Investigation of Seismic Activity and Subsurface Structure under Tectonic and Anthropogenic Settings

Qiong Zhang
University of Miami, qzhang@rsmas.miami.edu

Follow this and additional works at: https://scholarlyrepository.miami.edu/oa_dissertations

Recommended Citation
Zhang, Qiong, "Investigation of Seismic Activity and Subsurface Structure under Tectonic and Anthropogenic Settings" (2014). Open Access Dissertations. 1279.
https://scholarlyrepository.miami.edu/oa_dissertations/1279

This Embargoed is brought to you for free and open access by the Electronic Theses and Dissertations at Scholarly Repository. It has been accepted for inclusion in Open Access Dissertations by an authorized administrator of Scholarly Repository. For more information, please contact repository.library@miami.edu.
UNIVERSITY OF MIAMI

INVESTIGATION OF SEISMIC ACTIVITY AND SUBSURFACE STRUCTURE
UNDER TECTONIC AND ANTHROPOGENIC SETTINGS

By

Qiong Zhang

A DISSERTATION

Submitted to the Faculty
of the University of Miami
in partial fulfillment of the requirements for
the degree of Doctor of Philosophy

Coral Gables, Florida

August 2014
UNIVERSITY OF MIAMI

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy

INVESTIGATION OF SEISMIC ACTIVITY AND SUBSURFACE STRUCTURE UNDER TECTONIC AND ANTHROPOGENIC SETTINGS

Qiong Zhang

Approved:

Guoqing Lin, Ph.D.
Assistant Professor of Marine Geology and Geophysics

Falk Amelung, Ph.D.
Professor of Marine Geology and Geophysics

Shimon Wdowinski, Ph.D.
Research Associate Professor of Marine Geology and Geophysics

Christopher Harrison, Ph.D.
Professor of Marine Geology and Geophysics

Timothy H. Dixon, Ph.D.
Professor of Geophysics
University of South Florida

M. Brian Blake, Ph.D.
Dean of the Graduate School
To study interaction of tectonics and earthquakes, we need to address two fundamental questions, Earth's internal structure and the cause of triggered/inhibited seismicity. High-resolution seismic velocity models and analyses of high-precision relocated earthquake have proven to provide useful information from global to regional scale. In areas of active tectonics and/or anthropogenic activities, the stress heterogeneity could largely result from lateral and vertical variation of structure in local scale, which accounts for the non-uniform earthquake distribution. We have mainly conducted two regional studies in Puerto Rico, the plate boundary zone, and the active Coso geothermal field. By inverting for three-dimensional velocity model and relocating earthquakes, we find out the active seismicity in the southwestern island of Puerto Rico and the seismic quiescence in the north could be separated by a high velocity body underneath the Great Southern Puerto Rico Zone dipping to the north. Beneath the Coso geothermal field, our results reveal an anomalous low velocity zone corresponding to the ductile behavior at depths from 6 to 12 km. We suggest that the pervasive melts are not likely to exist in the upper crust and the identified magmatic system could be frozen, felsic, with the inclusion of water. In Coso, by thorough analyses of spatiotemporal distribution of earthquakes, we present a novel observation that the geothermal field is less susceptible to remote triggering than the surrounding areas. Combining with the observations of an abrupt drop of background seismicity with respect to the geothermal operation, the continuous net production rate, the strong
subsidence for the reservoir, we imply that the absence of remote triggering could result from loss of pore pressure accompanying the long-term loss of pore fluid. This phenomenon has also been recognized in the Salton Sea geothermal field, which share similar tectonic and anthropogenic settings as the Coso geothermal field. Our results have also shown that the geothermal production process has altered the stress orientation by adding more normal-faulting components. To further understanding how the stress state and failure conditions are affected by the anthropogenic activity, it is necessary to conduct systematic case studies in different areas with presence of both natural and induced earthquake and include other studies such as independent stress measurements, hydraulic response to stress, and structure of faults and fractures.
Acknowledgements

I have many people to thank for all kinds of help during my Ph.D. study. First, I would like to owe special thanks to my advisor, Guoqing Lin. She has supported my study in all ways, such as leading me to the field of seismology, sponsoring me to different training and conferences, and always available to help me in the shortest time. I also thank my other committee members. Shimon Wdowinski has provided me a lot of constructive comments and invoked my interests for physics of earthquakes. Falk Amelung has shown me to have a broad interest for geophysics and make presentation clear and simple. Chris Harrison has taught me the knowledge of geophysics and helped me revise the dissertation proposal. Tim Dixon always encourages me to continue my research and provides the insightful comments regarding the right direction for my projects. I am also grateful to other faculty and research scientists in MGG, who offer a variety of interesting geophysical and geological courses, especially Gregor Eberli and Ralf Weger, for answering me a lot of geology questions and creating a friendly environment.

My Ph.D. study has been a joyful journey with the company of many friends in MGG. I am grateful to Amanda Piggot. She helped me revise a paper sentence by sentence, which greatly improved my writing. She always encourages and believes me
when I deal with difficulties in work or life. I must also thank my two wonderful office-mates, Estelle Chaussard, my role model of research work, who has been motivating, encouraging and helping me to be a good scientist, and Heresh Fattahi, a good friend in work and life, who always supports me and provides the InSAR results for Chapter 4. I also gain a lot of inspirations from discussion with other group members, Peng Li, Wenliang Zhao, Fernando Greene, and Emanuelle Feliciano. I am thankful to the great company of Yula Hernawati, Irena Maura, and Qian Yang, for all the joy and movement they brought.

Finally, I thank my dear family. Without their support, I could not have completed my study in time. I appreciate the continuous encouragement and help from Zhongwen. I thank my sister Yan for encouraging me to come to the U.S. for study. My parents have been believing and comforting me in every moment.
Contents

List of Figures ix

List of Tables xii

1 Introduction 1

2 Seismic structure beneath the island of Puerto Rico inferred from local earthquake tomography and relocation 6

2.1 Summary 7

2.2 Background and Motivation 7

2.3 Data and Method 11

2.3.1 Data Set 11

2.3.2 1D Model Inversion 11

2.3.3 1D Earthquake Relocation and Relative Location Error 15

2.3.4 3D Model Inversion 16

2.4 Results 21

2.4.1 Final Vp and Vp/Vs Models 21

2.4.2 Earthquake Relocation 24

2.5 Concluding Remarks 28
3 Three-dimensional Vp and Vp/Vs models in the Coso geothermal field, California: seismic characterization of the magmatic system 31

3.1 Summary 32

3.2 Background and Motivation 33

3.3 Data and Methods 37

3.3.1 Data Set 37

3.3.2 One-dimensional Starting Model 38

3.3.3 Three-dimensional Tomographic Inversion 40

3.4 Results 42

3.4.1 Regional Coarse Model 42

3.4.2 Earthquake Relocation 43

3.4.3 Model Resolution Tests 44

3.4.4 Map Views of Final Velocity Models 50

3.4.5 Cross-sections of Final Velocity Models 56

3.4.6 In-situ Vp/Vs Ratios in Similar Event Clusters 57

3.5 Discussion 59

3.5.1 Geothermal Reservoir 61

3.5.2 Magmatic System 61

3.5.2.1 Comparison with Previous Seismic Studies 66

3.6 Conclusion 67
4 Absence of remote earthquake triggering in the Coso and Salton Sea geothermal fields

4.1 Summary

4.2 Overview

4.2.1 Background and Motivation

4.2.2 Tectonic and Anthropogenic Settings in Geothermal Fields

4.3 Data and Method

4.3.1 Data Set

4.3.2 Estimate of Catalog Completeness and b-value

4.3.3 Declustering Catalog

4.3.4 Statistical Analysis of Seismicity Rate Change

4.4 Results

4.4.1 Seismicity Rate Change in Coso

4.4.2 Correlation with Geothermal Production

4.4.3 Stress Rotation

4.4.4 Seismicity Rate Change in Salton Sea

4.4.5 Background Seismicity and Conceptual Model

4.5 Conclusions

4.6 Acknowledgements

5 Conclusions and Future Work
List of Figures

2.1 Seismicity and station distributions in the vicinity of Puerto Rico . . . 10
2.2 Optimal 1D Vp and Vp/Vs models . . . . . . . . . . . . . . . . . . . . 14
2.3 3D view of events used for inversion . . . . . . . . . . . . . . . . . . . 17
2.4 Trade-off curves of normalized data misfit and model variance . . . . 18
2.5 Distribution of Derivative Weighted Sum values for P rays . . . . . . 19
2.6 Distribution of Derivative Weighted Sum values for S-P rays . . . . . 20
2.7 Checkerboard test of Vp model . . . . . . . . . . . . . . . . . . . . . . 22
2.8 Checkerboard test of Vp/Vs model . . . . . . . . . . . . . . . . . . . . 23
2.9 Map view of the final 3D Vp model for the entire island . . . . . . . 25
2.10 Map view of the final 3D Vp/Vs model for the entire island . . . . . 26
2.11 Cross-sections of the final 3D Vp and Vp/Vs models . . . . . . . . . . 27
2.12 Depth distribution of earthquake locations . . . . . . . . . . . . . . . . 29
3.1 Geologic map of the Coso area and close-up of our study areas . . . . 35
3.2 Comparison of different 1-D velocity models . . . . . . . . . . . . . . . 39
3.3 Our regional-scale velocity model for study area I at 3 km depth . . . . 41
3.4 Map view of different catalogs in the Coso region . . . . . . . . . . . . 45
3.5 Distribution of Derivative Weighted Sum values for the Vp and Vp/Vs models in our study area I .................................................. 46

3.6 Distribution of the spread values for the final Vp model in study area II ................................................................. 47

3.7 Distribution of the spread values for the final Vp/Vs model in study area II ............................................................... 48

3.8 Checkerboard test for the Vp model ...................................................... 51

3.9 Checkerboard test for the Vp/Vs model ............................................. 52

3.10 Map view of the final Vp model for study area II ............................... 54

3.11 Map view of the final Vp/Vs model for study area II ............................ 55

3.12 Cross-sections of the Vp, Vp/Vs and Vs structures ............................ 58

3.13 Comparison of In-situ Vp/Vs ratios and the tomographic results ............ 60

3.14 Recovery test for the Vp/Vs anomaly in the CGF ................................. 63

3.15 Conceptual model of the subsurface structure for the Coso geothermal field .................................................................. 65

4.1 Tectonics and location of the Coso geothermal field ............................ 74

4.2 Tectonics and location of the Salton Sea geothermal field ..................... 75

4.3 Estimate of catalog magnitude completeness and b-value in the Coso area 78

4.4 Example of declustering catalog in the area of Rose Valley ................... 80

4.5 Spatial distribution of seismicity rate change and time series in the Coso area .................................................................. 83

4.6 Map view of the seismicity rate change between different time windows in the Coso area ................................................... 84
4.7 Monthly fluid injection and extraction volume within the Coso geothermal field .................................................. 86
4.8 Stress inversion results for seven subareas in the Coso area ........ 87
4.9 Map view of seismicity rate change and time series in the Salton Sea geothermal field and the vicinity ......................... 89
4.10 One-year time series in the Salton Sea geothermal field and its vicinity 90
4.11 Seismicity distribution and schematic model of stress states within the geothermal field .................................................. 92
List of Tables

2.1 Comparison of P-wave velocities in three models . . . . . . . . . 13

4.1 Statistical analysis for seven subareas . . . . . . . . . . . . . 83
Chapter 1

Introduction

To unravel the cause of earthquakes is the ultimate goal for seismologists. According to the stick-slip model (Brace and Byerlee, 1966), earthquakes occur when the loading stress in the fault zone exceeds the failure threshold. However, in the real solid earth, earthquakes distribute far from uniformly even in the local scale, due to large stress heterogeneity and various properties of geologic material. For example, while earthquakes are often located on planar faults, such as around the southern San Andreas Fault in southern California (Lin et al., 2007a), we have also observed a large amount of microearthquakes that are not aligned with well-defined surface fault traces. Besides the natural tectonic force, the anthropogenic activity can also produce earthquakes. To infer the causes of earthquakes in regions with active tectonics or human activity, it is necessary to understand high-resolution seismic subsurface structure and high-precision spatiotemporal distribution of earthquakes.

Seismic tomography has been used to understand the Earth’s interior for decades. It has contributed to study the deep Earth’s structure, such as the upper-mantle discontinuities and the subducting slab, and also the crustal structure. The simplified one-dimensional (1D) velocity model, Preliminary Reference Earth Model (PREM)
(Dziewonski and Anderson, 1981), has been used in different geophysical studies. In the last two decades, with the rapid increase of digital broadband stations and improved inversion algorithm, seismic tomography has been applied from the global to regional scale to approach different tectonic questions. High seismic activity is usually observed in areas with active tectonics, such as plate boundaries and Quaternary volcanoes. With enough seismic data recorded, it is possible to invert for reliable regional three-dimensional (3D) velocity models in these areas. Then we can infer the possible property of geological materials, such as the mineral composition, lithology and fluid content by investigating the lateral and vertical velocity anomalies and combining with other geophysical and geochemical studies.

Study of earthquakes is closely related to the seismic tomography, as the locations of earthquakes are coupled with the velocity structure. Although microearthquakes may not pose severe hazard as large earthquakes, study of microearthquakes can help us understand the physics of earthquakes, especially earthquake interaction. A large earthquake can generate aftershocks at the near field within one or two fault length and also trigger earthquakes at the far field with thousands of kilometers away. The near field triggering stems from the change of permanent stress loading, which is called static triggering, whereas the far field triggering, i.e. dynamic triggering is caused by change of transient stress carried by seismic surface waves. The static triggering has been well studied for a century since the theory of Omori’s law (Omori, 1895). However, the phenomenon of dynamic triggering was brought into attention only two decades ago and the mechanisms of dynamic triggering are not fully understood.
More constraints need to be added by accumulating various observations.

