DEFORMATION PROCESSES THROUGHOUT
THE EARTHQUAKE CYCLE

A Dissertation in
Geosciences
by
Matthew W. Herman

Submitted in Partial Fulfillment
of the Requirements
for the Degree of
Doctor of Philosophy

August 2017
The dissertation of Matthew W. Herman was reviewed and approved* by the following:

Kevin P. Furlong  
Professor of Geosciences  
Dissertation Advisor  
Chair of Committee

Charles J. Ammon  
Professor of Geosciences

Donald M. Fisher  
Professor of Geosciences

Christelle Wauthier  
Assistant Professor of Geosciences

Karl M. Reichard  
Research Associate and Assistant Professor of Acoustics

Rob Govers  
Associate Professor of Geophysics  
Utrecht University

Timothy Bralower  
Professor of Geosciences  
Associate Head for Graduate Programs and Research

*Signatures are on file in the Graduate School
Abstract

This dissertation presents observational and modeling analyses of deformation processes acting before, during, and after large earthquakes. Both seismic and geodetic observations of earthquakes and their foreshock/aftershock sequences are used to determine event locations and source parameters. Geodetic observations also provide constraints on aseismic processes, such as slow slip, inter-seismic loading, and post-seismic relaxation. Case studies include the 2011 Mw 9.0 Tohoku, Japan, the 2013 Mw 8.0 Santa Cruz Islands, the 2014 Mw 6.3 Mae Lao, Thailand, the 2014 Mw 8.2 Iquique, Chile, the 2015 Mw 8.3 Illapel, Chile, and the 2016 Mw 5.7 Christchurch, New Zealand earthquakes. Geophysical observations of these events are applied in conjunction with a variety of numerical modeling approaches to develop improved earthquake cycle deformation frameworks. Models of fault slip in an elastic half-space are applied during the co-seismic stage to compute surface displacements and stress changes. These stress changes are interpreted to trigger subsequent seismicity in nearby regions and induce other types of post-seismic deformation. Some of the observations are not compatible with these elastic half-space models, so finite element modeling approaches are used to further explore the deformation in more rheologically realistic systems. These models include simulations of inter-seismic deformation around a heterogeneous frictional plate boundary interface, accumulation and release of strain throughout each stage of the earthquakes cycle, and the spatial patterns of loading in a rheologically layered upper plate. By integrating observations and models, these studies highlight the importance of rheological heterogeneity, strain rate, and deformation history in earthquake cycle processes.
# Table of Contents

List of Figures ........................................................................................................... viii
List of Tables .............................................................................................................. xii
Preface ......................................................................................................................... xiii
Acknowledgements ..................................................................................................... xiv

Chapter 1 Introduction ................................................................................................. 1

Chapter 2 Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, Earthquake ................................................................. 3

Chapter 3 Revisiting the Canterbury Earthquake Sequence After the 14 February 2016 Mw 5.7 Event ................................................................................. 4

Chapter 4 Modeling Slip Deficit Accumulation Around Locked Regions on the Subduction Megathrust ........................................................................ 4

Chapter 5 Interactions Between Asperities Over Multiple Earthquake Cycles ................................................................................ 4

Chapter 6 Integrated Geophysical Characteristics of the 2015 Illapel, Chile, Earthquake ................................................................................ 5

Chapter 7 Patterns of Loading and Unloading in the Upper Plate Throughout the Earthquake Cycle in Japan .............................................. 5

Chapter 8 Related Work ............................................................................................... 5

Chapter 2 Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, Earthquake ................................................................. 6

Abstract ......................................................................................................................... 6
1. Introduction ................................................................................................................. 6
   1.1. Overview of Foreshock Seismicity ....................................................................... 7
2. Methods and Results ................................................................................................. 8
   2.1. Foreshock ΔCFS ................................................................................................. 10
   2.2. Slow Slip ............................................................................................................ 14
   2.3. Slow Slip Modeling ............................................................................................ 18
3. Discussion .................................................................................................................... 20
Acknowledgements ....................................................................................................... 22
References ....................................................................................................................... 22
Figures ............................................................................................................................. 25

Chapter 3 Revisiting the Canterbury Earthquake Sequence After the 14 February 2016 Mw 5.7 Event ................................................................. 32

Abstract ......................................................................................................................... 32
1. Introduction ................................................................................................................. 32
2. Methods ....................................................................................................................... 33
3. Results ......................................................................................................................... 36
4. Discussion .................................................................................................................... 39
<table>
<thead>
<tr>
<th>Acknowledgements</th>
<th>40</th>
</tr>
</thead>
<tbody>
<tr>
<td>References</td>
<td>40</td>
</tr>
<tr>
<td>Figures</td>
<td>43</td>
</tr>
</tbody>
</table>

Chapter 4  Modeling Slip Deficit Accumulation Around Locked Regions on the Subduction Megathrust ......................................................... 47

1. Introduction ........................................... 47
2. Methods ............................................. 49
3. Results ............................................. 51
   3.1. Single Asperity .................................. 51
   3.2. Multiple Locked Zones ........................... 56
4. Discussion .......................................... 57
   4.1. Interpreting Coupling Models .................. 57
   4.2. Earthquake Implications ......................... 60
5. Conclusion ........................................... 63
References ............................................... 63
Figures .................................................. 68

Chapter 5  Interactions Between Asperities Over Multiple Earthquake Cycles .......... 79

1. Introduction ........................................... 79
2. Model Setup .......................................... 81
3. Modeling Results .................................... 83
   3.1. Continuous Asperity Model ...................... 83
   3.2. Isolated Asperities ............................... 85
   3.3. Up-dip Unlocked .................................. 86
4. Discussion ............................................ 87
References ............................................... 89
Figures .................................................. 92

Chapter 6  Integrated Geophysical Characteristics of the 2015 Illapel, Chile, Earthquake..... 101

Abstract .................................................. 101
1. Introduction .......................................... 101
   1.1. Overview ......................................... 101
   1.2. Historical Seismicity ............................. 102
   1.3. Coupling Models .................................. 103
   1.4. This Study ....................................... 104
2. Methods and Results ................................ 104
   2.1. Mainshock Moment Tensor ....................... 104
   2.2. Seismic Finite Fault Model ..................... 106
   2.3. Geodetic Slip Model ............................. 107
   2.4. Aftershock Moment Tensors ..................... 108
   2.5. Hypocentral Relocation ......................... 109
   2.6. Aftershock Catalog ................................ 110
   2.7. Coulomb Failure Stress Changes ................ 112
3. Discussion ............................................ 114
Introduction........................................................................................................................................... 182
Double Source Moment Tensor Inversion............................................................................................. 182
Seismic Finite Fault Model.................................................................................................................... 183
Geodetic Finite Fault Modeling............................................................................................................. 183
Coulomb Stress Dependence on FFM .................................................................................................... 183

Appendix E Exploring the Nature of Transient Waves Observed at High-Rate GPS
Stations During the Tohoku Earthquake................................................................................................. 202

D.1. High-rate GPS Dataset .................................................................................................................. 202
D.2. Modeling High-Rate GPS Records ............................................................................................... 203
List of Figures

Figure 2-1. Tectonic setting of the 2014 Iquique earthquake sequence. ........................................ 25
Figure 2-2. Detailed view of 2014 Iquique sequence seismicity......................................................... 26
Figure 2-3. Evolution through time of the Coulomb failure stress change................................. 27
Figure 2-4. Cross-sections of ΔCFS resolved onto the megathrust ................................................. 28
Figure 2-5. Daily east-west and north-south components of GPS displacement.......................... 29
Figure 2-6. Results of the grid search for a rectangular slow slip patch................................... 30
Figure 2-7. Effects of an aseismic slip patch on the ΔCFS on the plate boundary immediately before the April 1st mainshock. ................................................................. 31
Figure 3-1. Seismicity in the Canterbury earthquake sequence from September 2010 through April 2016........................................................................................................... 43
Figure 3-2. Relocation results and moment tensor solutions.......................................................... 44
Figure 3-3. Coulomb stress change in the Canterbury Plains......................................................... 45
Figure 4-1. Finite element model setup........................................................................................... 68
Figure 4-2. Interface slip around a 40 km square locked patch.................................................... 69
Figure 4-3. Surface displacements for a 40 km square asperity..................................................... 70
Figure 4-4. Stresses on an along-dip cross-section through the middle of a 40 km square asperity. .......................................................................................................................... 71
Figure 4-5. Comparison of fault slip around different sized asperities..................................... 72
Figure 4-6. Comparison of short and long asperity slip distribution......................................... 73
Figure 4-7. Stresses on an along-dip cross-section through the middle of a 40 km by 500 km asperity, as in Figure 4-4.................................................................................................... 74
Figure 4-8. Slip deficit around adjacent 40 km square asperities................................................. 75
Figure 4-9. Surface displacements around adjacent 40 km square asperities............................. 76
Figure 4-10. Effect of rupturing an asperity adjacent to an unruptured locked zone................. 77
Figure 4-11. Comparison of our FEM-generated inter-seismic deformation with results from Metois et al. (2016).

Figure 5-1. Examples of subduction zones that ruptured in both single- and multiple-segment earthquakes.

Figure 5-2. Earthquake cycle finite element model setup.

Figure 5-3. Co-seismic slip on the interface during single- and mult-segment ruptures.

Figure 5-4. Co- and post-seismic slip for a continuous seismogenic zone with the up-dip region locked during the earthquake.

Figure 5-5. Total slip at each stage of the earthquake cycle on a down-dip transect through the continuous seismogenic zone model with up-dip region co-seismically locked.

Figure 5-6. Co- and post-seismic slip for isolated seismogenic zones with the up-dip region locked during the earthquake.

Figure 5-7. Total slip at each stage of the earthquake cycle on a down-dip transect through the isolated seismogenic zone model with up-dip region co-seismically locked.

Figure 5-8. Co- and post-seismic slip for models with the up-dip region unlocked throughout the entire earthquake cycle.

Figure 5-9. Total slip at each stage of the earthquake cycle on down-dip transects through the models with up-dip region always unlocked.

Figure 6-1. Tectonic setting and historical seismicity near Illapel, Chile.

Figure 6-2. Moment tensor solutions for Illapel earthquakes.

Figure 6-3. Map of the seismic finite fault model solution determined in this study.

Figure 6-4. Map of the geodetic slip model solution determined in this study.

Figure 6-5. Results of hypocenter relocation.

Figure 6-6. Map of relocated aftershocks, scaled by magnitude and colored by time.

Figure 6-7. Illapel earthquake sequence catalog statistics, using the same set of events as shown in Figures 6-1 and 6-6.

Figure 6-8. Coulomb failure stress change analysis.

Figure 6-9. Results of a grid search over the geographical area of the Illapel aftershock sequence for Omori decay constant within a circular area of radius 40 km.
Figure 6-10. Coulomb failure stress change generated by subduction zone thrust faulting resolved on normal faulting planes. ................................................................. 137

Figure 6-11. Relationship between inter-seismic coupling and Illapel mainshock slip. ....... 138

Figure 6-12. Schematic interpretation of Illapel sequence seismotectonics. .................. 139

Figure 7-1. Schematic representation of subduction zone material properties. .......... 149

Figure 7-2. Surface motion of Japan before, during, and after the 2011 Mw 9.0 Tohoku earthquake. ........................................................................................................ 150

Figure 7-3. Finite element model setup of inter-seismic loading in a subduction zone. .... 151

Figure 7-4. Cross-sectional stresses through the center of the rheologically layered finite element model. ........................................................................................................ 152

Figure 7-5. Cross-sectional stresses through the center of a finite element model with a fully elastic upper plate. .............................................................................. 153

Figure 7-6. Fit between observed and synthetic static displacements. ......................... 154

Figure 7-7. Cross-sectional stresses through the center of a fault with 1 meter of slip in an elastic half-space. ..................................................................................... 155

Figure 7-8. Sum of co-seismic and inter-seismic stresses. ......................................... 156

Figure 7-9. Schematic representation of stresses immediately after the earthquake. ....... 157

Figure 8-1. Tectonic setting of the 6 February 2013 Mw 8.0 Santa Cruz Islands earthquake. ................................................................................................................. 164

Figure 8-2. Coulomb failure stress changes (ΔCFS) resolved on aftershock focal mechanisms, modified from Hayes et al. (2014a). .................................................. 165

Figure 8-3. Regional tectonic setting and stress changes for the 2014 Iquique earthquake. ... 166

Figure 8-4. Regional moment tensor solutions for events in the Chiang Rai sequence. .... 167

Figure 8-5. Coulomb stress change caused by the mainshock resolved onto east-striking left lateral strike-slip faults. ................................................................. 168

Figure A-1. Evolution of stress changes throughout foreshock sequence. .................... 171

Figure B-1. Seismicity in the Canterbury Plains over time. ........................................... 172

Figure B-2. Results of sensitivity test on the ΔCS distribution. ...................................... 174
Figure D-1. Double source moment tensor inversion solutions applying different constraints ................................................................. 185

Figure D-2. Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b............................................. 186

Figure D-3. Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b... 192

Figure D-4. Comparison of predicted and observed GPS displacements from a selection of six representative seismic finite fault model inversions ........................................ 198

Figure D-5. Fits to co-seismic InSAR interferograms and GPS displacements generated by the geodetic slip model from this study. ............................................................... 199

Figure D-6. Coulomb failure stress changes from four Illapel earthquake slip models. ....... 200

Figure E-1. GEONET network and observed displacement time series. ......................... 205

Figure E-2. Moveout velocity modeling scheme................................................................. 206

Figure E-3. Moveout velocity fits to displacement time series ............................................. 207

Figure E-4. Particle motions during the growth of static offset. ........................................ 208

Figure E-5. Surface wave depth sensitivity kernels for the western United States velocity model used in U.S. Geological Survey analysis (Herrmann et al., 2014). ......................... 209
List of Tables

Table 3-1. Regional moment tensor solutions for the nine events from December 2013 through March 2016..........................................................46

Table D-1. 1-dimensional velocity model used in finite fault model inversion (Section 2.2). .................................................................................................201

Table D-2. Coulomb failure stress statistics resulting from randomly perturbing aftershock locations and source parameters within their uncertainty bounds........201
Preface

The results presented in this dissertation were produced with contributions from co-authors outside my dissertation committee. In addition, several Chapters have been published in scientific journals; these manuscripts have been reprinted in this dissertation with permission from the copyright holders as specified here.

Chapter 2 was reprinted from *Earth and Planetary Science Letters*, Volume 447, Matthew W. Herman, Kevin P. Furlong, Gavin P. Hayes, and Harley M. Benz, Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, earthquake, pages 119-129, Copyright 2016, with permission from Elsevier. The computation of stress changes and surface displacements, sensitivity tests, and the interpretations of the results are entirely my own. G. Hayes and H. Benz relocated the earthquake and determined event source parameters, including finite fault models for the largest earthquakes.

Chapter 3 was reprinted from *Geophysical Research Letters*, Volume 43, Matthew W. Herman and Kevin P. Furlong, Revisiting the Canterbury earthquake sequence after the 14 February 2016 Mw 5.7 event, pages 7503-7510, Copyright 2016, with permission from the American Geophysical Union.

Chapters 4, 5, and 7 have not yet been submitted for publication and the work is my own, applying software developed by R. Govers.

Chapter 6 has been accepted for publication in *Journal of Geophysical Research: Solid Earth*, Volume 122 (Integrated geophysical characteristics of the 2015 Illapel, Chile, earthquake by Matthew W. Herman, Jennifer L. Nealy, William L. Yeck, William D. Barnhart, Gavin P. Hayes, Kevin P. Furlong, and Harley M. Benz). It has been reprinted here with permission from the American Geophysical Union. The stress change analysis, synthesis of the various geophysical approaches, all of the figures, and seismotectonic interpretations are my own. J. Nealy and G. Hayes determined the mainshock moment tensor and the seismic finite fault model. W. Barnhart determined the geodetic slip model. W. Yeck and H. Benz relocated the aftershock hypocenters. Each author composed the section corresponding to their specific analysis.

In Chapter 8, I describe results from papers in which I am a co-author. In Hayes et al. (2014a), I provided the stress change software. In Hayes et al. (2014b), I computed stress changes and produced figures for the manuscript. In Pananont et al. (2017), I determined the earthquake source parameters, computed stress changes, produced many of the figures, and composed and edited the manuscript.

Fortran software and shell scripts I developed to compute displacements, strains, and stresses in an elastic half-space using the equations of Okada (1992) are available on Github (https://github.com/mherman09). Tutorials for applying these codes can be found on the Penn State Geodynamics website (https://geodyn.psu.edu/geodyn_software.html). Versions of the software and tutorials current as of the submission of this dissertation are available on Penn State ScholarSphere (https://scholarsphere.psu.edu).
Acknowledgements

Completing this dissertation would not have been possible without the mentorship, advice, assistance, collaboration, and support from a huge number of people. First and foremost, I would like to thank my advisor, Kevin Furlong. I think that the best supervisors combine two characteristics; they challenge their students to avoid being satisfied with previous accomplishments and they support their students in pursuing research and their career goals. Kevin filled these roles admirably. I was able to avoid scientific complacency over the past few years in no small part because of his keen insights and his omnipresent red pen. He also encouraged me to leverage his broad scientific network, giving me every opportunity to learn new technical skills, interact with scientists across the globe, and develop broader scientific perspectives. I think this dissertation shows the large strides I have taken over the duration of my graduate career towards becoming a better scientist; the person most responsible for this progress is Kevin.

