz as a function of trace separation (lag). Prior to making these measurements, the shot gather was passed through a wave-equation datuming process to remove $S_g$ and surface waves as well as to remove near-surface traveltime perturbations (Larkin and Levander, in press) (Appendix A) [75]. A coherence value of 1.0 corresponds to two identical traces, and a coherence value of 0.0 corresponds to two completely uncorrelated traces. A consistent increase in coherence with offset is apparent out to 20 km offset, a 1 to 1 target-depth to offset ratio. This offset-dependent trend is predicted by theoretical and empirical results of a wavefield back-scattered from a random, heterogeneous velocity field (Levander and Gibson, 1991) [77], and has been confirmed on synthetic seismograms from the models presented below. The amplitudes of the reflections northeast of this point decrease significantly, a feature we believe is due to a change in the impedance of the target zone, so longer offsets were not used.

Another test of the applicability of stochastic velocity models involves frequency tuning. If the heterogeneity responsible for the backscattered field is self-similar within the bandwidth of the source probe (a basic characteristic of a von Kármán distribution), then no tuning or frequency enhancement should be produced (Holliger et al., 1994) [62]. Figure 3.3 displays the amplitude spectra of $P_g$
Figure 3.1  a) Shot gather from Shotpoint 106 showing near offset reflectivity in the midcrust. Data has been bandpass filtered and corrected for geometric spreading. b) Amplitude decay curve of ensemble average after NMO correction. Midcrustal reflectivity corresponds to an amplitude increase of $\sim 10$ db.
Figure 3.2 Coherence versus offset for the midcrustal reflections of SP106. The increase of coherence with offset is consistent with the wavefield originating from a stochastic zone of velocity heterogeneity.
and the midcrustal reflective zone at near offsets. No frequency selection is evident from the midcrustal spectra relative to the Pg spectra, which is consistent with a self-similar representation of crustal heterogeneity.

3.3 Stochastic velocity models from the RISC experiment

The previous analysis has shown that the midcrustal reflective zone has the seismic signature of a stochastic medium. I now attempt to quantify features of the velocity distribution which are responsible for the reflectivity. If the reflections are due to the presence of horizontal mafic intrusions, an estimate of the total volume of reflecting bodies would constrain the volume of mafic intrusion within the reflective crustal column. This would provide direct evidence for the amount of mantle-derived magma injected into the crust during extension.

3.3.1 Surface model and upper crust

Because the target zone of interest is approximately 20 km deep, it is important to include the effects of the near surface and upper crustal scattering in the simulations. A deterministic near-surface velocity model across the study area was developed by inverting the traveltimes of shallow refracted arrivals (Zelt and Smith, 1992) [121] from the 10 inline shots. Horizontally oriented high and low velocity layers were then placed onto this deterministic velocity model. The aspect ratio of these bodies is 10:1, with the vertical scale corresponding to the finite-difference grid spacing of 20 m. Thus the high and low velocity sediments have characteristic scales of 200x20 m, and correspond to interlayered lacustrine and playa sediments and basalts, all known to exists within the basins of the western United States. The sediments have a fractal dimension of 2.5, the smallest possible for a binary fractal media (Goff et al. 1994) [42], which allows for layers of variable thickness and length to exist within the basin.

Between the sediments and the reflective zone at 20 km, a relatively transparent crust is observed. As discussed in Chapter 2, the basement outcrop within the Chocolate Mountains is composed of the Orocopia schist and the Precambrian/Mesozoic gneisses. It is unclear which of these two terranes compose the bulk of the upper crust at depth. I therefore have modeled both to determine if velocity fields corresponding to these outcrops would produce the observed transparent backscattered field. The “Orocopia schist” velocity model is displayed in
Figure 3.3  Amplitude spectra of $Pg$ and the midcrustal zone of reflections. No frequency selection is apparent in the spectra from the midcrust, which is consistent with the wavefield originating from a stochastic zone of velocity heterogeneity.
Figure 3.4. This model is defined by a von Kármán covariance function with a fractal dimension of 2.75 and horizontal and vertical characteristic scales of 1400 m and 2600 m, respectively. This spatial fabric is taken from digitized maps of the gneissic/granitic upper crust of the Strona-Ceneri Zone in northern Italy (Holliger and Levander, 1994) [61]. While the location and lithology of the Strona-Ceneri upper crust is different from that of the Orocopia schist, it is the only digitized map which suggests a continuous and Gaussian pdf, and hence is a convenient end-member upper crust. I envision that the Orocopia schist is a structurally deformed anisotropic material which now exhibits a continuous, heterogeneous velocity distribution. A continuous pdf also agrees with Dillon's [1990] [24] comments on the homogeneity of the exposed Orocopia Schist. The velocity fluctuations in the Orocopia schist upper crustal model therefore might obey a smooth Gaussian pdf. The velocities were chosen to coincide with laboratory measurements of a rock sample from the Orocopia schist which shows 15% anisotropy (Table 3.1), as well as the deterministic velocity gradient obtained from traveltime modeling. This gradient is 0.25 s$^{-1}$ above 5 km depth, and 0.018 s$^{-1}$ beneath 5 km depth.

The Precambrian/Mesozoic gneiss model (Figure 3.5) also contains an upper crust defined by a von Kármán covariance function, but with a fractal dimension of 2.5 and horizontal and vertical characteristic scales of 770 m and 910 m, respectively. The characteristic scales correspond to the spatial distribution of Mesozoic intrusions within the Precambrian gneisses exposed in the Chocolate Mountains, and were determined by statistical analysis of 1:25000 geologic maps (Dillon, 1976) [23]. The Precambrian unit is generally a quartz-diorite to diorite, with abundant meta-gabbro, anorthosite gabbro, meta-pyroxenite and meta-peridotite present. The Mesozoic unit is generally a quartz monzonite to granodiorite orthogneiss. It is not possible to differentiate the details within these two units with the available geologic maps, so I group them separately and assign a homogeneous velocity to each, recognizing that the smaller-scale bodies within these units, especially within the Precambrian rocks, may produce a significant scattered field. The amount of upper crustal scattering from the binary upper crustal model is of course dependent upon the impedance contrast between the two media. The assigned velocities are consistent with laboratory measurements of similar rock types from cratonic regions (e.g., Fountain and others, 1990) [32], with the Mesozoic units given a velocity of a felsic gneiss, and the Precambrian units a velocity of an intermediate gneiss. Figure 3.6 displays a digitized outcrop map of the Precambrian/Mesozoic
Figure 3.4  
a) Stochastic velocity model of an upper crust with a continuous (Gaussian) pdf and a von Kármán spatial fabric. The velocities used in the Gaussian pdf are consistent with velocities obtained from laboratory measurements of a rock sample from the Orocopia Schist.
b) 1-D velocity profile at the 8 km location in the model, showing velocities and gradient.
basement rocks and the statistical realization used in the finite-difference simulations. The size, distribution, complexity, and roughness of the geologic map are reasonably well-preserved in the realization. These velocity perturbations are superimposed upon the deterministic velocity gradients described above.

A comparison of the synthetic seismograms from the Orocopia Schist model and the Precambrian/Mesozoic gneiss model is shown in the amplitude decay curves in Figure 3.7. These curves were obtained by summing the decay curves from 200 near-offset traces which have recorded the backscattered field from these two models. At 5 s TWT, the backscattered field from the Precambrian/Mesozoic gneiss model is 20 db above that from the Orocopia schist model. In addition, the 20 db decrease from 1 to 5 s TWT from the Precambrian/Mesozoic gneiss model is consistent with the observed record from Shotpoint 106 (Figure 3.1b). For this reason, the Precambrian/Mesozoic gneiss model is used in the following simulations.

Rheological and buoyancy arguments suggest that at mid and lower crustal levels, intrusions may pond with a horizontal orientation (McCarthy and Thompson, 1988; Glazner and Ussler, 1989; Parsons et al., 1992; Holliger and Levander, 1994) [85, 41, 95, 60], providing horizontal impedance contrasts likely to produce high reflectivity. The mid and lower crust is therefore modeled as horizontal (17-22 km depth) and vertical (22 km to the Moho) elongate bodies, with the horizontal bodies producing the reflectivity between 5.5 and 7.0 s and the vertical bodies producing the less reflective zone between 7.0 and 8.5 s. A normal incidence reflection coefficient of 0.08 is held roughly constant for the sills and dikes in all simulations. This contrast corresponds to an intermediate rock in contact with a mafic rock, a geologic model in which the lower velocity Basin and Range crust determined from the PACE refraction spread is intruded by material whose velocities are consistent with that of the high velocity lower crust beneath the Salton Trough. The fractal dimension for these intrusions was held fixed at 2.7, consistent with measurements of extended and intruded crustal rocks in the Ivrea Zone (Holliger and Levander, 1992) [59]. Body size will be discussed in the following section on volume estimation.

Finally, a homogeneous high velocity wedge is placed above the crust-mantle interface to represent the high velocity lower crust identified in the PACE refraction profile. Though the homogeneity of this layer underpredicts the backscattering observed in the lower crust, its velocity does help produce a Moho reflection with
Figure 3.5  a) Stochastic velocity model of a bimodal upper crust with a spatial fabric corresponding to the distribution between the Precambrian (lighter) and Mesozoic (darker) basement rocks exposed in the Chocolate Mountains. b) 1-D velocity profile through a single column at the 8 km location in the model, showing velocity steps and gradient.
Figure 3.6  a) Digitized map of Precambrian/Mesozoic basement in Chocolate Mountains. b) Statistical realization of a) used in finite-difference simulations. Both maps are 5 km x 2.5 km in area.
Figure 3.7 Amplitude decay curves of backscattered fields from the Orocopia schist upper crustal model and Precambrian/Mesozoic gneiss upper crustal. The Precambrian/Mesozoic gneiss decay curve matches the decay curve from Shotpoint 106, with a 20 db decrease from 1 to 5 s. Seismograms were corrected for geometrical spreading before the computation of the decay curves.
an amplitude similar to that observed. The upper mantle is also assumed to be homogeneous with a velocity of 7.95 km/s. Figure 3.8 displays the combined deterministic and stochastic velocity model used in the simulations that follow. Tables 3.2 and 3.3 summarizes the rock properties used in the various simulations. Densities are assigned using the empirical velocity-density relationship of Nafe and Drake [1957] [91]. Shear wave velocities are assigned assuming a 0.29 Poisson’s ratio for the mafic rocks and a 0.25 Poisson’s ratio for all others.

3.3.2 Estimating volume of heterogeneity in the midcrustal reflective zone

Given a volume percentage of intrusion, there will be more individual scatterers in a media with smaller characteristic scales than with larger ones. Hence, body size must be addressed before volumes can be estimated from the backscattered field. Pullammanappallil et al. [1995] [97] have shown that a wave-number time domain analysis of a wavefield backscattered from a 2-D, horizontally oriented von Kármán media can provide accurate measures of the horizontal characteristic scales of the media. When applied to the reflective zone beneath Shotpoint 106, this method estimates a horizontal characteristic scale of 290 ± 50 m. Therefore, a horizontal characteristic scale of 300 m is used in the proceeding simulations. Because this measurement is likely a minimum estimate in the presence of travel path heterogeneity, simulations were also run with horizontal characteristic scales of 800 m. Because the crustal seismic experiment is limited to a surface-to-surface recording geometry, and resolution is modulated by the source signature, estimation of the vertical scale has proven more elusive. For this reason, two aspect ratios were modeled as well, 4:1 and 2:1. Two characteristic scales and two aspect ratios result in four distinct body sizes for a specific volume of intrusion. Five different intrusion volumes have been simulated: 1%, 5%, 15%, 25% and 50%, resulting in a total of 20 simulations.

To compute synthetic seismograms from the above models, a 2D visco-elastic finite-difference wave propagation simulator (Robertsson et al., 1994) [99] was used to compute synthetic seismograms from the stochastic models. The dominant frequency of the Ricker wavelet source is 18 Hz. Figures 3.9a-3.13a display the results of the 300x75 m intrusion simulations. All record sections and decay curves have been corrected for geometrical spreading. As the volume of intrusion increases,
Figure 3.8 Combined deterministic and stochastic velocity model in a) 2-D and b) 1-D. % volume of intrusion in the middle to lower crust is 5%.
and hence the number of impedance contrasts increases, the level of reflectivity increases. This is particularly true between 1% and 15%. Above 15%, the wavefield becomes saturated with reflections, and little difference is visibly observable. This qualitative observation is confirmed by the amplitude decay curves plotted beside the synthetic gathers (Figures 3.9b-3.13b). Between 1% and 15%, the amplitude step at 5.5 s increases from 0 db to over 15 db, but above 15% the amplitude no longer increases. Clearly, the model which best matches the observed data in Figure 3.1 is the 5% model, with a moderately dense reflectivity and a 10 db amplitude step at 5.5 s. A quantitative estimate of volume of heterogeneity based upon reflection density is described in Appendix B. This measure indicates that the 5-15% models best predicts the observed reflectivity.

Figures 3.14-3.16 display synthetic seismograms from the models with 5% volume of intrusion, but with varying body size. As the size of the bodies becomes larger, the number of individual impedance contrasts per unit area decreases, and the density of reflections therefore decreases. Thus, as the characteristic scales increase, the volume of intrusion must increase to maintain a given level of reflection strength and density. With the body sizes used in these models, however, 5% intrusion remains sufficient to produce the highly reflective crust.

Deriving quantifiable conclusions from amplitude studies is potentially wrought with circular arguments, due to the unknown absolute source amplitude and path effects to the target. Two aspects of the velocity models influence the amplitude response of the reflective zone: the heterogeneity of the upper crust and the impedance contrasts of the reflective bodies. If the upper crustal velocity fields used in these simulations produce insufficient backscattered energy within the upper crust, then a higher percentage of scattering material would need to be present within the midcrust. However, the amplitude decay from 1 to 5 s in the observed data and the synthetics is the same, indicating that the heterogeneity is sufficient. If the impedance contrasts in the sill region are too high, the above comparison would be an underestimate of the volume of intrusion. With the available data, it is not possible to make an exact measure of the impedances present within the midcrust. However, the close proximity of the reflectivity to the high velocity lower crustal layer suggests that these features are related, so it is assumed that the velocities of the reflective bodies coincide with the velocities of the lower crust. Given this, it is estimated that little more than 5% intrusion is responsible for the highly reflective midcrust beneath the Chocolate Mountains.
Figure 3.9  a) Synthetic seismogram for the 300x75 m 1% intrusion model and b) its corresponding amplitude decay curve. This amount of intrusion, even for the smallest body size, does not reproduce the level of reflectivity observed beneath Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.10  a) Synthetic seismogram for the 300x75 m 5% intrusion model and b) its amplitude decay curve. This amount of intrusion best reproduces the level of reflectivity observed beneath Shotpoint 106, both in the density of reflectors and the amplitude response. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.11  a) Synthetic seismogram for the 300x75 m 15% intrusion model and b) its amplitude decay curve. This amount of intrusion overpredicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.12  a) Synthetic seismogram for the 300x75 m 25% intrusion model and b) its amplitude decay curve. This amount of intrusion overpredicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.13  a) Synthetic seismogram for the 300x75 m 50% intrusion model and b) its amplitude decay curve. This amount of intrusion overpredicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.14  a) Synthetic seismogram for the 300x150 m 5% intrusion model and b) its amplitude decay curve. This amount of intrusion predicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.15  a) Synthetic seismogram for the 800x200 m 5% intrusion model and b) its amplitude decay curve. This amount of intrusion predicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Figure 3.16  a) Synthetic seismogram for the 800x400 m 5% intrusion model and b) its amplitude decay curve. This amount of intrusion still predicts the density and amplitude response of Shotpoint 106. The record section and decay curve have been corrected for geometrical spreading.
Finally, because the above finite-difference simulations assume a 2-D velocity field, it is necessary to consider the possible effects of a more realistic 3-D velocity field. At first glance, out-of-plane heterogeneity would simply add more reflections and diffractions to the recorded wavefield. This would mean that the above volume estimates would be over-estimates of the actual volume of scatterers within the reflective zone. However, this assumes that all of the scatterers observed in a 2-D velocity model would scatter energy only in the 2-D recording plane, even if the bodies were actually 3-D. This is clearly not the case. Therefore, the extension to 3-D will scatter some energy out of the recording plane as well as add scattered energy back from out of the plane. The relative contributions of these two competing effects are difficult to quantify, but it is possible that they may actually cancel each other. In this case, if the scattering velocity field is indeed stochastic, a 2-D velocity model would suffice in this instance.

3.3.3 Effects of stochastic heterogeneity on Moho signature

The stochastic velocity distributions within the upper and mid crust are reasonable representations of velocity distributions within the crust, so it is therefore of interest to observe the effects of such heterogeneity on reflections which originate below the heterogeneity, such as the Moho reflection. In the models described above (Figure 3.8), the crust-mantle boundary is a simple, horizontal first-order discontinuity between the homogeneous high-velocity lower crustal wedge and the homogeneous upper mantle. Although such a simple interface may not actually exist, it does serve to demonstrate clearly the path effects of wavelength-scale heterogeneity within the crust.

The reflection Moho in all of the simulations consists of a 0.5 s zone of discontinuous reflections (Figure 3.9-3.16). Much of this scattering takes place in the near surface and upper crust, for even the 1% intrusion model produces this signature. More intriguing still is that the upper crustal reflectivity would be termed "transparent" or "unreflective", yet it obviously plays a prominent role in deeper reflectivity. As the percentage of midcrustal heterogeneity increases, the reflections within the reflection Moho become shorter and more discrete. Such zones of discontinuous reflectivity in the Moho reflection are often interpreted as evidence for compositional layering at the crust-mantle boundary (eg., Hale and Thompson, 1982; Hauge et al., 1987; Hauser et al., 1987; Valasek et al., 1989) [45, 47, 48, 116].
These simulations suggest that such a signature should be expected for even a first-order discontinuity given the wavelength-scale crustal heterogeneity observed at the surface. This result underscores the importance of acquiring as much information as possible about the near surface and upper crustal velocity distributions when interpreting deeper reflections.