This thesis is focused on studying the seismic structure and microearthquakes through the methods of earthquake travel time tomography, earthquake relocation, and earthquake catalog analysis. I have been working in two areas with active microearthquakes: (1) the Puerto Rico region, which has been regarded as a microplate and undergoes the convergence from the North American plate to the north, and the Caribbean plate to the south; (2) the Coso area, southeastern California, with long-term geothermal production process and possible crustal magmatic system.

In Chapter 2, I focus on investigating the cause of active and diffuse seismicity in southwestern island of Puerto Rico. On the entire island, the seismicity is mainly distributed to one side of the Great Southern Puerto Rico Fault Zone but not confined to the main fault zone. To understand how the seismicity offsets the faults, I invert for the 3D velocity structure and relocate the earthquakes with both 1D and 3D velocity models. Our results reveal a high velocity body from the surface to 20 km depth beneath the major fault. A similar high velocity body is also found in the eastern island where a cluster of seismicity tends to occur at the boundary. These high velocity bodies corresponding to the plutonic rocks, composed of quartz diorite and granodiorite (Monroe et al., 1978), seem to separate regions with high seismicity and seismic quiescence. The delineated seismotectonic structure and the distribution of earthquakes suggest that the material in the southwestern island is weak, easy to break into failure, and different from the material in the north.

The research work of the Coso geothermal area is presented in Chapter 3 and 4.
With the net capacity of 270 MW, the Coso geothermal field is one of the largest geothermal fields in the USA and has been generating the electricity since 1987. The surface expression of hot springs, fumaroles, and lava dome suggests the geothermal field could be fed by a magmatic system. However, whether an active magmatic system still exists and how deep it could be if any has been the long-standing debate. Thus, one objective of my study is to image the fine-scale subsurface structure beneath the geothermal field and determine the location, geometry and state of the possible magma body. To this end, I invert for the 3D tomography model and re locate earthquakes based on the waveform cross-correlation and cluster analysis. Our results reveal an anomalous low velocity zone corresponding to the ductile behavior at depths from 6 to 12 km. Combining with other geophysical and geochemical studies, we explain this to be a frozen felsic magmatic system with the inclusion of water. We suggest that the pervasive melts are not likely to exist and the magmatic body can stand as the form of crystal mush.

With the advantage of the high-precision relocated catalog in the Coso geothermal field, I conduct a comprehensive study of remote triggering after large earthquakes, which is the main content of Chapter 4. Geothermal fields are regarded to be susceptible to dynamic triggering, mainly because of the high background seismicity rate and the presence of active geothermal fluid. To provide more constraints for mechanisms of dynamic triggering, I analyze the earthquake catalog and map the spatiotemporal distribution of remotely triggered earthquakes after the 1992 Landers earthquake. Our results show that dynamic triggering is absent in the actual geothermal area com-
pared to the rigorous response in the surround areas. In this chapter, I also present the results of dynamic triggering in the Salton Sea geothermal field after the 1999 Hector Mine earthquake. Similar results are obtained in these two geothermal areas. We explain the surprising observation in geothermal fields to be caused by the loss of pore pressure due to the long-term geothermal production process.
Chapter 2

Seismic structure beneath the island of Puerto Rico inferred from local earthquake tomography and relocation\textsuperscript{1}

\textsuperscript{1}This chapter will be resubmitted to the BSSA.
2.1 Summary

Sitting at the boundary of the Caribbean and North American plates, Puerto Rico is a seismically active zone of several historic large earthquakes and frequent small events. Here we focus on understanding the seismotectonic crustal structure of the island by obtaining and interpreting the three-dimensional (3D) compressional-wave velocity (Vp) and Vp/Vs models, and analyzing the distribution of relocated earthquakes. We first invert for one-dimensional (1D) Vp model for the island of Puerto Rico. With the minimum velocity model, we use the source-specific station term location method to relocate the seismicity on the island recorded by the Puerto Rico Seismic Network (PRSN) between 1986 and 2009. Using the obtained minimum 1D Vp model and the constant Vp/Vs ratio of 1.75, we invert for the 3D Vp and Vp/Vs velocity models. Two main features are shown by our 3D velocity models: (1) a 5% higher velocity body up to 20 km depth beneath the northwestern portion of the Great Southern Puerto Rico Fault Zone; (2) two higher Vp ($\sim 5\%$) and lower Vp/Vs ($\sim 3\%$) bodies beneath the areas with batholiths at the surface. Our 3D earthquake relocations show that the relocated earthquakes are distributed off these high velocity bodies. They may represent cool intrusive bodies, much harder than the surrounding wall rocks.

2.2 Background and Motivation

Puerto Rico lies at the boundary of the Caribbean and North American plates and accommodates motions relative to both plates (Figure 2.1). The North American
plate moves WSW towards Puerto Rico with 16.9±1.1 mm/yr (DeMets et al., 2000; Jansma and Mattioli, 2005). The excessive seismicity at the 19°N Fault Zone near the Puerto Rico Trench represents the relative motion between the subducting North American plate and the Puerto Rico microplate (McCann, 1985; Jansma and Mattioli, 2005). The Caribbean plate is moving towards the Puerto Rico microplate at a rate of 2.4 mm/yr (Jansma and Mattioli, 2005). However, whether the Caribbean plate subducts Puerto Rico is still controversial (Byrne et al., 1985; Dolan et al., 1998; Granja Bruña et al., 2010), due to lack of clear seismological evidence. The previous seismological analysis suggested a northward dipping zone related with the Caribbean plate at Puerto Rico but the diffuse seismicity pattern could only poorly define the slab (e.g., Dolan et al., 1998; Granja Bruña et al., 2010). In Dolan et al. (1998), the dip angle was estimated to be around 15°, but the proposed low-angle slab did not fit the gravity model (Granja Bruña et al., 2010).

Under active tectonic regimes, the island of Puerto Rico and the offshore area experience frequent seismicity. Four large (Mw >7.0) historical earthquakes have occurred since 1887. According to the record of the Puerto Rico Seismic Network (PRSN), more than 20,000 events were monitored between 1986 and 2009 and about four to seven earthquakes are reported to be felt (usually M ≥ 2.5) per month. Within the island, most of the seismicity is distributed over the southwestern part (Figure 2.1), which has been related to the faulting and the presence of serpentinite in previous studies (Mann et al., 2005; Huerfano et al., 2005). Two main strike-slip faults, the Great Southern Puerto Rico Fault Zone (GSPRFZ) and Great Northern Puerto
Rico Fault Zone (GNPRFZ), traverse the west and east of the island respectively (Figure 2.1). Some smaller on-land faults are hard to identify, probably due to high erosion rate or low rates of slip (McCann, 1985). From the current distribution of mapped faults and the diffuse seismicity, it is hard to attribute the seismicity in the southwestern portion to the faults. Especially near GSPRFZ, seismicity are not evenly distributed at two sides of the faults. In addition, the limited serpentinite belts cannot account for all the excessive seismicity.

Earthquake relocations and accurate velocity model often provide insights into probing deep structure of the lithospheric plate and help assess the seismic hazards. However, accurate three-dimensional (3D) velocity structure of the Puerto Rico region has not been published yet. In this study, we derive a reliable one-dimensional (1D) compressional-wave velocity (Vp) model for the island of Puerto Rico and invert for the 3D Vp and Vp/Vs models. With the improved velocity models, we relocate the seismicity beneath the island. Combining the subsurface velocity structure and the improved relative earthquake relocations in the island of Puerto Rico, we infer the possible cause of the unevenly distributed seismicity and help address whether the the arc island is subducted by the Caribbean plate.
Figure 2.1: Seismicity and station distributions in the vicinity of Puerto Rico. Dots denote the earthquakes between 1986 and 2009 and triangles represent stations used in this study. Two arrows indicate the plate movement directions relative to the Puerto Rico microplate measured by GPS data (DeMets et al., 2000; Jansma and Mattioli, 2005). Two major fault zones in our study area, the Great Northern Puerto Rico Fault Zone (GNPRFZ) and the Great Southern Puerto Rico Fault Zone (GSPRFZ), are marked by thick lines. Inset map shows the location of Puerto Rico, lying between the North American plate and the Caribbean plate.
2.3 Data and Method

2.3.1 Data Set

The Puerto Rico Seismic Network (PRSN) has been monitoring and processing earthquakes generated in the Puerto Rico region since 1974, and has been operated by the Geology Department of the University of Puerto Rico since 1987. The PRSN operates a network of short period and broadband stations throughout Puerto Rico (Figure 2.1). In our study area, a total of 21 stations are used consisting of 13 broadband stations and 8 short-period seismic stations.

The crustal depth for the island was estimated to be around 30 km from gravity and seismic data (Officer et al., 1957; Talwani et al., 1959; McCann, 2007). A total number of 2709 in-land crustal events are recorded by the PRSN between 1986 and 2009. These events have 14,598 P picks and 13,342 S picks in total. We then removed the picks with large travel time residuals (>1 s) and obtained 1072 events with more than 5 P and 5 S picks.

2.3.2 1D Model Inversion

In order to obtain a robust minimum 1D Vp model, we selected a subset for the inversion. To ensure the events are evenly distributed in space and have enough P picks, we adopted the criteria of 5 km spacing between master events along x, y, 2 km spacing in depth and each event with more than 9 P picks to obtain 265 events. The 1D Vp model inversion is based on the VELEST algorithm (Kissling, 1988; Kissling
et al., 1994, 1995). The inversion is solved by damped least square matrix, thus can be greatly dependent on the choice of damping parameters. We followed the method suggested by Aki et al. (1977) for the velocity damping parameter. After several tests on model variance and data variance, we chose optimal damping parameters for the velocity, event origin time and hypocenter. Because solving station corrections could impair the velocity structure, we set damping of station corrections in VELEST as 999 to keep them fixed. The PRSN daily operation model was used as the starting model for the inversion. We assigned 10 layers at -3, 0, 2, 5, 8, 14, 20, 25, 30, 35 km depths, which are relative to mean sea level. The inversion process is stopped when the velocity values for certain layers do not vary significantly and rms (root-mean-square) values of events are reduced by significant amount compared to the initial events (Kissling et al., 1994). The output minimum model for the island has data variance of 0.03 s and model variance of 1.69 km/s, compared to the starting model with data variance of 0.21 s and model variance of 3.18 km/s.

Overall, the minimum 1D model is smoother and slower than input model except at 8 km depth (Figure 2.2a). We tested the sensitivity of the final model to the starting model by constructing two other input models of faster and slower velocities than the PRSN model and comparing the difference of the output models (Figure 2.2b). Although at near surface and deepest depths, the output models are close to input models, from 5 to 30 km, three output models show the same trend and converge to similar results. The sensitivity test in Figure 2.2b demonstrates that the input model does not affect the minimum model given good quality data and optimal
Table 2.1: Comparison of P-wave velocities in three models

<table>
<thead>
<tr>
<th>Depth</th>
<th>McCann</th>
<th>PRSN</th>
<th>This study</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>6.45</td>
<td>4.30</td>
<td>4.43</td>
</tr>
<tr>
<td>2</td>
<td>6.45</td>
<td>4.84</td>
<td>4.97</td>
</tr>
<tr>
<td>5</td>
<td>6.45</td>
<td>5.44</td>
<td>5.53</td>
</tr>
<tr>
<td>8</td>
<td>6.45</td>
<td>6.00</td>
<td>6.40</td>
</tr>
<tr>
<td>14</td>
<td>6.45</td>
<td>6.80</td>
<td>6.86</td>
</tr>
<tr>
<td>20</td>
<td>7.13</td>
<td>7.60</td>
<td>7.21</td>
</tr>
<tr>
<td>25</td>
<td>7.13</td>
<td>7.70</td>
<td>7.56</td>
</tr>
<tr>
<td>31</td>
<td>8.01</td>
<td>7.81</td>
<td>7.80</td>
</tr>
</tbody>
</table>

The three models are the PRSN, the McCann and our minimum models. We show the comparison of P-wave velocity (km/s) at 8 layers.

parameters. According to our model, the crust for Puerto Rico Island is around 32±2 km, thicker than other continental crusts, which is consistent with the crust depth of 30 km derived from gravity and seismic refraction data (Talwani et al., 1959).

A previous 1D model for the island of Puerto Rico was obtained by McCann (2007), in which similar inversion algorithm was used but with different data. We chose the in-land events between 1986 and 2009, whereas McCann (2007) used events from 1974 to 2000, including offshore seismicity. The number of events in Puerto Rico increases significantly after 2006 and each year has almost double events than any year before, mainly because 15 more stations were deployed and earthquake detection algorithms have been improved by the PRSN since then. Compared to the layer-cake McCann model, our minimum 1D Vp model is continuous at adjacent layers and shows the same velocities at depths of 17 km and 31 km, which are the depth of two layers in the McCann model. Table 2.1 shows the actual values of velocity models.
Figure 2.2: Optimal 1D Vp and Vp/Vs models. (a) Comparison of the Puerto Rico Seismic Network (PRSN) model and our minimum model. (b) Sensitivity tests on different starting models. The solid lines mark three different input models and dashed lines represent three output models. (c) Distribution of S-P travel time versus P-wave travel time recorded from the PRSN catalog. The best-fitting value of 1.75 is used as starting model for 3D Vp/Vs inversion.
2.3.3 1D Earthquake Relocation and Relative Location Error

By using the minimum 1D model, we implemented the relocation process by running the COMPLOC package *Lin and Shearer* (2006), which calculated the source-specific station terms (SSST) (*Richards-Dinger and Shearer*, 2000; *Lin and Shearer*, 2005) to smooth the residuals from nearby events to mitigate the anomalies caused by velocity structure variation. A total of 1072 events from 1986 to 2009 are relocated. Because the heterogeneity of velocity model can be larger at longer distance, particularly when 1-D velocity model is used, we performed the station-event distance cutoff tests by choosing different cutoffs to compare if the depths of events change significantly and determined 100 km as the optimal cutoff of event-station distance. Events were chosen with depth between 0 and 32 km, which is the crust depth suggested by our minimum 1D velocity model. 10 iterations were conducted to compute static station terms and 20 for SSST. One thousand nearby events and 50 km distance cutoff were chosen for the first iteration of the SSST computation and 20 nearby events and 5 km distance cutoff for the last iteration. The average rms of travel time residuals is reduced from 0.34 s to 0.19 s after SSST relocation. However, a relatively smaller value of rms does not mean a small error in earthquake locations, because picking errors can also affect the locations.