I would also like to thank the rest of my Ph.D. committee: Charles Ammon, Donald Fisher, Christelle Wauthier, Karl Reichard, and Rob Govers. Your interrogation at my defense (as well as at my other examinations), your careful reading of and comments on my dissertation, and our other conversations were invaluable. I deeply appreciate your help in improving the science within this dissertation and successfully communicating my ideas.

On a similar topic, I do not have enough positive words to say about the faculty in the Department of Geosciences at Penn State, particularly about the job they did teaching me in graduate school. This education was not strictly limited to the classroom, although the coursework I took at Penn State gave me a wide, robust foundation on which most of this dissertation research is built. Conversations with my professors and listening to their talks at events like the Department Colloquium and Geodynamics Seminar were transformative, helping me gain a more thorough understanding of various facets of my science that I may not have considered, as well as more generally learning how to ask big, important questions.

I have been extremely lucky to collaborate with outstanding scientists around the world. Many of these mentors and colleagues are my co-authors throughout this dissertation, but I wanted to acknowledge them here as well because they have been instrumental in facilitating my growth as a scientist. The researchers at the U.S. Geological Survey National Earthquake Information Center taught me an incredible amount about global seismology, the tools used to understand earthquakes, and communicating earthquake information to the public. Harley Benz and Gavin Hayes have essentially been secondary advisors to me. It has been a joy to collaborate with and learn from them, their colleagues, and the post-doctoral researchers (especially Will Yeck and Jen Nealy) at the NEIC. Robert Herrmann at Saint Louis University inspired me to improve my shell scripting and Fortran skills, which have turned out to be among the most valuable tools I developed in graduate school. This dissertation would have been significantly shorter if I had not learned finite element modeling from Rob Govers. I now appreciate how Rob rarely just gave me direct answers to my questions about tectonic processes, modeling, or the details of the software; this approach challenged me to come up with solutions for myself and I came out more self-sufficient because of it. I also had the great pleasure to work with Passakorn Pananont and his research group at Kasetsart University. I look forward to many more years of interesting collaborations on topics related to earthquakes and tectonics in southeast Asia.

The staff at Penn State also saved my skin numerous times. Even when I made errors on my paperwork, turned it in late, or otherwise made logistics more difficult for them, they always helped me with a positive attitude. Thank you for your hard work, which allowed me to focus on
the things I find most interesting, rather than spending my entire time figuring out Department and Graduate School logistics and bureaucracy.

Anyone who did not endure the crucible of the Geodynamics Lab will likely not appreciate how much personal and professional growth comes from being cloistered in a windowless room with a demanding advisor who likes to stop in every few hours. To all of the people who helped make each working day a bit brighter (irrespective of the weather outside), I sincerely thank you: Rachel, Matt, Beth, Jamie, Thamer, Haley, Kirsty, and all of the undergraduates and post-docs who dedicated time to the windowless space.

Perhaps just as critically, I discovered an awesome collection of friends at Penn State who have stuck up for me, put up with me, managed to keep me out of trouble, and kept my ego in check as much as humanly possible. Katie, Leah, TJ, Paul, Matina, John, Claire, Tim, Rosie, Matt, and everyone else from Penn State: you are all fantastic, and kinder than I deserve.

To my family: I know this graduate school endeavor began with uncertainties about my interests, my motivation, and my job prospects. From the beginning, through all of the highs and lows, and up to the moment this dissertation was finally finished, you remained completely supportive. More than anything, your willingness to listen to me and my complaints when computers were misbehaving or I was irritable due to fatigue and hunger, shows me just how much you love me. I continue to strive every day to make you proud of me, Mom, Dad, Becca, and Lilly.

And last, but certainly not least: Kate. I am simply astonished at your unwavering commitment to my success over the last four years, despite all the trials and tribulations to which you were subjected. As you know, and I admit, this Ph.D. has taken as much time and effort from me as several personal relationships. But I am so very glad that I invested the energy to cultivate a relationship with you, in addition to my science. Your relentless work ethic, resilience, and patience are daily inspirations. I am incredibly thankful to have a partner who was not simply satisfied with me finishing this dissertation, but driven to spur me on to greater heights every moment. I am lucky and privileged to call you my partner.
Chapter 1

Introduction

Over the past half-century, there have been significant improvements in the density and information content of observations of the slow accumulation and rapid release of strain during the earthquake cycle. Seismometers capture the rapid ground motions associated with co-seismic fault slip, while geodetic observations, including Geographic Positioning System (GPS) stations and Interferometric Synthetic Aperture Radar (InSAR), directly measure the ground position during both the rapid co-seismic and slower inter- and post-seismic stages.

A primary area of focus for these geophysical observations has been subduction zone plate boundaries, where many of the largest earthquakes over the past decade have occurred. These great megathrust earthquakes (moment magnitude Mw 8.0 and larger) produce severe shaking over a broad geographical region and often generate tsunamis that can be even more destructive than the shaking effects. Smaller, but also destructive earthquakes have also occurred in settings farther from major plate boundary structures; the physical processes operating in these earthquake sequences may inform our understanding of earthquake cycle deformation in subduction and other tectonic settings.

Although current geodynamic models provide reasonable fits to these observations, they do not fully exploit the information content of the geodetic and seismologic data. Recent great subduction megathrust earthquakes highlight the need for and potential benefits of developing enhanced approaches to modeling the subduction earthquake cycle. In this dissertation, I have developed improved methodologies to combine seismic, geodetic, and other relevant geophysical information, along with appropriately realistic numerical models of lithospheric deformation to better understand the processes responsible for a range of observed earthquake behaviors. Specifically, I have addressed the roles rheological structure and heterogeneity play in the acquisition and subsequent release of earthquake-cycle strain. This heterogeneity affects the distribution of elastic stresses during the loading process and how these stresses are released in an earthquake rupture. In an earthquake, stresses transferred onto nearby structures may promote subsequent seismic activity, so understanding the processes by which earthquakes are triggered by previous deformation is critical to constraining seismic hazards.
The dynamics of plate interactions at subduction zones depend strongly on the material properties throughout the lithospheric plates and at their interface. To first order, the plates accumulate elastic strain through motion across a frictionally coupled interface. However, the megathrust appears to be free to slide in some locations, which affects the spatial distribution of slip deficit on the fault and the loading stresses around the fault. In addition, aseismic or slow slip can have an important role in the subduction zone earthquake cycle. In some subduction zones, slow slip is observed during the inter-seismic loading stage down-dip or up-dip of the locked megathrust. This slow slip in turn may load the locked fault and change the timing or kinematics of the eventual megathrust earthquake. Also, slow slip may occur contemporaneous or following the co-seismic phase, as the subduction boundary becomes unlocked, changing how the system is deformed from the expected co-seismic deformation.

Another aspect of rheologic variation in the subduction system is the interaction of the time-scales of strain accumulation and release in the earthquake cycle and the rate-dependent properties of the lithospheric plates. Strain and stress accumulate over $10^2\text{-}10^3$ years ($10^9\text{-}10^{10}$ seconds; the “inter-seismic” period), but are released over $10^1\text{-}10^3$ seconds (the “co-seismic” earthquake rupture). During the inter-seismic period, the plate boundary zone consists of a locked fault patch and continuous, stable sliding (or creep) elsewhere along the fault. Additionally, at depth the upper plate will deform viscously, relaxing the stresses associated with plate motion deformation instead of accumulating elastic strain. As a result, only the upper parts of the plates, which are cold and able to store strain elastically, become significantly stressed during the inter-seismic period, accumulating quasi-uniform strain like a compressed beam.

During the co-seismic rupture, the previously locked fault zone slips freely while the system is locked down-dip (and up-dip?). At these more rapid co-seismic strain rates, the entire upper plate behaves elastically: the upper part of the plate rebounds and relaxes, while the previously relaxed lower part of the upper plate may actually become elastically loaded by the rebound. These rate-dependent properties produce quite different geometries, boundary conditions, and patterns of strain accumulation and release. There is also a poorly understood potential link between the rheologically-dependent patterns of elastic strain storage in the volume of the inter-seismically loaded lithosphere, and the role that seismic waves radiated during slip on the fault may play in aiding the release of this strain.

Despite these known rheological complexities, between earthquakes, models of inter-seismic behavior at subduction zones typically treat the subduction system as a uniform elastic half-space. These current models of subduction zone loading and megathrust locking geometries
do not consistently anticipate these subtleties of strain accumulation and release. In this dissertation, I have developed improved inter-seismic loading models and interpretations that link observed surface displacements throughout the earthquake cycle to lithospheric deformation at depth, and apply these results to more accurately anticipate the eventual earthquake behavior and magnitude.

During the co-seismic rupture, the lithospheric deformation, including surface displacements and stress transfer, is well represented by deformation of a completely elastic medium. As part of my dissertation research, I have created a suite of Fortran codes to compute these fault-generated elastic displacements and stresses, which have been applied in a variety of publications to predict GPS and InSAR observations as well as determine stress changes. In addition, this software is used at the U.S. Geological Survey National Earthquake Information Center to automatically and rapidly compute surface displacements for large earthquakes. These codes as they exist at the time of this publication are available on Penn State ScholarSphere.

This research addresses the relationship between lithospheric structure and strain accumulation and release during the earthquake cycle. The potential ability to resolve the details of both co- and inter-seismic deformation requires our thinking to expand beyond the current elastic half-space into a multi-rheology system. In this way can we resolve the importance of the mechanical/rheological details with respect to the broader strain accumulation and release processes at subduction zones. In turn, these will allow us to refine our hazard models and better anticipate earthquake kinematics and magnitudes at subduction zones globally.

**Chapter 2: Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, Earthquake**

In this Chapter, I introduce the concept of co-seismic stress transfer and the earthquake triggering hypothesis. I apply this analysis to the sequence of events that preceded the 2014 Iquique earthquake and conclude that the mainshock was triggered by the foreshocks. GPS observations of surface displacements prior to the mainshock cannot be explained by the seismicity alone, so I determined the dimensions and slip that would account for the extra displacements. This aseismic slip contributed to the loading of the mainshock region.
Chapter 3: Revisiting the Canterbury Earthquake Sequence After the 14 February 2016 Mw 5.7 Event

The concept of co-seismic stress transfer and earthquake triggering can be applied in many different seismotectonic settings. The 2016 New Zealand earthquake provided an opportunity to test these methods in an intraplate strike-slip setting. This event appears to have been triggered as part of the extended sequence of events following the 2010 Mw 7.0 Darfield earthquake. In addition, earthquake slip has been promoted in the region near the city of Christchurch and should be considered as a future seismic hazard location.

Chapter 4: Modeling Slip Deficit Accumulation Around Locked Regions on the Subduction Megathrust

Modeling of subduction zone inter-seismic coupling is typically performed using equations for fault slip in an elastic half-space, but these solutions do not accurately represent the frictional conditions on the subduction interface. In this Chapter, I develop simple, elastic-only finite element models with more realistic frictional characteristics on the megathrust. The key result is that regions that have only partial slip are being restricted by proximity to fully locked sections. These models indicate that apparently complex patterns of strain accumulation can be explained by relatively straightforward distributions of locked and unlocked regions. In addition, patterns of slip in subduction zone megathrust earthquakes fit in with this locked/unlocked framework.

Chapter 5: Interactions Between Asperities Over Multiple Earthquake Cycles

While Chapter 4 deals primarily with inter-seismic loading, in Chapter 5, I describe how adjacent asperities of different sizes with different earthquake recurrence intervals interact throughout the entire earthquake cycle. When a single segment of the fault embedded in a locked region ruptures, it may have less slip than expected by the amount of plate motion. Subsequently, if an earthquake ruptures multiple segments simultaneously, this larger event may have large slip in the segment that has already had an earthquake. These results are important for understanding the seismic hazard in locations like Cascadia and South America, where previous events may not have released the entire accumulated slip deficit.
Chapter 6: Integrated Geophysical Characteristics of the 2015 Illapel, Chile, Earthquake

The 2015 Illapel earthquake provided an opportunity to constrain earthquake rupture process and the relationship between mainshock slip and aftershocks in excellent detail. Along with several co-authors, I led a study in which we analyzed the mainshock and its aftershocks through a variety of geophysical approaches and combined these results into a seismotectonic framework for the rupture. The sequence revealed heterogeneous slip and aftershock behavior on the interface consistent with the modeling results presented in Chapters 4 and 5.

Chapter 7: Patterns of Loading and Unloading in the Upper Plate Throughout the Earthquake Cycle in Japan

The models and interpretations in Chapters 2-6 are primarily based on slow or static processes, without considering the transient deformation associated with wave propagation from a slipping fault. In Chapter 7, I combine these two approaches to explain the differences in surface displacement patterns in Japan before, during, and after the 2011 Tohoku earthquake. I compute the deformation prior to the earthquake with a finite element model that represents a rheologically more realistic upper plate. I then explore the observations recording the rebound during the earthquake and compute the stress changes associated with this elastic process. Finally, I consider the residual stress field after the earthquake and conclude that it is consistent with the post-seismic observations.

Chapter 8: Related Work

As part of my Ph.D. research, I have been involved with other studies related to the topics in my dissertation. In this Chapter, I describe three of these studies, my contributions to the research, and how the conclusions connect to my dissertation goals.
Chapter 2

Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, Earthquake

Abstract

On April 1st, 2014, a Mw 8.2 (U.S. Geological Survey moment magnitude) earthquake occurred in the subduction zone offshore northern Chile. In the two weeks leading up to the earthquake, a sequence of foreshocks, starting with a Mw 6.7 earthquake on March 16th and including three more Mw 6.0+ events, occurred predominantly south of the April 1st mainshock epicenter and up-dip of the area of significant slip during the mainshock. Using earthquake locations and source parameters derived in a previous study (Hayes et al., 2014) and a Coulomb failure stress change analysis of these events, we assess in detail the hypothesis that the earthquakes occurred as a cascading sequence, each event successively triggering the next, ultimately triggering the rupture of the mainshock. Following the initial Mw 6.7 event, each of the three largest foreshocks (Mw 6.4, 6.2 and 6.3), as well as the hypocenter of the mainshock, occurred in a region of positive Coulomb stress change produced by the preceding events, indicating these events were brought closer to failure by the prior seismicity. In addition, we reexamine the possibility that aseismic slip occurred and what role it may have played in loading the plate boundary. Using horizontal GPS displacements from along the northern Chile coast prior to the mainshock, we find that the foreshock seismicity alone likely does not account for the observed signals. We perform a grid search for the location and magnitude of an aseismic slip patch that can account for the difference between observed signals and foreshock-related displacement, and find that a slow slip region with slip corresponding to a Mw ~6.8 earthquake located coincident with or up-dip of the foreshock seismicity can best explain this discrepancy. Additionally, such a slow slip region positively loads the mainshock hypocentral area, enhancing the positive loading produced by the foreshock seismicity.

1. Introduction

On April 1st 2014, a Mw 8.2 earthquake ruptured the subduction zone plate boundary offshore of the city of Iquique, Chile, generating strong shaking and a tsunami along the coast of
southern Peru and northern Chile. This earthquake nucleated at a depth of 25 km, and the rupture propagated to the southeast, with the centroid located ~30 km southeast of the hypocenter at a depth of ~35 km (Hayes et al., 2014). Finite fault models of the event (e.g., Hayes et al., 2014; Lay et al., 2014; Ruiz et al., 2014; Schurr et al., 2014; Duputel et al., 2015) indicate peak slip of 5-10 m, and a rupture area (with slip greater than 1 m) extending ~80 km along strike and ~60 km along dip (Figure 2-1). This event was followed two days later by a Mw 7.7 aftershock ~100 km SSE of the mainshock hypocenter.

The Peru-Chile subduction zone, where the Nazca plate descends eastward beneath the South America plate at a rate of 70-80 mm/yr, has hosted numerous great (Mw 8.0+) earthquakes historically. Prior to April 2014, the segment of the subduction zone offshore northern Chile, just south of the Peru-Chile border, was recognized as a seismic gap, with its last great earthquake (M ~8.8) occurring in 1877 (Comte and Pardo, 1991, Figure 2-1). Studies using geodetic observations of surface deformation on the coast of South America also recognized the seismic potential of the region, having identified a locked section of the megathrust interface near where the April 1st event occurred (Chlieh et al., 2011; Bejar-Pizarro et al., 2013).