3.4 Discussion and implications

The wide-angle and near-offset seismic data presented here have illuminated the large and small-scale crustal structure of the edges of the Salton Trough. A high velocity layer at the base of the crust beneath the Salton Trough is constrained to pinch out beneath the Chocolate Mountains. This layer is identified in the PACE 1992 velocity model based on the analysis of seismic refraction data, and its lateral extent is constrained by the RISC reflection and wide-angle data. This layer is interpreted to be the result of magmatic additions to the base of the crust during Basin and Range extension and/or Salton Trough rifting. The RISC seismic reflection data does not shed light on the age of emplacement of the mafic lower crust; however, the highly reflective midcrust beneath the Chocolate Mountains is spatially confined to the region above the high velocity lower crustal layer. For this reason, regardless of the time of emplacement, the midcrustal reflectivity is believed to be genetically linked to the emplacement of the high velocity lower crust beneath the Chocolate Mountains.

The high amplitude reflections adjacent to this high velocity layer have been modeled as horizontal intrusions with the same material properties as the lower crust. These intrusions form a midcrustal zone of mafic sills 5 km thick. The emplacement of horizontal intrusions within an otherwise extensional tectonic region may at first seem contradictory. However, recent rheological arguments have shown that such orientations can be expected at zones of contrasting strength, such as at the crust-mantle boundary or in the midcrustal brittle/ductile transition (McCarthy and Thompson, 1988; Glazner and Ussler, 1989; Parsons et al., 1992; Holliger and Levander, 1994) [85, 41, 95, 60]. As little as 5% of these horizontally oriented intrusions can produce the highly reflective lower crust. What is intriguing about this result is that such a small volume component can dominate the reflection signature, and potentially obscure important large-scale features, such as the high-velocity layer below. In addition, such a small amount of in-
trusion will have virtually no effect on the average crustal velocity, yet produce both vertical incidence and wide-angle "reflections". Exactly how much mantle derived material is emplaced into the crustal column during an extensional event such as Basin and Range rifting is still a topic of debate (Gans et al., 1989; Best and Christiansen, 1991) [36, 9]. Such studies involve assumptions of original crustal thickness, amount of stretching, and the ratio between extruded and trapped magma. The seismic reflection method probes these intrusive bodies as they exist within the present-day crust, and therefore can be used to estimate volumes of intrusion. The use of stochastic velocity models based upon geologic maps may be a necessary tool to measure this quantity, and others related to it, such as body size and aspect ratio. Such information will prove necessary for future interpretations of deep crustal seismic data.

3.5 Summary and conclusions

The results from the stochastic analysis of the RISC near-offset seismic data can be summarized as follows:

1) Velocity models of the upper crust based upon the distribution of Precambrian and Mesozoic gneisses in the Chocolate Mountains reproduces the observed backscattered field.

2) Amplitude modeling with visco-elastic finite-difference simulations indicates that the high amplitude reflective midcrust beneath the Chocolate Mountains can be due to little more than 5% horizontal intrusion of material with similar velocities as that observed in the high velocity lower crust.

3) Wavelength-scale stochastic heterogeneity within the upper and middle crust can produce a discontinuous zone of reflections from a first-order discontinuity. Such reflectivity is often interpreted as evidence for compositional layering at the crust-mantle boundary.
Table 3.1  Compressional wave seismic velocities of Orocopia Schist from laboratory measurements. Measurements are post maximum pressure. A-velocity is parallel to foliation and perpendicular to lineation. B-velocity is parallel to both foliation and lineation. C-velocity is perpendicular to both foliation and lineation.

<table>
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<th>B-velocity (km/s)</th>
<th>C-velocity (km/s)</th>
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<td>6.470</td>
<td>5.639</td>
</tr>
<tr>
<td>4.860</td>
<td>6.182</td>
<td>6.479</td>
<td>5.657</td>
</tr>
</tbody>
</table>

Table 3.2  Covariance function parameters used to create spatial fabric in the viscoelastic finite-difference simulations.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Model</th>
<th>$D$</th>
<th>$a_x$ (m)</th>
<th>$a_z$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basins</td>
<td>None</td>
<td>2.5</td>
<td>200</td>
<td>20</td>
</tr>
<tr>
<td>Upper crust</td>
<td>Orocopia schist</td>
<td>2.75</td>
<td>1400</td>
<td>2700</td>
</tr>
<tr>
<td></td>
<td>Pc/Mz gneiss</td>
<td>2.5</td>
<td>963</td>
<td>817</td>
</tr>
<tr>
<td>Lower crust</td>
<td>Ivrea zone</td>
<td>2.7</td>
<td>300</td>
<td>75</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>300</td>
<td>150</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>800</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>800</td>
<td>400</td>
</tr>
</tbody>
</table>
Table 3.3  Rock properties statistics used in the viscoelastic finite-difference simulations. OS - Orocopia schist model, PMG - Precambrian/Mesozoic gneiss model, IZ - Ivrea Zone, H - Homogeneous.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Model</th>
<th>pdf</th>
<th>$V_p(\text{avg})$ km/s</th>
<th>$\sigma$</th>
<th>$\rho(\text{avg})$ g/cm$^3$</th>
<th>$Q_p, Q_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basins</td>
<td>None</td>
<td>Binary</td>
<td>2.4, 2.9</td>
<td>0.25</td>
<td>1.16-1.35</td>
<td>50, 17</td>
</tr>
<tr>
<td>Fractured basement</td>
<td>OS</td>
<td>Gaussian</td>
<td>5.2-5.5</td>
<td>0.25</td>
<td>2.22-2.34</td>
<td>200, 100</td>
</tr>
<tr>
<td></td>
<td>PMG</td>
<td>Binary</td>
<td>5.2, 5.5</td>
<td>0.25</td>
<td>2.22-2.34</td>
<td>200, 100</td>
</tr>
<tr>
<td>Upper crust</td>
<td>OS</td>
<td>Gaussian</td>
<td>5.95-6.25</td>
<td>0.25</td>
<td>2.51-2.62</td>
<td>1000, 500</td>
</tr>
<tr>
<td></td>
<td>PMG</td>
<td>Binary</td>
<td>5.9, 6.2</td>
<td>0.25</td>
<td>2.49-2.60</td>
<td>1000, 500</td>
</tr>
<tr>
<td>Lower crust</td>
<td>IZ</td>
<td>Binary</td>
<td>6.2, 6.8</td>
<td>0.25, 0.29</td>
<td>2.60, 2.83</td>
<td>1000, 500</td>
</tr>
<tr>
<td>High-velocity wedge</td>
<td>H</td>
<td>N/A</td>
<td>6.9</td>
<td>0.29</td>
<td>2.87</td>
<td>1000, 500</td>
</tr>
<tr>
<td>Upper mantle</td>
<td>H</td>
<td>N/A</td>
<td>7.95</td>
<td>0.29</td>
<td>3.27</td>
<td>1000, 500</td>
</tr>
</tbody>
</table>
Chapter 4

Modeling Wide-angle Data from the RISC Experiment Using Combined Deterministic and Stochastic Velocity Fields

4.1 Introduction

The complicated wavefields recorded in both near-vertical and wide-aperture deep-crustal seismic experiments indicate that a significant fraction of the source energy is scattered from velocity heterogeneities within the crust. This scattering can occur within the sedimentary cover, the crystalline crust, and/or within the zone producing the reflection itself (e.g., the crust/mantle boundary). In Chapter 3, I discussed the response of the high-frequency, near-offset reflected wavefield. The target was a reflective zone with a relatively small amount of heterogeneity, and path effects were modeled by introducing a heterogeneous upper crust and basin structure. The upper crustal velocity structure was based upon rock velocities and distributions exposed at the surface within the Chocolate Mountains. As discussed in Chapter 2 and 3, these rocks are presumed to form the basement rock beneath most of southeastern California and western Arizona.

In this chapter, I investigate the path effects of this heterogeneous crust on the lower frequency wide-angle experiment by developing crustal-scale models for the southern Basin and Range province of southeastern California and the Salton Trough. I then compare synthetic seismograms from these models with data from the wide-angle RISC experiment. A quantitative analysis of both the recorded and the synthetic seismic records provides information on the effects of realistic velocity heterogeneity on the wide-angle seismic wavefield and also the potential effects of the near-surface, which in the case of crustal seismology consists of the sedimentary basins.
4.2 Description of wide-angle data

In addition to the inline shots analyzed in Chapter 2, a number of off-end shots were also recorded into the reflection spread, providing offset ranges of up to 180 km for wide-angle reflections from beneath the Salton Trough and the southern Basin and Range (Figure 2.2). The northern shots were recorded by the southern half of the reflection spread, and the southern shots were recorded by the northern half in a separate deployment. This recording geometry was chosen so that reflected energy from the off-end shots would have bottoming points beneath the vertical-incidence spread. The usefulness of this design has already been demonstrated in Chapter 2, where the identification of a high-amplitude diffraction (Figures 2.5 and 2.6) provided further evidence for the termination of a high velocity body beneath the Chocolate Mountains.

The recording geometry of the wide-angle experiment is in the form of a walk-away spread. The nominal shot spacing is 10 km. Each shot is recorded by a single deployment of 435 three-component receivers spaced 50 m apart, resulting in a 22 km recording aperture. Because the aperture is greater than the shot spacing, we can combine the record sections from each shot into a single supergather. Figure 4.1 displays the geometry for both the northern and southern supergatherers. The supergather constructed from the northern shots is termed the Basin and Range supergather because the sources and midpoints lie within the Basin and Range province. By the same logic, the supergather constructed from the southern shots is termed the Salton Trough supergather.

The utility in constructing these supergatherers lies in their resemblance to typical shot gatherers. If the deterministic part of the crust is 1-D, this supergather will "appear" as a large shot gather spanning the entire offset range of the experiment, with the same amplitude with offset characteristics as a simple shot gather. The bottoming points for the wide-angle reflections from northern shots lie between shotpoints 108 and 20 (Figure 4.1b). As noted in Chapter 2, the Proterozoic/Mesozoic basement rocks (and/or possibly the Orocopia schist) exposed within the Chocolate Mountains is believed to form the basement terrane beneath a large part of southeastern California and western Arizona (Barth and Ehlig, 1988; Glazner and O'Neil, 1989; May and Walker, 1989; Tosdal and others, 1989) [4, 40, 84, 115]. Thus, the energy recorded in the Basin and Range supergather remains in a single tectonic domain (the southern Basin and Range).
Figure 4.1 Schematic ray diagram showing geometry of wide-angle experiment. a) Geometry of the southern shots into the northern deployment. The resulting supergather is termed the Salton Trough supergather, because the source and midpoints are within the Salton Trough. Note the 2-D structure of the lower crust, which causes travelt ime discrepancies in the supergather. b) Geometry of the northern shots into the southern deployment. The resulting supergather is termed the Basin and Range supergather because the sources and midpoints are within the Basin and Range. The approximate 1-D structure of the crust causes this supergather to appear as a single wide-angle shot gather.
and also within a crustal column with a similar reflective signature. By creating a supergather, only a single finite-difference simulation is necessary to reproduce the data from numerous shots, provided the deterministic model is 1-D. The Basin and Range vertical component supergather is displayed in Figure 4.2. Note how well the gather mimics a single shot gather, even though it is a composite gather of 6 shots.

The Salton Trough supergather (Figure 4.3) has bottoming points within the Salton Trough, where the lower crustal structure is dipping to the northeast at approximately 20 degrees (Figure 4.1a). With this geometry, the individual phases within the supergather will have a traveltime mismatch from shot to shot. This also complicates the amplitude with offset characteristics, so caution is necessary when analyzing this gather as a single shot record.

One final note regarding the analysis of supergatheres must be made. Because the gather is composed of several shots, an analysis of offset-dependent amplitudes and structures is difficult. Each shot will produce a different $P/S$ amplitude ratio and a different amount of surface waves. In addition, amplitude balancing is difficult because the offset ranges which overlap between shots correspond to different surface locations. These site effects can dominate the amplitudes recorded, and thus inhibit the scaling of the shot gathers with respect to each other. For these reasons, all of the wide-angle shot gathers, both observed and synthetic, are plotted with a full trace normalization.

### 4.2.1 Vertical component

The vertical component supergatheres for both the Basin and Range and Salton Trough (Figure 4.2 and Figure 4.3) show a weak $Pg$ crustal refraction and a strong, reverberatory $PmP$ phase. The Basin and Range supergather also shows a weak $Sg$ and a strong $SmS$, phases not observed in the Salton Trough supergather. Little or no $PmS$ energy is observed.

### 4.2.2 Radial component

The radial component supergatheres for both the Basin and Range and Salton Trough (Figure 4.4 and Figure 4.5) show a weak $Pg$ crustal refraction and a strong, reverberatory $PmP$ phase. The Basin and Range supergather also shows a stronger $Sg$ and a very strong $SmS$, phases not observed in the Salton Trough supergather.
Figure 4.2  The vertical component Basin and Range wide-angle supergather. Range of contribution for each shotpoint is displayed at the bottom.
Figure 4.3 The vertical component Salton Trough wide-angle supergather. Range of contribution for each shotpoint is displayed at the bottom.
A very weak $PmS$ arrival is observed in the Basin and Range gather, but not in the Salton Trough gather.

4.3 Creation of velocity models

As in Chapter 3, I begin by defining the deterministic structure upon which I will superimpose the small-scale heterogeneity. The deterministic velocity model consists of four parts: basin, crust, crust-mantle boundary, and mantle. Each layer is homogeneous with a linear velocity gradient (two gradients in the case of the crust) defined by traveltime modeling. Wavelength-scale stochastic media are then superimposed onto this deterministic background velocity field which will reproduce the observed data as accurately as possible, and in so doing, determine which features of the velocity model are responsible for different characteristics observed in the data. To simulate the wide-angle experiment, the grid spacing within the finite-difference model must be increased to 60 m so that offsets of 150 km (the distance between shotpoints 101 and 21) can be calculated with available computer memory. This reduces the dominant frequency of the Ricker wavelet source to 5 Hz, which is an appropriate frequency band for a wide-angle crustal experiment. The top of the model is a horizontal free-surface, and the sides and edges are low-$Q$ absorbing boundaries.

4.3.1 Basin velocities, geometry, and near-surface $Q$

When making quantitative measurements of the seismic wavefield returning from deep crustal depths, it is important to address the scattering effects of the near-surface on the measurements. In mantle tomography experiments using earthquake sources, the near-surface is often the entire crustal column (e.g. Wagner and Langston [1992] [117]). In deep crustal seismics, near-surface generally means topography, basin geometry, and heterogeneity within the basin itself. The velocity structure of the crystalline crust and upper mantle is taken from the PACE 1992 velocity model of Parsons and McCarthy [in press] [94]. This model was obtained using a finite-difference traveltime modeling technique, and is composed of 1 km square grid cells. This coarse grid spacing is adequate for the modeling of the bulk velocity structure of the crystalline crust. However, depth to basement along the RISC receiver spread ranges from 0 to 2 km, which is on the order of the grid spacing of the velocity model. Thus, the resolution of the sedimentary basin structures
Figure 4.4 The radial component Basin and Range wide-angle supergather. Range of contribution for each shotpoint is displayed at the bottom.
Figure 4.5 The radial component Salton Trough wide-angle supergather. Range of contribution for each shotpoint is displayed at the bottom.
is poor, and is not sufficient for full-wavefield simulations with wavelengths less than a kilometer.

For the wide-angle simulations presented here, the structure and velocities of the basins beneath the receiver spread were determined from traveltime analysis (Zelt and Smith, 1992) [121] of first-break refractions recorded by the RISC profile from the inline shots (Shotpoints 105-110 and 126-129). The near-surface basin consists of low velocity material \( (P_{\text{avg}} = 3.1 \text{ km/s}) \) with a high \( P \)-wave velocity gradient of 0.8 s\(^{-1}\). Because the finite-difference program does not model a non-planar free-surface, the vertical distance from topography to basement is used as the basin structure for the simulations.

Northeast of the receiver spread, this near-surface velocity model is repeated to the end of the model, with minor alterations added based upon the location of basins and ranges at the surface. Southwest of the receiver spread, beneath the Salton Trough, the sediments are believed to extend all the way to the high velocity body at 14 km (Fuis et al., 1984; Herzig and Jacobs, 1994) [33, 56]. By 5 km depth, due to high velocity gradients within the sedimentary column, the upper crustal velocities are no longer different from the upper crustal velocities outside the Salton Trough. Therefore, the Salton Trough sediments are modeled with the same sedimentary velocity and gradient as observed beneath the recording spread to a depth at which this velocity function intersects the crystalline velocity function to the north. This depth occurs at approximately 3.5 km. Below this depth, the bulk velocity function is the same as to the north.

The effects of the near-surface basins are modeled by including low \( Q \) sediments above the 2-D basin structure described above at the top of the finite-difference grid. These layers correspond to interlayered lacustrine and playa sediments and basalts, all known to exist within the basins of the western United States (Okaya and Thompson, 1985) [92]. A velocity field was generated consisting of horizontally oriented bodies with characteristic scales of \( a_z = 600 \) m and \( a_z = 60 \) m, a fractal dimension of 2.5 (the smoothest possible binary fractal fabric), and a binary pdf with positive and negative velocity steps of 0.5 km/s. The net result of this statistical distribution is to produce high and low velocity planar layers with an average thickness of 60 m which occasionally pinchout horizontally.