Since the true locations are unknown, we estimated the relative location errors. The relocations are obtained by the grid search algorithm and it is difficult to estimate the error ellipses as in the classical least square methods. We simulated picking errors as Gaussian distributed random noises and added them to perturb the phase picks
at best-fit locations and we relocated the events with perturbed picks, repeating the previous relocation procedures. The mean of P and S picks travel time residuals are close to 0 and standard deviation are 0.13 and 0.14. The Gaussian noises were randomly chosen with the mean of noise 0 and standard deviation 0.13, 0.14 s to perturb P and S picks respectively. By estimating the relative locations of scatters around the best-fit locations, we can understand how much picking errors are involved. To estimate the uncertainties for relative locations, we advocated the bootstrap method (Efron, 1979) to resample 100 times. In the statistic sense, the true information will be biased if resampling number is too large (e.g. 1000) and information is not reliable if number too small (e.g. 10).

2.3.4 3D Model Inversion

To obtain 3D Vp and Vp/Vs model, we implemented the simul2000 algorithm Thurber (1983); Thurber and Eberhart-Phillips (1999) to invert for the 3D model and the earthquake relocation simultaneously. Our minimum 1D Vp model was used as the starting Vp model. We plotted the Wadati diagram (Kisslinger and Engdahl, 1973) to find the best-fitting Vp/Vs ratio to be 1.75 (Figure 2.2c), which was used for the starting Vp/Vs model. To ensure high quality events to be inverted, we chose master events with more than 7 P picks and 7 S picks. We obtained 564 master events from the original 2709 in-land crustal events. By running a series of inversions with different grid sizes, we chose the final grid spacing at 20 km,15 km, and 5 km along x, y, and depth respectively. The distribution of grid nodes and our master events can be seen
Figure 2.3: (a) Map view of events used for inversion. Each event includes at least 7 P picks and 7 S picks (red dots). Black diamonds denote the grid nodes in the inversion. (b) Map view of all 2709 events from 1986 to 2009 recorded by the PRSN. (c) Depth distribution of events used for inversion and grid nodes. The symbols are the same as (a).

Since the inversion is damped least-squares, the damping value is an important parameter especially for relatively non-uniform data. We used different damping parameters for inversions and chose the optimal damping values based on the trade-off curve of data misfit and model variance (Eberhart-Phillips, 1986). According to the trade-off curve shown in Figure 2.4, we chose 100 and 70 as the optimal damping values for Vp and Vp/Vs inversions, respectively.

Because our goal is to achieve reliable 3D velocity models, it is essential to assess the model resolution. The most direct way to estimate the resolution is the Derivative Weighted Sum (DWS), as they indicate the weighted ray density. Figure 2.5 shows
Figure 2.4: Trade-off curves of normalized data misfit and model variance for Vp and Vp/Vs. The optimal damping values for Vp and Vp/Vs are chosen to be 100, 70 respectively.

that the island has high number of DWS values (>2000) for P rays at the depths of 5, 10, and 15 km, especially in the southwestern portion. S-P rays have similar distribution as shown in Figure 2.6. Thus, the ray density is enough for us to resolve the subsurface structure, especially in the southwest.

Along with the inversion process, the resolution matrix is also computed and each row contains the information for each model parameter. Thus, the Diagonal elements of resolution matrices (RDE) is another measure to help assess the resolution for each model parameter. The RDE values are dependent on grid spacing and damping values. The values greater than 0.1 is often regarded to represent good resolution, which can also be seen from the following synthetic tests.

We performed checkerboard test to evaluate if the chosen parameter and data can recover the true velocity model. We constructed checkerboard patterns following the current grid spacing, with every 20, 15 km along x and y. Positive and negative 5% velocity perturbation was added to 1D model. Using the synthetic travel times, we
Figure 2.5: Distribution of Derivative Weighted Sum (DWS) values for P rays at the top six layers in our study area. The high numbers of the DWS values mainly occur at the depths of 5, 10, and 15 km, especially in the southwest of the island.
Figure 2.6: Distribution of Derivative Weighted Sum (DWS) values for S-P rays at the top six layers in our study area. The high numbers of the DWS values mainly occur at the depths of 5, 10, and 15 km, especially in the southwest of the island.
followed the same parameterization in the real data set to invert for the 3D $V_p$ and $V_p/V_s$ models. The output models are shown in Figures 2.7 and 2.8. Similar to what the DWS values indicated, most part of the island has recovered the true model at the depths of 5, 10, and 15 km. We also added the RDE contour lines with values greater than 0.1, which correlate with the well-resolved checkerboard patterns (Figures 2.7 and 2.8). Thus, we use the RDE contour line to represent well-solved areas in the following figures.

2.4 Results

2.4.1 Final $V_p$ and $V_p/V_s$ Models

We show our final 3D $V_p$ and $V_p/V_s$ models from 0 to 25 km depth (Figures 2.9 and 2.10). With limited resolution at the surface, the map view of $V_p$ model reveals two high velocity bodies in the north and east of the island (Figure 2.9), which correspond to the presence of batholiths (Bawiec, 1998). Except the areas of batholith, the zones between the the major two faults, GSPRFZ and GNPRFZ, show low $V_p$ anomalies at the depths of 5 and 10 km. A striking high velocity body (up to 5%) appears beneath the GSPRFZ from 5 to 20 km depth. It begins to be smeared by a low $V_p$ anomaly in the southwestern part of the island from the depth of 15 km. The low $V_p$ anomaly continues to the deeper depth of 25 km, although there is a lack of good resolution due to limitation of data.

Since the numbers of P and S picks are comparable, $V_p/V_s$ model has similar
Figure 2.7: Checkerboard test of Vp model at the top six layers. White contour lines denote the diagonal elements of resolution matrices (RDE) values greater than 0.1.
Figure 2.8: Checkerboard test of Vp model at the top six layers. White contour lines denote the diagonal elements of resolution matrices (RDE) values greater than 0.1.
resolution quality as the Vp model (Figure 2.10). The Vp/Vs is less heterogenous than the Vp model as we have shown above. At the upper 10 km, corresponding to the low Vp, low Vp/Vs anomalies occur between the GSPRFZ and GNPRFZ. At the greater depth, Vp/Vs ratio becomes more homogeneous and shows higher values than the background 1.75.

To better investigate the velocity anomalies we have observed from map views, we plot cross sections beneath three profiles (Figure 2.11). One of them goes along the GSPRFZ, and the others across the southwest and the east of the island. From the cross sections (Figure 2.11a), we can see that a high velocity body occurs beneath the GSPRFZ. This anomaly in the near surface features low to normal Vp/Vs. In the northwest, the anomaly can be seen till deeper than 25 km (Figure 2.11b). The relocated earthquakes with 3D velocity models are also added to the cross sections. The seismicity in the upper crust seems to occur off the high velocity body. Similar feature is also seen in the east of the island. Although the seismicity is relatively sparse in the east, some swarms occur below the high velocity body (Figure 2.11c). The two velocity bodies with high Vp, low to normal Vp/Vs correspond to the presence of batholiths. It is possible that these igneous rocks are hard to break and earthquakes occur off the batholiths.

### 2.4.2 Earthquake Relocation

We obtained 1D relocation of 1072 earthquakes with more than 5 P and 5 S picks by using the SSST method. For each pair of two events, the standard errors of relative
Figure 2.9: Map view of the final 3D Vp model for the entire island. Vp perturbations are shown relative to the layer-average values. White lines denote the RDE values greater than 0.1.
Figure 2.10: Map view of the final 3D Vp/Vs model for the entire island. White lines denote the RDE values greater than 0.1.
Figure 2.11: Cross-sections of the final 3D Vp and Vp/Vs models. The three profiles are shown by the black lines in (A). Black dots represent 3D relocation. White lines denote the RDE values greater than 0.1.
locations are computed in horizontal and vertical directions. The mean value for horizontal and vertical locations with minimum 1D model is 42 m and 71 m. The horizontal errors fall in the range of 0−0.3 km while a majority of vertical errors are in the range of 0−0.5 km. We also simultaneously inverted for the 3D relocation of these 1072 events after we obtained the final 3D velocity models.

The comparison of epicenters from the PRSN catalog, relocated seismicity using the minimum 1D model, and relocated seismicity using 3D model are shown in Figure 2.12. We can see that earthquakes in the west occur in all the depths, whereas seismicity forms swarm-like features in the east of the island. Both the 1D relocation and 3D relocation show more liner features than the PRSN catalog, such as a swarm in the east of the island around 5 km depth. Although the relative locations of the 1D relocations have some similar features as the 3D relocations, the 3D locations show different absolute locations.

### 2.5 Concluding Remarks

In this study, we inverted for minimum 1D Vp models for the island of Puerto Rico. The new continuous 1D model for the island agrees with a previous layer-cake model, but shows different structure from the PRSN model. Using the minimum 1D model, we relocated the earthquakes on the island between 1986 and 2009. For these relocated earthquakes, the mean standard errors of relative locations reach tens of meters. We also inverted high quality data for 3D Vp and Vp/Vs models. Our 3D velocity models and relocations provide insights into the subsurface structure within well resolved
Figure 2.12: Depth distribution of earthquake locations. Comparison of the PRSN catalog, 1D relocation and 3D relocation.

areas beneath the island of Puerto Rico.

An important tectonic problem with Puerto Rico is whether the active subduction from the Caribbean plate exists. From the cross sections (Figure 2.11b), we can see the seismicity occurs in a dipping plane, which has low Vp and normal Vp/Vs. Previous studies have used this dipping feature to argue the subduction exists (Mendoza and McCann, 2005). However, we cannot rule out the other possibility that the material in the northwest is harder than in the southwest in the upper 20 km depth, and the earthquakes occur below the hard material in the north, thus forming dipping feature from south to north. Our map view shows that the seismicity is mostly concentrated at the southern side of the GSPRFZ (Figure 2.11A). We detected a large high velocity anomaly up to 20 km depth beneath the northern side of the GSPRFZ, where batholith is found at the surface. Thus, the high velocity body could possibly
represent the plutonic rocks, hard to break into failure, whereas the materials in the southwest is relatively weak and seismicity occurs in all depths. To better address the question whether the Caribbean plate subducts the island of Puerto Rico, we need to investigate whether the dipping plane continues to greater depth, especially below 100 km. Since this study has focused on imaging the velocity structure and relocating the earthquakes in the crust, to detect the deeper structure will be included for the future work.

2.6 Data and Resources

The seismic data in this study are obtained from the Puerto Rico Seismic Network (PRSN). The maps are created by using the public domain Generic Mapping Tools (GMT) (Wessel and Smith, 1991) and MATLAB. The VELEST program by Kissling et al. (1995) is used for the inversion of the minimum 1D velocity model. Earthquakes are relocated by applying the COMPLOC program (Lin and Shearer, 2006). The algorithm of simul2000 Thurber (1983) is used for 3D model inversion.

2.7 Acknowledgments

We thank C. Mendoza for the Puerto Rico seismic tomography report and helpful discussions. We give our special thanks to Amanda Piggot and Monica Arienzo for English grammar revision. This work is supported by the National Earthquake Hazards Reduction program, under USGS awards G10AP00020 and G10AP00021.
Chapter 3

Three-dimensional Vp and Vp/Vs models in the Coso geothermal field, California: seismic characterization of the magmatic system.¹

¹This chapter has been published in the Journal of Geophysical Research.
3.1 Summary

We combine classic and state-of-the-art techniques to characterize the seismic and volcanic features in the Coso area in southern California. Seismic tomography inversions are carried out to map the variations of $V_p$, $V_s$ and $V_p/V_s$ beneath Coso. The velocities in the top layers of our model are correlated with the surface geological features. The Indian Wells Valley, with high silica content sediment strata shows low-velocity anomalies up to 3 km depth, whereas the major mountain ranges, such as the south Sierra Nevada and the Argus Range show higher velocities. The resulting three-dimensional velocity model is used to improve absolute locations for all local events between January 1981 and August 2011 in our study area. We then apply similar-event cluster analysis, waveform cross-correlation, and differential time relocation methods to improve relative event location accuracy. A dramatic sharpening of seismicity patterns is obtained after using these methods. We also estimate high-resolution near-source $V_p/V_s$ ratio within each event cluster using the differential times from waveform cross-correlation. The in situ $V_p/V_s$ method confirms the trend of the velocity variations from the tomographic results. An anomalous low velocity body with low $V_p$, $V_s$, and $V_p/V_s$ ratios, corresponding to the ductile behavior underlying the Coso geothermal field from 6 to 12 km depth, can be explained by the existence of frozen felsic magmatic materials with the inclusion of water. The material is not likely to include pervasive partial melt due to a lack of high $V_p/V_s$ ratios.
3.2 Background and Motivation

The Coso geothermal field (CGF) is located between the Sierra Nevada batholith and the Basin and Range Province in southeastern California (Figure 3.1a). As one of the largest geothermal fields in USA, it has been exploited to generate power through over 100 production wells since 1987 (Adams et al., 2000). The maximum heat flow of the geothermal field was estimated to be 10 times of the value from the background Basin and Range (Combs, 1980) (Figure 3.1b). A crustal magma body has been assumed to provide the primary heat source for the present surface geothermal system (Combs, 1980; Bacon et al., 1980; Duffield et al., 1980).