The 2014 Iquique earthquake was immediately preceded by a rich foreshock sequence, beginning on March 16th with a Mw 6.7 event 40 km south of the mainshock epicenter, and including three more Mw 6.2-6.4 events prior to April 1st. In the two weeks between the initial large foreshock and the mainshock, foreshock seismicity migrated northward from the March 16th epicenter towards the nucleation point of the April 1st mainshock. This sequential progression of seismicity appears to reflect cascading failure on the megathrust, in which stress changes associated with each foreshock helped trigger the next, ultimately leading to the main rupture (Hayes et al., 2014; Gonzalez et al., 2015). In this study, we assess this cascading failure scenario in improved detail. Slow slip contemporaneous with the foreshock sequence may also have played a role in promoting the mainshock rupture; however, the occurrence of such slip is debated (e.g. Kato and Nakagawa, 2014; Ruiz et al., 2014; Schurr et al., 2014; Bedford et al., 2015). Therefore, we test whether slow slip is likely to have occurred, and if it did, how it might have affected the nucleation process of the Iquique earthquake.

1.1. Overview of Foreshock Seismicity

This foreshock sequence began with a Mw 6.7 earthquake on March 16th 2014 at 21:16:29 UTC (16:16:29 local time) located at -70.772°E, -19.982°N, and a depth of 17 km (nodal
planes from the W-phase CMT solution for this event are strike=277º, dip=18º, rake=63º and 126º, 74º, 98º; Hayes et al., 2014; Figure 2-2). The location and faulting kinematics of this event are unusual in the context of the South America subduction zone geometry; although the event has one shallow-thrust nodal plane, the strike of that plane is rotated relative to the subducting Pacific slab (strike=335º), and its hypocentral depth implies it occurred in the upper plate, not along the plate interface, which at the epicenter is at a depth of approximately 25 km (Hayes et al., 2012; Figure 2-2). Gonzalez et al. (2015) observed that the strikes of the Mw 6.7 nodal planes match the average orientations of upper plate faults mapped onshore, consistent with the interpretation that the earthquake occurred in the upper plate on an offshore extension of these structures. However, there are few constraints on which of the two nodal planes is the fault plane, suggesting that either plane is a viable choice. Although this initial event occurred within the upper plate, the majority of subsequent foreshocks (63 events Mw 4.0 and larger from March 16th to April 1st) had locations and faulting kinematics compatible with their occurrence on the megathrust interface (Hayes et al., 2014; Figure 2-2). The foreshock seismicity migrated systematically northward at ~3 km/day along strike towards the hypocenter of the Mw 8.2 mainshock, remaining dominantly up-dip of the main slip region of the mainshock (Figure 2-2B).

2. Methods and Results

In this study, we assess whether the foreshock sequence can be considered a cascade of sequentially triggered events, that is, whether each foreshock made the next earthquake more likely to occur, leading up to the mainshock. To explore this scenario, as well as the role this sequence of foreshocks might have had in helping trigger and/or constrain the extent of the mainshock rupture, we have used relocated hypocenters and moment tensor kinematics of the foreshocks from March 16th to April 1st to compute Coulomb failure stress changes (\( \Delta CFS \)). Hayes et al. (2014) and Gonzalez et al. (2015) both previously proposed this cascading failure hypothesis, and used similar analyses to demonstrate that (a) most of the subsequent foreshocks occurred in areas loaded by the initial Mw 6.7 foreshock; and (b) the cumulative foreshocks loaded the mainshock hypocenter. We enhance these analyses by examining the event-by-event accumulation of \( \Delta CFS \) both on the megathrust interface and at the location of each event.

Earthquakes from March 16th through April 11th, including foreshocks and aftershocks of the Mw 8.2 mainshock and the April 3rd Mw 7.7 aftershock, were relocated by using a multiple-event, hypocentroidal decomposition algorithm (Jordan and Sverdrup, 1981) with local,
regional, and global seismic phase data. Earthquake source parameters (depth, moment magnitude, strike, dip, and rake) were constrained by W-phase inversion (Kanamori and Rivera, 2008) if they were large enough, or by computing regional moment tensors (RMTs) following the approach of Herrmann et al. (2011) for smaller events. A complete description and list of references for the techniques used to determine foreshock locations and source parameters can be found in Hayes et al. (2014). Similar to Hayes et al. (2014) and Gonzalez et al. (2015), we compute φCFS on the megathrust (as defined in Slab1.0; Hayes et al., 2012) since most of the foreshocks and the mainshock are interpreted to occur on that plate boundary interface. In addition, to leverage the hypocentral relocations and moment tensor solutions and avoid biases associated with defining events as plate boundary or intraplate, we resolve the φCFS directly onto foreshocks with moment tensor solutions. These results are compared to the φCFS assuming the same events occurred directly on the plate boundary interface. The cascading sequence hypothesis implies that most of the events, especially the larger ones, should occur in response to large, positive φCFS from preceding seismicity.

To compute φCFS, each event is transformed from a double couple moment tensor, i.e., a point source approximation, to a finite rectangular shear dislocation centered at the relocated hypocenter, with uniform slip. The exceptions are the Mw 8.2 mainshock and the Mw 7.7 aftershock, for which finite fault models (FFMs) were used to define the co-seismic slip. Fault dimensions are computed using empirical scaling relations between magnitude and fault size (e.g., Wells and Coppersmith, 1994; Mai and Beroza, 2000; Blaser et al., 2010; Yen and Ma, 2011), and slip is determined from the seismic moment equation,

\[ M_0 = \mu A d \]  

where \( A \) is the fault area, \( d \) is the slip, and \( \mu \) is the shear modulus (assumed to be 35 GPa). This shear modulus is consistent with the velocity models used for W-phase, RMT, and FFM solutions, as well as local velocity models (e.g. Dorbath et al. 2008). By treating the medium hosting these events as a uniform, isotropic, elastic half-space, the stress field produced by each dislocation can be computed (Okada, 1992). At each point in the foreshock sequence, the total change in stress field is the sum of the stress fields produced by all the previous earthquakes, starting with the March 16th Mw 6.7 foreshock. This stress field is resolved onto the plate boundary interface, assuming pure reverse faulting (i.e. a rake of 90°), to obtain traction vectors, which are partitioned into shear (\( \tau \)) and normal (\( \sigma_n \)) components to compute the Coulomb stress change (Reasenberg and Simpson, 1992),

\[ \Delta CFS = \tau - \mu \sigma_n \]  

9
In addition, we resolve the stress field onto the strike, dip, and rake of each event at its hypocenter. Using this approach, the evolution through time of ΔCFS on the megathrust plate boundary and at each successive event can be determined (Figure 2-3). Regions of positive ΔCFS are commonly associated with heightened seismicity; ΔCFS ≥ 0.01 MPa (0.1 bar) is thought to be a minimum threshold for triggering earthquakes (e.g., Stein et al., 1992). We consider any region with ΔCFS ≥ 0.01 MPa “positively loaded.” This analysis indicates that ~70% of the foreshocks were positively loaded. By taking into account the location uncertainties of each event (Hayes et al., 2014), ~80% of the foreshocks were positively loaded. To test whether this represents an increase in positively loaded seismic activity, we follow the procedure of Hardebeck et al. (1998) (c.f., Toda et al., 2011; Hayes et al., 2013), and also resolve the ΔCFS generated by the March 2014 foreshocks onto the nodal planes of 24 events from the Global Centroid Moment Tensor (GCMT) catalog occurring in the foreshock region shown in Figure 2-3A-E from January 2000 through February 2014 (Ekstrom et al., 2012; http://www.globalcmt.org). We find that only 11 of these GCMT events prior to the March 2014 foreshock sequence are positively loaded, implying that the rate of positively loaded seismic activity increased significantly during the foreshock sequence. In Section 2.1 we describe the ΔCFS evolution throughout the foreshock sequence (Mw 6.0+) in more detail.

2.1. Foreshock ΔCFS

The initial Mw 6.7 event on March 16th was the largest of the foreshocks, and as a result its changes to the stress field dominate throughout the foreshock sequence (Figure 2-3A). In the following analysis, since the choice of fault plane is poorly constrained, we tested both the shallowly dipping and steeply dipping nodal planes for the Mw 6.7 source fault. This choice of fault plane has a small but significant impact on the following analysis, especially near the Mw 6.7 hypocenter. Because the event appears to have been located in the upper plate and is oblique with respect to the plate boundary, the resulting pattern of ΔCFS on the plate boundary is not symmetric or particularly intuitive. The positive ΔCFS footprint on the interface (assuming the shallow-thrust fault plane) wraps around the event on its up-dip side, and only a narrow lobe of negative ΔCFS extends down-dip. In a dip-parallel cross-section through the Mw 6.7 epicenter (Figure 2-4A), the down-dip negative ΔCFS lobe is apparent as a large, negative ΔCFS zone beneath the event, but up-dip the slab is put into a lobe of large, positive ΔCFS. If instead we use the steeply dipping plane for the Mw 6.7 foreshock, the megathrust up-dip of the event becomes...
positively loaded all the way to the epicenter; this change is clear in the cross-section (Figure 2-4A,B). In general, the shallower the Mw 6.7 foreshock is, the smaller the positive ΔCFS region resolved on the megathrust interface, but the pattern of ΔCFS is mostly unchanged with depth. The precise position of the up-dip positive ΔCFS lobe also depends on the parameterization of the Mw 6.7 rectangular fault patch. The Wells and Coppersmith (1994) scaling relation for reverse earthquakes yields a fault 29 km along-strike and 14 km along-dip; varying these fault dimensions shifts the location of the positive ΔCFS lobe. Under most assumptions, much of the plate interface up-dip of the Mw 6.7 earthquake is positively loaded (Figure 2-4), in addition to the areas of positive ΔCFS to the north and south (Figure 2-3A). Importantly, the region of the mainshock hypocenter is not significantly loaded by this initial event, and specifically both the normal stress decrease (i.e. unclamping) and shear stress increase are less than 0.01 MPa in the hypocentral region.

In the two days following the initial Mw 6.7 foreshock, most of the earthquake activity occurred near the plate boundary interface within 15 km of the Mw 6.7 epicenter, dominantly up-dip to the west and south of the initial event (Figure 2-3B); very little activity is observed north of the Mw 6.7 epicenter towards the mainshock hypocenter. On March 17th, a Mw 6.4 earthquake occurred approximately 10 km west of the initial event, possibly near the up-dip edge of the Mw 6.7 inferred rectangular fault segment. Using (1) the shallowly dipping nodal plane for the Mw 6.7 fault, (2) the relocated hypocenters, and (3) the Wells and Coppersmith (1994) fault parameterizations, the March 17th Mw 6.4 earthquake was negatively loaded by the initial Mw 6.7 foreshock (Figure 2-3A,B; Figure 2-4A). However, the Mw 6.4 event occurred adjacent to the edge of the Mw 6.7 rectangular fault segment; as a result, the sign of the ΔCFS at the March 17th Mw 6.4 hypocenter is strongly dependent on the dimensions of the Mw 6.7 fault and the relative location of the two earthquakes. A small decrease (~15%) in the length of the Mw 6.7 fault, or increase (~5 km) in the distance between the two hypocenters moves the Mw 6.4 event into a positively loaded region, well within the uncertainties in both fault dimension and earthquake location. We tested three other empirical fault-scaling relations (Mai and Beroza, 2000; Blaser et al., 2010; Yen and Ma, 2011) to explore the range of reasonable fault dimensions for an earthquake of this size, with a specific focus on the along-strike length. While the Blaser et al. (2010) model yields a fault with similar length (28 km along-strike and 17 km along-dip), the other two scaling relations yield shorter lengths (Mai and Beroza, 2000: 18 km by 15 km; Yen and Ma, 2011: 23 km by 20 km), and move the Mw 6.4 event into a positively loaded region (Figure 2-4C). Alternatively, if we choose the steep nodal plane for the Mw 6.7 foreshock, the
March 17th Mw 6.4 event moves to a positive ΔCFS region irrespective of the scaling relation chosen (Figure 2-4B). Farther than ~15 km from the Mw 6.7 epicenter, the choice of fault plane does not significantly change the stress distribution, so without other constraints, in subsequent analyses we choose the shallowly dipping fault plane. The earthquakes in the days following the initial March 16th Mw 6.7 foreshock expand the spatial extent of the positive ΔCFS footprint on the megathrust (Figure 2-3B).

On March 22nd, a Mw 6.2 plate interface earthquake ruptured northwest of the Mw 6.7 epicenter in a region with positive ΔCFS, followed by several nearby Mw 5.0+ aftershocks over the next 24 hours (Figure 2-3C). These events expand the zone of positive ΔCFS on the plate interface along-strike to the north and down-dip to the east. One day later, a Mw 6.3 earthquake, the last of the Mw 6.0+ foreshocks, ruptured down-dip to the east of the March 22nd event, due north of the March 16th Mw 6.7 epicenter (Figure 2-3D). Although the March 23rd Mw 6.3 earthquake occurred within the footprint of positive ΔCFS on the megathrust interface, its relocated hypocenter is at 35 km depth, significantly deeper than the interface at this location (~25 km), placing it in a negative ΔCFS region. Other catalogs place the event at a shallower depth (< 30 km), in a positive ΔCFS region; for example, the GFZ hypocenter has a depth of 27 km (http://geofon.gfz-potsdam.de), the Global Centroid Moment Tensor project has a centroid 22 km deep (Ekstrom et al., 2012), and Schurr et al. (2014) relocate the event to a depth of 22 km. This is also the only event in the foreshock sequence that relocates more than ~5 km from the megathrust into the subducting plate (Figure 2-2C). Although we want to avoid selecting a depth to fit with our interpretation, the evidence suggests that this earthquake may have occurred closer to the megathrust, in a positive ΔCFS region. At either depth, this event adds to the positive loading at the mainshock hypocenter (Figure 2-3D).

For most of the other foreshocks, the deviation in depth from the megathrust interface does not have a significant impact on the ΔCFS for that event. As a test of this claim, we assume that all of the foreshocks with shallow thrust mechanisms and a northerly strike occurred on the megathrust interface, similar to Gonzalez et al. (2015). Moving these plate boundary foreshocks to the interface does not qualitatively change the ΔCFS footprint resolved onto the megathrust interface, but does tend to increase the magnitude of the stress change (Figure A-1). The majority of foreshocks do not change the sign of their ΔCFS after this vertical shift; 41 events remain positively loaded and 13 remain negatively loaded. Only 20 events experience a change in ΔCFS sign: 9 go from negative to positive (including the Mw 6.3 discussed above, as well as Mw 5.6 and 5.2 foreshocks), and 11 change from positive to negative (the largest of which is Mw 5.2). Of
these 20 events with changes in $\Delta CFS$ sign, 15 are Mw 5.0 and smaller, and these do not reflect a significant portion of the total deformation during the foreshock sequence. Therefore, our interpretation of the $\Delta CFS$ evolution remains unchanged irrespective of the precise choice of foreshock depths.

In the week before the April 1st Mw 8.2 mainshock, several foreshocks with magnitudes up to Mw 5.5 occurred near the mainshock hypocenter, a location with positive $\Delta CFS$ from the preceding events (Figure 2-3E). The cumulative effect of the foreshocks before April 1st, when resolved onto the plate boundary, was to significantly increase the $\Delta CFS$ at the mainshock hypocenter, consistent with previous analyses looking at the cumulative effect of the March 2014 foreshocks (Hayes et al., 2014; Gonzalez et al., 2015). We interpret this to indicate that the mainshock nucleation was likely promoted by the accumulated stress change of the preceding two weeks of seismicity. Although the foreshock sequence produces positive $\Delta CFS$ in the vicinity of the initial rupture of the mainshock, the region of greatest slip during this event has essentially no correlation with either the foreshock-produced $\Delta CFS$ or with the location of the foreshocks themselves. The main slip patch of the April 1st earthquake straddles lobes of both positive and negative $\Delta CFS$, and extends farther down-dip into a segment of the interface unaffected by the foreshock stress changes, suggesting that the down-dip part of the system was loaded by a mechanism other than the foreshock sequence. Therefore, we interpret that the foreshocks positively loaded the mainshock rupture nucleation site, potentially triggering its initiation. Subsequently, the rupture propagated down-dip into a region already loaded by other processes, with significant slip deficit, which was released during the Mw 8.2 mainshock.