The next step is to determine a satisfactory value of \( Q \) for the sedimentary layers. The spectral ratio method is one of the most widely used methods for estimating \( Q \) (e.g., Zelt and Ellis, 1989; Scheirer and Hobbs, 1990) [120, 103]. A
large number of spectral ratios can be calculated quickly, allowing for the use of statistical methods in the data analysis. However, a disadvantage in the method is that its reliability is strongly dependent upon the magnitude of the time difference between the data windows (Tonn, 1989) [114]. To estimate near-surface $Q$, only the first-break direct arrival over a short offset range ($< 5$ km) remains entirely within the sedimentary column. Unfortunately, arrival times between traces for this phase often differ by little more than a sample in time. However, if we can determine a large enough number of spectral ratios, inaccuracies in picking arrival times will hopefully average out. $Q$ can be defined as

$$A(r, f) = A(0, f) \exp\left(-\frac{\pi f}{Q v} r\right),$$

(4.1)

where $A(0, f)$ is the initial wave amplitude, $A(r, f)$ is the amplitude of the attenuated wave after travelling along the raypath $r$, $f$ is the frequency, $v$ is the velocity of the media, and $Q$ is the quality factor. Traditionally, this equation is Fourier transformed to remove the $r$ dependence (Tonn, 1989) [114], resulting in

$$\frac{-1}{\pi f} \ln \frac{A(r, f)}{A(0, f)} = \frac{1}{Q f}.$$

(4.2)

The slope of Equation 4.2 is $1/Q$. We can therefore estimate $Q$ by finding the slope of the best-fit line through the values determined from the left side of the equation.

To test the robustness of the method on direct arrivals, synthetic seismograms were computed using the viscoelastic finite-difference algorithm (Robertsson et al., 1994) [98] which can model the effects of anelasticity on the wavefield given a specified $Q$ value. This is the same algorithm used in the simulations of crustal data throughout this study, and thus it is essential that the method used to measure $Q$ gives the correct results for the synthetic seismograms. The experimental geometries for the tests are displayed in Figure 4.6. The first test was for a constant velocity medium (6.0 km/s), with sources and receivers embedded within the medium, and a $Q$ of 20. Source-receiver offsets range from 1 km to 8 km, and only the horizontal component was recorded. The second simulation was designed to determine the effects of velocity gradients on the measurements. This model consists of sources and receivers (vertical component) placed along the top boundary with a strong near-surface velocity gradient of 1 s$^{-1}$. In this model, $Q$ was held fixed at 50. The source for both models was a Ricker wavelet with a dominant
frequency of 18 Hz. A Ricker wavelet has approximately 1 octave on either side of the dominant frequency, thus providing 2 octaves of bandwidth over which to compute spectral ratio slopes.

The synthetic seismograms from both tests are displayed in Figure 4.7. In the first simulation, the effect of the low \( Q \) is readily apparent in the decrease in amplitude with offset. In the second simulation, a similar amplitude decay is present in the first break. The later arrivals are the surface direct wave (separating from \( Pg \)) and a converted \( S \) diving wave (at 1 s on the near trace). A spectral ratio for every possible pairing of traces is computed by letting each trace be the reference trace for every subsequent trace. For example, for a shot gather of 5 traces, \( 4 + 3 + 2 + 1 = 10 \) spectral ratios would be used to estimate \( Q \) for that gather. Unreasonable values for \( Q \), defined as \( Q < 0 \) and \( Q > 1000 \), are ignored, and an average of all other \( Q \) values is computed. Figure 4.8 displays the \( Q \) values determined from all possible trace pairings. The first value is from the ratio of the first two traces, the second value is from the first and third trace, and so on. The constant velocity model produces very little scatter in individual \( Q \) values and an overall average of 20.4328. The increase in scatter on the right side of the plot indicates a decrease in bandwidth in both traces in the ratio, resulting in larger errors in the best-fit line to the ratios. The gradient model produces much more scatter in the determined \( Q \)-values (Figure 4.9), but an accurate average \( Q \) of 50.6149. The increase in scatter towards the right side of the graph is not only a result of a decreased bandwidth, made even worse by the slower velocities, but also due to the curvature of the gradient arrival which decreases the time difference between adjacent traces with offset. Despite the large amount of scatter, the average \( Q \) value is quite accurate.

To determine an approximate \( Q \) for the basins in the observed data, spectral ratios of \( Pg \) arrivals were measured within 5 km of the source for Shotpoint 106. Before determining the spectra for each trace, top and bottom mutes were applied to the data to isolate the first 250 ms of the first break. I then followed the exact same procedures as described above for the synthetic examples, with the results displayed in Figure 4.10. A large amount of scatter is present as expected, but the vast majority of values plot below \( Q = 100 \). The average \( Q \) for this data is 46.0058. Based upon this result, the near-surface sediments within the basins of the crustal finite-difference models are assigned a \( Q_p \) of 50.
Figure 4.6  Experimental geometry used in viscoelastic finite-difference simulations to test robustness of $Q$-estimation technique. a) Constant velocity media with a $v_p$ of 6.0 km/s and $Q = 20$, with the source and the horizontal component receivers embedded within the model. b) Vertical velocity gradient model with a near-surface velocity of 2.3 km/s and a vertical gradient of 1 km/s/km and $Q = 50$, with the source and vertical component receivers at the surface.
Figure 4.7 Synthetic shot records used to test spectral ratio method of determining near-surface $Q$. a) Seismogram from Model 1. b) Seismogram for Model 2.
**Figure 4.8** $Q$ determinations for constant velocity model using spectral ratios of all possible pairs of traces. The horizontal axis represents the number of ratios computed. Despite some scatter in individual measurements due to limited bandwidth and picking resolution, the average $Q$ remains accurate.
Figure 4.9  $Q$ determinations for gradient model using spectral ratios of all possible pairs of traces. The horizontal axis represents the number of ratios computed. Despite the large scatter in individual measurements due to limited bandwidth and picking resolution, the average $Q$ remains accurate.
Figure 4.10 $Q$ determinations from spectral ratios of all possible pairs of traces for Shotpoint 106 of the RISC experiment. The vast majority of values fall below $Q = 100$, and the average $Q$ is 46.0058, and we therefore use a $Q$ of 50 for the basins in our simulations of crustal data.
The crust

The deterministic model for the crystalline crust consists of two homogeneous layers with distinct linear velocity gradients. The upper layer is between the basin bottom and 5 km depth and has a $P$-wave velocity gradient of 0.25 s$^{-1}$. This higher gradient near-surface crystalline layer is commonly observed in velocity models of continental crust, and is indicative of the depths at which cracks can still remain open, thus accounting for the slower velocities. The lower layer extends from 5 km down to the crust-mantle boundary and has a gradient of 0.018 s$^{-1}$. The velocities for both of these layers are obtained from the PACE 1992 refraction modeling of Parsons and McCarthy [in press] [94], and the traveltine inversion of the RISC first breaks.

The stochastic velocity field superimposed upon this deterministic background velocity model is the same as that used for the upper crust in Chapter 3, the statistical parameters for which are listed in Tables 3.2 and 3.3. This field is a representation of the spatial distribution of Precambrian and Mesozoic gneisses exposed in the Chocolate Mountains which are known to form the basement beneath much of southeastern California and western Arizona (Barth and Ehlig, 1988; May and Walker, 1989; Tosdal et al., 1989) [4, 84. 115].

The crust-mantle boundary and upper mantle

The crust-mantle boundary beneath the southern Basin and Range is modeled both as a first-order discontinuity and as a gradient layer at the base of the crust. This is done to address the question of whether the presence of such an intermediate layer is necessitated by the data in the presence of a heterogeneous crustal column above. Beneath the Salton Trough, the high velocity layer at the base of the crust is modeled as a homogeneous body with a linear gradient of 0.018 s$^{-1}$. Due to the variable data quality of the southern shots, modeling of the internal structure of this layer is not possible, nor is the modeling of the details of the crust-mantle boundary. The mantle in all simulations is modeled as a homogeneous halfspace with a velocity determined from the PACE 1992 model [94].
4.4 Modeling of the RISC supergathers

Because of the computational expense of forward modeling crustal seismic data using a viscoelastic finite-difference algorithm, it is important to define particular features within the recorded data which one believes is reproducible and is affected to varying degrees by the parameters which define the velocity model, be they deterministic or stochastic. A few basic features of the supergathers are so prominent that a satisfactory model must reproduce these characteristics. First, the model must produce the main phases observed, and not produce others. This, in effect, tests the effect of wavelength-scale velocity heterogeneity on the phases arising from the deterministic interfaces in the velocity model. Because the only prominent reflected phases recorded are from the crust-mantle boundary, the analysis is restricted to these phases. Second, the model should reproduce the reverberatory nature of the recorded wavefield. The cause of the reverberations can then be isolated as due to near-surface heterogeneity or crystalline heterogeneity. This latter part involves the analysis of $PmP$ and $SmS$ coda waves.

Due to the 2-D crustal structure, variable shot quality of the southern shots, and lack of knowledge of the small-scale velocity structure beneath within the Salton Trough sediments, the analysis of the Salton Trough supergathers is limited to the creation or omission of the main phases within the gather. The simplicity and consistency of the northern shots allows for more detailed analysis of seismic coda. I begin with the analysis of the northern shots or Basin and Range supergathers, followed by an analysis of the southern shots or Salton Trough supergathers.

4.4.1 Northern shots

Path effects on Moho phases

The vertical component Basin and Range supergather displayed in Figure 4.2 exhibits a strong $PmP$ arrival at offsets beyond 60 km, but which can be traced back to near-vertical incidence. $SmS$ is also strong beyond 60 km. $Pg$ and $Sg$ are both weak beyond 50 km offset. $PmS$ (or $SmP$) is barely noticeable. The horizontal component supergather (Figure 4.4) displays a similar response, but with the shear wave arrivals ($Sg$ and $SmS$) enhanced relative to the compressional phases. Again, $PmS$ is barely noticeable. The above observations require that 1) both $P$ and $S$ wave energy must be created at or near the source, 2) the crust-mantle boundary
structure must not produce significant converted energy, and 3) precritical PmP and SmS must be weak relative to other phases.

A starting velocity model for the Basin and Range crust is displayed in Figure 4.11. The sedimentary cover consists of high and low velocity layers with a low Qp value of 50 and a Qs of 25. Qp is obtained from the spectral ratio analysis discussed above, and Qs is simply assumed to be half of Qp in the simulations. In the modeling to the RISC data, no constraint was found for Qt. The crystalline crust consists of the Precambrian/Mesozoic gneiss velocity distribution with Qp of 1000 and a Qs of 500. The crust-mantle boundary is a first-order discontinuity.

The synthetic seismograms for a shot at the northeast edge of this model are shown in Figures 4.12 and 4.13. The reverberatory nature of PmP and SmS is well produced in the simulations. However, some features do not coincide with the observed data. Both the vertical and horizontal components have a crustal refraction (Pg) which appears more reverberatory. No model was found which produced a satisfactory near-offset Pg phase. The effects of topography and a more complicated basin-basement interface are obviously important in modeling this phase, and are likely oversimplified in the models presented. In addition, the composite nature of the supergather makes any amplitude decay with offset analysis impossible over the whole gather. Also, the scattered field after the arrival of the surface wave is also poorly modeled. Again, no model was found which reproduced the complicated wavefield following the surface-wave arrival. This also is likely due to the lack of topography in the simulations and the smooth basin-basement interface.

Finally, unlike the observed data, a strong PmP phase is observed on the vertical component synthetic at precritical offsets, a phase which is only slightly identifiable on the supergather. This is also true for the PmS phase of the horizontal component. It is well known that by changing a step discontinuity to a thin gradient zone, pre-critical reflected energy is suppressed. To test this possibility, a simulation was run through a model with a 1 km thick gradient zone at the crust-mantle boundary. The results of this addition to the model can be seen in Figures 4.14 and 4.15. As expected, the precritical energy in both PmP and PmS is suppressed.

Before concluding that a lower crustal gradient layer must be present beneath the RISC profile, it is once again important to consider the effects of the nearsurface upon the identification of reflected phases. As noted above, the compli-
Figure 4.11  Basic velocity model for southern Basin and Range. The crust-mantle boundary is a first-order discontinuity.
Figure 4.12 Vertical component synthetic seismogram for the velocity model shown in Figure 4.11. Though the reverberatory nature of $PmP$ is well produced, $Pg$ is stronger than in the observed data. In addition, precritical $PmP$ is also stronger than observed.
Figure 4.13 Horizontal component synthetic seismogram for the velocity model shown in Figure 4.11. The strong $PmS$ phase is not observed in the recorded data.
Figure 4.14 Vertical component synthetic seismogram for a shot from the left edge of the velocity model shown in Figure 11, but with a 1 km thick transitional layer at the crust-mantle boundary. Pre-critical $PmP$ is now suppressed.
Figure 4.15  Horizontal component synthetic seismogram for the velocity model shown in Figure 11, but with a 1 km thick transitional layer at the crust-mantle boundary. The strong $PmS$ phase is now suppressed.
icated scattered wavefield which follows the surface waves in the observed data is poorly modeled in the simulations. This phase arrives before $PmP$ out to 30 km offset. The high-amplitude surface waves observed in the supergather might easily obscure the identification of precritical reflected energy, and thus suggest a gradient zone when none exists. This hypothesis is given more credence when the observed supergather is high-pass filtered to remove the surface wave energy. This gather is displayed in Figure 4.16. With the low-frequency surface waves totally removed, a nearly continuous $PmP$ phase is observed from near-offsets out to post-critical distances. This indicates that the presence of surface waves in low-frequency refraction data may influence the modeling of the crust-mantle boundary toward a gradient structure. The previous chapter also showed that a simple near-surface basin structure causes a first-order crust-mantle boundary to appear as a zone of short, discontinuous reflections. Therefore, although there is some evidence for a high gradient layer at the crust-mantle boundary (mainly the precritical $PmP$ amplitude at offsets greater than 30 km), the effects of near-surface structure and phases can influence this conclusion. In addition, the following chapter demonstrates that a rough crust-mantle interface has the same effect on $PmP$ amplitudes as a high gradient transition. Nonetheless, if this gradient layer does exist, it likely consists of thin, horizontally oriented bodies which reflect the shorter wavelengths (75-100 m) and not the longer wavelengths (1.5-2.0 km).

Coda analysis

Analysis of earthquake coda waves is often used to determine information about the scattering properties of the crust and mantle (Aki and Chouet, 1975; Langston, 1989) [1, 74]. In these studies, coda is defined as the tail of a seismogram after the arrival of $P$, $S$ and surface waves. In this study, I define coda as the less coherent energy in a seismogram after and between prominent coherent phases. With this definition, $PmP$ coda is the energy between $PmP$ and $Sg$ at wide angles. Analysis of such coda waves in deep crustal seismic data is complicated, due to the fact that $S$ waves and surface waves have not yet arrived, and the time window between coherent phases where coda can be analyzed is usually < 10 s. However, because the scattered field forms such a large part of the seismic section, an investigation into the source of such coda (basin scattering or scattering
Figure 4.16  Vertical component supergather from the southern Basin and Range with a high-pass filter > 16 Hz applied. A continuous $P_mP$ phase is now observed from vertical incidence to wide-angles, suggesting that the crust-mantle gradient zone is composed of thin interlayered bodies.
within the crystalline crust) is an important step in understanding the effects of wavelength-scale heterogeneity on the deep crustal wavefield.

In the Basin and Range supergathers, three different coda decay rates are observed (Figures 4.17 and 4.18). These particular curves are for an offset range of 90-95 km for Shotpoint 21. The first distinct coda (Coda A) slope follows directly behind $PmP$ and has a relatively steep decay rate (in dB/s). The second (Coda B) has a flat decay rate and lasts until the arrival of $Sg$. The third (Coda C) follows directly behind $SmS$. Insufficient recording time prevents later coda decays from being measured. Figures 4.19 and 4.20 display the amplitude decay curves from the synthetic seismograms in Figures 4.2 and 4.4. These curves are strikingly similar to the decay curves of the supergathers, indicating that the characteristics of the velocity field which created the recorded data is present in the stochastic velocity model. An explanation for these three coda decays is the goal of the next two sections.

**Decay rates with offset.** It is important to be able to separate the effects of scattering within the near-surface basin structure from scattering within the crystalline crust, which is the target of crustal studies. To do this, amplitude decay rates of the initial $PmP$ coda were measured at incremental offsets for each northern shot. Because of the variable basin thickness within an individual recording spread, the coda decay rate should change from shot to shot and with position within the recording spread if the coda is dominated by basin reverberation and energy diffusion out of the basin. On the other hand, if the decay rate does not change with offset, or changes with offset regardless of the shot or the position within the recording spread, then the coda is probably due to path effects within the crystalline crust.

The initial $PmP$ coda is observed for 1-4 s and decays as an exponential function with time in the form $Ce^{-kt}$, where $t$ is time, $k$ is the decay constant, and $C$ is an amplitude constant. To measure these decay rates $k$, the wide-angle shot gathers were time shifted with a linear moveout correction for the wide-angle shots which recorded a wide-angle $PmP$ phase at offsets greater than 40 km, amplitude decay curves were calculated summed over 5 km groupings. Before summation, the traces were time shifted with a linear moveout correction which flattened the $PmP$ reflection. This linear correction was adjusted for each offset range, so that the onset of $PmP$ over the 5 km aperture was even. Figure 4.21 is a plot of the exponential constant (the $k$ value of $e^{-kt}$) versus offset for each shot. The larger the
Figure 4.17 Decay curve for Shotpoint 21 $PmP$. 