Several lines of evidences favor the existence of a magma body beneath Coso. First, Coso is in a transtensional tectonic regime and the extension facilitates the ascent of magma. Coso sits in an extensional stepover between the dextral striking faults to the north and south. The geothermal field is surrounded by three main valleys with a series of strike-slip and normal faults (Reasenberg et al., 1980). Lithospheric extension causes crustal thinning beneath the geothermal field, which favors the intrusion of basalt from the mantle into the crust, sustaining a long-lived magma reservoir in the crust (Duffield et al., 1980). The crustal thinning at Coso is supported by the geochemical result of the isotopic composition of rocks, which are strongly influenced by the asthenosphere (Monastero et al., 2005).

The geothermal field is characterized by young volcanic rocks, mainly Pleistocene rhyolite dome, flanking basalt flows and cinder cones. The active volcanoes erupted around 4 Ma and the Pleistocene bimodal eruptions continued from 1.04 to 0.03 Ma,
while the latest active volcanism is estimated to be 0.03 Ma (Duffield et al., 1980; Bacon, 1982; Manley and Bacon, 2000). It is believed that the magmatic activity beneath the dome field triggered the Coso geothermal system around 0.2 Ma (Adams et al., 2000; Duffield et al., 1980). Combs (1980) showed that the geothermal field is associated with 38 rhyolite domes, surface hydrothermal manifestations, and higher heat flows than the surrounding area. The highest heat values correspond to the geological features, such as the Sugarloaf Mountain (SM, the largest rhyolite dome in the area), the Devil’s Kitchen (DK) and the Coso Hot Springs (CHS) with the presence of hot springs and fumaroles. The present geothermal production area lies in a $6 \times 10 \text{ km}^2$ north-south trending zone between the SM and CHS (Bishop and Bird, 1987; Feng and Lees, 1998; Fialko and Simons, 2000). The well log data up to 500 m show that the high-temperature system has been developed beneath the flank of the present production area, with temperatures up to $328^\circ\text{C}$ (Adams et al., 2000). The production wells are concentrated above 3 km depth. To 10 km south of the SM and CHS, the area has not been exploited widely, although the heat flow value is as high as 120 mW/m$^2$ (Figure 3.1b).

Exploring the existence of a residual magma chamber beneath the geothermal field has been the focus of numerous geophysical, geochemical, and geological studies (e.g., Duffield et al., 1980; Wu and Lees, 1996; Manley and Bacon, 2000; Lees, 2002; Wilson et al., 2003; Monastero et al., 2005). The extensive studies reveal that a long-lived rhyolitic magma reservoir presumably emplaced around 0.03 Ma and provided the heat flux for the overlying hydrothermal system. However, the accurate location of
Figure 3.1: Geologic map of the Coso area. It is bounded on the north by the northwest-striking Owens Valley Fault while in the south dominated by the dextral Little Lake fault zone (LLFZ) and Airport Lake fault zone (ALFZ). (a) Map showing tectonic features and our study areas I and II. Blue dots represent the earthquakes between 1981 and 2011. Stations are shown by the red triangles. Black lines denote Quaternary faults. Our study area II is marked by the red box. In the inset map, red star shows the location of the Coso area in California and the rectangle shows our study area I. (b) A close-up of the study area II. Black dots denote the grid nodes for inversion. Blue dots are the master events for the tomographic inversions. Diamonds are colored by the heat values from Combs (1980) and Saltus and Lachenbruch (1991). High heat flow values within the geothermal field are present in the Sugarloaf Mountain (SM), the Devil’s Kitchen (DK) and the Coso Hot Springs (CHS). The geothermal field is surrounded by three main valleys, the Rose Valley (RV), the Coso Wash (CW) and the Indian Wells Valley (IWV). Other abbreviations are SSNFZ, Southern Sierra Nevada Frontal Fault Zone; WCF, Wilson Canyon Fault; and AHF, Ash Hill Fault.
the magma reservoir varies between different studies. Most previous geochemical and petrological studies suggest that the top of the magma begins below 6 km depth, based on analyses of mineralogy and ages of erupted volcanic rocks (e.g., Duffield et al., 1980; Bacon et al., 1981; Manley and Bacon, 2000). The gravity data also suggest a low resistivity and low density zone 5 km beneath the CGF (Wamalwa et al., 2013). With the advantage of mapping the subsurface structure, seismic studies have also pointed out that the anomalous body exists around 6 km depth from the teleseismic receiver function analysis (Wilson et al., 2003) and the ambient noise tomography (Yang et al., 2011). However, these large-scale studies failed to map the lateral variation of the anomalies on a fine scale comparable to the geothermal field. The local earthquake body-wave tomography study by Hauksson and Unruh (2007) also detected a low P-wave velocity zone ranging from 5 to 10 km depth beneath Coso, but they argued that the slightly varied ratios of the P- and S-wave velocity (Vp/Vs) cannot support the existence of a magma body shallower than 10 km.

In this study, we apply local earthquake body wave tomography to invert for three-dimensional (3-D) P-wave velocity (Vp) and Vp/Vs model in the Coso area. The objective is to image the high-resolution subsurface velocity structure underneath the Coso geothermal field and outline the location, geometry, and depth of the magma body. Combining these models with other geophysical studies, we try to understand how the geothermal and magmatic systems operate to generate the surface manifestations. By analyzing our tomography models and distribution of earthquake relocations, we aim to determine the accurate location of the magma body and answer
the unsolved questions related to the magmatic system, such as the upper bound of
the reservoir depth, the interface between silicic and mafic magma, and whether a
large magma reservoir serves as the main heat source or several small heat sources
exist.

3.3 Data and Methods

3.3.1 Data Set

We develop the regional and local velocity models for Coso (Figure 3.1) using data
from the Southern California Earthquake Data Center (SCEDC). The regional study
area I includes Coso and the adjacent areas, such as the Garlock Fault, the Sierra
Nevada, and the Death Valley, and is referred as the Coso region. Study area II
focuses on the vicinity of the Coso geothermal field. We obtain P- and S-wave first
arrival times and waveform data in our study area I from 159,295 events between
January 1981 and August 2011 from the SCEDC. Seismic stations are more densely
distributed in the vicinity of the geothermal area than other parts of the study area I,
which ensures the resolution and reliability of the tomographic models for our study
area II (Figure 3.1a). The completeness magnitude for the entire study area II is
estimated to be M 1.3, and the maximum magnitude is Mw 5.75. The waveform data
are resampled at 100 Hz sample rate and filtered by applying a bandpass 1-10 Hz
for waveform cross-correlation. These data processing steps are similar to those in
previous studies for southern California (e.g., Shearer et al., 2005; Lin et al., 2007b).
In local earthquake tomography, well-recorded and evenly distributed events are usually used in the inversion. We choose 1,893 master events for study area I from the entire data set by applying the criteria of 5 km spacing in horizontal plane and 2 km in depth between master events and each event having more than 14 P and 9 S picks. As a result, a total number of 40,859 P and 20,321 S-wave phase picks are selected for the inversion for the regional model. To invert for a finer model for study area II, we select 1263 master events with 25, 248 P picks and 13, 884 S picks with the criteria of 2 km spacing in horizontal and 1 km in depth between master events, and each master event having more than 14 P picks and 9 S picks (Figure 3.1b).

### 3.3.2 One-dimensional Starting Model

Three-dimensional velocity model inversions typically start with one-dimensional (1-D) models. We use the layer-average velocity model for southern California by Lin et al. (2007b) as the starting model and invert for the minimum 1-D model for Coso by using the program VELEST (Kissling et al., 1994, 1995). The model damping parameters are chosen based on the tradeoff tests of optimizing data misfit and model variance. We set the station corrections to be zero during the inversion to avoid tradeoff between the velocity structure and station corrections. The model variance from our minimum model is decreased by 35% from the starting model and the data variance is reduced by 26%. Hauksson and Unruh (2007) also inverted for a 3-D velocity model for Coso using with the average model of Hauksson (2000) for southern California as the 1-D starting model. Compared with the 1-D model by Hauksson...
Figure 3.2: Comparison of different 1-D velocity models. Blue line shows the model by Hauksson (2000) for southern California, which was used as the starting model to generate the 1-D Vp model for the Coso region by Hauksson and Unruh (2007) shown by the black line. Our minimum Vp model for the Coso region denoted by the red line is produced by using the model of Lin et al. (2007b) for southern California as the starting model (pink one).
and Unruh (2007), our minimum model is 3-14% faster above 9 km depth, which mainly results from distinct starting models, data coverages, and different inversion parameters. These 1-D models are shown in Figure 3.2.

3.3.3 Three-dimensional Tomographic Inversion

The goal of our study is to obtain an accurate velocity model for interpreting the crustal structure beneath the geothermal field. The inversion for Vp/Vs ratios, which is indicative of both lithology and rheology of subsurface materials, depends on both P and S-wave ray paths. Since there are fewer S-wave picks than P-wave picks and the quality of S-wave data is not as good as P wave, Vs models are usually poorly resolved compared to Vp and the method of deriving Vp/Vs ratio from Vp divided by Vs is not reliable (Thurber and Eberhart-Phillips, 1999). In this study, we solve the Vp/Vs model directly by using the S-P travel time differences. In the simul2000 algorithm (Thurber, 1983; Thurber and Eberhart-Phillips, 1999), ray paths are selected from the fastest travel time between the source and receiver and calculated by approximate ray tracing. These ray paths are curved non-planar (Eberhart-Phillips and Michael, 1998) and the S ray paths are approximated by P paths. The algorithm is a damped least-squares inversion and the optimal damping parameters are chosen based on the trade-off curve of data misfit and model variance (Eberhart-Phillips, 1986). We run a series of inversions using different damping parameters and choose the damping values of 150 for Vp, 80 for Vp/Vs in the inversion for the regional model.

We first solve the velocity model for study area I and then invert for the final 3-D
Figure 3.3: Our regional-scale velocity model for study area I at 3 km depth. Vp perturbations are shown relative to the layer-average value. The thick black line encloses the area with the diagonal element of the resolution matrix greater than 0.1, which are considered well-resolved. Most of the observed velocity anomalies correlate well with the known geological features. For example, the Sierra Nevada, the Argus Range, the Granite Mountains and the core of Mojave Desert are represented by high-velocity anomalies, whereas the Eastern California Shear Zone and the Indian Wells Valley show low-velocity anomalies. Coso is located at the north end of the Indian Wells Valley.
velocity model for study area II, the vicinity of the CGF. The inversion follows the two steps, 1) start with the minimum 1-D model of the Coso region and invert for the 3-D Vp and Vp/Vs models in study area I with a coarser uniform horizontal grid spacing of 6 km; and 2) use the resulting 3-D model for the Coso region to invert for the final 3-D model of the CGF (study area II) with a finer horizontal spacing of 3 km. By doing this, the final model for the CGF shows more detailed velocity anomalies than the regional model. A constant Vp/Vs ratio of 1.73 is used for the starting Vp/Vs model inversion for the region based on the Wadati diagram (Kisslinger and Engdahl, 1973). The depth distribution of seismicity suggests that the seismicity is focused in the upper 15 km depth, and we set up the depth layers at -5, 0, 3, 6, 9, 12, 15, 20, and 25 km. Note that in this study all depths are relative to mean sea level.

3.4 Results

3.4.1 Regional Coarse Model

The regional model is obtained after 6 iterations when the reduction of data variance becomes insignificant. Compared to the initial models, the data variance is reduced by 69% and 45% for final Vp and Vp/Vs models, respectively. The root-mean-square (RMS) of the travel time residuals is reduced from 0.24 to 0.11 s. We show a representative velocity image at 3 km depth in Figure 3.3 with the Vp perturbations relative to the layer-average value of the inverted model. The inverted Vp anomalies reflect the near-surface geological features. The Indian Wells Valley and the Eastern California
Shear Zone mainly exhibit low Vp. The mountain areas such as the Sierra Nevada, the Argus Range, and the Granite Mountain with the Mesozoic granitic rocks, and the stable Mojave Desert show high Vp anomalies.

### 3.4.2 Earthquake Relocation

After obtaining the regional 3-D velocity model, we invert for the 3-D earthquake relocations in the simul2000 by fixing the velocity model. The absolute earthquake relocations have been improved by taking into account the biasing effect of velocity structure. To improve the accuracy of relative earthquake locations, we implement the waveform cross-correlation, similar event cluster analysis, and differential time relocation methods by Lin et al. (2007a) to relocate all the events in the Coso region.

Both the absolute and relative locations are improved compared to the SCSN catalog locations. The improvement is demonstrated by the sharpening of the relocated seismicity in Figure 3.4. Our relocations are consistent with the previous relocation catalogs for southern California, such as the SHLK catalog (Shearer et al., 2005), the LSH catalog (Lin et al., 2007a), and the latest HYS location catalog by Hauksson et al. (2012) for events between 1981 and 2011. We use similar criteria for cross correlation to those for the LSH catalog such as the correlation coefficient cutoff, station-event distance range, and minimum average of the maximum correlation coefficient. The main difference between our catalog and previous ones stems from the different absolute locations and we use the produced 3-D locations from simul2000 (Figure 3.4b). Our catalog chooses 0.65 as correlation coefficient cutoff and encom-
passes the highly correlated local events in Coso, whereas the HYS catalog uses the correlation coefficient cutoff of 0.6.