The stress change resulting from the mainshock overwhelms the cumulative stress change from the foreshocks, expanding the footprint of positive $\Delta CFS$ on the megathrust from approximately 100 km along-strike by 100 km along-dip to 400 km by 400 km (Figure 2-3F). Many aftershocks followed the Iquique earthquake, occurring dominantly south of the epicenter coincident with the foreshocks, and south of the slip region (20.5º-21.0ºS). Most of the aftershocks ruptured in areas of increased $\Delta CFS$ generated by the mainshock slip. In particular, the largest aftershock (Mw 7.7) appears to have occurred in response to the mainshock static stress change. Prior to April 1st, the location of the April 3rd Mw 7.7 aftershock was unaffected by the foreshock $\Delta CFS$ (Figure 2-3A-E); in contrast, the Iquique mainshock generated a large positive $\Delta CFS$ in the vicinity of this largest aftershock. Thus, the April 3rd aftershock was not brought closer to failure by the foreshocks, but likely occurred in response to the mainshock.
2.2. Slow Slip

The underlying tectonic loading mechanism for the great 2014 Iquique earthquake is the long-term accumulation of strain and stress resulting from relative Nazca-South America plate motion over a locked portion of the plate boundary. The analysis above demonstrates how moderate seismicity in the subduction zone may perturb the effects of this loading process and potentially aid in the initiation of rupture of a great megathrust event. Another possible triggering mechanism for megathrust events is slow slip and related phenomena (Peng and Gomberg, 2010). Some recent laboratory studies of earthquake nucleation have identified a slow precursory slip phase prior to faster rupture, which might occur in the days to weeks before an earthquake and promote the future rupture (Latour et al. 2013). Previous analyses of the surface displacements and seismicity observed during the March 16th –April 1st foreshock sequence disagree as to whether aseismic slip occurred contemporaneous with the foreshock sequence (Kato and Nakagawa, 2014; Ruiz et al., 2014; Schurr et al., 2014; Bedford et al., 2015). If a large amount of aseismic slip did occur, depending on its location, timing, and magnitude, it could have significant consequences for the stress state on the megathrust and on the resulting sequence of earthquakes.

Surface displacements recorded onshore at several nearby GPS stations are the primary constraints on aseismic deformation in the months leading up to the Mw 8.2 mainshock (Ruiz et al., 2014; Schurr et al., 2014). These GPS data and their uncertainties have been digitized from those publications for analysis here (Figure 2-5). In both cases, long-term linear trends and seasonal periodic trends were removed, in addition to other processing steps. Both studies published displacement time series from stations IQQE and PSGA, and these time series overlap within their uncertainties at most points. Several daily positions do differ by more than their formal uncertainties, perhaps reflecting day-to-day variance, but in general the positions and their trends match well. This suggests that the signal is not strongly dependent on the difference in processing parameters between the two studies. The stations closest to the foreshock cluster and mainshock epicenter recorded displacements of up to 15 millimeters between March 16th, before the initial Mw 6.7 foreshock, and April 1st, immediately before the Mw 8.2 mainshock. Using this GPS dataset, Schurr et al. (2014) argued that the foreshock seismicity could account for all of the horizontal surface displacement that occurred. Bedford et al. (2015) expanded on this argument, demonstrating that the seismicity accounts for the observed displacements within the uncertainty in the modeling parameters, but they also identify two aseismic transients from March 16th-18th and 22nd-25th, which they acknowledge could be aseismic afterslip associated with the
larger foreshocks. In contrast, Ruiz et al. (2014) concluded that the foreshock seismicity only accounts for less than half the accumulated surface displacements, and located a patch of slow slip on the megathrust interface coincident with the site of peak slip in the Mw 8.2 mainshock, down-dip of the foreshock locations. Kato and Nakagawa (2014) also inferred slow slip using different observations; they located small repeating earthquakes in the foreshock sequence, and suggested that these events correspond to slow slip occurring up-dip of the mainshock region, which subsequently migrates towards the mainshock nucleation point.

We compare these pre-Mw 8.2 GPS datasets with displacements predicted using the relocated foreshock hypocenters and RMT kinematics to assess whether the foreshock seismicity accounts for the observed displacements, and attempt to quantify the uncertainty in the modeled displacements, similar to the analysis in Bedford et al. (2015). The mean displacement from January 1st to March 16th is removed from the digitized GPS time series to set the initial observed displacements to zero; likewise, modeled surface displacements are set to zero prior to March 16th. Using the same rectangular faults and elastic half-space properties as in the ΔCFS analysis above, we compute horizontal surface displacements at the GPS station locations resulting from these dislocation sources using Okada (1992) solutions for static displacements in an elastic half-space. These static displacements accumulated from March 16th through April 1st and are compared directly with the observed horizontal GPS time series (Figure 2-5).

Despite noise in the observed daily GPS time series, the displacements predicted from the seismicity appear to be systematically smaller than the observed GPS displacements. At most stations, horizontal displacements from foreshock seismicity account for only 50-75% of the observed signal. We consider the possibility that the misfit might be a consequence of events missing from the foreshock catalog; however, it is unlikely that earthquakes larger than M ~5.5 are missing, and any smaller events located in or around the foreshock region make a negligible contribution to the predicted surface displacements.

In order to rigorously test the robustness of the systematic misfit between observed and synthetic GPS displacements, we explore how various uncertainties affect the predicted displacements. The primary uncertainties are in the (a) earthquake source parameters; (b) earthquake locations; and (c) elastic half-space parameters, particularly the value of the shear modulus.
2.2.1. Earthquake source parameter uncertainty

We find that the predicted displacements are largely independent of the fault dimensions. With source-receiver distances (50 km and larger) much larger than the largest fault dimensions (15 km and smaller), the foreshocks effectively behave as point sources. To test this claim, the displacements were calculated using the Okada (1992) equations for point sources instead of finite rectangular sources, and the resulting displacements were unchanged. The predicted surface displacements do show slight sensitivity to changes in the kinematics of the foreshocks, especially the March 16th Mw 6.7 event, because it is the largest foreshock and closest to the coast. The displacements generated by the GCMT solution for the March 16th Mw 6.7 foreshock (strike=358°, dip=56°, rake=74°; Ekstrom et al., 2012) are 5-10% larger than those generated by the U.S. Geological Survey (USGS) W-phase solution, whereas the displacements generated by the USGS body wave moment tensor solution (strike=258°, dip=24°, rake=44°; http://earthquake.usgs.gov/earthquakes/eventpage/usc000ndnj) are 5-10% smaller (Figure 2-5). Variations in the kinematics of the other, smaller events in the sequence have less of an effect on the surface displacements; for example, altering the kinematics of the next largest foreshock (Mw 6.4) within its uncertainty changes the predicted surface displacements by 2-3%. Accounting for uncertainties from all the events, the total uncertainty due to kinematics is less than 15%.

2.2.2. Earthquake location uncertainty

The location uncertainty of the foreshocks determined using the relocation algorithm is typically 3-6 km horizontally and ~4 km vertically, which is small compared to the source-receiver distance, and has a minimal impact on the predicted displacements. We quantified the effect of location uncertainty by randomly perturbing the foreshock positions within their location uncertainty ellipses. This test indicates that the modeled displacements could vary by up to ~5% (Figure 2-5). This is particularly evident at station PSGA, but is almost negligible at the other stations. Therefore, it seems that location uncertainty alone cannot account for the discrepancy between modeled and observed surface displacements.

2.2.3. Modeling uncertainty

The previous sections suggest that uncertainties in the foreshock source parameters are unlikely to account for the systematic misfit between predicted and observed GPS displacements,
leaving only uncertainties in the modeling parameters. One possible source of misfit is not including appropriate rheological complexity such as elastic heterogeneity or visco-elasticity. The effects of lateral elastic heterogeneity have been explored in transform settings (e.g., Schmalzle et al., 2006; Lindsey and Fialko, 2013), suggesting that even elastic modulus variations of a factor of 2 lead to relatively small displacement differences. Viscous flow also should not play a significant role in observed displacements over the relatively short time scale of two weeks, assuming realistic subduction zone viscosities (Turcotte and Schubert, 2002).

We find that the dominant uncertainty is the assumed value of the shear modulus. Specifically, the displacement generated by a fault in an elastic half-space is proportional to the product of the fault slip and fault area. Fault area is determined by empirical relations (and was shown above to have essentially no effect on the displacements), and fault slip is inversely proportional to the shear modulus (Equation 1). The shear modulus in the lithosphere is not precisely known, and therefore the fault slip is unknown. The shear modulus used in lithospheric-scale deformation models typically ranges from 25 to 50 GPa; we choose a value of 35 GPa, consistent with the velocity models used for source inversion (Hayes et al., 2014). This implies that our fault slip estimates, and hence our seismically modeled GPS displacements, could potentially be up to ~35% larger for a shear modulus of ~25 GPa (Figure 2-5). This would account for the entire misfit between observed and modeled displacements. Although independent studies also indicate a higher shear modulus for this part of the Peru-Chile subduction zone (35-40 GPa; e.g., Dorbath et al., 2008), we cannot entirely rule out a lower shear modulus.

One argument for a low shear modulus, and hence large fault slip, is that strongly sheared fault zones typically have lower elastic moduli than the undeformed rock around them. We have run several numerical models with a relatively weak fault zone embedded inside a stronger surrounding medium, and indeed these require greater fault slip to match observed surface displacements. However, when this same large slip value is applied to the fault in an elastic half-space model, it produces displacements that are too large. We find that the appropriate fault slip value to use in half-space modeling is the value corresponding to the higher shear modulus surrounding material.

For completeness, we also test the sensitivity of the modeled surface displacements to the Poisson’s ratio of the material. Crustal rocks are often assumed to have Poisson ratios ranging from 0.20 to 0.30 (Vutukuri et al., 1974). We find that the predicted surface displacements vary by no more than 5% with respect to Poisson’s ratios in this range (Figure 2-5).
2.2.4. Comparison with Previous Studies

Three previous studies have compared observed GPS displacements with displacements predicted by seismicity (Ruiz et al., 2014; Schurr et al., 2014; Bedford et al., 2015). Of these, only Bedford et al. (2015) explored earthquake and modeling uncertainties. They used foreshock locations and source parameters from the GEOFON catalog (http://geofon.gfz-potsdam.de), and applied uniform uncertainties to the earthquake locations (±15 km north and east, ±10 km vertical), source kinematics (±15º strike and rake, ±10º dip), and half-space shear modulus (28-44 GPa). In this study, we use foreshock source parameters from Hayes et al. (2014); these focal mechanisms agree with the GEOFON solutions within the range proposed by Bedford et al. (2015), suggesting the focal mechanism uncertainties are reasonable. We enhance the earthquake location estimates by using hypocenters relocated by Hayes et al. (2014); this yields improved absolute locations and tighter constraints on these locations (±5 km north and east, ±4 km vertical). We also provide a more detailed assessment of how each parameter uncertainty maps into the predicted surface displacement uncertainty; the contributions are the earthquake locations (<5% displacement uncertainty), focal mechanisms (~15%), half-space shear modulus (~35%), and Poisson’s ratio (<5%). It is clear from these values that the choice of shear modulus is the most likely modeling parameter to account for the observed GPS. But, as we showed in Section 2.2.3, the shear modulus is unlikely to be low enough to account for the entire observed-predicted GPS misfit.

2.3. Slow Slip Modeling

Based on the analysis presented above, we interpret the difference between observed GPS and displacements predicted by the seismicity to be meaningful. This suggests that there is additional, aseismic deformation occurring contemporaneous with the foreshock seismicity, which we suggest to be slow slip on the megathrust interface, consistent with Ruiz et al. (2014) and Kato and Nakagawa (2014). It is possible that this slow slip could have a significant impact on the stress conditions in the vicinity of the foreshocks and the mainshock, since it accounts for ~25% of the observed surface deformation. Therefore, constraining this possible aseismic slip location and magnitude is important in determining its impact on the ∆CFS evolution of the slab interface.

Here, we assume that aseismic slip accounts for the entire difference (Δu) between predicted and observed horizontal surface displacements and solve for its probable location.
Recognizing that the relatively sparse data cannot support a complex model, the slow slip is assumed to be a single, rectangular patch with uniform slip that produces displacements \( u_{ss} \) at each station. A grid search is performed to find the location, dimensions, and slip amplitude of this patch that minimizes the sum of the magnitudes of residual horizontal displacement vectors:

\[
e(\vec{x}, L, W, s) = \sum_{i=1}^{n} \left| \Delta u_i - u_{ss,i} \right|
\]

where \( e \) is the misfit function to be minimized, as a function of slow slip position \( (x) \), dimensions \( (L, W) \), and slip \( (s) \). Modeled slow-slip displacements are computed using Okada (1992) solutions for rectangular faults in an elastic half-space. The center of the slow slip rectangle is varied over the region shown in Figure 2-6, in increments of 0.04 degrees, and the event depth, strike, and dip are taken from Slab 1.0 at each location (Hayes et al., 2012). The rake angle for the aseismic patch is held constant at 90°, consistent with slip in the foreshocks and mainshock. At each location, the along-strike fault length is varied from 25 to 150 km, with the down-dip fault width held constant at 25 km. Preliminary searches demonstrated that the quality of the fit was mostly insensitive to fault width. For each grid point, the slip on the patch is scaled to minimize the sum of the residual displacements.

The results of this grid search suggest that the center of the aseismic slip rectangular fault patch is likely coincident with or up-dip of the foreshock region, and that the slow slip patch extends ~100 km along-strike, roughly the same along-strike extent as the foreshock sequence (Figure 2-6). We find that the misfit, \( e \), is more sensitive to the location of the slow slip center than the length of the patch; moving the center by 30 km to the north or south of the preferred region reduces the quality of the fit (increasing \( e \) by ~10 mm) comparably to the maximum effect of varying fault length at any location. The slip computed for an aseismic patch 100 km long and 25 km wide, located just up-dip of the foreshocks, is ~0.20 m (Figure 2-6). Assuming a shear modulus of 35 GPa, these dimensions and slip yield a moment of \( 1.75 \times 10^{19} \) Nm, equivalent to Mw 6.8. At this size, the inferred aseismic slip should have a significant impact (slightly greater than the initial March 16th Mw 6.7 event) on the state of stress on the megathrust during the foreshock sequence. The GPS observations suggest that the slow slip occurs essentially contemporaneously with the earthquakes making up the foreshock sequence, perhaps during two pulses associated with the largest foreshocks (Bedford et al., 2015). We propose two scenarios for the specific location of the slow slip consistent with the timing indicated by Bedford et al. (2015) and our grid search results: (1) the slow slip occurs coincident with the foreshocks as a series of slow slip events associated with each of the foreshock seismic events; or (2) it occurs spatially
and temporally independent of the foreshocks, likely farther up-dip than the seismicity. The grid search suggests the latter interpretation is preferred, but does not preclude slow slip being coincident with the foreshocks (Figure 2-6), and a combination of the two scenarios above is also possible.

The two scenarios have slight differences in how they affect the Coulomb stress evolution on the megathrust interface (Figure 2-7). In scenario 1, the effect of adding a coincident patch of slow slip to each foreshock is to enhance the amplitude of ∆CFS, without significantly changing the ∆CFS distribution from the pattern computed using the seismicity alone (Figure 2-7B). In scenario 2, slow slip occurring up-dip of the foreshocks loads the entire area hosting foreshocks (Figure 2-7C). Both scenarios result in increased positive ∆CFS at the mainshock hypocenter, but differ in the loading produced up-dip of the foreshock seismicity; scenario 1 results in a positively loaded plate boundary up-dip of the seismicity, whereas scenario 2 produces negative ∆CFS on the up-dip interface. The lack of moderate seismicity in this up-dip region is more consistent with the aseismic slip occurring up-dip. Although we cannot resolve the fine scale spatial patterns of ∆CFS resulting from aseismic slip, we can say that the apparent aseismic slip increases ∆CFS at the Mw 8.2 hypocenter by 0.1-0.2 MPa. In addition, in both scenarios, there is an increase in the down-dip extent of positive ∆CFS, placing a larger portion of the mainshock slip region into a positively loaded stress condition (Figure 2-7).

3. Discussion

Although the first event of the 2014 Iquique sequence, on March 16th, occurred within the upper plate, it had a significant influence on the state of stress of the plate boundary. Taking into account various location, scaling, and modeling uncertainties, this event appears to be the catalyst for the series of sequentially triggered events on the plate interface, and its ∆CFS footprint was a dominant control on the location of subsequent seismicity. Although we might expect heightened aftershock activity in the upper plate following the initial Mw 6.7 earthquake, the upper plate became relatively seismically quiescent compared to the plate boundary. There is small-scale heterogeneity and uncertainty in the three-dimensional ∆CFS distribution related to event locations and fault scaling, but the vast majority of earthquakes in the foreshock sequence occurred near the megathrust interface, within the broadly positive ∆CFS footprint established mostly by the Mw 6.7 and three other Mw 6.0+ foreshocks. This supports our interpretation that
the foreshocks can be considered to represent a cascading sequence, in which each event was triggered by the events preceding it.

Similar to its Mw 6.0+ foreshocks, the April 1st Mw 8.2 mainshock nucleated in a location that was positively loaded by preceding seismicity, suggesting its initiation was also triggered by the previous foreshocks. Even without any contribution from slow slip, the cumulative $\Delta CFS$ resulting from the foreshock sequence was sufficient to load the Mw 8.2 hypocenter. However, the pattern of $\Delta CFS$ resulting from foreshock seismicity does not correlate with the region of dominant slip during the April 1st Mw 8.2 mainshock, suggesting additional loading mechanisms. One possible loading mechanism is the stress change from the initial slip at the mainshock hypocenter. Most of the kinematic finite fault models for the Iquique mainshock (e.g., Hayes et al., 2014; Lay et al., 2014; Ruiz et al., 2014; Schurr et al., 2014; Duputel et al., 2015) indicate that the rupture began at the north end of the foreshock seismicity, and grew east (down-dip) and south (along-strike) into the main slip region down-dip of the foreshocks ~30 seconds later (Figure 2-3). Although we do not perform a full dynamic simulation of the stresses during the rupture, we estimate the effect of a small preliminary slip event at the hypocenter on the stress state of the megathrust interface. If we treat the initiation of the mainshock as a M ~6.5 thrust earthquake at the hypocenter, it adds significant loading down-dip to the east of the hypocenter, in the northern part of the main rupture patch. In particular, stresses generated near the Iquique mainshock hypocenter are not reduced by the negative stress lobe from the initial Mw 6.7 foreshock (Figure 2-3). We also note that moving the 17, 22, and 23 March Mw 6.0+ foreshocks to the plate boundary produces a similar stress effect, generating an area of positive $\Delta CFS$ on the subduction interface immediately east of the hypocenter. Both of these scenarios are consistent with the mainshock beginning as a triggered event similar to the previous Mw 6.0+ foreshocks, and growing eastward into the down-dip region that was initially loaded, before expanding along strike.