**Observed data from RISC experiment, southern Basin and Range**
Figure 4.18 Decay curve for Shotpoint 21 SmS.
Figure 4.19  Decay curve for $PmP$ for Basin and Range supergather simulation.
Figure 4.20  Decay curve for $S_mS$ for Basin and Range supergather simulation.
\( k \) value, the steeper the decay. Two features are immediately obvious in the plot. First, the decay rates are different for each shot. Second, regardless of the shot, the pattern of decay rate over the recording spread is consistent. The higher rates of decay correspond to the thinnest basin cover in the Chocolate Mountains, and the slower decay rates correspond to the thicker sedimentary cover on the flanks of the Chocolate Mountains. This suggests that the thickness of the sedimentary cover, at both the shot and receiver, controls the decay rate of the initial \( PmP \) coda. To test this idea further, a simulation was run on the 2-D deterministic velocity model alone, without the presence of wavelength-scale heterogeneity. The amplitude decay curve for this simulation (Figure 4.22) at 90 km offset does not flatten out like those from the stochastic models (i.e., only the initial \( PmP \) coda is present). The steep decay of this \( PmP \) coda is due to diffusion of energy out of the basin as the energy reverberates within the low-velocity surficial layer.

\textit{Frequency-wavenumber analysis.} To further constrain the cause of \( PmP \) coda, the wide-angle shot gathers and the synthetics were transformed into the frequency-wavenumber domain. Such an analysis has been used in the study of scattering coda for teleseismic waves to identify energy arriving at varying velocities. Using this technique, Levander and Hill [1984] [78] identified the presence of Rayleigh waves in the coda of synthetic seismograms. These surface waves were created at the rough basement interface. Wagner and Langston [1992] [117], also using synthetic data, showed that the apparent velocity of arrivals within coda waves can be used to differentiate the sources of such energy.

The frequency-wavenumber spectra of the initial \( PmP \) coda A is plotted in Figure 4.23. It is readily apparent that the coda energy is dominated by phases with apparent velocities > 4 km/s. With the \( P \) wave sedimentary velocity averaging 3.1 km/s, and Rayleigh and \( S \) wave sedimentary velocity < 2.0 km/s, it is clear that basin scattering or body-to-surface wave mode conversion is not a prominent component of the initial \( PmP \) coda waves. Instead, as suggested in the previous section, the initial \( PmP \) coda is the result of the \( PmP \) reflection reverberating within the low velocity basin layer, and slowly diffusing back into the crust. This agrees with the results of Wagner and Langston [1992] [117], who showed that coda waves of energy travelling through an extremely anisotropic media (in this case, the basin layer itself) are a result of layer reverberation. Figure 4.24 displays the frequency wavenumber of the initial \( PmP \) coda of the finite-difference simulation shown in Figure 4.2. Again, the bulk of the energy arrives at velocities well above
Figure 4.21  Exponentials of initial $PmP$ coda decay for individual shots. Basin thickness thins with offset for each shot. The decay rate for each shot increases with decreasing basin thickness. Diffusion out of the basin is responsible for this initial coda.
Figure 4.22 Amplitude decay curve for 2-D, homogeneous model at 90 km offset. Only the initial $PmP$ coda is present. This decay is due to diffusion of energy as the waves reverberate within the low-velocity surficial layer.
those of the sedimentary fill, indicating that basin scattering and surface wave conversion are not important. This suggests that the inclusion of surface topography is not crucial in modeling wide-angle seismic data, a simplification which is often of concern in the analysis of scattered waves.

The frequency-wavenumber spectra of the flat $PmP$ coda B is plotted in Figure 4.25. The frequency-wavenumber spectrum for the flat $PmP$ coda in the synthetic seismogram in Figure 4.2 is displayed in Figure 4.26. The stochastic, wavelength-scale heterogeneity determined from the geologic outcrop satisfactorily predicts the characteristics of this coda as well. This coda has no measurable decay rate in this time window regardless of the shot or position within the recording spread, indicating that the basin structure does not influence this part of the $PmP$ coda. Like the initial $PmP$ coda, the flat coda is also composed of energy with apparent velocities greater than 4 km/s, eliminating basin scattering and surface wave conversions as the source of this coda. This coda is therefore due to either 1) near-surface planar reverberations and/or 2) scattering and $P$ to $S$ conversions from heterogeneities within the crystalline crust.

### 4.4.2 Southern shots

The wide-angle data from the southern shotpoints (Figures 4.3 and 4.5), those within the Salton Trough, are distinct from the northern shot records in a number of important ways. Because the crust-mantle boundary is much shallower than in the southern Basin and Range, the $PmP$ phase arrives much closer to the $Pg$ phase. More important, even though the $PmP$ phase is as strong, if not stronger than that of the northern shots, little or no $SmS$ is observed. Understanding the cause of the lack of shear wave energy is the main focus of this section.

There are two endmember explanations for the lack of shear wave phases in the Salton Trough shot records. Either the shear waves are preferably attenuated within the thick sedimentary cover in the Salton Trough, or shear waves are not created at the source. Both explanations have some intuitive appeal. The first explanation, that of preferred attenuation, may exist within a thick (up to 14 km) sedimentary column with active hydrothermal activity. On the other hand, it is well known that strong $S$ wave energy in marine seismic data is created at the water bottom, and hence the lack of an abrupt basin-basement interface may prevent the
Figure 4.23  Frequency-wavenumber spectrum of initial \( PmP \) coda in southern Basin and Range supergather.
Figure 4.24 Frequency-wavenumber spectrum of initial $PmP$ coda in synthetic seismogram from preferred southern Basin and Range stochastic velocity model.
Figure 4.25  Frequency-wavenumber spectrum of flat $PmP$ coda in southern Basin and Range supergather.
Figure 4.26 Frequency-wavenumber spectrum of flat $PmP$ coda in synthetic seismogram from preferred southern Basin and Range stochastic velocity model.
conversion of the $P$ wave source, and hence the lack of an abrupt basin-basement interface may prevent the conversion of the $P$-wave source.

One important observation from the Salton Trough supergathers (Figure 4.3 and 4.5) suggests that the lack of $SmS$ energy in the observed data is due to insufficient production of shear wave energy at the source. Only Shotpoint 100, 104 and 105 produce significant surface waves and/or $Sg$ arrivals. From the map in Figure 2.2, these shots lie at the edges of the Salton Trough. Shotpoints 104 and 105 are responsible for the strong Surface wave and $Sg$ arrival in the supergather at offsets less than 40 km. Shotpoint 100 produces the relatively strong $SmS$ arrival at 100-120 km offset. The shotpoints within the center of the Salton Trough (Shotpoints 101, 102 and 103) produce only a $P$-wave response. The obvious conclusion is that the presence of a basement contact below a shot is necessary to produce a $P$ to $S$ conversion near the source. Analysis of the Basin and Range supergather synthetics has already shown that such a contact can produce large converted shear waves.

### 4.5 Discussion and conclusions

Velocity models which include both large and small-scale structures have been used to create synthetic seismograms which reproduce the main phases and character of the recorded data from the RISC wide-angle profiles. The effect of wavelength-scale heterogeneity on both the specular phases and the scattered field was analyzed. With respect to the specular phases, it was necessary to introduce a high gradient zone at the the base of the crust to suppress the PmS converted phase and pre-critical $PmP$ phase. However, it was also noted that high-pass filtering the data creates a $PmP$ phase which is relatively continuous at all offsets. This indicates that near-source surface waves may contribute to the lack of precritical $PmP$ in narrow-band refraction data (5-10 Hz). On the other hand, in the following chapter, it will be shown that a rough crust-mantle boundary has the same effect as a gradient layer. It is therefore possible that a realistic rough interface could be produced which acts as a gradient layer on long wavelengths and a back-scatterer on shorter wavelengths.

Analysis of the coda waves which follow $PmP$ and $SmS$ (Coda A) shows that an initial (1-4 s) steeply decaying coda is due to reverberation and diffusion of energy out of the near-surface basin layer. The decay rates of this coda are depen-
dent upon the thickness of the basins at the source and receiver locations, with the thicker basins producing a slower decay rate. Frequency-wavenumber spectra of this coda contain little energy at slow velocities, indicating that conversion of incident body waves to surface waves is not important at long offsets. A reverberatory $PmP$ phase is often cited as evidence for a layered lower crust (e.g. Sandmeier and Wenzel, 1986) [102]. The above results, however, indicate that the initial reverberation is more likely due to a combination of energy diffusion of $PmP$ out of the basin layer and wavelength-scale scattering along the entire crustal path length, both sedimentary and crystalline. The coda which lies between the initial $PmP$ coda and the $Sg/SmS$ phases (Coda B) also contains little slow energy, and is relatively flat. These characteristics are consistent with coda generated as scattered waves and $P$ to $S$ body-wave conversions from wavelength-scale heterogeneities within the crust.

The wide-angle shots within the Salton Trough are unique because of the lack of shear waves recorded. Only the shotpoints near the edges of the basin produce significant shear waves. This suggests that the presence of a basin/basement contact near a shot is important in producing near-source converted energy like that observed in the Basin and Range shots. Shotpoints 104 and 105 lie just northeast of the projected Sand Hills fault, a relict transform within the Salton Trough, indicating that this fault may represent a major bounding structure for the basin.

The effects of the near-surface play an important role in the creating the complicated seismograms modeled in this chapter. Better knowledge of the details of the near-surface layer, including bassement topography and velocity structure, will aid in differentiating the effects of near-surface heterogeneity from crustal heterogeneity. Inclusion of surface topography in the finite-difference scheme should create more realistic surface waves and help address the effects of surface waves on the identification of near-offset $PmP$. 
Chapter 5

Small-Scale Crustal Structures of the Northern Basin and Range from the PASSCAL 1986 Seismic Experiment

5.1 Introduction

Possibly the most geophysically studied region within the Basin and Range province is the area around the Carson Sink and Dixie Valley in northern Nevada. It has been the site of numerous refraction and reflection surveys from 1963 to the present as well as the site of a passive seismic array for earthquake recordings. It has been the locus of recent seismicity. All of this complimentary and multi-scale seismic data provide an excellent opportunity to study the effects of wavelength-scale heterogeneity on the seismic wavefield. I have developed models which reproduce the recorded reflected and refracted wavefields from this region, and have tested the important parameters within the models to provide constraints on the velocity field at depth. In particular, I am interested in developing a single model which can explain the higher-frequency reflective character beneath the PASSCAL 1986 array as well as the traveltime and amplitude characteristics of the lower-frequency wide-angle wavefield. As an aside, I also discuss how large-scale structural anisotropy can explain discrepancies in published velocity models of the northern Basin and Range.

Before presenting the velocity models, however, I first briefly review the tectonic history of western Nevada and in particular the region around the Carson Sink and Dixie Valley. As in Chapter 2, I pay special attention to the geologic events which are most likely to have produced features which would be imaged seismically, such as the presence of both felsic and mafic rocks within the crust. I then review the geophysical data upon which the models are based.
5.2 Tectonic and magmatic history

Like all parts of the Basin and Range Province, the region surrounding the Carson Sink and Dixie Valley in northwestern Nevada has experienced a complicated geological evolution ranging from subduction related magmatism to active continental rifting (Speed et al., 1988) [108]. The oldest rocks exposed in western Nevada are of island arc affinities and were accreted onto North America during the Devonian-Mississippian Antler and the Permian-Triassic Sonoman orogenies (Speed, 1977; Speed, 1979; Speed and Sleep, 1982) [106, 107, 109]. For the northern Dixie Valley region, the basement is comprised of Late Triassic to Early Jurassic mudstones, sandstones and carbonates deposited in a marginal basin (Speed, 1976) [105]. Beginning in the Early Jurassic, an Andean-type margin formed along the western U. S. margin, resulting in extensive arc-related magmatism as far east as Montana (Dickinson, 1977; Coney, 1978) [22, 17]. In the Stillwater Range, which separates the Carson Sink from Dixie Valley, this magmatism is expressed as the Humboldt lopolith which is composed of gabbros, anorthosites, and diorites (Speed, 1977) [106].

The distinctive basin and range topography which characterizes this area today has formed from the Oligocene to the present. The Carson Sink-Dixie Valley area lies along the N-S trending Nevada Seismic Belt, which has been the site of many large historic earthquakes (Ryall et al., 1966; Doser, 1986) [101, 25], including the magnitude 7.1 Fairview Peak and magnitude 6.8 Dixie Valley earthquakes of December 16, 1954 (Tocher, 1957) [113], both on normal faults. Estimates of the total amount of extension within the northern Basin and Range vary, and range from 20% in Dixie Valley based on fault offsets and gravity modeling (Okaya and Thompson, 1985) [92] to 200-300% within the highly extended terranes to the east (Gans, 1987; Speed et al., 1988) [35, 108]. Extension varies temporally as well as spatially. Geodetic measurements indicate a recent rate of extension for the Nevada Seismic Belt at \( \geq 2.5 \text{ mm/yr} \), whereas geologically determined rates range from 0.38-1.0 mm/yr (Thompson and Burke, 1973; Okaya and Thompson, 1985) [111, 92].

The Basin and Range is topographically high (roughly 1.4 km) on average (Eaton, 1982) [26], and regional Bouguer gravity is inversely correlated with topography, indicating that regional topography is compensated at depth (Eaton et al., 1978) [27]. Within this regionally high province, the Carson sink-Dixie
Valley area is at the center of a regional topographic low, known as the Lahontan depression (Thompson, 1985) [110]. Reduced heat flow throughout the Basin and Range province is significantly higher than that for stable cratons, with the Carson Sink-Dixie Valley area on the 2.5 HFU contour of the Battle Mountain High (Lachenbruch and Sass, 1978) [72].

A considerable amount of magmatic material was added to the crust during Basin and Range extension (Anderson, 1989) [3], with the most recent activity consisting of bimodal compositions (Christiansen and McKee, 1978) [14]. The youngest volcanic rocks in the Carson Sink-Dixie Valley area are the 8 Ma rhyolites and basalts within Dixie Valley (Hastings, 1979) [46]. However, the recent activity along the Nevada Seismic Belt has led some workers to speculate that magmatic intrusion is occurring at depth even today (Eddington et al., 1987; Zoback, 1989) [29, 122]. In addition, data from the PASSCAL 1986 seismic experiment are believed to have identified a small in situ magma body at the base of the crust (Jarchow et al., 1993) [69].

5.3 Previous seismic models

The earliest controlled-source crustal seismic data collected within northwestern Nevada was a refraction line between Fallon and Eureka, NV, recorded by the United States Geological Survey in the early 1960's (Eaton, 1963) [28]. The Fallon-Eureka refraction profile was first interpreted by Eaton [1963]. Two alternate models were derived from the data, the first consisting of a uniform 6.0 km/s crust, the second including a 6.6 km/s lower crust (Figure 5.1a). Crustal thickness for the first model was determined to be 23 km beneath Fallon and 33 km beneath Eureka. The second model increased the crustal thickness by 2-3 km. Both of these models overpredict the observed Bouguer gravity anomaly, when these velocities are converted to densities. Prodehl [1979] [96] reinterpreted this data using 1-D techniques which allow for only velocity gradients (i.e. no first-order discontinuities), and obtained a model which does match the observed Bouguer gravity. This model consisted of crustal thicknesses of 30 km beneath Fallon and 35.5 km beneath Eureka, and a crustal velocity which ranged from 4.0 km/s to 6.8 km/s (Figure 5.1b). The base of the crust was modeled as two high gradient layers, 6.8-7.0 km/s and 7.0-7.9 km/s.
Figure 5.1  Velocity models determined from the Fallon-Eureka refraction line. Note lack of high velocity layer at the base of the crust. a) Models proposed by Eaton [1963]. b) Model proposed by Prodehl [1979].
Between 1982 and 1984, the Consortium for Continental Reflection Profiling (COCORP) conducted a $>1000$ km long seismic reflection transect across the entire Basin and Range province at about $40^\circ$ deg N (Figure 5.2) (Allmendinger et al., 1987) [2], using a Vibroseis source and a 9.8 km receiver, 96-channel receiver array. Line 1 and Line 2 of this transect are of particular interest, as they coincide with the 1986 PASSCAL northern Nevada reflection and refraction survey discussed below. Two important observations can be made from the COCORP data which have influenced the development of geologic models of the Basin and Range province. The first is that, despite the striking topographic feature of alternating mountain ranges and valleys and the presence of rocks from 10 km depths now at the surface, the Moho remains relatively flat, varying no more than 4 km across the entire province (Klemperer et al., 1986; Allmendinger et al., 1987) [71, 2]. Second, a transparent upper crust above a more reflective middle crust is common, especially in the west (Hauge et al., 1987) [47].

In 1986, two seismic refraction profiles were recorded in northern Nevada as part of the PASSCAL northern Nevada seismic experiment. The two lines were perpendicular to one another, one parallel and one perpendicular to the predominant strike of the basins and ranges (Figure 5.2). The north-south line (NS) extends 200 km from Fallon, NV, to Winnemucca, NV, and runs mainly along the edge of the Carson Sink and the Buena Vista Valley. The east-west line (EW) extends 280 km from Gerlach, NV, to Austin, NV, crossing nine mountain ranges. These refraction lines provided the first modern refraction data for this region. Instrument spacing was approximately 1 km. Shot spacing was roughly 50 km, except in the center of the EW line as part of the “piggyback” reflection spread. Figures 5.3-5.5 show three separate velocity models of this dataset based upon analyses of traveltimes and amplitudes of interpreted crustal phases.