Quantitive estimates of location uncertainties indicate that both absolute and relative location accuracies are significantly improved. The absolute location error estimates are provided by the simul2000 algorithm (Thurber and Eberhart-Phillips, 1999). The mean horizontal uncertainty is 120 m, and 300 m for the vertical. The relative location uncertainties are estimated by a bootstrap method (Efron and Gong, 1983), similar to Lin et al. (2007a). After resampling for 15 times, we obtain the median of the relative location uncertainties of 11 m in horizontal, and 22 m in depth. In study area I, about 66% of events fall into 1,225 clusters with at least 5 events. Three distinct clusters with more than 5,000 events are observed from the distribution of the relocated hypocenters within the Rose Valley, to the east of the CGF, and around the Airport Lake Fault Zone, respectively.

3.4.3 Model Resolution Tests

Several parameters are used to examine the model quality, which mostly depends on the geometry and density of rays. Ray-dependent measurements and synthetic tests are conducted together to assess the resolution of the tomographic models. In this paper, we focus different resolution estimates on our study area II. We plot resolution estimates to assess the ray coverage including the Derivative Weighted Sum (DWS) values, diagonal elements of resolution matrices (RDE) and spread function. The DWS values indicate the weighted ray density directly. The weight scheme is based
Figure 3.4: Map view of different catalogs in the Coso region. The red dots represent the common events that are relocated by different methods. (a) Southern California Seismic Network (SCSN) catalog, covering events in Coso from 1981 to 2011. Grey dots show the events that are not relocated but remain in the original SCSN catalog. (b) 3-D relocations produced by the simul2000 algorithm (Thurber, 1983; Thurber and Eberhart-Phillips, 1999). (c) HYS catalog, the relocation catalog for southern California between 1981 and 2011 by Hauksson et al. (2012). (d) Highly correlated events in this study. See Figure 1 for the abbreviations.
Figure 3.5: Distribution of Derivative Weighted Sum (DWS) values for the Vp and Vp/Vs models at the top four layers in our study area I. The high numbers of the DWS values mainly occur in our study area II.
Figure 3.6: Distribution of the spread values for the final Vp model in study area II. Colored nodes denote spread values below 3. Values greater than 3 are not shown. Black lines denote the diagonal elements of the resolution matrix greater than 0.1, which are also shown in Figures 6-11.
Figure 3.7: Distribution of the spread values for the final Vp/Vs model in study area II. Colored nodes denote spread values below 3. Black lines denote the RDE values greater than 0.1.
on the distance of rays from each node. Based on the distribution of DWS values (Figure 3.5), high numbers of the DWS values (>2000) with the maximums of 16,000 for P rays and 10,000 for S-P rays are present in the vicinity of the Coso geothermal field and the adjacent areas, which make up of our study area II. The deficiency of the DWS measurement is that the ray directions are not taken into account, thus smearing cannot be estimated.

The model resolution matrix gives the information for all the nodes and each row represents the averaging vector of a model parameter. The RDE reflects the resolution for each node and provides relative measures of the ability of the data for detecting anomalies in different locations. Resolutions greater than 0.1 are considered to indicate good quality of the models in this study based on the synthetic data tests. However, the RDE values depend mainly on the grid spacing and damping parameters.

The spread function investigates the dependence of a model parameter on the other grid nodes. Ideally, the velocity at each node is independent of other nodes, but common ray geometries can link neighboring model parameters and can lead to artificial smearing of anomalies across multiple nodes. Thus, the spread value will be zero in an ideal case. For the real data sets, we regard spread function values smaller than 3 as good values indicating that the peaked resolution is achieved at the grid and the distant nodes have no significant contribution. We choose the cutoff of 3 to represent low spread values with good resolution because the distributions of the spread function values smaller than 3 for the Vp and Vp/Vs models are consistent with the RDE contours with resolutions greater than 0.1 (Figures 3.6 and 3.7). The
cutoff of 2.5 or 3 has been chosen by previous tomography studies (e.g., Sherburn et al., 2006; Reyners et al., 2006) to show the nodes without too much smearing.

Resolution tests, i.e., checkerboard tests in this study, are also performed to compare with the ray-dependent measurements. Checkerboard models are constructed to assess the amount of image blurring. Five percent of synthetic velocity perturbations are assigned to blocks with dimensions of $6 \times 6$ km at all layers. The synthetic travel times are then inverted to recover the velocity anomalies using the same parameterization method as in the real data set. The synthetic anomalies are reconstructed well underneath the geothermal field, especially at depths of 3 km and 6 km for both $V_p$ and $V_p/V_s$ models (Figures 3.8 and 3.9). The RDE contour lines with resolutions greater than 0.1 are correlated with the well-resolved checkerboard patterns. For $V_p/V_s$, the ability of reconstructing checkerboard patterns is similar to that of $V_p$.

We assess the reliability of the velocity features by considering the RDE, spread values and well-resolved checkerboard patterns. The area around the geothermal field shows reliable resolutions, especially at depths of 3 and 6 km. The RDE contour lines correlate mostly with the spread values and checkerboard test results. In the following map views and cross-sections, we use the RDE contour lines of 0.1 to enclose the well-resolved area of the velocity structure.

3.4.4 Map Views of Final Velocity Models

With the regional 3-D model as input, we invert for the velocity structure in study area II. Figures 3.10 and 3.11 show the velocity images above 9 km depth, below which
Figure 3.8: Checkerboard test for the Vp model with blocks of 6×6 km. The velocity perturbations relative to initial 1-D model are shown in grey scale. (a-d) The true velocity model. (e-h) The inverted velocity model. White lines denote the RDE values greater than 0.1.
Figure 3.9: Checkerboard test for the Vp/Vs model with blocks of 6×6 km. (a-d) The true velocity model. (e-h) The inverted velocity model. White lines denote the RDE values greater than 0.1.
the seismicity is sparse and rays are insufficient to resolve the velocity structure well. The lateral heterogeneities are large at near-surface depths of 0 and 3 km for both Vp and Vp/Vs. At the surface, a notable feature of low Vp (ranging between 3.6 and 4.6 km/s) is observed within and around the CGF. The nearby faults and valleys show low Vp, such as the Little Lake Fault Zone (LLFZ), the Airport Lake Fault Zone (ALFZ) and the Indian Wells Valley (IWV). The intensely high Vp anomalies lie along the Southern Sierra Nevada Frontal Fault (SSNFZ) and the Argus Range. At 3 km depth, the southeastern part of the CGF exhibits high Vp. Low Vp values are prominent in the upper 3 km near the IWV and the adjacent LLFZ and ALFZ. The ALFZ forms pull-apart basins and cuts the IWV (Monastero et al., 2005). The low Vp anomaly is consistent with the geological study that the 3 km high-silica-content sediment exists beneath the Valley (Monastero et al., 2002). Below 6 km depth, the Vp lateral variations become small and a relatively low Vp is apparent at the CGF.

Spatial correlation between the Vp and Vp/Vs anomalies is observed for most parts of our study area II. In the Argus range with strong high Vp, high Vp/Vs anomaly is seen at the surface. However, we do not see high Vp/Vs corresponding to the high Vp beneath the SSNFZ. There is a variation of Vp/Vs anomalies across the CGF in the upper 3 km depth. The northern part of the geothermal field shows low Vp/Vs whereas the other areas within and proximity to the CGF show high Vp/Vs. The model in the vicinity of the CGF mainly shows low Vp/Vs features below 6 km depth, but a notable high Vp/Vs body (1.78-1.80) is observed around the ALFZ at 6 km depth.
Figure 3.10: Map view of the final Vp model for study area II. Vp perturbations are shown relative to the layer-average values, which are also given for each slice. Dashed circle represents the location of the Coso geothermal field. Black dots show the grid nodes used in the tomographic inversions. Two red stars represent the geologic sites, SM and CHS. White lines denote the RDE values greater than 0.1. See Figure 1 for the abbreviations of the major geological faults and valleys around the geothermal field.
Figure 3.11: Map view of the final Vp/Vs model for study area II. The symbols are the same as in Figure 9.
3.4.5 Cross-sections of Final Velocity Models

The geometrical shapes of the velocity anomalies are easier to track in cross-sections. Figure 3.12 shows the cross-sections of the inverted velocity models beneath the CGF and surrounding geologic regions. The Vs model is obtained from Vp divided by Vp/Vs model. We combine the Vp, Vp/Vs, and Vs models with the relocated seismicity to find out the prominent features around the geothermal field. Beneath the profiles B, D, E, which pass through the CGF, an extensive velocity anomaly of low Vp, low Vs, and low Vp/Vs from 6 to 12 km depth is observed. The size of this anomaly is about 10 km in lateral and 6 km in depth. The distribution of the relocated seismicity underneath the CGF indicates that the brittle-ductile depth is around 5 km compared to 10 km in adjacent areas. This shallower brittle-ductile transition depth has also been mentioned by previous studies (e.g., Monastero et al., 2005; Hauksson and Unruh, 2007). The anomalous zone of low velocity anomalies between 6 and 12 km is aseismic. The combination of the ductile behavior and low velocity features suggests the circumstance in this depth range is different and we will discuss the possible conditions in the following section.

Comparing the profiles D and E, we can see that the Vp/Vs ratio varies at the surface inside the CGF, while profile D passes through the geothermal exploitation area and E is roughly 7 km away. The Vp/Vs ratio is lower in the northern part, the main exploitation areas with steam and hot water at the surface, than the southern part of the geothermal field. The other striking features include low Vp, low Vs, and low Vp/Vs at the upper 3 km of the LLFZ, the IWV and the ALFZ. The Vp/Vs
ratio changes to be around 1.81 at depth of 6 km beneath the ALFZ and the adjacent eastern part. Some small velocity bodies of low Vp/Vs are also visible in the cross-section b and c (Figure 3.12), which may come from smearing of the low Vp/Vs for the CGF.

3.4.6 In-situ Vp/Vs Ratios in Similar Event Clusters

In order to complement our tomographic results, we estimate in-situ Vp/Vs ratios within similar earthquake clusters using the demeaned P- and S-wave differential times from waveform cross-correlation by applying the technique presented in Lin and Shearer (2007). This technique assumes that the scale length of changes in Vp/Vs ratios is greater than the size of the similar event clusters and all the correlated events within each individual compact cluster have the same local Vp/Vs ratio. It provides higher resolution for near-source Vp/Vs ratios than typical tomographic inversion methods by using high-precision differential times and a robust misfit function method and has been applied to study the near-source structure for the entire southern California (Lin and Shearer, 2009), the rupture zone of the 1989 Loma Prieta earthquake (Lin and Thurber, 2012), and Mammoth Mountain at the southwest rim of Long Valley caldera (Lin, 2013a). We applied this approach to all the 1,225 similar event clusters in the study area I and estimated standard uncertainties in the in situ Vp/Vs ratios. These uncertainties are computed using the bootstrap approach (e.g., Efron and Gong, 1983), in which the pairs of differential P and S times in the same cluster are randomly resampled 1000 times. In order to obtain the most
Figure 3.12: Cross-sections of the Vp, Vp/Vs and Vs structures across and proximity to the CGF. The geothermal field is denoted by the dashed circle with the red stars marking the locations of the SM and CHS. The five profiles are shown by the black lines with profile B, D and E passing through the geothermal field, A and C off the CGF. Relocated earthquakes within 5 km of both sides of the profiles are projected to the Vs cross-sections, denoted by grey dots. White contour lines represent the RDE values greater than 0.1. See Figure 1 for the abbreviations.
robust results, we select 227 event clusters with uniformly distributed events and uncertainties of $V_p/V_s$ ratios less than 0.03.

To compare with our tomographic results beneath the CGF, we project the in-situ $V_p/V_s$ ratios along the profile B in Figure 3.12. The in-situ $V_p/V_s$ ratios vary from 1.5 to 1.8 for the 36 clusters around the CGF (Figure 3.13). The mean value of 6 clusters with slightly higher $V_p/V_s (> 1.7)$ is 1.786 and these clusters mainly focus around the area with $V_p/V_s$ ratio of 1.77 inverted from the tomography. The other 30 clusters with low in-situ $V_p/V_s$ ratios ($\leq 1.7$) are mainly distributed along the low velocity zone within 20 km away from the CGF resolved by the tomographic inversion. To further verify whether the observed low $V_p/V_s$ anomalies from 6 to 12 km from tomography is reliable, we plot the in-situ $V_p/V_s$ ratios for each earthquake near the CGF within similar event clusters within this depth range. The median value of the in situ $V_p/V_s$ ratios is 1.656, consistent with the estimate of 1.667 from our tomography model. Therefore, the near source in-situ $V_p/V_s$ supports our tomographic results.