After the rupture, the Mw 8.2 mainshock clearly loaded the area that hosted the largest aftershock (Mw 7.7) two days later, a section of the megathrust that was unaffected by the foreshocks. Such a large aftershock generates additional strong shaking, which can be especially dangerous to structures weakened by the main event. In cases like the Iquique sequence, $\Delta CFS$ analysis can help identify areas that are more likely to host these large subsequent events. For example, the February 2011 Mw 6.1 Christchurch earthquake occurred in an area positively loaded by the September 2010 Mw 7.0 Darfield earthquake, and the more recent April 2014 Mw 7.8 Gorkha, Nepal, earthquake appears to have enhanced the stress at the location of its May 2014
7.3 aftershock. Although ∆CFS itself is not a predictive tool, these cases illustrate that it can constrain which nearby faults have been loaded and may present future hazards.

Finally, although the formal uncertainties between modeled and observed GPS overlap, the modeled displacements generally fit the observed well only if the shear modulus becomes very low, which is somewhat inconsistent with other constraints on the local shear modulus. This suggests some aseismic slip occurred during the foreshock sequence. By adding a segment of slow slip slightly up-dip of the foreshocks, the loading at the mainshock hypocenter (and at the foreshocks) is enhanced, as is the loading of the main slip patch of the mainshock. The correlation between the down-dip loading extent and the location of the main slip patch is striking, leading us to speculate that the stress contribution from the inferred aseismic slip may have exerted some control over the segment of fault that slipped the most in the mainshock. This analysis suggests that observations of slow or aseismic slip events at subduction zones might be critical for understanding the evolution of loading on the megathrust interface, and may help place constraints on the locations of subsequent great events.

Acknowledgements

This work was supported by NASA Earth and Space Science Fellowship 15-EARTH15R-0096. Many of the figures in this manuscript were created using the Generic Mapping Tools (Wessel and Smith, 1991). We thank J. Nealy and two anonymous reviewers for constructive reviews that helped improve this manuscript.

References


Figure 2-1. Tectonic setting of the 2014 Iquique earthquake sequence.

Rupture areas of historic earthquakes in the subduction zone are outlined by dotted pink lines. Foreshock seismicity from March 16th through the April 1st Mw 8.2 mainshock is shown in red, seismicity from the mainshock to the April 3rd Mw 7.7 aftershock is shown in orange, and seismicity from April 3rd to June 6th is shown in yellow. The circles are scaled by event magnitude. The rupture area of the mainshock (slip greater than 1 meter) is shown in grey, and shows most of the slip occurring dominantly down-dip of the foreshock events.
Figure 2-2. Detailed view of 2014 Iquique sequence seismicity.

(A) Focal mechanisms in map view colored and scaled the same as in Figure 2-1. Most of the earthquakes have focal mechanisms consistent with occurring on the plate interface (light gray dashed lines, from Hayes et al., 2012), except for the initial Mw 6.7, which is rotated significantly counterclockwise from the other focal mechanisms. The dashed blue line indicates the location and orientation of the cross-section (Panel C). (B) Earthquake latitude versus time. The red arrow highlights the northward migration of foreshock seismicity towards the mainshock hypocenter. (C) Cross-section through the subduction zone, with focal mechanisms shown in side view looking from the south. All earthquakes from Panel A are projected onto this cross-section. The dashed line is the subduction plate boundary from Slab 1.0 (Hayes et al., 2012), and the solid line is ETOPO1 seafloor bathymetry along the cross-section in Panel A (Amante and Eakins, 2009).
Figure 2-3. Evolution through time of the Coulomb failure stress change.

$\Delta CFS$ is resolved onto the megathrust interface (assuming a rake of 90°) and onto the fault plane of each event (with the rake given by the focal mechanism). Focal mechanisms are colored by the sign of the $\Delta CFS$ resolved onto the fault plane, with positive colored red and negative blue. The stars indicate the epicenters of the Mw 8.2 and Mw 7.7 earthquakes, and the grey lines are the 1 m slip contours of the same events. (A) The Mw 6.7 earthquake on March 16th (choosing the shallowly dipping fault plane) establishes the initial pattern of $\Delta CFS$ on the megathrust, which extends lobes of positive $\Delta CFS$ to the north and south. (B) A Mw 6.4 earthquake on March 17th updip of the initial foreshock occurs on the plate interface and expands the zone of positive $\Delta CFS$ to the north and farther updip. (C) A Mw 6.2 earthquake on March 22nd occurs on the plate boundary north of the initial set of foreshocks. This event pushes the region of significant positive $\Delta CFS$ to the mainshock hypocenter. (D) The final large foreshock (Mw 6.3) occurs on March 23rd, putting the mainshock hypocenter squarely within a positively loaded area. (E) Several foreshocks occurred near the nucleation site of the mainshock in the week preceding the event. (F) The April 1st Mw 8.2 earthquake has a larger impact on the $\Delta CFS$ on the plate interface than all of the foreshocks combined, so this map is shown at a smaller scale. Prior to the mainshock rupture, the Mw 7.7 aftershock site was not significantly affected by the foreshock seismicity, but the mainshock increases the $\Delta CFS$ at the location of the Mw 7.7 event (c.f. Panels E and F).
Figure 2-4. Cross-sections of ∆CFS resolved onto the megathrust.

∆CFS is shown for different source parameters of the March 16th Mw 6.7 foreshock. The cross-section is drawn through the hypocenters of the Mw 6.7 and Mw 6.4 foreshocks, as shown in the inset map. Because the nodal planes of the initial foreshock are oblique to the cross-section, we shade the portion of the fault on the near side of the cross-sectional plane and outline the section on the far side with a dashed line to show perspective. (A) Using the shallowly dipping nodal plane and Wells and Coppersmith (1994) scaling, the Mw 6.7 fault is 29 km along-strike and 14 km along-dip. The Mw 6.4 foreshock occurs near the up-dip edge of this segment, and in this parameterization is negatively loaded. (B) If the steeply dipping plane is used for the Mw 6.7 source fault, then the Mw 6.4 event is positively loaded for any scaling relation (Wells and Coppersmith, 1994, is used here). (C) With a different scaling relation (Mai and Beroza, 2000) applied to the shallow-thrust fault plane, the source fault becomes more compact (18 km by 15 km), moving the Mw 6.4 event into a positively loaded region.
Figure 2-5. Daily east-west and north-south components of GPS displacement.

Data (stations shown in Figure 2-6) digitized from Ruiz et al. (2014) (red) and Schurr et al. (2014) (blue). The time of the initial March 16th Mw 6.7 foreshock is indicated by a dashed line 16 days before the mainshock. Superimposed are modeled seismic displacements: the dark green line is the displacement generated by placing each foreshock at its relocated hypocenter, using the fault plane solutions from Hayes et al. (2014), and giving the elastic half-space a shear modulus of 35 GPa; lighter green lines represent the effect of perturbing the foreshock locations and source parameters; light blue lines represent the effect of perturbing locations, source parameters, and elastic properties of the half-space; the pink line represents using the GCMT solution for the March 16th Mw 6.7 foreshock; and the orange line represents using the USGS body wave moment tensor solution for the Mw 6.7 foreshock. Notice that the seismically derived displacements systematically under-predict the observed GPS displacements, particularly for the east-west component.
Figure 2-6. Results of the grid search for a rectangular slow slip patch.

The search is for a patch that best fits the extra horizontal displacements (green vectors) at GPS stations (grey) along the northern Chile coast. Better fits (a lower value of the misfit function, $e$) are indicated by darker blue colors, and the most likely location for the excess slip is coincident with or up-dip of the foreshock seismicity. A representative fault solution with a good fit is shown in the black outline, and with a slip of 0.20 m produces the displacements shown in orange.
Figure 2-7. Effects of an aseismic slip patch on the $\Delta$CFS on the plate boundary immediately before the April 1st mainshock.

(A) The cumulative effect of seismicity alone on the $\Delta$CFS (Figure 2-3E). (B) Adding slip at the location of each foreshock amplifies the $\Delta$CFS pattern from panel A, but does not change the distribution of positive and negative $\Delta$CFS. (C) In contrast, adding a rectangular aseismic slip patch up-dip of the foreshocks, as shown in Figure 2-6, with ~0.2 m of displacement, positively loads the foreshock and mainshock hypocentral region, as well as a portion of the mainshock slip region. It also creates a large area of negative $\Delta$CFS up-dip of the foreshocks.
Chapter 3

Revisiting the Canterbury Earthquake Sequence After the 14 February 2016 Mw 5.7 Event

Abstract

On 14 February 2016, an Mw 5.7 (GNS Science moment magnitude) earthquake ruptured offshore east of Christchurch, New Zealand. This earthquake occurred in an area that had previously experienced significant seismicity from 2010 to 2012 during the Canterbury earthquake sequence, starting with the 2010 Mw 7.0 Darfield earthquake and including four Mw ~6.0 earthquakes near Christchurch. We determine source parameters for the February 2016 event and its aftershocks, relocate the recent events along with the Canterbury earthquakes, and compute Coulomb stress changes resolved onto the recent events and throughout the greater Christchurch region. Because the February 2016 earthquake occurred close to previous seismicity, the Coulomb stress changes resolved onto its nodal planes are uncertain. However, in the greater Christchurch region, there are areas that remain positively loaded, including beneath the city of Christchurch. The recent earthquake and regional stress changes suggest that faults in these regions may pose a continuing seismic hazard.

1. Introduction

On 14 February 2016, an Mw 5.7 (GNS Science moment magnitude; http://www.geonet.org.nz/quakes/region/newzealand/2016p118944) earthquake ruptured offshore east of Christchurch on the South Island of New Zealand. This event occurred in the vicinity of the earlier Canterbury earthquake sequence, which began with the 3 September 2010 Mw 7.0 Darfield earthquake and included Mw ~6.0 events on 21 February, 13 June, and 23 December 2011 (Figure 3-1) (Quigley et al., 2016, and references therein). Following the decay of the aftershocks associated with the 23 December 2011 earthquake, seismicity in the region returned to a level of relative quiescence (~1.8 events ML (GNS Science local magnitude) 2.0+ per day from June 2012 to February 2016; this rate is still substantially higher than the ~0.2 events per day before the 2010 earthquake; Figures 3-1b and 3-1c and B-1 in Appendix B). From June 2012
to December 2014, there were no earthquakes larger than ML 4.7, and from January 2015 up to the February 2016 event, there were no earthquakes larger than ML 4.0.

Although the 2016 earthquake did not cause significant damage or casualties, it generated significant shaking, with measured peak ground accelerations of up to 36% g (http://info.geonet.org.nz/display/appdata/Strong-Motion+Data). It also suggests that residual increased seismic rupture probability remains in the general region of the Canterbury earthquake sequence, even years after the initial main shock and more than 4 years since the occurrence of the last significant event.

Placing this recent earthquake into context requires us to look back on the Canterbury earthquake sequence, revisit interpretations of kinematics, and reevaluate the implications for future seismotectonic activity in the Christchurch vicinity described during or soon after the completion of the main sequence (e.g., Fry and Gerstenberger, 2011; Sibson et al., 2011; Beavan et al., 2012; Ristau et al., 2013; Herman et al., 2014; Steacy et al., 2014). In this study, we explore the location of this recent event to assess whether it occurred in a seismic gap along one of the previously active fault strands of the Canterbury sequence or, alternatively, if it occurred on a distinctly separate, previously inactive structure. Earthquakes occurring in seismic gaps on active structures are generally thought to be constrained in size and location by the dimensions of the gap and the remaining slip deficit (e.g., McCann et al., 1979; Fialko, 2006). In contrast, if significant subsequent earthquakes occur on structures other than those associated with previously active faults, i.e., on faults that have not yet hosted significant seismic activity, then the population of faults that could possibly host a later earthquake is increased. Thus, the residual hazard after an earthquake sequence like the Canterbury sequence could remain higher than previously anticipated and be partly controlled by regional stress conditions and the number and size of available faults.

2. Methods

To analyze the February 2016 event and its aftershocks in comparison with earlier seismicity in the Canterbury Plains from 2010 to 2014, we determine earthquake fault parameters (strike, dip, rake, seismic moment/magnitude, and centroid depth) for 11 earthquakes from January 2014 to March 2016. For consistency with previous results, we implement the regional moment tensor (RMT) inversion technique described in Herman et al. (2014) and references therein. This algorithm uses the Computer Programs in Seismology software (Herrmann, 2013) to
perform a grid search for the double-couple point source that best fits regional broadband waveforms from seismic stations on South Island, New Zealand (Herrmann et al., 2011). These waveforms are processed by: (a) deconvolving to ground velocity in m/s; (b) rotating to vertical, radial, and transverse components; (c) truncating from 10 s before the P wave arrival to 120 s after the P wave arrival; and (d) bandpass filtering typically from 0.02 to 0.0625 Hz (50-16 s).

The 14 February 2016 earthquake was large enough that it required a longer period passband (0.01-0.05 Hz) to maintain a point source approximation. Synthetic waveforms are generated using a regional South Island, New Zealand velocity structure and are processed identically to the observed traces. This velocity model is the same as in our previous study and was derived from surface wave dispersion measurements at the same regional seismic stations (Herman et al., 2014). We are able to determine RMT solutions for earthquakes as small as Mw 3.6 and consistently determine solutions for nearly all events Mw 4.5 and larger. Fault planes are interpreted for all 161 RMT solutions based on the orientation of the nodal plane most closely aligned with nearby aftershock cluster orientations, although fault planes for some events remain ambiguous, particularly in the complex deformation regions near 172.4ºE and 172.8ºE (Figure 3-2; Herman et al., 2014).

The double-difference hypocenter relocation software hypoDD (Waldhauser and Ellsworth, 2000) is used to determine relative locations of all 161 earthquakes with RMT solutions, from the September 2010 Darfield earthquake to the most recent aftershock of the February 2016 event in March 2016. We picked 3833 P wave arrivals from the vertical components of local strong motion and regional broadband stations and 1784 S wave arrivals from the transverse components of regional broadband stations. P wave picks are assigned a weight of 1.0, while S wave picks are assigned a weight of 0.2, reflecting greater uncertainty in the shear wave arrival time. The locations used to initialize hypoDD are the GNS Science hypocenters. Multiple configurations of the hypoDD event-linking and inversion schemes were used to test the robustness of the relative relocation solutions.

Static stress changes are computed by placing the earthquakes in an elastic half-space and using a code written by the authors to apply the equations of Okada (1992). We treat all of the sources as having finite fault areas. The 2010 Darfield mainshock geometry and slip distribution is defined by published finite fault models (e.g. Hayes, 2010; Beavan et al., 2012). The aftershocks are converted from point source moment tensor solutions to finite rectangular shear dislocation sources of equivalent moment with uniform slip. Empirical relations between magnitude and fault dimensions appropriate for South Island, New Zealand earthquakes define
the fault dimensions (Yen and Ma, 2011; Stirling et al., 2013), and slip is computed from the seismic moment equation, assuming the half-space has a shear modulus of 40 GPa (and a Poisson’s ratio of 0.25). To compute ΔCS (Reasenberg and Simpson, 1992), shear and normal stresses are resolved at the location of a target structure, onto the geometry of the structure, with a coefficient of friction of 0.5. At each location, the stress change value is the cumulative sum of the stress change contributions from all of the preceding seismicity, beginning with the 3 September 2010 Darfield mainshock. In particular, ΔCS is resolved onto both nodal planes of the February 2016 Mw 5.7 event. The earthquake is placed at its relocated epicenter (172.76ºE, 43.50ºS) at the centroid depth derived from the RMT analysis (9 km). This depth is very similar to the relocated hypocenter depth (8 km); therefore, the choice of centroid or hypocenter depth does not make a significant difference in the calculated ΔCS value for this event.