These models are in general agreement on average crustal velocity and crustal thickness. They also agree on the presence of a first-order velocity discontinuity at approximately 20 km depth. In detail, however, the models differ due to the interpretational nature of phase identification and modeling techniques. In particular, Catchings and Mooney [1991] [13] model a high velocity layer above the crust mantle boundary, Holbrook [1990] [57] interprets no such layer, and Benz et al. [1990] [5] model instead a low velocity upper mantle layer. An explanation for these differing interpretations is given later in this chapter, as well as a hypothesized starting model which may help explain these discrepancies.
Figure 5.2 Basemap of northern Nevada region showing COCORP and PASSCAL seismic lines. Dotted lines indicate the PASSCAL refraction experiment receiver spread. Stars indicate location shotpoint locations. Dashed lines indicate location of COCORP reflection survey. The solid line delineates where the PASSCAL and COCORP lines coincide.
Figure 5.3 Published velocity models from the 1986 PASSCAL northern Nevada refraction experiment based on 2-D raytracing (from Catchings and Mooney [1991]). Note the presence of the 7.4 km/s layer at the base of the crust. a) North-south line. b) East-west line.
Figure 5.4 Published velocity models from the 1986 PASSCAL northern Nevada refraction experiment based on 2-D raytracing (from Holbrook [1990]). Note the lack of a high velocity layer at the base of the crust. a) North-south line. b) East-west line.
Figure 5.5  Published velocity models from the 1986 PASSCAL northern Nevada refraction experiment based on 1-D reflectivity (Benz [1991]).
5.4 Data to be simulated

An example of one of the refraction shot records (shot point 4) is displayed in Figure 5.6. The source was 2000 lbs of chemical explosives, recorded by 120 seismographs with 2 Hz seismic sensors at an average spacing of 1 km and a maximum offset range of 145 km. In addition to the first break $P_g$ phase, a prominent $PmP$ phase is observed at wide angles between 60 and 120 km, with little or no precritical $PmP$. Numerous other smaller amplitude phases have also been interpreted as velocity discontinuities within the crust by previous workers (Holbrook, 1990; Catchings and Mooney, 1991) [57, 13]. However, as mentioned above, the differences between these models indicates that the modeling of these smaller crustal phases is subjective.

At the intersection of these refraction lines, a piggyback reflection spread recorded the large refraction shots. The reflection spread consisted of 8 and 10 Hz geophones with a group spacing of 60 m, and thus recorded a higher resolution near-offset image of the crust. A high-quality characteristic shot gather (Shotpoint 4B) is displayed in Figure 5.7. The onset of reflectivity at 4 s below a transparent upper crust is readily observed. In addition, a bright set of reflections interpreted as the reflection Moho occurs at 9-10 s. Between these two zones is a less reflective lower crust, a feature previously downplayed in previous interpretations of the COCORP datasets.

The refraction and reflection records displayed in Figures 5.6 and 5.7 represent the data to be simulated in this chapter. Once again, the goal is to produce a single model which will reproduce both the reflected and refracted wavefields.

5.5 Stochastic velocity models from the PASSCAL experiment

Lacking other information, I used the same upper crustal model as used to model the RISC data. The basin structure was obtained by digitizing the basin geometry obtained from forward modeling of the first-break traveltimes of the refraction data (Holbrook, 1990; Catchings and Mooney, 1991) [57, 13]. While not a detailed image of the basin structure, this model provides the gross basement structure under basins and ranges and creates the undulatory travelttime anomalies characteristic of the observed data. Thin layers with 10:1 aspect ratios and alternating high and
Figure 5.6  Refraction shot record for shot point 4 from the PASSCAL 1986 northern Nevada experiment.
Figure 5.7  Reflection shot record for shot point 4B from the PASSCAL 1986 northern Nevada experiment.
low velocities fill the basins. As in the RISC model, \( Q_p \) within the basins is held fixed at 50 (\( Q_s = 25 \)).

The deterministic model for the northern Basin and Range is obtained by averaging the various velocity-depth functions which have been obtained by previous workers (Holbrook, 1990; Catchings and Mooney, 1991; Benz et al., 1990; Hawman et al., 1990; Zelt and Smith, 1992) [57, 13, 5, 50, 121]. A graph of a few of these functions is shown in Figure 5.8. Despite some differences within the models resulting from phase interpretation and 2-D structure, the bulk velocity function is relatively consistent. This function has a linear velocity gradient of 0.02 s\(^{-1}\) to \( \sim 20 \) km depth, which then grades to 6.8 km/s at the base of the crust. The presence and/or size of the velocity step at 20 km varies. Also, the high velocity layer (> 7.2 km/s) at the base of the crust is not present in all of the models and will be part of the subject of inquiry in the modeling below. Within 5 km of the surface, I model a steeper velocity gradient of 0.25 s\(^{-1}\) corresponding to the crustal depths where cracks have not yet closed under lithostatic pressure.

Detailed pertophysical maps and laboratory measurements were not available within the local basement exposures. However, like the Chocolate Mountains in southeastern California, the rocks exposed at the surface consist of an older gneissic basement intruded by Mesozoic arc-related magmas. Without further information on the rock distributions within the basement, I have chosen to model the upper crust with the same statistics as those obtained from the Chocolate Mountains described in Chapter 3. This juxtaposition of roughly isotropic felsic and intermediate bodies produces a transparent upper crust characteristic of the northern Basin and Range.

The high-amplitude reflectivity observed at midcrustal levels suggests higher impedances and/or horizontal orientations within the midcrust. This reflectivity is modeled as horizontally oriented mafic dikes within a more felsic matrix. The statistics describing these bodies are similar to those used for the mafic intrusions in the RISC data and are listed in Table 5.1. In Chapter 3, it was shown that a relatively small volume (\( \geq 5\% \)) of properly oriented impedances can produce a highly reflective wavefield. For the PASSCAL data, an upper limit can be placed on the volume of intrusion by traveltime modeling of \( P_g \). Holbrook [1990] [57] suggests that approximately 10% basaltic intrusion could be responsible for the reflective midcrust and be undetectable by refraction methods. Given the bulk velocities within the model, and assuming that bulk velocity differences of less
PASSCAL Basin & Range 86
Refraction Velocity

Figure 5.8 Characteristic velocity-depth functions obtained from analysis of the PASSCAL refraction data.
than 0.1 km/s would be undetectable at these depths, I estimate that between 5% and 25% intrusion could be present. Lack of knowledge of the actual upper crustal scattering field prevents further constraining these limits. In the following simulations, I use the upper extreme of 25% intrusion.

### Table 5.1 Covariance function statistics used to create spatial fabric in the viscoelastic finite-difference simulations.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Model</th>
<th>$D$</th>
<th>$a_x$ (m)</th>
<th>$a_z$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basins</td>
<td>None</td>
<td>2.5</td>
<td>200</td>
<td>20</td>
</tr>
<tr>
<td>Upper crust</td>
<td>Pc/Mz gneiss</td>
<td>2.5</td>
<td>817</td>
<td>963</td>
</tr>
<tr>
<td>Mid crust</td>
<td>Ivrea zone</td>
<td>2.7</td>
<td>240</td>
<td>60</td>
</tr>
</tbody>
</table>

Below 7 s two-way-time, or roughly 2 km depth, the reflectivity beneath shot point 4B decreases in amplitude. Two endmember models can explain this amplitude decrease: 1) a change in fabric orientation from predominantly horizontal to vertical, or 2) a decrease in impedance contrast between the different rock types.

The first model was suggested by Holliger and Levander [1994] [60] to explain similar features in other collapsed Phanerozoic orogenic belts. They explain the change in orientation by suggesting that horizontal intrusions would be preferred in rheologically weak zones. Such a model for the northern Basin and Range is displayed in Figure 5.9. In this model, no bulk changes in velocity are required to produce the reflectivity observed in the data. Such a change in orientation requires, according to the Holliger and Levander [1994] [60] model, a two-layer crustal rheology, such as that shown in Figure 5.10. A quartz-dominated rheology above 20 km depth produces a midcrustal zone of horizontal intrusions at the base of this layer (10-20 km). Above this depth, either mafic bodies are oriented vertically, or a level of neutral buoyancy is achieved at 10 km depth for the mafic intrusives. Below this depth, a stronger rheology (diabase or plagioclase) would promote the formation of dikes in an extensional environment, until such depths, presumably just above the crust-mantle boundary, where even this stronger layer
becomes weak. Figure 5.11 displays the finite-difference simulation from the model in Figure 5.9. Note the less reflective lower crust between the more reflective midcrust and Moho.

While the velocity model in Figure 5.9 does not require a velocity step at \( \sim 20 \) km, the requirement of a two-layer rheology would suggest the presence of one. In addition, most of the published velocity models have identified a velocity discontinuity at \( \sim 20 \) km depth in this region. If we assume that the mid-crustal reflections are due to the presence of horizontal mafic intrusives, and that these intrusives arise from the mantle and are evenly distributed between 10 and 30 km depth, then an increase in the background crustal velocity between 20 and 30 km depth would decrease the impedance contrasts in the lower crust and thus decrease the reflectivity. This decrease would take place regardless of the orientation of the intrusives. If no change in orientation is present, the horizontal orientations would be buoyancy controlled, with the time-scale of the emplacement of the intrusives much smaller than the time constant for lower crustal rheologies. This model was proposed by Holbrook [1990] [57]. Figure 5.12 displays such a velocity model, and the corresponding synthetic is shown in Figure 5.13. The decrease in reflectivity between 7 and 9 s is readily apparent.

Figures 5.11 and 5.13 reproduce the near-offset reflection data reasonably well. However, such a simple crust-mantle interface does not predict the wide-angle response. Figure 5.14 displays a wide-angle simulation from the same model as shown in Figure 5.12. Note the strong precritical \( PmP \) energy between 0 and 60 km offset, a feature not observed in the recorded data. The traditional solution to this problem, and that used by Benz et al. [1990] [5], is to insert a transitional high-gradient layer at the crust-mantle boundary. This high-gradient layer focuses reflected energy at the critical distance. Benz et al. [1990] [5] found the details of this layer difficult to constrain, but did require its presence in some form to fit the amplitude characteristics of \( PmP \). Figure 5.15 displays a wide-angle simulation with a 4-km-thick transitional layer between 26 and 30 km depth. As expected, precritical \( PmP \) is attenuated.

Before concluding that a gradient layer must exist at the base of the crust, however, I present an alternative model of a rough crust-mantle boundary. While investigating the cause of \( P \)-wave coda from regional earthquakes, Dainty and Schultz [1995] [20] calculated the reflection coefficients for reflected plane waves at varying incident angles for a range of scattering angles from a rough crust-
Figure 5.9  Sills and dikes model for northern Basin and Range.
Figure 5.10  Strength curve for two-layer crustal rheology.
Figure 5.11  Finite-difference synthetic seismogram for the sills and dike model for the northern Basin and Range.
Figure 5.12  Velocity model for northern Basin and Range with a higher velocity lower crust.
Figure 5.13 Finite-difference synthetic seismogram for the higher velocity lower crustal model for the northern Basin and Range.
Figure 5.14  Wide-angle synthetic seismogram for the higher velocity lower crustal model for the northern Basin and Range. Note the strong precritical $PmP$ energy.
Figure 5.15  Wide-angle synthetic seismogram for a model with a 4 km thick high gradient transitional layer at the crust-mantle boundary. Note the lack of precritical $PmP$ energy.
mantle boundary. They tested two model interfaces: one with an rms slope of 10 degrees, and one with an rms slope of 30 degrees. The results are displayed in Figure 5.16. As the roughness of the interface increases, the vertical-incidence response decreases, while the energy scattering at the critical angle increases. This produces essentially the same effect as a high-gradient transitional layer, indicating that a rough crust-mantle boundary might also explain the amplitude character of $PmP$.

Two rough Moho models have been studied, and are displayed in Figure 5.17. The first rough crust-mantle interface has a characteristic scale of 15 km, a maximum deviation from the mean of 2 km, and a fractal dimension of 1.0, resulting in an rms slope of 12 degrees. The second rough interface has a characteristic scale of 5 km, a maximum deviation from the mean of 2 km, and a fractal dimension of 1.0, resulting in an rms slope of 37 degrees. The smoother topography was designed to be similar to the basement topography of the near surface. Figure 5.18 displays basement topography plotted above the smoother (12 degree rms slope) crust-mantle interface. Basement topography was obtained from Hauge et al. [1987], and combines surface elevations within the ranges with basin depths obtained by analysis of COCORP reflection data. It is a deterministic feature. The resolvable structural detail is limited and variable, dependent on the data quality over the basins. Nevertheless, the two interfaces are quite similar in amplitude and dominant wavelength. The second, rougher (37 degree rms slope) model was used to test the potential impact of different levels of roughness on the reflected wavefield. A priori knowledge of the roughness of the crust-mantle boundary does not presently exist. The best exposure of possible crust-mantle boundaries, the Ivrea Zone in northern Italy, reveals only a few kilometers of this contact, which unfortunately is now a strike-slip fault related to the Alpine orogeny. Therefore, little evidence exists regarding the fine-scale structure of the crust-mantle boundary.

The wide-angle simulations for the rough Moho models are displayed in Figures 5.19 and 5.20. As predicted, precritical $PmP$ is suppressed, with the rougher interface removing more of the precritical $PmP$ energy. The higher-frequency reflection simulations from these models are shown in Figures 5.21 and 5.22. Note the similarity of the reflection Moho due to the horizontal ultramafic intrusions just above the crust-mantle boundary. This reflection signature is the same as that in Figure 5.13, which is from a model with a flat crust-mantle discontinuity. This underscores the ability of a small amount of heterogeneity to
Figure 5.16  Reflection coefficient versus scattering angle curves for plane waves of varying incident angle upon a rough interface (from Dainty and Schultz [1995]. The calculations assume that the characteristic wavelength (horizontal scale) of the interface is equal to the dominant wavelength of the incident wave. Roughness is altered by changing the deviation about the mean depth assuming Gaussian statistics. a) Results from an interface with a 10 degree rms slope. b) Results from an interface with a 30 degree rms slope. Note how the near-offset reflection coefficients decrease while the critical angle reflection coefficients increase with increasing roughness.
Figure 5.17  Velocity models with a rough crust-mantle boundary. a) A rough Moho interface with a characteristic scale of 15 km, a maximum deviation from the mean of 2 km, and a fractal dimension of 1.0, resulting in an rms slope of 12 degrees. b) A rough Moho interface with a characteristic scale of 5 km, a maximum deviation from the mean of 2 km, and a fractal dimension of 1.0, resulting in an rms slope of 37 degrees.
Figure 5.18 Comparison of basement topography to the rough Moho topography with a 12 degree rms slope. Basement topography was obtained from Hauge et al. [1987], and combines surface elevations within the ranges with basin depths obtained by analysis of COCORP reflection data. Note that the two interfaces are quite similar in amplitude and dominant wavelength.
obscure important details of structures below. To test whether these rough crust-mantle interfaces could by themselves produce the reflective Moho, the ultramafics were removed, and a reflection gather again calculated. These results are are displayed in Figures 5.23 and 5.24. Clearly these interfaces are unable to predict the high-amplitude Moho reflections in the observed data. In fact, despite the larger first-order velocity step and at the crust-mantle boundary (6.9 km/s to 8.0 km/s), the Moho is essentially indetectable. This suggests that in regions where a near-offset reflection Moho is not observed yet a wide-angle $PmP$ is present, a degree of roughness on the crust-mantle boundary may be indicated.

The above results confirm that a rough crust-mantle boundary can predict the observed low-frequency, wide-angle amplitude response of $PmP$. However, the rough interfaces alone could not also reproduce the observed high-amplitude Moho reflectivity. Nonetheless, it is intriguing to speculate that a rough interface could be developed which would produce both the high-frequency reflectivity at near offsets and the lower-frequency AVO response of $PmP$ at wide angles. Dainty and Schultz [1995] [20] mention that different levels of roughness would be required to satisfactorily scatter different wavelengths. Their rough crust-mantle topography was created assuming a Gaussian distribution about a mean depth and a characteristic wavelength equal to the dominant wavelength of the incident wave. With these assumptions, the roughness was determined only by the deviation from the mean, and this number can be increased until the observed amplitude effects (i.e., decrease in precritical $PmP$) are recreated. In the above simulations, the topography was defined using statistics based upon the von Kármán function, and hence has multi-scale roughness. The roughness was altered by adjusting the characteristic scale rather than the deviation from the mean, which is held constant. Assuming a lower crustal velocity of 7.0 km/s, the lower-crustal wavelengths in the above simulations range from 175 m in the reflection gathers to 3000 m in the refraction gathers. These wavelengths are less than the characteristic scales used to create the topography (5 and 15 km), and hence, due to the self-similar nature of a von Kármán distribution, should all experience a similar degree of roughness when interacting with the interface. This explains the ability of the above two models to scatter both high and low frequency waves. Further modeling should determine whether a single rough interface can indeed be defined which would predict both the near-offset and wide-angle wave fields. This interface may or may not conform to a von Kármán distribution.
Figure 5.19  Wide-angle synthetic seismogram for a model with a rough crust-mantle boundary, rms slope = 12 degrees. Note the lack of precritical $PmP$ energy.
Figure 5.20  Wide-angle synthetic seismogram for a model with a rough crust-mantle boundary, rms slope = 37 degrees. Note the lack of precritical $PmP$ energy.
Figure 5.21  Near-offset synthetic seismogram for a model with a 12 degree rms slope rough crust-mantle interface and ultra-mafic lower crustal intrusions to produce the characteristic reflection Moho.
Figure 5.22  Near-offset synthetic seismogram for a model with a 37 degree rms slope rough crust-mantle interface and ultra-mafic lower crustal intrusions to produce the characteristic reflection Moho.
Figure 5.23  Near-offset synthetic seismogram for a model with a 12 degree rms slope rough crust-mantle interface but no horizontally-oriented ultramafic bodies between 26 and 30 km. Note the lack of reflection Moho, despite the large impedance contrast from crust to mantle.
Figure 5.24  Near-offset synthetic seismogram for a model with a 37 degree rms slope rough crust-mantle interface but no horizontally-oriented ultramafic bodies between 26 and 30 km. Note the lack of reflection Moho, despite the large impedance contrast from crust to mantle.
As a final comparison between the various models of the crust-mantle boundary, amplitude versus offset (AVO) curves were calculated for the four models presented. These curves, along with the observed AVO curve from shot point 4, are displayed in Figure 5.25. The observed AVO curve from shot point 4 shows a great deal of scatter at all offsets. The critical distance appears to be somewhere between 80 and 100 km offset. The high amplitudes at offsets less than 50 km are likely due to surface wave contamination. These surface waves are observed at similar times as $PmP$ within this offset range in Figure 5.6. For the synthetics, the highest near-offset amplitudes are from the flat Moho model, as expected. The scatter in the measurements underscores the impact of crustal and near-surface heterogeneity on a simple interface. The amplitudes from the flat Moho can be up to twice that for the gradient Moho model. The rough Moho models lie between these two extremes. At wider angles, the 12 degree Moho model produces the most scatter, a result of focusing of reflected energy from the curved interface. The relative consistency of the 37 degree Moho curve indicates that as roughness increases, the interface focuses more energy toward the critical distance. This curve also resembles more closely the gradient Moho curve, with the highest amplitudes occurring just beyond 100 km offset.