### 3.5 Discussion

With the finer 3-D seismic velocity model, we resolve the subsurface structure beneath the geothermal field. The velocity of seismic waves can be affected when they propagate through different materials due to changes in lithology, mineral composition, presence of fracture, temperature, pore pressure, and fluid saturation. Here we combine the velocity structure, the seismicity distribution, and the geological and geochemical knowledge to infer the dominating factors for the observed anomalies.
Figure 3.13: (a) In-situ Vp/Vs ratios for the clusters around the CGF. Clusters are projected along the profile B in Figure 3.12. The tomography results are the same as in Figure 11(b2). Clusters with in-situ Vp/Vs below 1.7 are shown by red dots and clusters with higher Vp/Vs shown by green dots. (b) Comparison of the in-situ Vp/Vs and tomography results for each event near the CGF between 6 and 12 km depth. Black dots show the in-situ Vp/Vs ratios and red dots denote tomography results. The median values of the Vp/Vs at 1 km depth intervals are given by the stars.
3.5.1 Geothermal Reservoir

Our tomographic results map the feature of the geothermal reservoir in the shallow depth. In map views shown by Figure 3.10a and 3.11a, low Vp and low Vp/Vs anomalies dominate the geothermal production area at the surface. Our velocity structure reflects the vapor-dominated geothermal field in high-temperature system. Because seismic velocities reduce with increased temperatures (e.g., Mueller and Raab, 1997; Fielitz, 1976; Sato et al., 1989; Ito et al., 1979), we observe low Vp at the surface. Similar low Vp/Vs features within the production area were observed by previous studies (Walck, 1988; Lees and Wu, 2000) and were interpreted to be affected by vapor.

3.5.2 Magmatic System

Although upwelling magma beneath the geothermal field has been proposed in previous studies (e.g., Wu and Lees, 1996; Lees and Wu, 1999, 2000; Lees, 2002), the velocity anomalies we observe beneath the CGF are low Vp, low Vs, and low Vp/Vs between 6 and 12 km depth, which may be a candidate for magmatic materials but refute the possibility of a magma chamber with a large amount of partial melt. The low velocity zone with ductile behavior is the typical feature associated with magma, but the low Vp/Vs ratios suggest a lack of pervasive partial melt. To validate the low Vp/Vs structure, we conduct the recovery test for the observed low Vp/Vs anomaly from the real data inversion underneath the CGF shown in Figure 3.12(b2). Cross section of the true and recovered Vp/Vs structure is shown in Figure 3.14. The anom-
lous body between 6 and 12 km depth is recovered well, although slight horizontal smearing is observed.

We take into account the chemical composition of magma from other studies in Coso to further detect if the anomalous velocity zone can refer to magmatic material. It has been argued that different silica contents of magma, crack aspect ratios and volume percentage of magma can make magmatic system behave with both low or high Vp/Vs ratios (Nakajima et al., 2001; Hauksson and Unruh, 2007; Patane et al., 2006). In our tomographic results, the Vp and Vs anomalies show the similar lateral and vertical geometry. This means that the lithology influences both Vp and Vs, but affects Vp more to result in low Vp/Vs. The anomalies do not vary with depth, and we may assume a conductive temperature gradient and hydrostatic pore pressure in this anomalous zone. The mineral composition and fluid content may play an important role in affecting the seismic velocity ratio among all the other factors. The study by Christensen (1996) suggests that the ratio is decreased with increased silica contents for the rocks with 55-75% of silica contents. Previous studies (e.g., Sanders et al., 1995; Nakajima et al., 2001; Lin and Shearer, 2009) reported that cracks with a small volume percent of H₂O could decrease Vp, Vs and Vp/Vs. From the estimate of Nakajima et al. (2001), 2% of water with aspect ratio of 0.1 can reduce Vp from the original 6.31 km/s to 6.1 km/s and Vp/Vs as low as 1.67 compared to the reference value of 1.71, which is consistent with our observations in the upper and middle crust. The presence of gas can also decrease Vp and Vp/Vs ratios because of the very low bulk modulus (Husen et al., 2004a; Lin, 2013a). The geochemical study by Manley
Figure 3.14: Recovery test for the Vp/Vs anomaly in the CGF. The top is true model and bottom is the recovered model. The zero distance refers to the location of the CGF. The Vp/Vs anomaly beneath the CGF corresponds to the anomaly at depths between 6 and 12 km shown in Figure 3.12(b2). The anomaly is recovered well in depth with some horizontal smearing.
and Bacon (2000) pointed out that the erupted magma in Coso was felsic with 66.5 wt% of silica contents, saturated with an H$_2$O-rich fluid, and with the presence of the vapor phase in melt inclusions. The percentage of H$_2$O in the melt inclusion can be around 4.5-6.2 wt% in different dome groups (Manley and Bacon, 2000), which is larger than the typical percentage of gas dissolved in magma, 0.2-3 wt%. Thus, we suggest that the low Vp, low Vs and low Vp/Vs body lacking of seismicity may indicate the molten material is felsic, rich in gas or with the inclusion of water. Although we cannot link the large quantities of hot water near the surface to the presence of water at 6 km depth, it is possible that water is rich in the anomalous velocity zone and high content of water can facilitate magma crystallization (Ritchey, 1980), which will further lower Vp/Vs ratios.

We show a conceptual model for subsurface structure in Figure 3.15. The content of the erupted magma suggests that the material is felsic, which is much more viscous than mafic magma. The felsic magma originates from the basaltic magma in deeper depth, which ascended to crystallize and expelled gas to induce fluid accumulation. Between 6 and 12 km depth beneath the CGF, the magmatic material is likely to exist in the form of magma mush, with a small percent of melt trapped among small crystals. The silica-rich magma mush might also contain a high portion of water. Our model shows that the velocity anomaly up to 12 km depth and it could be the interface of the silicic magma and mafic magma or the mafic magma could be deeper. However, the ray coverage is insufficient to resolve the structure well at those depths.
Figure 3.15: Conceptual model of the subsurface structure for the Coso geothermal field. The schematic diagram shows the main features derived from our results: (1) low velocity zone beneath the geothermal field shown by Vp cross-section; (2) shallow brittle-ductile depth (around 5 km depth) beneath the geothermal field; (3) crystallizing magma body with the inclusion of water between 6 and 12 km depth; and (4) water that may result from the underlying crystallizing basaltic magma. Lateral scale is approximate.
3.5.2.1 Comparison with Previous Seismic Studies

The low velocity body in the middle crust below the geothermal field has been observed in previous tomography studies (e.g., Reasenberg et al., 1980; Walck and Clayton, 1987). In the study by Hauksson and Unruh (2007), variation in Vp/Vs model is too small to confirm the existence of the magma body and they interpret the low Vp, normal Vp/Vs features to be the possibility of brine. We include more S wave arrival times for the inversions and our Vp/Vs model shows low value of 1.667 beneath the geothermal field, which may be associated with the magmatic system. Other factors causing differences between these two models are the two-step inversion scheme and grid spacing of 3 km in this study. The 10 km grid used in Hauksson and Unruh (2007) may hide some local anomalies with short wavelengths. Our Vp model is generally consistent with their model but shows more variations in the vicinity of the geothermal field. Some features are revealed by both models, such as the low Vp, low Vp/Vs anomalies beneath the IWV.

A low Vp anomaly beneath the CHS with the top at 5 km depth was detected by calculating teleseismic receiver functions (Wilson et al., 2003). They interpreted this as the top of the magma chamber and estimated the amount of melts to be 1.5-5% by assuming a Vp/Vs ratio of 2.5. Their Vp/Vs model was obtained by dividing the preferred 1-D P- and S- velocity models, which were used for the receiver function analysis. Our high-resolution Vp/Vs model shows that the ratio is estimated to be around 1.667 and does not agree with such an amount of melt.

The 3-D shear-wave velocity model resolved by ambient noise tomography revealed
low shear velocity zone between 6 and 12 km depth beneath the Coso geothermal field (Yang et al., 2011). Their tomographic results showed Vs structures at grid spacing of 0.25°×0.25°. The low Vs body was estimated to be at the size of 0.5° × 0.5°, which covers an area larger than the entire geothermal field. Our Vp and Vp/Vs models show the velocity varies inside the geothermal field and the prominent low Vp, low Vs and low Vp/Vs anomaly exist in the similar depth range as they indicated.

3.6 Conclusion

We developed 3-D high-resolution Vp and Vp/Vs models for the Coso region in southern California. A by-product of this study is a high-precision earthquake relocation catalog. We focus our interest on exploring the 3-D velocity structure in the vicinity of the Coso geothermal field. The tomographic results reveal the features of the geothermal reservoir and the magmatic materials. Our observations suggest a lack of a large amount of remnant melt from the Pleistocene volcanic activity underneath the geothermal field but the detected low velocity anomalies could be associated with the magmatic system. The tomography results reveal a low velocity body of 10 km diameter in the depths between 6 and 12 km. Another technique of estimating near source in-situ Vp/Vs was used to compare with our tomography models. The low in-situ Vp/Vs around the geothermal field is consistent with the low velocity anomalies derived from the tomography models. We interpret the 6×10 km² low velocity anomaly to be a region of hot and weak felsic rocks, trapped with a series of small silicic magma chambers, sills or dikes under the cooling phase with inclusion of wa-
ter. The data cannot show good resolutions below 15 km depth and whether the low
velocity anomalies extend to the deeper depth needs further investigation with more
seismic data.

3.7 Acknowledgments

We thank the SCEDC for maintaining the seismic data and making them available.
We are grateful to Steven Sherburn and an anonymous reviewer for their constructive
comments. The work has benefited from discussions with J. Ole Kaven, Mel Erskine,
Michael Hasting, and Zhongwen Zhan. We give our special thanks to Ralf Weger for
the petrology discussion. Funding for this research was provided by the National Sci-
ence Foundation grant EAR-1045856. Plots are made using the public GMT (Generic
Mapping Tools) software.
Chapter 4

Absence of remote earthquake triggering in the Coso and Salton Sea geothermal fields

\(^3\text{This chapter has been submitted to the Nature Geoscience.}\)
4.1 Summary

Recent increases in induced seismicity raise public concerns over damaging earthquakes (Ellsworth, 2013). High induced seismic activity is often assumed to indicate a critical stress level. However, the physical mechanism of induced seismicity is not fully understood yet. Recently, earthquake triggering by the seismic waves of large remote earthquakes was utilized to detect critically loaded faults (van der Elst et al., 2013). Here we examine the remote triggering phenomena in geothermal fields and their surrounding areas to assess the anthropogenic effects on the stress state. We find that the geothermal production areas, although with high seismicity rate, are less susceptible to remote triggering, which suggests a less-critical stress state than in the surrounding areas. We interpret this absence of triggering as a result of reduced pore pressure and lifted effective fault strength due to the loss of geothermal fluids. We also propose that the induced seismicity is caused by highly heterogeneous stress perturbations at shallow geothermal production depths, thus does not necessarily indicate an overall critical stress state.

4.2 Overview

4.2.1 Background and Motivation

High induced seismic activity has been documented in many geothermal production areas (e.g., the Geysers (Segall and Fitzgerald, 1998), Coso (Feng and Lees, 1998), Salton Sea (Brodsky and Lajoie, 2013), and Basel (Switzerland) (Deichmann and Gi-
ardini, 2009)). However, injected fluid volume is usually less than the extracted volume during production. The apparent paradox of extraction-induced seismicity and reduced pore pressure was explained by stress change due to contraction of geothermal reservoirs (Segall, 1989; Brodsky and Lajoie, 2013; Ellsworth, 2013). This model successfully explained the observed spatial distribution of induced seismicity, stress orientations, and surface subsidence over reservoir and its vicinity. However, it is also well recognized that highly heterogeneous and variable stress perturbations within the production area must exist. For example, Segall et al. pointed out that the thermoelastic and poroelastic stresses could possibly be dominant near production and injection wells, but are difficult to model (Segall and Fitzgerald, 1998). With the increased public concerns over the possibility of inducing large earthquakes, it has become more critical to assess the heterogeneous component of stress field and its contribution to induced seismicity. A recent stimulation experiment (Schoenball et al., 2014) showed that the local stress perturbation near the injection wells changed the background stress field significantly.

In this study, we apply the remote triggering to estimate the stress state within the geothermal production areas. Since seismic waves radiated from remote earthquakes carry a small amount of transient stress, e.g., at the level of 0.02 MPa after 3000 km away (Prejean et al., 2004), initiation of triggered earthquakes requires triggering sites to be close to the critical state before the arrival of seismic waves. The first recognized dynamic triggering case is the widespread triggering in the western U.S. after the 1992 Mw 7.3 Landers earthquake (Hill et al., 1993). If assembling the observation
for over decades, volcanic and geothermal areas are recognized to be susceptible to remote triggering, such as the Yellowstone (Hill et al., 1993; Husen et al., 2004b), Long Valley (Hill et al., 1993; Hill and Prejean, 2007), the Geysers (Hill et al., 1993; Gomberg and Davis, 1996; Prejean et al., 2004), Salton Sea geothermal field (Gomberg et al., 2001; Hough and Kanamori, 2002), and Coso geothermal fields (Hill et al., 1993; Prejean et al., 2004; Hill and Prejean, 2007; Peng et al., 2010). These fields usually lie within tectonically and seismically active regions with the presence of rich magmatic and/or geothermal fluids.

Owing to the observation that the stress change associated with seismic waves is temporary but the triggered seismicity can last weeks or even months in these sites (Husen et al., 2004b; Pankow et al., 2004), the present mechanisms of dynamic triggering involve the interaction of seismic waves and crustal fluid. The oscillation of seismic waves can increase the pore pressure within the cracks via unclogging fracture (Brodsky et al., 2003) or bubble nucleation (Manga and Brodsky, 2006), and the fluid takes some time to build up the effective pore pressure.

Since the observational triggering studies have revealed that geothermal areas are susceptible to dynamic triggering (Hill et al., 1993; Gomberg and Davis, 1996; Prejean et al., 2004), we aim to evaluate the stress state of active geothermal fields by comparing whether natural and anthropogenic settings react to remote large earthquake differently. We focus on a fine scale of tens of kilometers to compare the actual geothermal areas and their adjacent fault zones. To minimize the effect of the heterogeneity of the stress field, we choose the operational geothermal fields under active
tectonic settings, the Coso geothermal field in southeastern California, and the Salton Sea geothermal field in southern California. In this study, we test how these two sites responded to the 1992 Mw 7.3 Landers earthquake and the 1999 Mw 7.1 Hector Mine earthquake, respectively.