To explore what parts of the Canterbury Plains remain positively loaded, we also determine the cumulative ΔCS in the region from the 156 earthquakes from September 2010 up to (but not including) the February 2016 Mw 5.7 event, and consider the additional ΔCS generated by the February 2016 event and its aftershocks. This regional ΔCS is computed at a depth of 10 km, the mean depth of our RMT centroids. We resolve the ΔCS onto three different fault orientations representing the dominant kinematic styles of the 2010-2013 Canterbury earthquakes: E-W right lateral strike-slip, NW-SE left lateral strike-slip, and NE-SW reverse faulting. The ΔCS distributions resolved onto right lateral strike-slip and the conjugate left lateral strike-slip faults are very similar, except for the details of the location and amplitude of stress change in close proximity to the earthquakes of the Canterbury sequence (Figure B-2). These three fault orientations are also consistent with nearby structures imaged in reflection seismic surveys (e.g. Barnes et al., 2016; Figure 3-3). With constraints on the kinematics and locations of faults in the Canterbury Plains from earthquakes and fault mapping, we can use ΔCS resolved onto these known target orientations to identify structures and areas that may have been loaded by the Canterbury sequence. This approach differs from studies that resolve the ΔCS on optimally oriented planes, which may not reflect existing structures (e.g. Steacy et al., 2014). We focus on the eastern end of the Canterbury sequence near Christchurch, because the ΔCS near the mainshock rupture and in the broader Canterbury Plains region is dominated by the mainshock, and has been discussed in Herman et al. (2014). Finally, we explore the sensitivity of the ΔCS computation to variations in coefficient of friction (0.2-0.8), target fault depth (5-15 km), and mainshock slip model (Hayes, 2010; Beavan et al., 2012) (Figure B-2).
3. Results

Relocation of the earthquake hypocenters provides us with an internally consistent catalog and does not significantly alter the general pattern of seismicity from the initial GNS Science locations (Figure 3-2a). The most noticeable change in the relocated catalog is that the earthquakes near Christchurch exhibit slightly greater levels of clustering, highlighting the primary structures hosting seismicity throughout the sequence. One area that might be expected to have some improved spatial coherence indicative of internal structure is the stepover feature near 172.4ºE, which has been an area of intense and sustained activity throughout the Canterbury sequence (Herman et al., 2014). However, our relocated hypocenters do not appear to align in any systematic manner in the stepover region; this result suggests that whatever the internal structure, the fault segments are short and spaced closer together than our phase arrival picks can resolve (<2 km). This interpretation is consistent with the relocations performed by Syracuse et al. (2013), in which they found hypocenters in the stepover clustered into three N-S oriented, vertically dipping planes ~2 km apart.

A primary motivation for this relocation exercise was to obtain a location for the 14 February 2016 Mw 5.7 event and its aftershocks consistent with and relative to previous events. The U.S. Geological Survey places its epicenter on the east (oceanside) of the NE-SW trend of reverse earthquakes offshore of Christchurch primarily associated with the 23 December 2011 event, while the GNS Science location for the event is immediately west (landside) of the reverse faulting structures. Our relocation is consistent with the GNS Science epicenter, placing the event closer to the intersection of the strike-slip faulting beneath southern Christchurch and the reverse faulting east of Christchurch (Figure 3-2a). The relocated epicenter of the February 2016 event is distinctly offset ~3 km west of the trend of reverse faulting aftershocks.

The 150 regional moment tensor (RMT) solutions for earthquakes in the Canterbury Plains from September 2010 to November 2013 (Herman et al., 2014; grey focal mechanisms in Figure 3-2b) were dominated by strike-slip faulting (W-E right lateral or NW-SE left lateral), and also included many oblique-reverse to reverse mechanisms, particularly at the eastern end of the sequence. Most of these events were consistent with a sub-horizontal P-axis oriented at ~115º (Sibson et al., 2011; Holt et al., 2013). Here, we have determined RMT solutions where possible, for any earthquake reported by GNS Science as M 4.0 or larger (http://www.geonet.org.nz) from January 2014 to April 2016. We found 11 events with waveforms suitable for RMT inversion, including the 14 February 2016 Mw 5.7 event and 10 more earthquakes with RMT magnitudes ranging from Mw 3.7 to 4.1 (Figure 3-2b; Table 3-1).
Six of these 11 earthquakes occurred prior to February 2016. These six events include three strike-slip earthquakes near the stepover zone described in Herman et al. (2014), with magnitudes Mw 3.9 (January 2014; Event 1 in Figure 3-2b and Table 3-1), 4.0 (December 2014; Event 4), and 3.8 (January 2016; Event 6). An Mw 3.7 strike-slip event (Event 2) occurred in March 2014 under southern Christchurch, and an isolated Mw 3.9 reverse faulting earthquake (Event 3) occurred in September 2014 near the coast, north of Christchurch (northeast of the region shown in Figure 3-2b). An Mw 3.7 strike-slip earthquake (Event 5) occurred in May 2015 in the foothills of the Southern Alps, northwest of the main Canterbury sequence and associated with a cluster of seismicity that was active during the earlier sequence (north of the region shown in Figure 3-2b). These earthquakes have focal mechanisms that resemble the mechanisms of the earlier events in the Canterbury sequence: they are dominantly strike-slip, and also include oblique and reverse mechanisms.

The 14 February Mw 5.7 earthquake (Event 7) occurred offshore east of Christchurch, near the region where the E-W trending right lateral strike-slip structure under southern Christchurch (associated with the February 2011 event), the NW-SE left lateral strike-slip structure extending onto Banks Peninsula to the south (associated with the June 2011 event), and the NE-SW trending reverse structure offshore east of Christchurch (associated with the December 2011 events) all intersect. The mechanism for this event is dominantly strike-slip, with an orientation compatible with either of the conjugate strike-slip trends. Interpreting its fault plane is difficult because there is no clear surface evidence for the fault orientation, aftershocks do not clearly align with either nodal plane, and the event does not appear to fall on an extension of any of the previously ruptured structures. In particular, because the Mw 5.7 epicenter does not lie on a previously active trend of activity and has a strike-slip focal mechanism distinctly different from the nearby reverse faulting earthquakes, we interpret that it occurred on a separate structure from the ones that have already hosted seismicity in the Canterbury sequence. Four of its aftershocks were large enough for RMT inversion (Figure 3-2b). The 14 February Mw 3.9 earthquake (Event 8) five hours following the mainshock was a strike-slip event to the southeast of the mainshock, and the 18 February Mw 4.0 earthquake (Event 9) was a reverse faulting event north of the mainshock. Ten days later, on 28 February, an Mw 3.9 strike-slip faulting earthquake (Event 10) occurred beneath western Christchurch. The most recent event for which we determined an RMT occurred on 12 March, and was an Mw 3.9 reverse faulting earthquake (Event 11) offshore east of Christchurch. This earthquake had an unusual focal mechanism
rotated nearly 90º from the orientation of other reverse faulting earthquakes in this offshore segment.

The Coulomb stress change (ΔCS) resolved onto either nodal plane of the 14 February 2016 earthquake indicates that the event occurred in a region where previous events had reduced the stress levels. In general, the ΔCS is negative near locations where earthquakes have already occurred (Figure 3-3), however, it should be noted that the stress changes in the vicinity of previous ruptures vary from positive to negative over relatively short distances. Therefore, for either nodal plane, small changes to the event location, dimensions or complexities of the source fault slip, or changes in frictional or elastic parameters may place the event into a ΔCS state of the opposite sign (Figure B-2). This lack of resolvability of the ΔCS sign is common for events that occur very close to the source of stress change, especially for complex or multiple sources (e.g. foreshocks and aftershocks of the 2014 Mw 8.2 Iquique earthquake; Hayes et al., 2014; Herman et al., 2016). Likewise, during the Canterbury sequence, much of the seismicity occurred near the rupture zone of one of the main events (the 2010 Mw 7.0 Darfield and 2011 Mw ~6.0 Christchurch earthquakes), and as a consequence, half of the Canterbury earthquakes occur in areas where the cumulative ΔCS from the preceding seismicity was negative (Bebbington et al., 2016). However, we find that all of the main Christchurch aftershocks (besides the February 2016 Mw 5.7 event) had positive cumulative ΔCS resolved onto their interpreted fault planes. These observations suggest that, although the capability to resolve ΔCS near the Canterbury earthquakes is not currently sufficient to understand the near-rupture aftershock activity, ΔCS appears to be an appropriate tool for anticipating earthquakes away from the rupture zones.

This provides motivation to examine stress changes in the regions farther from the already ruptured areas. The particular distribution of positive ΔCS depends on the target fault orientation, i.e., the style of faulting in the subsurface (Figure 3-3). Unlike the areas near the Canterbury earthquakes, these ΔCS regions tend to be robust with respect to changes in source or modeling parameters (Figure B-2). One of the key areas of concern is under the city of Christchurch: whether potential fault structures here are strike-slip or reverse (although few are mapped; Barnes et al., 2016; Figure 3-3), they appear to have been positively loaded. Similarly, the region east of the offshore structures has been positively loaded for structures with any of the kinematics observed in the Christchurch sequence. The main difference lies to the north: if the target structures north of the offshore events are strike-slip, they have reduced ΔCS (Figure 3-3a), whereas reverse structures in this region have increased ΔCS (Figure 3-3b). Finally, adding the recent events does very little to change the general patterns of ΔCS, except in very close
proximity to the event epicenter (within ~3 km). Although ∆CS is not a predictive tool in the sense that it can tell us where and when the next significant earthquake will occur, it does suggest that Christchurch and the nearby vicinity should be considered as locations for possible continuing aftershock activity.

4. Discussion

A question about this sequence, that has been asked since the mainshock ruptured on 3 September 2010, is when the sequence will finish (e.g. Quigley et al., 2016). Fitting the seismicity in the 150 days after the 2010 Darfield mainshock with an Omori decay law (e.g. Utsu et al., 1995) yields a decay exponent of ~0.9, and extrapolating this Omori fit, the seismicity rate returns to its pre-Darfield rate (~0.2 events per day) 20-30 years after the Darfield earthquake. However, the large 2011 Christchurch earthquake sequences, as well as the occurrence of several moderate to large earthquakes near Christchurch in 2012 and 2013, and now this Mw 5.7 event, complicate this simple decay scenario, and suggest that the time frame for the completion of this intraplate earthquake sequence may be significantly longer (e.g. Reyners et al., 2013). This is consistent with aftershock duration-loading rate relationships: at the relatively slow deformation rate in the Canterbury Plains (16 nstrain/yr of maximum compression; Wallace et al., 2007), a longer aftershock period is expected (as compared to aftershocks in plate boundary settings; Stein and Liu, 2009). One possible mechanism for such sustained activity is that there are numerous, relatively small structures in the Canterbury Plains that have been brought closer to failure by the preceding seismicity, and some of these loaded faults eventually fail in earthquakes.

The locations and mechanisms of the Canterbury earthquakes, including the recent February 2016 event, are consistent with this proposed mechanism: the earthquakes highlight a complex network of faults, suggesting that we should not think of the subsurface as containing only a few dominant structures, but rather as densely populated with faults with multiple orientations, all of which can potentially slip in an earthquake. This interpretation of the subsurface structures from seismicity is also consistent with seismic reflection surveys in the Canterbury Plains that have imaged dense sets of faults both onshore (e.g. Dorn et al., 2010) and offshore (e.g. Barnes et al., 2011; Barnes et al., 2016; Figure 3-3). The seismically imaged faults have strikes ranging from E-W to NE-SW, and there is likely another set of faults to accommodate left lateral strike-slip on NW-SE faults, as in the June 2011 event. Although some of these imaged segments appear to be relatively short (~5 km), empirical relations between
magnitude and fault size suggest a Mw 5.7 earthquake rupture area in South Island, New Zealand has dimensions of 5 km by 5 km (Yen and Ma, 2011).

As the Coulomb stress change analysis demonstrates, the stress changes in the Canterbury Plains are significant, and vary over relatively short spatial scales (Figure 3-3). Although the details of the Coulomb stress change cannot be precisely resolved because of the uncertainties inherent in the earthquake source characterization, event locations, and elastic properties, it seems likely that complex stress perturbations in a region containing many faults with diverse orientations would bring some of these faults closer to failure. In particular, although a structure might lie in a region where right lateral strike-slip faulting is inhibited by stress changes from the Canterbury sequence, if it is a reverse fault, its kinematics might be favored by the stress changes. So although the sequence seems to be returning to relative quiescence, similar small to moderate earthquakes on nearby unmapped faults should be considered as a continuing hazard even when all of the primary structures, e.g., the Darfield fault, have apparently ruptured.

Acknowledgements

This work was supported by NASA Earth and Space Science Fellowship 15-EARTH15R-0096. Seismic data were obtained from the New Zealand GeoNet project and its sponsors EQC, GNS Science and LINZ. Many of the figures in this manuscript were created using the Generic Mapping Tools (Wessel and Smith, 1991). T. Parsons and an anonymous reviewer provided helpful feedback that improved the manuscript. P. Barnes and NIWA provided offshore fault data. The authors and NIWA accept no liability for any loss or damage (whether direct or indirect) incurred by any person through the use of or reliance on the data.

References


Figures

(A) Earthquakes ML 4.0 and larger, colored by rupture date and scaled by magnitude. The February 2016 event is indicated by a star. The mapped area is outlined in the inset. (B) Local magnitude versus time, starting with the Mw 7.0 (ML 7.2) Darfield earthquake on 3 September 2010 through April 2016. Note the gap in significant earthquakes from 2013 to 2015. (C) Seismicity rate over time. The black curve shows the seismicity rate (number of events ML 2.0 and larger per day), and the red curve shows the cumulative moment release of all recorded events following the Mw 7.0 Darfield mainshock.

Figure 3-1. Seismicity in the Canterbury earthquake sequence from September 2010 through April 2016.
Figure 3-2. Relocation results and moment tensor solutions.

(A) Relative relocation results. Grey dots indicate the GNS Science locations of the 161 earthquakes for which we have computed regional moment tensor (RMT) solutions. Black dots are the relocated hypocenters of the 150 RMT solutions from Herman et al. (2014). Red dots are the relocated hypocenters of the 11 RMT solutions determined in this study. (B) RMT solutions from Herman et al. (2014) in grey, and solutions from this study in red. Solutions from this study have been slightly enlarged to emphasize the focal mechanisms. The label numbers correspond to the labels in the text and in Table 3-1. Events 3 and 5 occurred off the northern end of the mapped region.
Figure 3-3. Coulomb stress change in the Canterbury Plains.

The stress field is generated by all 161 earthquakes in the RMT catalog from this study, resolved at 10 km depth. The 2010 Darfield mainshock slip is the Beavan et al. (2012) solution, which is based on geodetic observations. Using a different slip model, e.g. Hayes (2010) changes the details of the $\Delta$CS distribution near the fault, but not the general pattern (Figure B-2). White symbols are earthquakes with the same kinematics as the target faults. Note that many of the white symbols occur in blue regions because they have reduced the nearby $\Delta$CS as a result of their rupture. Transparent dark symbols are the earthquakes with different mechanisms in the sequence. Solid lines are mapped faults interpreted to have the same kinematics as the target faults, and dotted lines are interpreted as normal faults, but may accommodate oblique or strike-slip deformation (fault data courtesy of NIWA; Barnes et al., 2016). The star indicates the location of the 14 February 2016 Mw 5.7 earthquake, and an outline of Christchurch city is shown for reference. (A) $\Delta$CS resolved onto right lateral strike-slip faults with a strike of 80°, like one of the nodal planes of the February 2016 event. This pattern is nearly the same as the $\Delta$CS resolved onto left lateral strike-slip faults with a strike of 150° (Figure B-2). (B) $\Delta$CS resolved onto reverse faulting earthquakes with a strike of 40°.
Table 3-1. Regional moment tensor solutions for the nine events from December 2013 through March 2016.

The event labels correspond to those described in the text and in Figure 3-2b.