The presence of a rough Moho interface, rather than a smooth gradient transitional layer, results in a strikingly different view of the interaction of the mantle and crust. The gradient model has traditionally been interpreted as thin, interlayered intrusives and magmatic cumulates which have over time, due to buoyancy, created an increase in seismic velocity with depth. This model represents a crust-mantle boundary as a chemical transitional zone. In the rough interface model, however, there is no transitional layer. There may be intrusives within the lower crust enhancing a reflective zone, but the crust and mantle are essentially distinct. Two endmember interpretations for the roughness can be made: tectonic or magmatic. In the tectonic interpretation, the roughness would be due to faulting of the upper mantle, similar to what has taken place at the surface. The lack of upper mantle earthquakes in the Basin and Range makes this view suspect, but it is worth mentioning as a possibility. The second interpretation, that of a magmatic interface, suggests that the upper mantle, or parts or it, has at times been mobile, and has risen upwards into the weaker lower crust. Regardless of the interpretation, the fact remains that a rough interface can also explain the amplitude characteristics of $PmP$. 
Figure 5.25 Amplitude versus offset (AVO) curves for $PmP$. a) Observed data from PASSCAL shot point 4 (east). Trace were normalized prior to amplitude measurements using a two second pre-first break time gate. Note the high degree of scatter along the curve. b) AVO curves for the four rough crust-mantle interfaces described in the text. At near offsets, the flat Moho curve has the highest amplitudes, and the gradient Moho the lowest. At wide angles, the 12 degree Moho produces the most scatter.
5.6 An alternative model for the lower crust

Before concluding this chapter, I take the rough crust-mantle boundary hypothesis one step further, and propose a simple 3-D model which may explain some of the discrepancies in the published velocity models alluded to earlier. A variety of crustal velocity models have been obtained for northern Nevada since Eaton’s first attempts in 1963. Most of these models are in general agreement with respect to upper and middle crustal velocity within reasonable error estimates and considering the variety of modeling techniques used. However, one feature in the models is conspicuously inconsistent: the presence, or lack thereof, of a high velocity layer (7.4–7.6 km/s) at the base of the crust. Determination of whether such a layer does exist has important ramifications for models of Basin and Range rifting as well as for the growth of continental crust. Does mafic underplating contribute a significant volume of material to the crust during rifting (Furlong and Fountain, 1986) [34]? Alternatively, are magmatic additions to the crust broadly distributed throughout the crustal column in the form of sills and/or dikes? I propose a simple starting model which predicts the traveltimes of the observed refraction data and also explains the reason for the inconsistencies between the published models.

First, let us review again the published models and the data which has inspired these models. A shot gather from the early 1960’s (East-West) Fallon-Eureka refraction experiment is displayed in Figure 5.26. Figure 5.1 displays the velocity models of Eaton [1963] [28] and Prodehl [1979] [96]. Eaton [1963] [28] and Prodehl [1979] [96] observed no high velocity layer from analysis of this refraction line. Wide-angle $P_{m}P$ indicates an apparent velocity of about 6.6 km/s. Thompson et al. [1989] [112] provide an alternative interpretation of the Fallon-Eureka line using 2-D raytracing. In this model, a 7.4 km/s layer is present at the base of the crust. This points to the non-uniqueness inherent in traveltime modeling of deep crustal phases. Nevertheless, the Prodehl [1979] [96] model indicates that it is possible to model the traveltimes of the major phases without the presence of a high velocity layer at the base of the crust.

The 1986 PASSCAL refraction dataset was designed to try and address the inconsistencies mentioned above. However, the results, based on the various published interpretations of the data, are just as inconsistent. Holbrook [1990] [57] interpreted the refraction data using 2-D forward raytracing. The resulting model is displayed in Figure 5.4. No high velocity layer is present in the lower crust.
Figure 5.26  Shot gather of the Fallon-Eureka refraction line. Note the low apparent velocity of wide-angle $PmP$. 
Catchings and Mooney [1991] [13], using the same raytracing technique, do model a high velocity layer at the base of the crust (Figure 5.3). This again points to the nonuniqueness inherent in travelt ime modeling lower crustal phases. Another analysis is presented by Benz et al. [1990] [5] using a 1-D reflectivity method. These models are shown in Figure 5.5. The models are simple 1-D velocity profiles with few if any midcrustal interfaces. Surprisingly, the north-south model shows a high velocity layer and the east-west model does not. Benz et al. [1990] [5] interpret this layer as low velocity upper mantle, rather than high velocity lower crust, but offer no explanation for its lack of existence in the east-west profile.

An explanation for how the inconsistencies arise becomes apparent when one looks at the shot records used in these interpretations. Figure 5.27 shows two reversing shot records from the north-south line. The wide-angle reflection which closely follows $Pn$ beyond 140 km offset has an asymptotic velocity of approximately 7.5 km/s. This explains the inference of the fast velocities at the base of the crust in some of the models. This phase led Catchings and Mooney [1991] [13] to model a lower crustal high velocity layer, but was modeled by Holbrook [1990] [57] as originating from structures within the mantle (Figure 5.4). Figure 5.28 shows two reversing shot gathers from the east-west line. In these gathers, there is no wide-angle reflection closely following $Pn$. Instead, a much slower (6.8 km/s) wide-angle reflection is observed at later times. This phase does not support the presence of high velocities at the base of the crust.

One important observation must be noted before a consistent model for the northern Basin and Range can be developed. Within all of the models discussed thus far, the evidence for a high velocity lower crust comes only from north-south oriented lines, or lines oriented roughly parallel to the strike of the basins and ranges. Otherwise, the data is satisfactorily modeled with crustal velocities less than 7.0 km/s. There are two explanations for this observation: 1) mineralogic anisotropy or 2) structural anisotropy.

By mineralogic anisotropy, I mean anisotropy induced by preferred orientation of minerals within a otherwise homogeneous rock type, such as that observed within the upper mantle beneath oceanic crust (Morris et al., 1969) [89]. In the case of northern Nevada, a 4-5 km thick layer within the lower crust or upper mantle with a seismic anisotropy of 5% is necessary. By structural anisotropy, I mean wavelength scale or greater bodies of high and low velocity material, such as felsic and mafic rock, with a consistent orientation. For northern Nevada, large
Figure 5.27  Shot records from the north-south line of the PASSCAL 1986 northern Nevada seismic experiment (from Benz et al. 1990).
Figure 5.28 Shot records from the east-west line of the PASSCAL 1986 northern Nevada seismic experiment (from Benz et al., 1990).
ultramafic bodies elongated parallel to the basins and ranges might produce a fast velocity in lines oriented parallel to the ranges, but also produce an intermediate velocity in lines perpendicular to the ranges.

If the upper mantle beneath northern Nevada is composed of peridotite, the first hypothesis, that of mineralogic anisotropy, could be accomplished by a layer of properly oriented olivine crystals, with the fast axis aligned parallel to the predominant spreading direction of the region. Laboratory measurements of rocks high in olivine commonly produce seismic anisotropies on the order of 5% (Kern and Richter, 1981) [70], sufficient to produce the observed anomaly. However, below this layer, anisotropy is not observed, with mantle velocities in both north-south and east-west directions consistently measured at 8.0 km/s (Holbrook, 1990; Catchings and Mooney, 1991; Benz et al., 1990) [57, 13, 5]. If the preferred orientation of olivine crystals is responsible for the anisotropy above, then the velocity below this layer should be the average of the fast and slow velocities, or 7.8 km/s.

To test the structural anisotropy model, we first need to show that the travel-times from the major phases of the Nevada PASSCAL 1986 data can be satisfactorily matched by a model consisting of a lower crustal layer whose velocity in the north-south direction is fast (7.6 km/s), and whose velocity in the east-west direction does not exceed 7.0 km/s. Rather than repick the phases of interest, which would require an interpretation of the data, I choose to match the travel-times calculated from the 1-D models published by Benz et al. [1990] [5]. These travel-times match the data, and avoid any controversy which may result from reinterpretation of the data. The 1-D velocity models of shotpoint 8 for the north-south line and shotpoint 1 for the east-west line were input into a 2-D raytracing program (Zelt and Smith, 1992) [121] and travel-times for main phases were calculated. These times were then used to develop two models, one with a 7.6 km/s lower crust and one with a 6.9 km/s lower crust. Effort was made to maintain a consistent midcrustal velocity in both models (6.4 km/s). The results are displayed in Figures 5.29 and 5.30. Except for minor near-surface differences in velocity (necessitated by $P_g$ curvature), the only velocity difference between the two models is in the lower crust. It is readily apparent from Figures 5.29 and 5.30 that this alternative model matches the calculated travel-times of the Benz et al. [1990] [5] models, and hence the main phases of the PASSCAL refraction dataset.

I will now show that a 3-D velocity structure can also produce such travel-times. While full-wavefield 3-D finite-difference modeling of crustal-scale models
Figure 5.29 Ray diagram and traveltime fits to the north-south model of Benz et al. [1990].
Figure 5.30 Ray diagram and traveltime fits to the east-west model of Benz et al. [1990].
is not now possible due to computer memory limitations, a simple analysis of the expected refracted traveltimes from such a model can be accomplished using 3-D finite-difference eikonal solvers. Two suites of models were run. Both involve a sinusoidally undulating high velocity lower crustal layer, but whereas the first suite of models (Figure 5.31) is characterized by a high velocity lower crustal layer which periodically pinches out completely, the second suite of models (Figure 5.31) merely thins and thickens periodically. The models were 200x200x36 km. Because the finite-difference traveltime solver calculates only first arrival times, the mantle velocity was set to a low velocity so that the lower crustal arrivals could be observed. The maximum thickness of the high-velocity layer was held constant at 4 km. The wavelength of the sinusoid (the half-width of the high-velocity bodies) was varied from 8 km to 128 km.

For each model, traveltimes were calculated at the surface along linear line arrays oriented parallel and perpendicular to the strike of the high-velocity bodies. Figure 5.33 displays the refracted traveltime curves for the model shown in Figure 5.31. At offsets less than 130 km, both lines (parallel and perpendicular) have identical traveltimes because the first-arriving rays have not sampled to the depths of the high velocity bodies. Beyond 130 km, the traveltime curves diverge, with the line oriented parallel to the strike of the bodies arriving with a moveout velocity of 7.6 km/s, and the line oriented perpendicular to the bodies arriving later at a moveout velocity of 7.0 km/s. Thus, one line measures the fast velocity within the lower crust, and the other measures the average velocity of the lower crust, as is observed in the Nevada PASSCAL refraction data.

Figure 5.34 displays the traveltime curves for the model shown in Figure 5.32. In this case, because the high velocity bodies do not pinch out for a significant distance, both traveltime curves have moveout velocities of 7.6 km/s, with the curve for the perpendicular line delayed slightly. This is an important result, for it suggests that if elongate bodies are responsible for the observed anisotropy, these bodies must pinch out on the scale of the wavelengths of interest. Further work involving the relationship of seismic velocity to body size and wavelength will help constrain the actual body sizes necessary for this model to hold. In these examples, a 50-50% fast and slow rock velocity was assumed, but is not necessary. Less fast velocity bodies would merely result in a slower observed average velocity. This is likely the case in the northern Basin and Range, where the 6.8 km/s lower crustal velocity measured in the east-west direction would require a 6.0 km/s background
Figure 5.31 3D velocity model for a structurally anisotropic lower crust. Model used for simulations extended to 200 km in both directions, and the shot and receiver arrays were correspondingly centered.
Figure 5.32 3D velocity model for a structurally anisotropic lower crust which does not produce an observed anisotropy because the bodies do not pinch out at their base for infinite frequencies. Model used for simulations extended to 200 km in both directions, and the shot and receiver arrays were correspondingly centered.
velocity in the lower crust, which is probably unreasonably small. In the Benz et al. [1990] [5] velocity models, a velocity of 6.4 km/s exists at depths of 22 km. If this is taken as the background velocity, then about 30% of the lower crust would be in the form of elongate ultramafic bodies.

5.7 Summary and conclusions

Combined deterministic and stochastic velocity models for the northern Basin and Range have been presented which reproduce the observed reflected and refracted wavefields reasonably well. The near-offset reflection signature is dominated by the wavelength-scale impedance structure. The midcrustal reflectivity between 4 and 7 s is produced by 5-25% horizontally-oriented, wavelength-scale mafic intrusions. The most consistent explanation for the less reflective lower crust involves an increase in bulk velocity at ~ 20 km depth, which may result in a decrease in the impedance contrasts within the zone. The reflection Moho is also modeled as arising from horizontally-oriented bodies just above the crust-mantle boundary. The wide-angle refraction data is dominated by the kinematics of the average velocity structure and its first order features, such as the Moho. The refraction data require either a high-gradient transitional layer at the crust-mantle boundary and/or a crust-mantle boundary with topography.

A simple 3-D model is proposed to explain the inconsistencies in the published velocity models for the Nevada PASSCAL data. This model involves the presence of large, elongate ultramafic bodies at the base of the crust, oriented perpendicular to the spreading direction of the region. Such a model predicts the wide-angle traveltimes, but the actual wavefield response has not been calculated. A rough or undulatory crust-mantle boundary has been shown to produce the observed $PmP$ amplitude response, but the wavefield response parallel to such bodies necessitates a 3-D modeling technique requiring computer memory unavailable at this time.
Figure 5.33  Traveltime curves for receiver arrays perpendicular and parallel to the strike of lower crustal high velocity bodies which pinch out completely for a distance equal to their width.
Figure 5.34 Traveltime curves for receiver arrays perpendicular and parallel to the strike of lower crustal high velocity bodies which do not pinch out for a significant distance.
Chapter 6

Concluding Remarks

6.1 Summary of results

The complex waveforms recorded in deep crustal seismic records are testament to
the multi-scale velocity structure that exists within the earth's crust. Traditional
deterministic velocity models predict the main phases within a seismogram, but
generally assume a rather homogeneous crust. Stochastic velocity fields allow the
inclusion of wavelength-scale heterogeneity with these deterministic velocity mod-
els, so that a realistic full-wavefield simulation can be calculated. In this thesis,
I have combined deterministic and stochastic velocity models to investigate the
effects of wavelength-scale heterogeneity on seismic data from the Salton Trough-
Basin and Range transition in southeastern California and the Carson Sink-Dixie
Valley region of northern Nevada. The results shed light on varied subjects con-
cerning the nature of the earth's crust, and the effects of wavelength-scale hetero-
geney on the seismic methods used to investigate crustal structure.

The RISC/PACE reflection and refraction/wide-angle reflection data identify a
high-velocity body beneath the Salton Trough and Chocolate Mountains. Seismic
velocities and geochemical analyses indicate that this body is predominantly of gabbo-
ric composition, added to the base of the crust during rifting. This body extends
more than 35 km north of the closest active transform fault within the Trough,
the location of which, using the analogy of the oceanic ridge-transform systems,
should define the northeastward lateral extent of magmatic intrusion. However,
fossil transforms indicate that the locus of rifting within the Salton Trough has
moved over a broad area, to within 10 km of the edge of the high velocity identified
in the RISC data, and that a proto-Salton Trough may have existed as early as the
Miocene. Thus, the gabbroic body has probably been emplaced within the last 15
Ma, and its wide lateral extent is indicative of the ephemeral and mobile nature
of spreading loci during the initial stages of continental rifting.
The wide-angle and near-offset seismic data have illuminated the large and small-scale crustal structure of the edges of the Salton Trough. A high velocity layer at the base of the crust centered beneath the Salton Trough is constrained to pinch out beneath the Chocolate Mountains. This layer is identified in the PACE 1992 velocity model based on the analysis of seismic refraction data, and its lateral extent is constrained by the RISC reflection and wide-angle data. This layer is interpreted to be the result of magmatic additions to the base of the crust during Basin and Range extension and/or Salton Trough rifting. The RISC seismic reflection data does not shed light on the age of emplacement of the mafic lower crust; however, the highly reflective midcrust beneath the Chocolate Mountains is spatially confined to the region above the high velocity lower crustal layer. For this reason, regardless of the time of emplacement, the midcrustal reflectivity is believed to be genetically linked to the emplacement of the high velocity lower crust beneath the Chocolate Mountains. The high amplitude reflections adjacent to this high velocity layer have been modeled as horizontal intrusions with the same material properties as the lower crust. These intrusions form a midcrustal zone of mafic sills 5 km thick. The emplacement of horizontal intrusions within an otherwise extensional tectonic region may at first seem contradictory. However, recent rheological arguments have shown that such orientations can be expected at zones of contrasting strength, such as the Moho or the midcrustal brittle/ductile transition (McCarthy and Thompson, 1988; Glazner and Ussler, 1989; Parsons et al., 1992; Holliger and Levander, 1994) [85, 41, 95, 60]. As little as 5% of these horizontally oriented intrusions can produce the highly reflective lower crust. What is intriguing about this result is that such a small volume component can dominate the reflection signature, and potentially obscure important large-scale features, such as the high-velocity layer below. In addition, such a small amount of intrusion will have virtually no effect on the average crustal velocity, yet produce both vertical incidence and wide-angle “resections”. Exactly how much mantle derived material is emplaced into the crustal column during an extensional event such as Basin and Range rifting is still a topic of debate (Gans et al., 1989; Best and Christiansen, 1991) [36, 9]. Such studies involve assumptions of original crustal thickness, amount of stretching, and the ratio between extruded and intruded magma. The seismic reflection method probes these intrusive bodies as they exist within the present-day crust, and therefore can be used to estimate volumes of intrusion. The use of stochastic velocity models based upon geologic maps may be
the necessary tool to measure this quantity, and others like it such as body size and aspect ratio (e.g., Pullanmanappallil et al., 1995) [97]. Such information will prove necessary for future interpretations of deep crustal seismic data.