4.2.2 Tectonic and Anthropogenic Settings in Geothermal Fields

As one of the largest geothermal fields in the United States, the Coso geothermal field (CGF) has been in operation with a net capacity of 270 MW since 1987. It lies within the seismically active southern Walker Lane belt, which accommodates the relative motion between the Sierra Nevada block and the western edge of the Basin and Range Province (Figure 4.1a). The Coso area sits under the transtensional stress regime with major strike-slip faults and small normal faults (Unruh et al., 2002). With active tectonics and seismicity, the Coso geothermal field and its surrounding area have been studied for detection of remote triggering by the 1992 Landers (Hill et al., 1993), the 2002 Denali (Prejean et al., 2004), and the 2010 Chile earthquakes (Peng et al., 2010). These studies found remotely triggered earthquakes in a broad area surrounding the geothermal field. To investigate the anthropogenic effects on the stress state within geothermal reservoirs, we conduct a fine-scale case study in the Coso area (56×67 km², box in Figure 4.1a) to compare the remote triggering following the Landers earthquake in the Coso Geothermal Field (∼6×10 km², circles in Figure 4.1) and in its surrounding areas. We define the geothermal field based on the locations of the injection and production wells of the geothermal operation and the area of
Figure 4.1: Tectonics and location of the Coso geothermal field (CGF). Circles in the two maps outline the location of the CGF. (a) Tectonic map showing the background seismicity (yellow dots) between 1981 and 2011 and the surface traces of the faults (black lines). Purple box encloses our entire study area, including the CGF and its vicinity. Most of the faults shown here are strike-slip, including the Little Lake Fault Zone (LLFZ) and the Airport Lake Fault Zone (ALFZ). The beach ball represents the 1992 Landers earthquake. (b) Range of the CGF defined by the locations of the active injection and extraction wells for the geothermal operation and the observed subsidence. Locations of the wells (black triangles) are obtained from the California Oil, Gas, and Geothermal Resources (DOGGR). Background is the subsidence between July 1993 and September 1995 from InSAR data (Fialko and Simons, 2000).

The Salton Sea geothermal field (SSGF) is currently producing 340 MW of electricity from the geothermal power plant (Sass and Priest, 2002). The plant starts to operate in 1982 and expands in 1990s with continuously increasing net production volume. The SSGF lies in an extensional basin between the end of the San Andreas Fault and the Imperial Fault, which are the major right-lateral strike-slip
faults (Figure 4.2a). Characterized by Quaternary volcanism and high seismicity, as one of the seismically active zones in southern California (Lin, 2013b), the SSGF has been a natural laboratory for the remote triggering study. It especially showed energetic response after the 1999 Mw 7.1 Hector Mine earthquake (Gomberg et al., 2001; Hough and Kanamori, 2002). However, these studies worked on the scales of hundreds of kilometers and did not differentiate the actual geothermal area. Similar to outlining the CGF, we define the SSGF (Figure 4.2b) based on locations of wells and subsidence area.
4.3 Data and Method

4.3.1 Data Set

In this study, we used the latest relocation catalog between 1981 and 2011 in the Coso area (Zhang and Lin, 2014) for our remote triggering analyses. The original seismic data were obtained from the Southern California Earthquake Data Center (SCEDC) recorded by the Southern California Seismic Network (SCSN). In 1992, when the Landers earthquake occurred, 11 stations were available in our study area and 3 of them were located within 15 km of the geothermal field. Over 94% of the relocated events were recorded by 8 or more stations. Here we only describe the detailed analyses in the Coso geothermal field since the analyses for the Salton Sea geothermal field are similar.

4.3.2 Estimate of Catalog Completeness and b-value

The largest earthquakes in our study area of Coso and the CGF are Mw 5.75 and Mw 4.41, respectively. As the regional network may not be able to detect all the earthquakes, we first estimated the completeness magnitude Mc of the relocation catalog. Following the method of the maximum curvature (Wiemer and Wyss, 2000), we computed Mc as the magnitude with the maximum derivative of the frequency-magnitude curve without adding other smoothing parameters. The estimated Mc is 1.3 for the entire study area and is 1.0 for the CGF (Figure 4.3). Therefore, we use M 1.3 as the magnitude threshold in this study to ensure the catalog completeness.
within the entire study area. The Gutenberg-Richter b value is estimated together with the Mc (Figure 4.3). The larger b-value inside the geothermal field than in the entire study area indicates that the geothermal field is more dominated by events with smaller magnitudes.

4.3.3 Declustering Catalog

Statistical analysis of earthquake catalogs assumes that earthquakes are random, independent events. However, the real catalog consists of many sequences of mainshocks and aftershocks, which are strongly dependent on each other. The clustering patterns of aftershocks in time and space are especially obvious after moderate or large earthquakes (Mw ≥ 4.0).

To remove the effect of aftershocks of Mw ≥ 4.0 mainshocks, we follow the declustering method described by Reasenberg (1985) to decluster the full earthquake catalog. For each earthquake, the method first models the spatial and temporal extent of an earthquake interaction zone based on the source dimension of the mainshock and the modified Omori’s law. Then every event is cross-correlated with every other event to form clusters whenever two associated events are proximity to each other in location and time. The cluster grows with more cross-correlated events. The events within a single cluster are dependent on each other and thus regarded to be aftershock-related sequences. The largest event in a single cluster is considered as the mainshock and kept in the final declustered catalog.

We show an example of declustering in the area of Rose Valley in Figure 4.4. It
Figure 4.3: Estimate of catalog magnitude completeness and $b$-value for the entire study area (a) and the Coso geothermal field (b). The magnitudes of the largest earthquakes during our study period are Mw 5.75 and Mw 4.41 for the entire study area and the Coso Geothermal Field, respectively. Red crosses show the magnitude thresholds. The larger $b$-value inside the geothermal field than that for the entire study area indicates that the geothermal field is more dominated by events with smaller magnitudes.
clearly shows that the aftershocks of an Mw 4.1 mainshock in February 1992 have been removed. However, the increased seismicity after the Landers was much less affected, which means that the increase is not a coincident aftershock sequence after some local earthquakes.

After removing the clustered aftershocks, the remaining catalog contains independent events and is assumed to perform as a Poisson process (Reasenberg, 1985). Using the declustered catalog, we investigate the spatiotemporal distribution of seismicity in our study area following the Landers earthquake.

### 4.3.4 Statistical Analysis of Seismicity Rate Change

We calculated the statistical significance of the seismicity rate change by computing the $\beta$-statistic (Matthews and Reasenberg, 1988; Reasenberg and Simpson, 1992; Hill and Prejean, 2007), which compared the difference between the observed and expected seismicity in 30 days, normalized by standard deviation of the expected seismicity. It can be expressed as: (Matthews and Reasenberg, 1988; Hill and Prejean, 2007)

$$
\beta(n_a, n_b, t_a, t_b) = \frac{n_a - E(n_a, n_b)}{\sqrt{\text{var}(n_a, n_b)}}
$$

(4.1)

where $n_a$ and $n_b$ are the numbers of earthquakes in the time period of $t_a$ and $t_b$, respectively. $E(n_a, n_b)$ is the expected number of earthquakes in $t_a$ based on the sample of background seismicity rate in $t_b$. $\text{var}(n_a, n_b)$ denotes the variance of the number of earthquakes in $t_a$ based on the sample of the background seismicity rate in $t_b$. $\beta$-values larger than 2 or smaller than -2 indicate significantly higher or lower seismicity
Figure 4.4: Example of catalog declustering in the Rose Valley (subarea 3). Red dots represent microearthquakes with $1.3 \leq M_w < 4.0$ and green dots for $M_w \geq 4.0$. Blue line marks the onset of the Landers earthquake. (a) Time series of the full catalog, including a $M_w 4.1$ mainshock and its aftershock sequence. (b) Time series of the declustered catalog.
rates than the background (Reasenberg and Simpson, 1992; Hill and Prejean, 2007).

The study area was gridded with blocks of $5 \times 6 \text{ km}^2$, which was chosen to ensure relatively uniform seismicity in different blocks. We then calculated the seismicity rate change for 30 days after the Landers earthquake relative to the background seismicity from 1987 to 1993 using the $\beta$-statistic. We also computed the seismicity rate changes in other time windows (10, 100 days) relative to the same background period.

Because the declustered catalog is assumed to be a stationary process following the Poisson’s distribution, we also calculated the Poissonian probability of the observed seismicity 30 days following the Landers earthquake. When the probability is less than 0.05, we rejected the null hypothesis that the increased seismicity was a random occurrence.

4.4 Results

4.4.1 Seismicity Rate Change in Coso

We showed the spatial distribution of seismicity rate in different time windows (Figures 4.5a and 4.6). The areas outside the geothermal field showed high $\beta$-values (up to 6), indicating wide-spreading triggering within 30 days (Figure 4.5a). This increase can also be observed in other time windows (e.g., 10 days, Figure 4.6a) and became less significant after 100 days following the Landers earthquake (Figure 4.6b). With 95% of confidence, our results indicate that the increased seismicity outside the geothermal
field cannot be a random occurrence (Table 4.1). However, we also observed an absence of seismicity rate change inside the geothermal field. During the time windows from 10 to 100 days after the Landers earthquake, the maximum and minimum $\beta$ values in the CGF were 1.81 and -1, respectively, which fell in the typical statistical range of background seismicity and means that there was no significant increase or decrease of seismicity rate inside the geothermal field. This indicates that the active geothermal field is less susceptible to remote triggering than the surrounding areas.

In order to compare the spatiotemporal variation of the seismicity in consistent spatial scales, we assigned the geothermal field as subarea 1 and divided the adjacent area into 6 subareas (black boxes in Figure 4.5a) based on the distribution of the background seismicity from 1981 to 2011. We showed the time series for all the 7 subareas between 1991 and 1993 (Figure 4.5b). The time window must be chosen short enough to keep the detection criteria the same and long enough to compare the seismicity before and after the Landers earthquake. We observed an abrupt increase in the seismicity rate with respect to the Landers earthquake for all the 6 subareas outside the CGF, which is consistent with previous studies (Hill et al., 1993; Prejean et al., 2004; Hill and Prejean, 2007; Peng et al., 2010). In contrast, the geothermal field itself appeared unaffected by the Landers earthquake, which is consistent with our $\beta$-statistic analysis above. Although some smaller events may be missing from the regional catalog, they would not change the observation that the inside and outside the geothermal field responded to the large remote earthquake differently based on the same detection criteria.
Figure 4.5: (a) Spatial distribution of the declustered seismicity (green circles) within 30 days following the Landers earthquake and the $\beta$-statistic of the seismicity rate change (colored grids), calculated relative to the background period 1987-1993. Based on the distribution of the background seismicity between 1981 and 2011 (grey dots), we assigned the Coso geothermal field (CGF) as subarea 1 and divided the adjacent area into 6 subareas, including the Coso Range (CR), Rose Valley (RV), Centennial Flat (CF), Wilson Canyon Fault (WCF), and Airport Lake Fault Zone (ALFZ). We masked out those grid blocks with too sparse background seismicity between 1987 and 1993. (b) Number of events per day in the 7 subareas between 1991 and 1993. The three-year time window was chosen to keep the detection criteria identical before and after the Landers earthquake. Red dots represent microearthquakes with $1.3 \leq M_w < 4.0$ in the declustered catalog and green dots for $M_w \geq 4.0$. Blue line marks the onset of the 1992 Landers earthquake.

Table 4.1: Statistical analysis for seven subareas

<table>
<thead>
<tr>
<th>Subarea</th>
<th>Poisson Probability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coso Range</td>
<td>0.0102</td>
</tr>
<tr>
<td>Rose Valley</td>
<td>0.0139</td>
</tr>
<tr>
<td>Centennial Flat</td>
<td>0.0453</td>
</tr>
<tr>
<td>Wilson Canyon Fault</td>
<td>$10^{-6}$</td>
</tr>
<tr>
<td>Airport Fault Lake Zone1</td>
<td>0.0134</td>
</tr>
<tr>
<td>Airport Fault Lake Zone2</td>
<td>0.0101</td>
</tr>
</tbody>
</table>
Figure 4.6: Map view of the seismicity rate change between different time windows in the Coso geothermal field and the vicinity. All the legends are the same as those in Figure 4.5a, which shows the $\beta$-statistic of 30 days after the Landers earthquake relative to the background seismicity (1987-1993). (a) $\beta$-statistic of 10 days after the Landers earthquake. (b) $\beta$-statistic of 100 days after the Landers earthquake.
4.4.2 Correlation with Geothermal Production

The absence of remote triggering inside the CGF suggests that the geothermal production has altered the stress state within the reservoir to be less critical than the surrounding areas. The geothermal fluids in the reservoir have been extracted with monthly net production (i.e., extraction-injection) volume at a level of $10^6$ m$^3$ since 1987 (Figure 4.7). Loss of geothermal fluids can result in contraction of reservoir due to poroelastic and thermoelastic effects (Segall, 1989; Segall and Fitzgerald, 1998). The surface subsidence accompanying the reservoir contraction has been observed in the CGF, revealed by InSAR data (Fialko and Simons, 2000). Another effect of fluid loss is to reduce the pore pressure ($p$), hence increased the fault strength $f(\sigma - p)$, where $f$ is the frictional coefficient and $\sigma$ is the normal stress. The average stress state within the reservoir is affected by the reduced pore pressure competing against the stress change from reservoir contraction. Our observation implies that the stress state within the reservoir was brought away from failure, and the weak transient stress carried by surface waves from the Landers earthquake cannot trigger seismicity.