<table>
<thead>
<tr>
<th>Label</th>
<th>Origin Time</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Depth (km)</th>
<th>Strike</th>
<th>Dip</th>
<th>Rake</th>
<th>Magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2014-01-10 02:33:54</td>
<td>172.386</td>
<td>-43.565</td>
<td>13</td>
<td>180</td>
<td>70</td>
<td>-10</td>
<td>3.9</td>
</tr>
<tr>
<td>3</td>
<td>2014-09-12 00:18:11</td>
<td>172.785</td>
<td>-43.152</td>
<td>11</td>
<td>20</td>
<td>50</td>
<td>75</td>
<td>3.9</td>
</tr>
<tr>
<td>4</td>
<td>2014-12-12 13:37:04</td>
<td>172.366</td>
<td>-43.639</td>
<td>7</td>
<td>0</td>
<td>85</td>
<td>-35</td>
<td>4.0</td>
</tr>
<tr>
<td>6</td>
<td>2016-01-27 23:24:42</td>
<td>172.517</td>
<td>-43.539</td>
<td>8</td>
<td>175</td>
<td>60</td>
<td>25</td>
<td>3.8</td>
</tr>
<tr>
<td>7</td>
<td>2016-02-14 00:13:43</td>
<td>172.762</td>
<td>-43.503</td>
<td>9</td>
<td>155</td>
<td>65</td>
<td>5</td>
<td>5.7</td>
</tr>
<tr>
<td>8</td>
<td>2016-02-14 05:27:49</td>
<td>172.818</td>
<td>-43.550</td>
<td>11</td>
<td>180</td>
<td>75</td>
<td>20</td>
<td>3.9</td>
</tr>
<tr>
<td>9</td>
<td>2016-02-18 06:17:42</td>
<td>172.757</td>
<td>-43.479</td>
<td>5</td>
<td>210</td>
<td>40</td>
<td>85</td>
<td>3.9</td>
</tr>
<tr>
<td>10</td>
<td>2016-02-28 14:32:40</td>
<td>172.635</td>
<td>-43.564</td>
<td>6</td>
<td>150</td>
<td>65</td>
<td>0</td>
<td>3.9</td>
</tr>
<tr>
<td>11</td>
<td>2016-03-12 10:01:31</td>
<td>172.799</td>
<td>-43.490</td>
<td>7</td>
<td>150</td>
<td>45</td>
<td>75</td>
<td>3.9</td>
</tr>
</tbody>
</table>
Chapter 4

Modeling Slip Deficit Accumulation Around Locked Regions on the Subduction Megathrust

1. Introduction

Numerous great (Mw 8.0+) earthquakes have ruptured subduction zone plate interfaces (megathrusts) over the past decade. Anticipating the kinematic characteristics of these events, particularly their likely locations and slip distributions, is an important unresolved issue in subduction zone seismotectonics and seismic hazard assessment. A key component of this effort is constraining the processes by which elastic strain accumulates during the inter-seismic stage of the earthquake cycle. To first order, elastic deformation is concentrated around frictionally locked areas on the plate boundary (asperities), where the plates cannot slide past each other (Byerlee, 1970; Dieterich, 1972; Scholz et al., 1972; Lay and Kanamori, 1981). Regions outside these asperities appear to have frictional characteristics that allow relatively smooth sliding, and are commonly interpreted to accumulate low amounts of elastic strain and near-zero slip deficit, thus presenting low seismic hazard (Ruff and Kanamori, 1980; Peterson and Seno, 1984; Jarrard, 1986; Scholz and Campos, 1995). Improved geodetic and geological observations led to studies that indicated some areas of subduction megathrusts are only partially coupled, i.e., slip occurs, but at a rate slower than plate motion (e.g., Shen-Tu and Holt, 1996; Prawirodirdjo et al., 1997; Ozawa et al., 1999; Sieh et al., 1999; McCaffrey et al., 2000). Despite continuing improvements in geodetic observational capabilities and better resolution of coupling on the subduction megathrust, the physical process responsible for producing partial coupling remains poorly understood. Therefore, it is still unclear what role these areas play in inter-seismic loading and how they behave during great earthquake ruptures. In this study, we demonstrate that apparent partial coupling occurs around fully locked zones irrespective of the frictional state of the surrounding interface, and that geophysical observations are consistent with an interface consisting entirely of fully locked and completely frictionless sections. These low friction areas can still slip during the co-seismic rupture, potentially contributing to seismic moment release or other forms of deformation in the earthquake.
The current paradigm for plate boundary coupling processes is based on (a) the methodology for determining interface slip from geodetic observations and (b) experimental research on rock friction. A primary approach to infer slip deficit on the megathrust is based on inversions of GPS observations for a distribution of “back-slip” (Savage, 1983) on the subduction interface. High back-slip areas on the megathrust determined in such inversions correspond to regions of large slip deficit (defined as the amount of slip expected to occur on the interface due to relative plate motion minus the amount of slip that actually occurs) and low back-slip areas are interpreted to be slipping at near the full plate velocity. If back-slip is equal to the full plate velocity (or, conversely, is zero), it is straightforward to interpret that area as completely frictionally locked (conversely, completely unlocked). The physical mechanism of coupling is typically interpreted through the lens of rate-state friction (Dieterich, 1978; Ruina, 1983; Scholz, 1998), where the inter-seismically locked areas are velocity-weakening, leading to unstable stick-slip behavior, and creeping areas are velocity-strengthening, resulting in stable sliding. In this framework, zones of intermediate back-slip are often interpreted to reflect a rheological property of the interface at that location, specifically a parameterization of velocity-strengthening friction, that allows slip but not at the full rate (Hetland and Simons, 2010; Kaneko et al., 2010; Scholz and Campos, 2012). This velocity-strengthening behavior is thought to inhibit earthquake nucleation and rupture propagation (Tse and Rice, 1986; Marone and Scholz, 1988; Kaneko et al., 2008).

However, partial coupling will occur near locked zones irrespective of the specific frictional properties of these velocity-strengthening regions. This is a simple consequence of elasticity; rocks immediately adjacent to a high-stress, deformed zone must also be deformed unless they are physically decoupled from that adjacent region or undergo inelastic deformation. Since the materials on either side of the seismogenic portion of the megathrust behave elastically to first order (Turcotte and Schubert, 2002), deformation should be spatially continuous. Therefore, we expect the fault surface around an asperity to have a slip deficit immediately adjacent to the locked area that approaches the slip deficit within the asperity, independent of its frictional properties. Over some distance from the asperity, the slip increases back to full plate motion. Because these near-asperity regions of the plate boundary acquire slip deficit, even with zero friction, they may potentially contribute to slip during co-seismic strain release.

In this study, we utilize a finite element modeling (FEM) approach to assess the inter-seismic patterns of slip deficit and elastic deformation around a locked region on the subduction megathrust. Our results are compared with models of slip on the megathrust derived from a
variety of geophysical datasets (geodetic, seismic, and tsunami) that provide information about the deformation processes observed at different locations on the upper plate during both the inter-seismic and co-seismic stages of the earthquake cycle. The fully locked regions restrict the surrounding areas from full plate motion, generating a zone of partial slip around them that can accumulate several times larger slip deficit than within the locked area alone. In addition, observations made during several recent great megathrust earthquakes are consistent with the spatial patterns of loading predicted in our models. Specifically, earthquakes appear to be capable of rupturing any region that accumulates slip deficit, whether that slip deficit occurred through frictional locking or by its location adjacent to a locked patch. This has implications for interpreting co-seismic stress drops; if an earthquake ruptures an area of the megathrust with relatively low friction, then the shear stress value on the interface before and after the earthquake might be very similar. Likewise, the amount of heat produced during rapid sliding in a low-friction region might be much less than expected for typical frictional sliding.

If the characteristics of seismic waves radiated from the slipping fault depend on the mechanism responsible for the generation of the local slip deficit, it may be possible to identify during the co-seismic stage the regions that are inter-seismically locked or free to slip by mapping the style of seismic radiation to specific sections of the interface (Lay et al., 2012). Tsunami observations also fit into the framework of deformation accumulating around an interface that is locked at depth; although there may be essentially freely sliding conditions on the subduction interface up-dip of the seismogenic zone due to both velocity-weakening friction and low material strength, this region can accumulate a very large fraction of the full slip deficit and consequently have large surface (seafloor) displacements. However, our models demonstrate that this shallow slip deficit on the interface is not associated with significant elastic strain of the upper plate adjacent to the trench, potentially leading to large amplitude block motions of the upper plate near the trench that generate tsunami without producing significant seismic radiation.

2. Methods

To explore the distribution of slip deficit on the megathrust around a locked asperity, we have created a simple 3-D subduction FEM (Figure 4-1). We use the code GTECTON (version 2017.1, compiled with PETSc 3.4.2) to solve the static mechanical equilibrium equation for displacements at each nodal point of the finite element mesh (Govers and Wortel, 1993). The subducting plate is a planar slab 100 km thick with a 25º dip. The upper plate is a triangular
wedge that extends 400 km horizontally from the trench to the back-arc (the x-direction), and the model extends 1000 km along strike (y-direction). Deformation is driven by 1 meter of relative displacement between the two plates; the front and back of the subducting plate are displaced in the down-dip direction, while the upper plate backstop is held fixed. Since the system is elastic, the modeled displacements, strains, and stresses scale linearly with the amount of relative displacement (or velocity, as is typically reported in inter-seismic coupling models), assuming the model geometry is held constant. We isolate the effects of interface locking characteristics by assigning the same mechanical properties and homogeneous elastic rheologies (Young’s modulus=1x10^{11} Pa; Poisson’s ratio=0.25) to the two plates. The specific choice of Young’s modulus does not affect the modeled displacements (only the associated stress levels), because our model is driven by displacements and not forces.

The inter-seismic frictional properties of the megathrust surface are represented by one of two boundary conditions: fully locked (welded across the boundary, i.e. no discontinuous displacements allowed) and fully unlocked (free to slide parallel to the interface with zero resistance). For straightforward comparison with the solutions of Okada (1992) that describe deformation for a rectangular fault slipping in an elastic half-space, we define locked sections as rectangular areas on the megathrust with one side oriented parallel to strike (the fault length) and the other along dip (the fault width) (Figure 4-1). The rest of the surface is free to slide, using the “slippery node” formulation, which allows nodes on the interface to have discontinuous displacements in the plane of the fault (Melosh and Williams, 1989). These freely slipping areas of the megathrust are defined to have zero shear stress resolved on the interface, while interface-normal stresses remain continuous and out-of-plane stress components can be discontinuous (e.g., Burgmann et al., 2005).

Over the decades- to centuries-long time scales of interseismic loading, a shear stress-free boundary condition is a reasonable approximation. The shallow regions up-dip of the seismogenic zone on the megathrust, where “seismogenic zone” is defined as the depth range over which earthquakes nucleate, are thought to be frictionally weak (Scholz, 1998; Hardebeck, 2015). The regions down-dip of the seismogenic zone are envisioned to be narrow shear zones creeping continuously at low shear stresses due to higher temperatures and viscous rheology (Tichelaar and Ruff, 1993; Hyndman et al., 1997; van Keken et al., 2002). We also define regions outside of locked areas, within the seismogenic depth range, as shear stress-free, although it is unknown whether these areas slide stably because of mineralogy (Saffer and Marone, 2003), high fluid pressures (Audet et al., 2009), interface topography and geometry (Wang and Bilek, 2014), or
other processes. By assigning zero shear stress boundary conditions to all of these unlocked areas, our models generate the minimum amount of inter-seismic slip deficit and strain that can accumulate. Any additional resistance to sliding will result in larger slip deficits and elastic strains than the shear stress-free conditions in our models.

This model setup allows us to define an arbitrary number of locked regions with varying dimensions on the interface and compute the resulting deformation throughout the model. Multiple asperities with varying dimensions are often inferred in inter-seismic coupling models (e.g., Metois et al., 2016), and these regions can potentially interact with each other during loading and the subsequent earthquake. We test the effects of varying the size and location of a single locked patch and also explore how nearby locked regions interact via models that contain more than one locked patch.

The results from these models include the displacements at each nodal point and the discontinuous displacements at each point on the interface. Each element has a constant elastic strain and stress field, which is computed from the difference in displacements at its nodal points through the linear element shapefunctions. In addition to examining displacements throughout the model volume, we specifically evaluate the distribution of inter-seismic fault slip in our models by sampling the discontinuous displacements at each node on the megathrust. The crustal displacements associated with this slip distribution can be observed through geodetic observations at the surface, so we analogously sample displacements on the surface of the upper plate in our models. We compare these displacements to the distribution of strain and stress throughout the model; in particular, we evaluate the relationship between fault slip and elastic strain and how that differs between areas that are frictionally locked and areas that are free to slide.

3. Results

3.1. Single Asperity

For the initial model, we consider a 40 km by 40 km locked zone (the dimensions of a Mw ~7 earthquake) centered at a depth of 30 km. By definition, within the boundaries of this locked region the two plates have zero relative motion, i.e., it accumulates full or 100% slip deficit (Figure 4-2A). Surrounding the locked region, although the interface is free to slide, the plates are physically restrained from slipping at the full plate motion due to their proximity to the
locked region. In other words, partial slip deficit accumulates on the megathrust adjacent to the locked patch irrespective of the particular frictional characteristics at that location, even if they would allow continuous sliding at the full plate rate in the absence of locking. This effect decreases with distance from the edge of the locked zone and at a sufficiently large distance the plates move past each other at the full rate. Immediately adjacent to the 40 km by 40 km locked region, slip on the interface increases rapidly from zero at the edge to 50% of the full plate motion over a distance of ~15 km (Figure 4-2B-C). By ~60 km from the edge, slip is at 80% of the full plate motion, and by ~100 km from the edge, slip is back to 90%, making it essentially indistinguishable from the plate rate. This zone of partial slip deficit extends in all directions on the plate boundary from the locked zone, appearing as an annulus of slip deficit. Not only are the magnitudes of the relative motion vectors reduced on the megathrust around the locked region, but the direction of slip is also deflected around its boundaries (Figure 4-2A). The deviations from pure dip-slip motion defined in the model boundary conditions are minor (<20º), but are not simply an artifact of the rectangular shape of the locked zone; similar deflections persist around elliptical and irregularly shaped locked regions, although the precise orientations of the deflected fault slip vectors vary depending on the shape.

The partial slip deficit zone is symmetric along a transect running parallel to strike through the center of the locked zone, consistent with the symmetry of the model (Figure 4-2B). Down-dip of the locked patch, the functional shape of the slip as it recovers to full plate motion is similar to the shape in the along-strike direction (Figure 4-2B-C). The area of the megathrust between the up-dip edge of the locked area and the trench accumulates significant slip deficit regardless of its frictional state because the zone of partial slip deficit intersects the subduction trench. The tapering of the upper plate near the trench, combined with the effects of the free surface at shallow depths, leads to a more gradual return to full fault slip and the shallow slip immediately up-dip of the locked section never exceeding ~50% of the full plate motion.

The deformation driven by plate motions across the locked region generates displacements on the surface of the upper plate in the model, which are comparable to geodetic observations (Figure 4-3). The subducting plate drives the upper plate away from the trench in the positive x-direction and downward. The surface displacements are sensitive to the slip deficit over the entire plate boundary interface (Wang and Dixon, 2004), not just within the locked region. A back-slip model that only represents the slip deficit accumulating inside an asperity (implying full plate motion everywhere outside the asperity) does not produce the same surface deformation (Figure 4-4A, inset). The surface displacement distribution is sensitive to the full
distribution of slip deficit on the interface, whether that is accumulated through frictional locking, proximity to a locked zone, or other effects not explored here.

In Figure 4-4, we plot the stress field and superimpose modeled displacements (for 1 meter of relative motion between upper and subducting plates) projected on a vertical cross-section oriented perpendicular to strike through the center of the locked zone. Because our model is homogeneous and isotropic elastic, the pattern of strain distribution is identical to that of the stress distribution. We plot the “effective shear stress,” i.e., the second invariant of the deviatoric stress tensor (Figure 4-4A; derivation in Appendix C),

\[ \sigma_{eff} = \sqrt{\frac{1}{3} \left( \sigma_{xx}^2 + \sigma_{yy}^2 + \sigma_{zz}^2 - \sigma_{xx} \sigma_{yy} - \sigma_{xx} \sigma_{zz} - \sigma_{yy} \sigma_{zz} \right) + \sigma_{xy}^2 + \sigma_{xz}^2 + \sigma_{yz}^2} \]

the four non-zero independent components of the stress tensor (Figure 4-4B-E), and shear and normal stresses resolved onto the orientation of the megathrust (Figure 4-4F-G). The largest elastic strains and stresses are concentrated in the volume immediately adjacent to the locked area. Although we primarily focus on upper plate deformation, we note that there is also significant deformation in the subducting plate; the upper ~50 km of the subducting plate beneath the locked zone accumulates effective shear stress values greater than 0.1 MPa. In the upper plate above the locked region, effective stresses >0.2 MPa extend from the megathrust to the surface (Figure 4-4A). These large effective shear stresses extend all the way to the trench. An examination of the components of the stress tensor reveal that the dominant contribution to near-trench stresses is from \( \sigma_{yy} \) (Figure 4-4D), whereas the other stress components are an order of magnitude smaller within ~20 km horizontal distance of the trench. This large \( \sigma_{yy} \) is generated as displacements are deflected around the locked region (Figure 4-2A); on the up-dip side, the deflected displacements result in compressional deformation. The region within 10-20 km horizontal distance of the trench has large x and z displacements (~0.5 m), even though the associated \( \sigma_{xx}, \sigma_{zz}, \) and \( \sigma_{xz} \) stresses here are fairly small (≤0.02 MPa), implying that the near trench region is displaced without significant internal deformation. In the arc-ward direction from the locked zone, horizontal compressive stresses (\( \sigma_{xx} \)) are >0.02 MPa throughout much of the thickness of the upper plate (Figure 4-4B). The only other stress component that is large in the arc-ward direction from the locked region is \( \sigma_{xz} \), which has large positive values extending down-dip along the interface (Figure 4-4C). The normal and shear stresses resolved on the interface highlight the nature of the friction boundary conditions we assigned to the megathrust. Normal stresses are continuous across the megathrust whether the frictional properties are locked or free.
to slip (Figure 4-4F). The shear stresses are highest at the locked zone and zero on the interface outside this region (Figure 4-4G).