Velocity models which include both large and small-scale structures have also been used to create synthetic seismograms which reproduce the main phases and character of the recorded data from the RISC wide-angle profiles. The effect of wavelength-scale heterogeneity on both the specular phases and the scattered field was analyzed. With respect to the specular phases, it was necessary to introduce a high gradient zone at the base of the crust to suppress the PmS converted phase and precritical PmP phase. However, it was also noted that high-pass filtering the data creates a PmP phase which is relatively continuous at all offsets. This indicates that near-source surface waves may contribute to the lack of precritical PmP in the narrow-band refraction data. On the other hand, in Chapter 5, a rough crust-mantle boundary was shown to have the same effect as a gradient layer, so Moho topography may also play a role.

Analysis of the coda waves which follow PmP and SmS (Coda A) in the RISC wide-angle data shows that an initial (1-4 s) steeply decaying coda is due to diffusion of energy out of the near-surface basin layer. The decay rates of this coda are dependent upon the thickness of the basins at the source and receiver locations, with the thicker basins producing a slower decay rate. Frequency-wavenumber spectra of this coda contain little energy at slow velocities, indicating that conversion of incident body waves to surface waves is not important in this locale at long offsets. A reverberatory PmP phase is often cited as evidence for a layered lower crust (e.g. Sandmeier and Wenzel, 1986) [102]. The above results, however, indicate that the initial reverberation is more likely due to a combination of energy diffusion of PmP out of the basin layer and wavelength-scale scattering along the entire crustal path length, both sedimentary and crystalline. The coda which lies between the initial PmP coda and the Sg/SmS phases (Coda B) also contains little slow energy, and is relatively flat. These characteristics are consistent with coda generated as scattered waves and P to S body-wave conversions from wavelength-scale heterogeneities within the crust.

The RISC wide-angle shots within the Salton Trough are unique because of the lack of shear waves recorded. Only the shotpoints near the edges of the basin produce significant shear waves. This suggests that the presence of a basin/basement
contact near a shot is important in producing near-source converted energy like that observed in the RISC Basin and Range shots.

Combined deterministic and stochastic velocity models for the northern Basin and Range have been presented which reproduce the observed reflected and refracted wavefields reasonably well. The near-offset reflection signature is dominated by the wavelength-scale impedance structure. The midcrustal reflectivity between 4 and 7 s is produced by 5-25% horizontally-oriented, wavelength-scale mafic intrusions. The most consistent explanation for the less reflective lower crust involves an increase in bulk velocity at ~ 20 km depth, which may result in a decrease in the impedance contrasts within the zone. The reflection Moho is also modeled as arising from horizontally-oriented bodies just above the crust-mantle boundary. The kinematics of the wide-angle refraction data are dominated by the bulk velocity structure within the crust or at major interfaces, such as crust-mantle boundary. The refraction data require either a high-gradient transitional layer at the crust-mantle boundary and/or a crust-mantle boundary with topography.

A simple 3-D model is proposed to explain the inconsistencies in the published velocity models for the Nevada PASSCAL data. This model involves the presence of large, elongate ultramafic bodies at the base of the crust, oriented perpendicular to the spreading direction of the region, similar to the basin and range surface topography at the surface. Such a model predicts the wide-angle traveltimes, but the actual wavefield response has not been calculated. A rough or undulatory crust-mantle boundary has been shown to produce the observed PmP amplitude response, but the wavefield response parallel to such bodies necessitates a 3-D modeling technique requiring computer memory unavailable at this time.

6.2 Future work

Future work on the effects of wavelength-scale heterogeneity on the deep crustal wavefield can be divided into three categories: 1) scattering from discrete bodies, 2) scattering from rough interfaces, and 3) 3-D effects. The importance of the latter category is a forgone conclusion, but its level of importance is still unknown, and will remain in the realm of mathematical theory until modeling techniques become more efficient and computer memory of this magnitude becomes available. Despite the potential problems involved with ignoring out-of-plane effects, I believe
that the first and second categories can still be investigated further with significant benefits.

Scattering from discrete bodies involves scattering from isolated impedances within the crustal column (including the sedimentary basins) and upper mantle. This was the main focus of this thesis. One of the more important assumptions made in constructing the velocity models was the statistics which defined the sedimentary layers within the near-surface basins and the crystalline upper crust. In the basins, these statistics had to be assumed. A statistical description based upon an actual sedimentary basin could be beneficial. For the crystalline crust, realistic statistical descriptions are available, but are based on only a few maps. Often, as is the case for the Chocolate Mountains, these maps are not created to differentiate petrophysical properties, but to differentiate characteristic units which define the geologic history of the area. Thus, within a Precambrian gneiss, for example, the distribution of minor mafic and ultramafic bodies could be very important for the seismic response of the media. Analysis of more maps, as well, as more detailed maps, is therefore an important avenue of research.

Scattering from rough interfaces involves the interaction of seismic waves with the three main interfaces in the crustal column: surface topography, the basin-basement interface, and the crust-mantle boundary. Surface topography can create near-source shear waves, as well as generate or block surface waves. Both of these wave types, in particular surface waves, are a significant part of source-generated noise. Available finite-difference codes could not model a non-planar free surface, and hence the effects of topography were not modeled. As these codes become available, these effects will begin to be modeled and addressed.

The basin-basement interface is modeled in this thesis as a 2-D deterministic interface. However, roughness on this interface has been shown to produce measurable effects. It therefore seems prudent that future work should involve a rough basement interface. A statistical description of such an interface should be possible from geologic outcrop and/or detailed velocity analysis of reflection seismic data, so that reasonable effects created.

Finally, a rough crust-mantle boundary has been shown to produce the same effects as a high-gradient lower-crustal transitional layer. Though it will be rather difficult to obtain an independent statistical description of the crust-mantle boundary, it will be possible to determine what level of roughness is necessary to produce
the observed data. This will aid in the understanding of the interaction of the upper mantle and the continental crust during tectonic events.
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wavelet modeling of wave propagation: Optimal finite-difference stencil de-


Appendix A

Wave-equation Datuming for Improving Deep Crustal Seismic Images

A.1 Introduction

The quality of reflections in a shot gather are reduced due to a variety of effects associated with the near-surface. Air waves, surface waves, surface and basement topography and near-surface velocity heterogeneity all combine to deteriorate the quality of a seismic record. The standard method for removing the effects of topography and near-surface velocity variations is to apply a static time-shift to each trace, corresponding to the delay time due to the topography at the receiver station and the velocity directly below the station. For small changes in elevation, small velocity changes between the input and output datum, and vertically travelling energy, this may be sufficient. However, in areas where these simplifications are invalid, a more accurate technique must be employed.

One such technique involves the extrapolation of the recorded wavefield from the recording surface to another reference surface or datum. The basic conceptual premise behind this technique can be stated as follows: provided we have adequate information on the near-surface velocity distribution, we can downward continue the recorded wavefield below this complicated near-surface layer and then upward continue to a flat datum using the basement velocity as the replacement velocity, completely removing the effects of the surface layer and topography. Such techniques are termed wave-equation datuming [6]. Their suitability requires that 1) the near-surface velocity distribution is adequately known and 2) the recorded wave-field is not spatially aliased for the wavelengths of interest. Wave-equation datuming has been shown to be quite successful when applied to synthetic seismograms [6, 118, 104] and stacked and unstacked marine data to remove the effects of an irregular water bottom [7, 8, 119]. In addition, McMechan and Sun [1991] showed with some synthetic examples that horizontally travelling low-phase velocity energy which does not penetrate too deeply (i.e., direct body waves and
surface waves) are also removed by this technique, even if the near-surface velocity is poorly known. Therefore, wave-equation datuming is capable of removing many of the near-surface noise problems commonly associated with land-based crustal seismic experiments.

In an effort to remove the effects of the variable near-surface velocity layer and topography, we have used wave-equation datuming to downward-continue shot gathers to the base of the surface layer, and then we redatum upward to a flat surface using the basement velocity as the replacement velocity. Results from both synthetic and recorded data are presented.

A.2 Theory

The mathematical formulation can be described as follows. Given a wavefield \( U(P_0) \) recorded on a surface \( S_0 \), we would like to obtain the wavefield \( U(P_1) \) at some other arbitrary surface \( S_1 \), where \( P_0 \) and \( P_1 \) are positions on the recording surface and on the extrapolation surface (or datum), respectively (Figure A.1). We begin with the integral theorem of Helmholtz and Kirchhoff [44],

\[
U(P_1) = \frac{1}{4\pi} \int_{S_0} \left( \frac{\partial U(P_0)}{\partial n} G(P_0, P_1) - U(P_0) \frac{\partial G(P_0, P_1)}{\partial n} \right) ds,
\]

which states that given the wavefield and its normal derivative along a surface \( S_0 \), and the Green's function \( G(P_0, P_1) \) of the medium, we can predict the motion \( U(P_1) \) at any other point in the medium. For constant near-surface velocity, we choose the free-surface Green's function as the impulse solution to wave equation.

\[
G = \frac{e^{ikr}}{r} - \frac{e^{ik\hat{r}}}{\hat{r}}.
\]

This is generated by assuming point sources located at \( P_1 \) and \( \hat{P}_1 \), which are mirror images of each other about the recording surface. \( r \) and \( \hat{r} \) are the distances along the ray paths from \( P_1 \) and \( \hat{P}_1 \) to the recording surface \( S_0 \), respectively, and are equal in magnitude. The utility of choosing this Green's function is that on the integration surface \( G = 0 \), so that the normal derivative of \( U(P_1) \) is not needed. Though only an exact solution to the wave equation when the integration surface is flat (not necessarily horizontal), this Green's function is still adequate as long as the undulations in the integration surface are longer than the dominant wavelength. In the case of deep crustal seismic surveys, dominant near-surface
wavelengths are on the order of 100 m. Since basement topography in crustal velocity models is rarely known to this accuracy, the upward continuation part of the layer replacement is not a problem. For the downward continuation step from surface topography to basement, as long as dramatic changes in topography do not occur over a lateral distance of a hundred meters or less, this simple Green’s function should be sufficient. Taking the normal derivative of $G$ we obtain

$$\frac{\partial G}{\partial n} = \cos(\hat{n}, \vec{r}) \left[ ik - \frac{1}{r} \right] \frac{e^{ikt}}{r} - \cos(\hat{n}, \vec{r}) \left[ ik - \frac{1}{r} \right] \frac{e^{ikt}}{r}. \quad (A.3)$$

Substituting Equations (A.2) and (A.3) into Equation (A.1) and noting that $\cos(\hat{n}, \vec{r}) = -\cos(\hat{n}, \vec{r})$, we obtain

$$U(P_1) = \frac{1}{2\pi} \int_{S_0} U(P_0) \left( \cos(\hat{n}, \vec{r}) \left[ \frac{i\omega}{cr} - \frac{1}{r^2} \right] \right) e^{i\omega \tau} \, ds, \quad (A.4)$$

where $\omega = ck$ is the angular frequency, $\tau = r/c$ is the traveltime along the raypath from $P_0$ to $P_1$, and $c$ is the acoustic velocity.

To reduce computational effort, the $1/r^2$ term is usually neglected [118], under the assumption that $1/r$ is small compared to $\omega/c$. This approximation is adequate as long as the extrapolation distance is a few wavelengths from the recording surface. However, in crustal seismics, spread lengths are often quite long and therefore cross both sedimentary basins and mountain ranges, so that the bottom of the near surface layer (the basement interface) will often approach the topographic interface. We therefore retain this term in the calculation. Synthetic tests have shown that propagation distances as small as $\lambda/2$ yield excellent results when keeping the $1/r^2$ term. Smaller propagation distances produce phase distortions in the wavelet shape, but correct static time shifts are still achieved. Therefore, given a near-surface dominant wavelength of 100 m, phase distortions will occur directly below areas where basement velocities (not simply basement rocks themselves) are within 50 m of the surface. Though this may appear to be a substantial limitation, it should be remembered that even over mountainous terrain where basement rocks are at or near the surface, a weathering layer is usually present, and this is part of the near-surface layer which is to be removed by the datuming process.

Equation (A.4) is a surface integral derived using a point source Green’s function. However, most recording geometries involve linear arrays (we assume a 2-D Earth structure), and we record the wavefield along a line, so that a surface integration is not actually possible. We consequently need to modify the integral so
Figure A.1  Geometry for Kirchhoff extrapolation. $P_0$ is defined along the integration surface, $S_0$, and $P_1$ is defined on the extrapolation surface. $S_1$. $\hat{n}$ is the inward unit normal vector to the integration surface, and $\theta$ is the angle between $\hat{n}$ and the local raypath at $P_0$. For a constant wave speed of $c_0$, the raypaths are straight lines.

that an integration over the line array is scaled properly to account for the 3-D spreading from a point source into a line array. Lafond [1991] applied the method of stationary phase along the out-of-plane coordinate to generate the so-called 2.5-D integral for pre-stack migration. We employ the same method, retaining the $1/r^2$ term, and obtain

$$U(P_1) = \frac{1}{\sqrt{2\pi}} \frac{1}{\sqrt{\omega}} e^{iT} \int_{S_0} U(P_0) \left( \cos(\hat{n}, \tau) \left[ \frac{i\omega}{cr} - \frac{1}{r^2} \right] \sqrt{cr} e^{i\omega r} \right) dx \quad (A.5)$$

Equation (A.5) is the Kirchhoff integral we use in our implementation of the wave-equation datuming for deep crustal profiles. The constant velocity assumption significantly reduces computation time, and although gradients do exist in the near-surface, the examples later in this paper indicate that a significant improvement in data quality is still achieved. Other workers have developed similar extrapolation integrals for use in traditional industry seismic profiles and have discussed their computational advantages [6, 7, 118, 104, 10].
A.3 Method

A near-surface velocity model must first be developed before extrapolating the wavefield to a new datum surface. Using this velocity model, implementation of Equation (A.5) involves the calculation of the aperture width, the length of the raypath from each point of the aperture to the extrapolation point, the traveltime along the raypaths, and the angles between the local normal to the aperture ($\hat{n}$) and the local raypaths. Then, for each frequency in the wavefield, the disturbance (seismic trace) from each point on the aperture is fed into Equation (A.5) and the result is one point on the datumed surface. The process is then repeated at each new point on the datum surface until a complete extrapolated wavefield exists at the new surface.

A.3.1 The aperture

The aperture is the number of traces along the integration surface used in computing one extrapolated point. In standard reflection recording involving a moving spread along the profile, the aperture length used is often the entire spread length. However, for crustal-scale experiments, one deployment can be 50 km or more in length. Such a long aperture will increase the computation time significantly and is not necessary to construct an accurate extrapolated wavefield. In our implementation, the width of the aperture is chosen so that the half-width of the main lobe of its corresponding Fraunhofer diffraction pattern is less than the average station spacing. We choose the Fraunhofer diffraction pattern, as opposed to the Fresnel diffraction pattern, because of its simpler expression. It is simply the Fourier transform of the aperture. Although we recognize that in our case we are never, strictly speaking, in the Fraunhofer field, experience shows that this criteria is more than sufficient. Using a significantly smaller aperture may cause smearing from trace to trace in the extrapolated field, and a larger aperture needlessly increases computation time. The equation for the Fraunhofer intensity pattern of a rectangular aperture can be written [44]

$$I(x) = \frac{l_x^2}{\lambda^2 x^2} \sin^2 \left( \frac{l_x x}{\lambda x} \right).$$

The width of the main lobe is

$$\Delta x = 2 \frac{\lambda x}{l_x}.$$
Hence, with $\Delta x$ as the nominal station spacing, $\lambda_{\text{max}}$ the longest wavelength and $z_{\text{max}}$ the farthest vertical extrapolation distance, a proper choice for the aperture width would be

$$l_x = \frac{2\lambda_{\text{max}}z_{\text{max}}}{\Delta x}.$$  

A cosine taper is applied to the ends of each aperture to reduce edge effects.