4.4.3 Stress Rotation

To study further how the continuous production process can affect the stress field, we computed the average stress orientation in all the subareas. We selected events with more than 10 P-wave first motion polarities from a recent focal mechanism catalog (1981-2010) for southern California (Yang et al., 2012). Using the SATSI package (Hardebeck and Michael, 2006; Martínez-Garzón et al., 2013), all the subareas with
Figure 4.7: Monthly fluid injection and extraction volume. Red curve denotes the extracted fluid volume and blue curve represents the injected fluid. The net production (i.e., extraction-injection) volume is denoted by the green curve.

enough events have achieved robust stress orientation, indicated by the distribution of the bootstrap sampling dots. We find that the average stress fields in the subareas outside the geothermal field show strike-slip orientation, whereas it is distinctively normal faulting for the CGF (Figure 4.8). We also calculated the stress orientation for the CGF before the operation of the geothermal production (1981-1986), which showed strike-slip with larger uncertainties due to poor data coverage though. Similar results of stress orientation rotation have been observed in the Geysers and Soultz-sous-Forets (France) geothermal fields (Martínez-Garzón et al., 2013; Schoenball et al., 2014). The geothermal stimulation experiment (Schoenball et al., 2014) shows that the orientation of the stress field could be temporary and the normal faulting component decreased following the shut-in of the reservoir stimulation.
Figure 4.8: Stress inversion results for all the 7 subareas in the Coso area. The average stress orientation from 1981 to 2010 for these subareas are shown in seven panels and the first panel shows the stress orientation in the CGF before the geothermal operation (1981-1986). The orientations of three principal stresses ($\sigma_1 > \sigma_2 > \sigma_3$) are shown with clouds representing bootstrap points within 95% confidence intervals. The center corresponds to the vertical direction. Note that the CGF has normal faulting regime, whereas other areas share strike-slip regime. In the CGF, about 80% of the events with high-quality focal mechanisms occur at depths of 1.5-3 km.

Similar results of stress orientation rotation have been observed in the Geysers geothermal field and Soultz-sous-Forêts (France) geothermal system (Martínez-Garzón et al., 2013; Schoenball et al., 2014). The geothermal stimulation experiment (Schoenball et al., 2014) shows that the stress field could be perturbed by anthropogenic activity temporarily and normal faulting component decreases following shut-in of the reservoir stimulation. Their independent stress measures imply that the locally perturbed stress could be at the level of tens of MPa, coupling with active seismicity, but the high stress cannot represent the initial average stress field (Schoenball et al., 2014).
4.4.4 Seismicity Rate Change in Salton Sea

We used a relocated earthquake catalog (Lin, 2013b) to examine the remote triggering in the Salton Sea area, which has been shown to respond to the 1999 Mw 7.1 Hector Mine earthquake (Hough and Kanamori, 2002; Gomberg et al., 2001). We conducted similar analyses of seismicity rate change as above to show the $\beta$-statistic and one-year time series (Figures 4.9 and 4.10). We observed the increased seismicity within 5 km north and south of the SSGF, indicated by large $\beta$ values ($\sim$6) within 4 days after the Hector Mine earthquake. However, the seismicity inside the SSGF did not show an abrupt increase. The remotely triggered earthquake sequence identified by a previous study (Hough and Kanamori, 2002) fell outside the geothermal field. The observation in the Salton Sea area supports our inference that the production process has pulled the stress field away from the critical state by reducing pore pressure, making it less susceptible to remote triggering compared to the adjacent areas.

4.4.5 Background Seismicity and Conceptual Model

A less-critical stress state appears to contradict the high seismicity rate within the reservoir. Therefore, we investigated the spatiotemporal distribution of the seismicity from 1981 to 2011. We divided the seismicity into a shallow layer ($< 3$ km) that is associated with the production process, and a deeper layer confined between 3 and 10 km depth. Our results showed that the seismicity rates in the shallow and deeper layers both decreased immediately following the onset of the geothermal operation in 1987 (Figure 4.11a). The abrupt drop of the seismicity rates at all the depths is
Figure 4.9: Map views of the seismicity rate change in the Salton Sea geothermal field (SSGF) and its vicinity. $\beta$-statistic of 4 days (a) and 30 days (b) after the 1999 Mw 7.1 Hector Mine earthquake relative to the background seismicity (1996-2002). Polygon outlines the SSGF based on the location of the active injection and extraction wells. Boxes mark the subareas outside of the geothermal field, including the Brawley Seismic Zone (BSZ) and Imperial Fault (IF). Grey dots denote the background seismicity from 1981 to 2010. Inset shows the location of the Hector Mine earthquake and the SSGF.
Figure 4.10: One-year time series in the Salton Sea geothermal field (SSGF) and its vicinity. The first panel is the SSGF. The other three are the subareas outside the SSGF. Red dots represent microearthquakes with $1.7 \leq M_w < 4.0$ in the declustered catalog and green dots for $M_w \geq 4.0$. Blue line marks the onset of the Hector Mine earthquake.
consistent with our explanation that reduced pore pressure after production increased the fault strength within the CGF. As the geothermal production grew, the seismicity rate at the deeper layer remained low, whereas the shallow seismicity rate recovered and increased to a higher level (Figure 4.11a). The shallow geothermal induced seismicity has been proposed to result from reservoir contraction (Segall and Fitzgerald, 1998), and local effects such as diffusion of pore pressure from injection wells and temperature change (Majer et al., 2007). Considering the reservoir sitting in a less-critical stress state inferred above, we propose that the increase in shallow seismicity could be more dominated by the highly heterogeneous and variable (in time) stress perturbations at local scales (Figure 4.11b), possibly very close to the production or injection wells. Because these stress perturbations vary quickly in space and time, and the average stress level is not close to failure, it is less likely to produce large earthquakes or remotely triggered seismicity.

4.5 Conclusions

In summary, we studied the fine-scale remote triggering in geothermal fields and their vicinities to assess the anthropogenic effect on the stress state. We find that the geothermal production areas, although with high seismicity, are less susceptible to remote triggering than the surrounding areas. We attribute this absence of triggering to the reduced pore pressure and lifted effective fault strength due to fluid loss during geothermal production. We also propose that the induced seismicity correlated with production is mainly controlled by highly heterogeneous stress perturbations at
Figure 4.11: Seismicity distribution and schematic model of stress states within the geothermal field.  
(a) Number of shallow earthquakes (< 3 km depth) and deeper earthquakes (between 3 and 10 km depth) within the geothermal field for the time period of 1981 to 2011. The dots represent the number of earthquakes every two years. The reduction in the number of earthquakes occurred in 1987 when the production operation began.  
(b) Schematic model of stress states affected by anthropogenic activity. Net production of the geothermal fluids caused decrease of pore pressure and higher failure criterion for both shallow and deeper faults, which explains the absence of remote triggering. At the shallow depth, highly heterogeneous stress perturbations (blue wiggles) are responsible for induced seismicity. At larger depths, it may take longer time for the tectonic stress to overcome the higher fault strength.
shallow depths. The magnitude of these local stress perturbations could be at the level of tens of MPa (Schoenball et al., 2014). Therefore, it is difficult to infer the average stress state from induced seismicity alone. Although our observations indicate a stress state away from failure, tectonic stress will continue to accumulate to balance the decreased pore pressure (Figure 4.11b). In addition, change in production volume or hydrological system may also alter the pore pressure. Hence, it is important to monitor the average stress state inside the geothermal field at fine scales either by remote triggering or other stress measures.

4.6 Acknowledgements

We thank the SCSN for making seismic data available. We appreciate the communication with Hiroo Kanamori, David Hill, and Stephanie Prejean for constructive feedbacks. We also thank Estelle Chaussard for useful discussion and Heresh Fattahi for providing InSAR results. Funding for this research was provided by the National Science Foundation grant EAR-1045856. Most of figures are made using the public GMT (Generic Mapping Tools) software.
Chapter 5

Conclusions and Future Work

This thesis mainly presents how to better understand the causes of triggered/inhibited earthquakes by imaging the seismic structure and mapping spatiotemporal distribution of microearthquakes. My research work mainly solved three questions: (1) What is the possible cause of active seismicity in the southwestern island of Puerto Rico and seismic quiescence in the north; (2) Whether the active magmatic system still exists in the crust of the Coso geothermal field; (3) Whether geothermal areas are more susceptible to dynamic triggering.

In the Puerto Rico region, we have generated the first 3D crustal velocity model and new relocation catalogs, which can be used for the routine processing in the Puerto Rico Seismic Network. In the Coso area, we have obtained high-solution 3D velocity models and high-precision earthquake relocation, from which we have just extracted part of the information to address the questions regarding magmatic system and dynamic triggering. We believe the new 3D models together with the relocated seismicity can provide useful information for investigating other seismologic and tectonic features. In addition, we have only interpreted the upper crustal structure due to the limit of the data. With additional data sets, such as ambient noise, it is possible
to detect the deeper structure to better characterize whether the identified magma mush is fed by a deeper magma chamber.

We demonstrate that seismic tomography combined with relocated earthquakes can provide a unique access to probing solid Earth’s structure. However, to tackle seismotectonic puzzles, we could not solely depend on the seismological results. For example, we have included other geophysical and geochemical studies, the results of resistivity and density and the content of erupted magma, to interpret the magmatic system in the Coso geothermal area. To support our model that the loss of pore pressure results in the absence of dynamic triggering, we also integrate the geodetic measurements to show geothermal areas are undergoing the loss of fluid and/or thermal contraction.

Although my research work shows the consistent results in Coso and Salton Sea and we imply the loss of pore pressure from the anthropogenic activity is a dominating effect for the absence of remote triggering, a lot of questions remain unknown. How much pressure could serve as the threshold for remote triggering? Is there any quantitative estimate of how the stress is perturbed and released? How much the induced seismicity depends on the initial natural stress state? Further independent stress measurement and poroelastic numerical modeling could be useful for answering these questions.

After investigating the background seismicity in geothermal fields, I have realized that the anthropogenic activity can largely change the stress state, in the ways of changing the net pore pressure, perturbing the stress field with tens of MPa, and
rotating the stress orientation. Thus a large amount of induced seismicity has been observed in these active geothermal areas. Our research work and some other studies have suggested that the induced seismicity share the normal-faulting focal mechanisms in several geothermal fields (Martínez-Garzón et al., 2013; Schoenball et al., 2014). It is necessary to analyze the sequence of induced seismicity with better data coverage to characterize more different features from natural earthquakes. If adding more permanent broadband stations near the wells, it is possible to obtain the long-term earthquake sequence and help differentiate whether a certain earthquake is natural or induced. In addition, denser array will improve the detection of smaller earthquakes, relocation of earthquakes, and determination of magnitude and focal mechanisms.

Besides geothermal areas, increased seismicity has been observed in oil and gas production area, wastewater disposal regions, and areas with mining activity. Both natural and anthropogenic forces play roles in affecting the stress state in these areas. The detection of remote triggering has been proven to be one method to detect the stress state. Our results show that the geothermal production process can inhibit remote triggering and induce a great amount of shallow seismicity. To obtain the systematic observation, we need to dissect more cases in these different anthropogenic settings to analyze how anthropogenic activities affect the stress state in the ways of inhibiting and promoting the failure of seismicity.

Although the induced seismicity has been regarded to associate with the injection and production processes, their physical mechanism and possible seismic hazard are not well understood. To explain the induced seismicity, the current hypotheses in-
clude: (1) fluid injection induced earthquakes by reducing the effective normal stress 
(*Hubbert and Rubey*, 1959; *Nicholson and Wesson*, 1992); (2) well stimulation and 
reservoir volume change resulted in elastic compaction of brittle materials (*Segall*, 
1989; *Mossop and Segall*, 1999); (3) induced seismicity that occur at the undisturbed 
areas with kilometers away or deeper than wells is caused by the diffusion of pore 
pressure from the injection wells (*Simpson et al.*, 1988). However, there is no uniform 
theory to account for various observations from fields. For example, some larger earth-
quakes occur once the production rate is largely increased, while the fluid injection 
remains negligible (*Frohlich and Brunt*, 2013). Some earthquakes occur with respect 
to fluid injection, and in other areas the time lag is long (*Simpson et al.*, 1988). To 
advance the understanding of the induced seismicity and the controlling factors, we 
need to investigate more earthquake sequence in different settings closely combining 
with other geophysical studies, such as studies regarding the stress measurement, 
hydraulic conditions, and presence of pre-existing faults.
Bibliography


Lin, G. (2013a), Seismic investigation of magmatic unrest beneath Mammoth Mountain, California, USA, *Geology*.


Mann, P., C. Prentice, J. Hippolyte, N. Grindlay, L. Abrams, and D. La-Davila (2005), Reconnaissance study of Late Quaternary faulting along Cerro Goden fault zone, western Puerto Rico, *Active Tectonics and Seismic Hazards of Puerto Rico, the Virgin Islands, and Offshore Areas, 385*, 115–138.
Martínez-Garzón, P., M. Bohnhoff, G. Kwiatek, and G. Dresen (2013), Stress tensor changes related to fluid injection at the Geysers geothermal field, California, Geophysical Research Letters, 40(11), 2596–2601.


Mendoza, C., and W. McCann (2005), Improving the seismic hazard model for puerto rico through seismic tomography and a reliable microearthquake catalog with re-calculated magnitudes and calibrated hypoentral error estimates, U.S. Geological Survey External Research, 05HQGR0012, 20.


Yang, Y., M. Ritzwoller, and C. Jones (2011), Crustal structure determined from ambient noise tomography near the magmatic centers of the Coso region, southeastern California, *Geophysics, Geochemistry, Geosystems, 12*.