### A.3.2 Raypaths and travel-times

For a constant velocity medium, the raypaths are the straight line segments between $P_0$ and $P_1$, and $r$ is the length of the straight raypath. The traveltime is then simply $\tau = r/c$. In the examples below, a constant velocity near-surface layer is assumed, and is justified by the significant improvement in data quality after datuming. However, for many seismic datasets, a constant velocity assumption may not be adequate, and a raytracing technique can be implemented to define the raypaths and traveltimes from $P_0$ to $P_1$. For media with constant velocity gradients, the travel paths are arcs of circles, and these terms can still be calculated analytically. For heterogeneous velocity distributions, these terms are calculated numerically.

### A.4 Simple synthetic example

To test the algorithm, we created a synthetic shot record with a viscoelastic finite-difference wave propagation simulator (Robertsson and others, 1994). The model geometry is displayed in Figure A.2. The model consists of an irregular, low-velocity (3.0 km/s) near-surface layer overlying an intermediate velocity (5.0 km/s) basement. To create a target reflection, a high velocity (8.0 km/s) layer was placed at the bottom of the model at 4 km depth. The source was a Ricker wavelet with a dominant frequency of 18 Hz placed 20 m below the surface in the center of the model. No surface topography is present because of the inability of the finite-difference code to model a nonplanar free-surface boundary condition. Seismograms were calculated every 20 m along the free surface, resulting in a spatially unaliased shot gather. If spatially aliased energy is present, such as air waves, they should be attenuated before datuming or the aliased energy will be spread to adjacent traces.

The resulting seismic record displayed in Figure A.3a contains various reflected and refracted waves complicated by the irregular near-surface. For display pur-
poses, only every 10th trace is plotted. Present in the seismogram are two first-break arrivals from the near-surface layer and basement, a surface wave with a velocity of 1.6 km/s, reflections from the basement interface, and a reflection at 1.9 s from the horizontal interface at 4 km depth. The events at 2.5 s and 2.9 s are pegleg multiples.

Figure A.3b shows the seismogram after downward continuation through the 3.0 km/s layer to the basement interface. Already the slower surface wave phase has been removed, and “moved” off the section to negative times. The same is true of the 3.0 km/s direct wave. The target reflection has moved up in time as expected, and now shows the traveltimes delays due to basement topography alone. The second pass through the datuming routine returns the wavefield to the original surface, but with a replacement velocity of 5.0 km/s. This final result is displayed in Figure A.3c. The surface waves are still gone, the first break refraction is now a single phase with the basement velocity of 5.0 km/s, and the target reflection has the correct depth and symmetry of a reflection from a flat interface. The basement reflection/refraction which interfered with the target reflection at the right side of the shot gather has been shifted to earlier times and is now separated from the target reflection. Finally, the symmetry in amplitude for the target reflector is recreated, with only a slight underestimation of amplitude on the right side of the hyperbola at short offsets.

As a final test, we also processed the same synthetic seismic record combining velocity filtering and traditional statics using the exact velocity model to compute the static corrections. The result of this optimal traditional processing is displayed in Figure A.3d. The symmetric nature in both traveltimes and amplitude of the target reflection is not as clearly reproduced as in the wave-equation datumed gather. For example, slight delays still exist along the reflection directly above the thickest section of the basin resulting from the traditional static assumption of vertically travelling waves. Also, interference with the basement reflection still obscures the right side of the target reflection.

A.5 Real data example

In 1992, Rice University and the University of Southern California (RISC) collected a deep crustal seismic dataset consisting of 444 three-component recording stations with a 50 m group interval. Instrumentation was provided by ARCO.
PASSCAL, and Chevron. The instruments were deployed end-to-end twice, and many shotpoints were fired into both deployments, resulting in a total spread length approaching 45 km. Like most land crustal seismic datasets, the shot gathers of the RISC experiment are complicated by the presence of surface waves and irregularities in topography and the surface layer velocity distribution. Figure A.4 and A.5 display a typical shot gather at two different offset ranges. Note the complicated arrival times of the first break due to the topography at the surface and along the basement interface. Note also the surface waves interfering with reflections at both near and far offsets. Processing prior to wave-equation datuming included a bandpass filter and air-wave attenuation.

As mentioned above, the success of the datuming technique depends on the near-surface velocity model used by the datuming algorithm. A poor velocity model results in an inaccurate removal of surface related statics, but does not effect the removal of surface waves and other horizontally travelling low-velocity phases. The ability to determine a satisfactory near-surface velocity model is data dependent, and may be best measured by how well the character of the original data is preserved. A near-surface velocity model across the study area was developed by inverting the travel-times of shallow refracted arrivals [121] from 10 shot gathers with average spacing of 5 km. This coarse shot spacing limits the level of detail resolvable in the velocity model. It also prevents the wavefield extrapolation of receiver gathers because the shot space is spatially aliased for all waves of interest.
Figure A.3  a) Synthetic shot record containing main phases observed in a typical field record. The target reflection is present at 1.9 s. b) Synthetic shot record after downward continuation through a 3.0 km/s velocity layer to the basement interface. Note the attenuation of the surface wave phase. c) Synthetic shot record after upward continuation to the flat topographic surface at the basement velocity (5.0 km/s). The target reflection is intact, with symmetric moveout and amplitude. d) Synthetic shot gather with traditional statics and f/k filtering applied to remove near-surface effects. Note lack of symmetry in both traveltime and amplitude.
Figure A.4  Near-offset shot gather from Shotpoint 106 of the RISC deep crustal seismic experiment. Preliminary processing includes a bandpass filter and air-wave attenuation.
Figure A.5  Far-offset shot gather from Shotpoint 106. Preliminary processing includes a bandpass filter.
Before datuming, the near-surface velocity from the initial inversion was replaced by the average velocity of the layer, and the model was fed back into the inversion with the near-surface velocity held fixed and only the basement topography allowed to change. This final velocity model was used as the starting velocity model for the datuming process, and is displayed in Figure A.6. Other models using different velocities and basement depths were tested but did not produce the best static corrections (indicated by a flat first break), though all models tested removed the surface wave energy.

The Kirchhoff integral in Equation A.4 was used to downward extrapolate the recorded wavefield from the topographic surface to the modeled basement surface. The wavefield on the basement surface was then upward continued to a flat surface using the Kirchhoff integral with the basement velocity of 5.1 km/s as the extrapolation velocity, thus removing the surface layer from the data. The results of the wave-equation datuming applied to SP106 are shown in Figure A.7 and A.9. The most striking improvement observed in the datumed shot gather is the elimination of the surface waves and the subsequent enhancement of reflectivity. This occurs at both near and far offsets. Reflections which were clear on the field record are virtually unchanged on the datumed gather. The air wave in Figure A.4, which is aliased and imperfectly suppressed, produces smearing in the near-offset section. In addition to the removal of surface waves, the first break is flattened, indicating that we have adequately removed near-surface delays. This has resulted in an increase in reflector continuity in areas where these delays were most prevalent. Frequency spectra before and after datuming are identical, except for the removal of the high amplitude, lower frequency part of the surface waves. Again, as a further check, traditionally processed sections which include vertical static corrections based upon the velocity model in Figure A.6 and f/k filtering are displayed in Figures A.8 A.10 for comparison. The velocity filtering is not as successful in removing the surface waves from the data, and also results in smearing from trace to trace, most evident in the Moho reflection at 8.5 s in Figure A.8.

A.6 Conclusions

The application of wave-equation datuming by means of Kirchhoff extrapolation to crustal seismic data significantly improves the reflection image. Surface waves are removed, even if the near surface velocity field is not perfectly known. Phases
Figure A.6  Velocity model determined from the inversion of first-break refractions used in the datuming of the shot records in Figures A.4 and A.5

which are spatially aliased (e.g. air waves) must be suppressed prior to datuming. Time delays due to near-surface velocity heterogeneity can be removed if the near-surface velocity structure is adequately known. This is a significant improvement over traditional datum static corrections which assume a vertical time correction and do not honor wave propagation, and can be used on a variety of crustal datasets where the near-surface has a deteriorous effect on the recorded wavefield. As shot and receiver spacing increase in data aquisition, more detailed velocity models will be possible, and techniques to remove near-surface heterogeneity based upon wave propagation may become more valuable in the analysis of deep crustal seismic data.
Figure A.7  Near-offset Shotpoint 106 shot record after datuming. Surface waves have been removed to reveal reflectivity beneath the shot. First break is flattened indicating near-surface static problems have been reduced.
Figure A.8  Near-offset Shotpoint 106 shot gather with traditional statics and f/k filtering to remove near-surface effects.
Figure A.9  Far-offset shot gather from Shotpoint 106 after datuming. Surface waves have been removed to reveal wide-angle reflections.
Figure A.10 Far-offset shot gather from Shotpoint 106 with traditional statics and f/k filtering to remove near-surface effects.
Appendix B

Method of Estimating Volume of Heterogeneity Using a Measure of Reflection Density

B.1 Introduction

Determination of the amount of mantle derived basaltic magma within the continental crust of the Basin and Range province is crucial to understanding the interaction between the crust and upper mantle during continental extension. Unfortunately, geological estimates of the amount of mantle derived basaltic magma intruded into the crust during Basin and Range extension can range from 5 to 15 km of the crustal column (15-50% by volume) (e.g., Gans et al. [1989]; Best and Christiansen [1991]) [36, 9]. Such a large range in estimates is indicative of the uncertainties involved in making such estimates based upon surface exposures alone. Most of this material does not reach the surface as lava, but ponds at various depths within the crust, depending upon magma supply and density.

The distribution of mafic rocks within the crustal column is one of the targets of subsurface geophysical techniques such as reflection and refraction seismology. Refraction seismology attempts to obtain bulk velocity estimates of the crust and upper mantle. Such measurements are reasonably accurate within layers from which a refracted or diving body wave penetrates and arrives as a first break. This occurs in the upper crust and upper mantle. Estimates within the middle and lower crust are less unique, requiring the interpretation of relevant seismic phases. For example, in the northern Basin and Range province, mid-crustal velocity estimates range by 0.2 km/s (Holbrook, 1990; Catchings and Mooney, 1991; Benz et al., 1990; Hawman et al., 1990; Zelt and Smith, 1992) [57, 13, 5, 50, 121]. Assuming a background seismic velocity of 6.2 km/s, an addition of 5% magma with a 6.6 km/s seismic velocity results in a 0.02 km/s increase, an addition of 15% results in a 0.06 km/s increase, an addition of 25% results in a 0.10 km/s increase, and an addition of 50% results in a 0.20 km/s increase. The range in possible volume percentages within a 0.2 km/s velocity limit is the same as the
ranges determined from geologic studies of the surface. Looked at another way, given a lower crust with presumed gabbroic composition, as much as half can be of felsic composition and still be within the errors of the method. Combined with the variability of velocities given a particular rock type, volume estimates from bulk velocity estimates are even less precise. Clearly, a more precise measure would be useful.

In Chapter 3, synthetic seismograms computed from velocity models of an intruded crust indicate that a significant difference in the amount of reflectivity is observed when the percentage of the high-velocity bodies responsible for the reflectivity is changed, especially in the ranges from 0 to 25%. Specifically, the density of reflections in time and space is distinctly different. This is the range where bulk-velocity determinations provide the least information. Therefore, an estimate of the volume of intrusion based upon the density of reflections might be possible and would provide a more precise constraint than those from measurements of bulk velocity alone.

### B.2 Method

As a working definition of reflection, I simply mean the return to a geophone of a wave with an amplitude above a certain threshold value. This is a trace-specific definition, and does not consider lateral correlation between traces. Recorded data often changes markedly from trace to trace due to near-surface velocity heterogeneity and geophone site response. This complexity can obscure reflector continuity without some preprocessing of the data, and such preprocessing can prompt more questions concerning the validity of a measure than ignoring coherence altogether. With this single-trace definition of reflector in mind, the question remains how to count the number of reflections within a specified zone in a seismogram.

To simplify the seismogram, each trace is transformed into an energy envelope which preserves the location of the reflected energy in time and removes the wavelet shape and negative amplitudes. This is accomplished by adding the squared amplitude of a seismogram to its Hilbert transform (Bittner and Rabbel, 1991) [12]. This transformation of the data was accomplished using a standard seismic processing package.

Because the amplitude of a signal is the sole criteria for whether it is deemed a reflection or not, determination of a proper threshold value is essential. This
threshold value must remove the background scattered field due to path effects as well as preserve the target reflections. In the description below, "background field" is defined as that part of the seismogram just above the reflective zone of interest, and the "target field" is this reflective zone. Five different thresholds were tested, with varying degrees of success: 1) the peak amplitude of the background field, 2) the rms amplitude of the background field, 3) the average of the peak amplitudes in the background and target fields, 4) the average of rms amplitude of background and target fields, and 5) the average of the rms amplitude of the background field and the peak amplitude of the target field. This final threshold value has proven to be the most stable measure, and is used in the measurements made below.

Thresholding involves simply subtracting the threshold values for each trace from each energy envelope. After thresholding, each trace (energy envelope) is scanned within the target zone of reflectivity for remaining positive amplitudes. Each positive peak is counted as a reflection. The average number of reflections per trace should be related to the total volume of intrusion.

Before proceeding with the application of the method, some assumptions and limitations should be mentioned. First, the above method does not invert a recorded seismic section for a volume parameter. Rather, a suite of velocity models must be created with a range of body sizes and volumes which predict the time and amplitude character of the background and target fields. A comparison of the number of reflections calculated in the observed data with the number of reflections calculated from the synthetic seismograms identifies quantitatively which model (and hence what volume) best predicts the observed wavefield. Second, the method implicitly assumes that a binary (or a distinct bimodal) distribution of velocities is responsible for the reflected wavefield. Finally, because high-amplitude reflectivity only exists in the presence of vertically-oriented structures, a lower crust heavily intruded by dikes will not be an applicable target.

**B.3 Application**

Before measuring the reflected wave field, it is useful to count the actual number of reflections within the velocity models. This was accomplished by averaging the number of impedance contrasts in each grid column of the velocity model over the target zone (i.e., the horizontal intrusions between 17 and 22 km). The results are listed in Table B.1 and displayed in Figure B.1. The number of impedances
does not increase linearly with an increase in volume of heterogeneity. Instead, as the volume of heterogeneity increases, the high-velocity bodies begin to coalesce, and the number of potential reflectors no longer increases simply with volume. This is most evident in the 800 x 200 m model. This feature is seen visually in Figures 3.9-3.13, where saturation of the reflected wavefield begins to be observed above 15% intrusion.

Analysis of the number of impedances within the velocity models is an interesting starting point, but the determination of such an exact image with seismic data is not possible. The ideal seismic image can be represented by the primary reflectivity series (PRS) which consists of a series of 1-D, vertical-incidence, primaries-only synthetic seismograms (Claerbout, 1985) [15]. The PRS includes the effects of a bandlimited source wavelet. The PRS was computed for the stochastic velocity models, and the number of reflections was calculated using the thresholding technique described above. The results are listed in Table B.2 and displayed in Figure B.2. The number of reflections in the PRS is significantly fewer than the number of reflectors in the velocity models. Note also that, like the reflector count, there is not a linear relationship between volume and reflections, with the wavefield approaching saturation beyond 25%. The method, though not an absolute measure, does appear to predict the general shape of the curve with increasing volume and, as indicated earlier, is most robust at lower volume levels. Higher volumes (< 25%) result in a saturated wavefield.

The above method is used to estimate the volume of mafic intrusion within the reflective mid-crust by making use of the suite of velocity models and corresponding synthetic seismograms for shot point 106 described in Chapter 3. Figure B.3 displays the energy-envelope seismic section for shot point 106. The background field time gate was set to 4.5-5.0 s and the target field time gate was set to 5.5-7.0 s. The average number of reflections per trace after thresholding for this shot gather is 4.99. Using the same procedure, measurements were also made on the synthetic shot gathers from the 300x75 m intrusion models and the 800x200 m intrusion models. The results are listed in Table B.3 and displayed in Figure B.4. 10-15% intrusion is suggested by these measurements. Again, like the PRS seismograms, the wave field becomes saturated beyond a 25% volume. Note also that for a fixed density of reflections, larger body sizes require larger volumes of intrusion.
Figure B.1 Reflectors count from stochastic velocity models for various volumes of heterogeneity. Note that there is not a linear relationship between volume and number of impedances.
Figure B.2  Reflection count from PRS of stochastic velocity models for various volumes of heterogeneity. The number of reflections in the PRS is significantly fewer than the number of reflectors in the velocity models. Note also that, like the refector count, there is no linear relationship between volume and reflections, with the wavefield approaching saturation beyond 25%.
Figure B.3  Energy-envelope seismic section for shot point 106. The target zone of reflections is between 5.5 and 7.0 s.
Figure B.4 Reflection count from energy-envelope seismic sections computed from stochastic velocity models with varying volumes of heterogeneity.
B.4 Discussion and Conclusions

A detailed error analysis of the above method has not been pursued to date. The standard deviation of the synthetic measurements ranged from approximately one reflection for the low volumes to as much as three reflections for the larger volumes. These large relative standard deviations suggest that the results of the method as applied to the RISC data should be viewed with caution. Part of the problem may result from the relatively short time window over which the reflectivity is observed. A better result might be obtained for a data set with a reflective zone of a few seconds or more.

Nevertheless, saturation of both the impedance field and the reflected field at higher intrusion volumes is an interesting characteristic of stochastic velocity models. Furthermore, the highest potential resolution of reflection density occurs within the low-volume range, which is the range in which traditional velocity estimates would provide the least information. This indicates that further analysis of the statistics of the reflected wave field might provide needed constraints on important geologic questions such as volumes of intrusion.
**Table B.1** Reflector count from stochastic velocity models described in Chapter 3.

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<td>4.52</td>
<td>10.08</td>
<td>13.35</td>
<td>17.35</td>
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**Table B.2** Reflection count measurements the PRS of the stochastic velocity models.

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**Table B.3** Reflection count measurements for synthetic seismograms of shot point 106.

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