3.1.1. Locked Patch Dimensions

The spatial distribution of coupling seen in inversions of inter-seismic surface velocities and the diversity of earthquake sizes indicate that frictional asperities in subduction zones can take on a range of dimensions, from 100s of m (Mw 3.0) to 100s of km (Mw 8.0+). The underlying physical character of these variously sized asperities appears to be similar (Kanamori and Anderson, 1975; Ide and Beroza, 2001), so we treat locked zones of different sizes identically in our models and therefore each generates a region of partial slip deficit on the adjacent sections of the interface. The extent of the megathrust that accumulates partial slip deficit from asperities of different sizes correlates with the dimensions of the locked patch. This change in tapering distance of the partially coupled zone is not simply an artifact of the distance to the model edges and the boundary conditions enforced there; our model dimensions are sufficiently large that the differently sized asperities that we test do not interact strongly with the edges, unless otherwise noted. We examine the differences between square locked patches centered at 30 km depth (71 km from the trench along the megathrust) on the interface with side lengths 10, 40 (previously described in Section 3.1), and 80 km (Figure 4-5). We define 80% of the relative plate motion as the cut-off value between areas that are accumulating “significant” partial slip deficit and those that are effectively moving at near the plate rate. Although larger locked zones have greater areas around them that are restrained from having full plate motion, the effect does not scale linearly with the size of the locked area. In fact, the smaller locked zones have a proportionally larger effect, as measured by the ratio of the 80% partial coupling distance to the locked patch side length. The 10 km square locked zone generates a small partial slip deficit zone surrounding it; by ~25 km from its edge, the motion on the megathrust is back to 80% of the full value (partial slip width/locked dimension=2.5). In contrast, the zone of partial slip deficit around a 40 km locked zone extends ~60 km from the edge before slip on the interface recovers to 80% of the plate motion (partial slip width/locked dimension=1.5), and it takes ~80 km from the edge of an 80 km locked region to reach 80% of full plate motion (partial slip width/locked dimension=1.0). Therefore, at the scale that coupling models (and earthquake slip models) can resolve the slip distribution on the interface (~25 km; e.g., Wallace et al., 2004; Loveless and Meade, 2010; Hayes et al., 2013; Metois et al., 2016), a small (~10 km) asperity that could rupture in a Mw ~6.5
earthquake would likely be indistinguishable from its partial slip deficit area, instead appearing as a broader region sliding at an intermediate value less than the full slip rate. It may be possible to distinguish asperities from their partial slip deficit rings once they have dimensions larger than the current ~25 km resolution limit or with the addition of offshore geodetic networks. However, interference between the partial slip deficit zones of nearby asperities may complicate this resolvability (c.f., Section 3.2).

In many subduction zones, coupling and consequent earthquake slip extend much greater distances along-strike (>200 km) than down-dip (50-100 km). We compare a model with a locked patch having a 40 km down-dip width and a 500 km along-strike length to the reference 40 km by 40 km model described in Section 3.1 (Figure 4-6A). Near the along-strike edges of the 500 km long locked zone, the spatial distribution of recovery to full plate motion is similar to the 40 km square case. The primary difference in the 500 km long locked zone model is in the extent of its partial slip deficit region up- and down-dip from the locking; down-dip, slip recovers to 80% of the full plate motion at a distance of ~160 km down-dip from the edge of the locked region, compared to ~60 km for the 40 km square locked zone. We note that the rheology of the upper plate in these deeper regions may no longer be elastic and thus the expected style of deformation behavior likely deviates from our elastic-only assumption (for an exploration of the effects of rheological heterogeneity, see Chapter 7). Up-dip of the long, locked region, at the trench, the slip never exceeds 30% of the full value (compared to 50% in the square locked zone model). The surface displacements through the center of the 40 km square and 40 km by 500 km locked area models are qualitatively similar, although surface displacement magnitudes are ~30% larger in the model with a long locked zone (Figure 4-6B-C).

Likewise, the stresses in the short- and long-locked region models are qualitatively similar in cross-section, with a few critical significant differences (Figure 4-7). The effective shear stress in cross-section through the locked zone is significantly smaller when the locked region extends a large distance along strike. This is dominantly due to the reduction in the $\sigma_{yy}$ stress component (<0.04 MPa) in the upper plate (Figure 4-7D). Although $\sigma_{yy}$ has a small negative value throughout the entire upper plate in the model with a 500 km long locked region, this is a modeling artifact; at the along-strike edges of the locked zone, slip deflected around the locked region interacts with the zero y-displacement boundary conditions at the edges of the model, producing a uniform compression in the along-strike direction. Much of the near-fault region has effective shear stress values greater than 0.2 MPa. Up-dip of the fault, the region of the upper plate within ~30 km horizontal distance of the trench has conspicuously low effective shear
stress magnitudes (<0.1 MPa), indicating little elastic strain accumulation. The displacements in this region are a large percentage (~70%) of the relative plate motion, indicating that this region moves dominantly through block motion, i.e., without significant internal deformation.

3.2. Multiple Locked Zones

To evaluate how locked zones and their partial slip deficit regions interact, we place two square 40 km locked areas on the megathrust at the same depth (centered at 30 km) and separate them along strike (Figure 4-8). When the two locked zones are less than ~20 km apart, the modeled deformation is very similar to the deformation produced by a single, 100 km long locked zone (Figure 4-8B); slip in the intervening region is low, never exceeding ~20% of the plate motion and the down-dip extent of the slip deficit ring is expanded. As the locked zones become more separated, slip on the freely sliding interface between them increases (Figure 4-8B-C). Similarly, increasing the distance between locked regions causes the surface deformation footprint to separate along-strike (Figure 4-9). When the two locked regions are close, their surface displacements overlap significantly (Figure 4-9A), but as they separate, displacement footprints also become distinguishable (Figure 4-9B-C). Although it is possible to resolve the two locked zones once they are separated by more than ~25 km, they and their corresponding slip deficit rings overlap and interact up to much larger distances; the two partial slip deficit zones around these 40 km square locked zones are not fully distinct from each other until the separation distance between the locked zones is >200 km. It is uncommon for subduction zones to have creeping (i.e., unlocked) sections that extend more than 200 km along-strike (e.g., Sumatra: Chlieh et al., 2008; Japan: Loveless and Meade, 2016; Chile: Metois et al., 2016), so we anticipate that most subduction zone asperities have partial slip deficit regions that overlap and interact with neighboring asperities.

One important consideration that these models with multiple locked regions identifies is how much slip can likely occur when a single asperity fails. Superimposing slip deficit from a single locked zone onto the double locked region model, it is apparent that if one of the two locked areas slips, that ruptured region cannot recover the full accumulated slip deficit because slip is still restrained by the partial slip deficit region around the adjacent locked zone (Figure 4-8; Figure 4-10). As the locked regions become closer to each other, the amount of slip that can occur in the rupture of a single section decreases according to the slip deficit tapering pattern surrounding the unruptured locked zone. In the case of two 40 km square locked regions
separated by 20 km, the amount of slip available to occur in one square zone is \( \sim 70\% \) of the full plate motion (Figure 4-10). Because these locked regions are very close, the amplitude of slip available changes over the length of the rupture, i.e., on the side closest to the adjacent locked section, less slip is available to occur than on the side farthest away. In the situation with two 40 km square locked regions separated by 20 km, the side closest to the other locked zone can slip up to 60\% of the full slip deficit, while the far side can slip up to 80\% of the full slip deficit (Figure 4-10). This pattern of asymmetry in slip amplitude depends on there being no locked region on the other side of the rupturing asperity. More complex patterns can result from realistic distributions of variously sized and separated asperities.

4. Discussion

Our models demonstrate that significant inter-seismic deformation complexity can arise from a relatively simple distribution of locked and unlocked regions on the interface. This suggests that some of the apparent complexity in observed inter-seismic coupling on the interface may be misinterpreted as a consequence of local frictional properties. In addition, constraining the processes that generate inter-seismic deformation may help understand recent observations of the earthquake rupture process in subduction systems.

4.1. Interpreting Coupling Models

An important general result from our models is that adjacent regions on the subduction megathrust should interact, with locked zones restraining the amount of slip on nearby low-coupling areas. In contrast, standard coupling inversions typically utilize Green’s functions derived from individually slipping faults embedded in a continuous medium (i.e., surrounded by zero back-slip), which have no inherent interactions between nearby fault sections (Okada, 1992; McCaffrey et al., 2000; Meade, 2007). The benefit of this formulation is that these Green’s functions are linear and independent (unlike the multiple partial slip zones described in Section 3.2), making them well suited for kinematic inversions. In theory, the slip distribution associated with interacting faults can be recovered through such inversions (Wang and Dixon, 2004). However, issues of low spatial resolution and modeling assumptions such as smoothing, interface geometry, or subduction zone rheology, can obscure such effects except in special cases, where resolution of interface slip is unusually high (e.g., beneath the Nicoya Peninsula; Protti et al., 2013). We suggest that applying shear stress constraints in inter-seismic megathrust slip
inversions, with high back-slip zones having finite shear stress resolved onto the interface and partial to low back-slip zones having a low amount of shear stress resolved on the interface, would reflect a more physically appropriate description of the inter-seismic coupling process (Burgmann et al., 2005).

The inter-seismic slip deficit accumulation rate at a point on the megathrust has been interpreted in kinematic inversion studies to be a consequence of either a spatial average of fully locked and unlocked sections (McCaffrey et al., 2000; Wallace et al., 2004) or the rate-state frictional characteristics at that location, with parameters tuned to allow the observed amount of slip (Loveless and Meade, 2011; Metois et al., 2016). Some recent theoretical models that incorporate rate-state friction also take into account the deformation throughout the surrounding volume in the response of a particular location on the interface (Hetland and Simons, 2010; Kaneko et al., 2010; Lambert and Barbot, 2016), but these models tend to be complex, making isolating particular inter-seismic processes difficult. Our relatively simpler models demonstrate that the areas of the megathrust adjacent to locked asperities accumulate slip deficit irrespective of their specific frictional characteristics. Since we define the unlocked sections of the plate boundary as having zero frictional resistance, our models represent the minimum amount of slip deficit that accumulates in these regions. However, a comparison of our models with an inverted coupling distribution from the South America subduction zone suggests that treating the interface as a distribution of locked and freely sliding sections can substantially account for the observations, without invoking particular, varying frictional parameters other than locked and unlocked.

Metois et al. (2016) performed an inter-seismic coupling inversion for the 2000 km length of the South America subduction zone offshore of Chile. They identify multiple regions accumulating full slip deficit, some of which are fairly continuous along strike and some which are more separated. Here, we focus on their coupling solution from 20°S to 29°S (Figure 4-11A), which we interpret to have seven near-rectangular highly coupled zones surrounded by apparent partial coupling. Their inversion applies the solutions of Okada (1992), so it is appropriate to compare their results with our elastic-only model. We place the seven fully coupled zones we identified from their inversion into our forward FEM, defining them as locked, and assign the rest of the interface to have zero frictional resistance. To better represent the geometry and kinematics of this subduction zone, the dip of the interface in this model is 20° and the convergence azimuth is 75° (instead of perpendicular to the strike of the subduction zone as in our previous models). Since the deformation approaches the along-strike ends of the model, there are significant
interactions with the boundary conditions at the surfaces there; thus, we do not compare deformation within 100 km of the ends of the model.

The resulting forward model produces a distribution of slip on the interface and synthetic horizontal GPS velocities that resemble the results published by Metois et al. (2016), even without explicitly fitting the geometry or kinematics of the system (Figure 4-11). The inverted and FEM coupling distributions on the megathrust are most similar within ~150 km of the trench. The primary difference lies down-dip of the locked regions, where our model predicts a gradual transition to full plate motion sliding over a horizontal distance of ~200 km from the eastern edge of the fully locked sections, whereas the Metois et al. (2016) inversion has an abrupt transition from fully locked to fully sliding over a horizontal distance of ~50 km. Despite these differences, our models both produce similar horizontal surface velocities (Figure 4-11C). This similarity suggests that the Metois et al. (2016) inter-seismic coupling distribution can be interpreted as a distribution of fully locked and freely sliding regions. In this framework, the partial coupling that they see in their inversion may occur dominantly as a consequence of proximity to a fully locked area.

An important set of inter-seismic deformation observations that our models do not reliably account for are sites of vertical uplift. Significant inter-seismic uplift, with magnitude comparable to or greater than the amount of subsidence, is observed in geodetic and geological observations in Chile, as well as in Cascadia (Burgette et al., 2009), Japan (Aoki and Scholz, 2003), and Sumatra (Chlieh et al., 2008). In contrast, our models cause the surface to displace broadly downward in the inter-seismic stage. Even by adjusting the model boundary conditions to promote vertical motions (e.g., by allowing the backstop of the upper plate to move vertically, instead of being completely fixed as described in Section 2), the maximum amount of surface uplift we can introduce is less than 10% of the maximum subsidence. One way to introduce surface uplift through inter-seismic back-slip in an elastic half-space is to have a sharp down-dip transition from fully locked to fully sliding on the interface. This generates a pattern of subsidence above the locked region and uplift farther from the trench (e.g., the Okada-based vertical surface uplift curve in Figure 4-4). Many of these inversions, including the one presented by Metois et al. (2016), fit vertical motions in models consisting of an elastic half-space by having an abrupt transition from fully locked to full plate motion at the base of the seismogenic zone, typically under the coast (Figure 4-11A). However, geological materials are likely unable to support the large shear stresses on the interface caused by such an abrupt transition in inter-seismic slip on the interface, especially over the long time periods of inter-seismic loading.
Therefore, we suggest that other factors, including variations in megathrust geometry or rheological heterogeneity (Chapter 7), may be responsible for the observed uplift patterns.

4.2. Earthquake Implications

Earthquake seismic moment is equal to the product of the shear modulus, fault slip, and fault area. Here, we focus primarily on the product of fault slip and area, because the shear modulus is uniform throughout our models. Because of the partial slip deficit region surrounding each asperity, the slip-area product over the whole plate boundary is several times larger than that within the asperity alone; therefore, if the entire megathrust unlocks during the co-seismic rupture stage, the resulting earthquake magnitude could be significantly larger than predicted by the asperity dimensions. As an example, consider the 40 km square locked zone discussed in Section 2; the moment magnitude of an earthquake that ruptures a 40 km by 40 km square asperity with 1 meter of slip (assuming a shear modulus of 35 GPa, as in our models) is Mw ~7.2. Integrating over the shaded area of partial slip deficit in Figure 4-2a, the total slip deficit-area product around this 40 km square locked region is ~3 times larger, corresponding to moment magnitude Mw ~7.5. We do not necessarily expect an earthquake rupture to release all of this slip deficit by propagating through the entire partial slip region surrounding the locked zone, because some of the regions we define in the model as freely sliding during the inter-seismic period may have a rheology consistent with velocity strengthening friction that inhibits rupture propagation (Tse and Rice, 1986; Marone and Scholz, 1988; Kaneko et al., 2008). However, if the co-seismic slip rates are high enough, it may be possible for the rupture to penetrate outside the asperity boundaries and into the areas of partial slip (Di Toro et al., 2004). This may explain how several recent megathrust ruptures along the South America subduction zone had significant slip in regions interpreted as having only partial apparent coupling, including the 2010 Mw 8.8 Maule earthquake (Moreno et al., 2012), the 2014 Iquique earthquake (Ruiz et al., 2014; Schurr et al., 2014), and the 2015 Mw 8.3 Illapel earthquake (Herman et al., 2017; Chapter 6).

Because the inter-seismic frictional characteristics differ between asperity and non-asperity areas, we expect there may be differences in how these regions slip and radiate seismic energy as well. This approach has been previously used to identify and interpret frictional segmentation of the megathrust (e.g., Lay et al., 2012). The 2015 Mw 8.3 Illapel earthquake is a good example of an event that appears to have spatial variations in slip that may map to different frictional characteristics of the subduction plate boundary (Melgar et al., 2016; Tilmann et al.,
The earthquake ruptured a large width of the interface, with slip occurring as deep as 40 km and as shallow as 5-10 km, though studies disagree as to the shallowest extent of slip (see Chapter 6 and references therein). Irrespective of how close the Illapel rupture came to the trench, its shallow slip extends farther up-dip than the shallowest megathrust seismicity in the subduction zone prior to the earthquake, suggesting slip propagated farther up-dip than the region capable of nucleating earthquakes. In addition, multiple studies noted that the deeper part of the rupture (25-45 km depth) produced seismic radiation enriched in short periods (0.5-2.0 s), whereas the shallower parts of the rupture produced longer period (2-50 s) signals (Melgar et al., 2016). These two features, (1) co-seismic slip imaged in areas that did not host seismicity and (2) longer period radiation than is typical, may be diagnostic of co-seismic slip that occurs in regions that are loaded because of proximity to a locked zone rather than being locked themselves (Lay et al., 2012). Thus, in our inter-seismic deformation framework, we interpret the region from 20 to 45 km depth as the asperity, fully locked prior to the 2015 Illapel rupture. The slip deficit region on the up-dip edge of this asperity resulted in the near-trench portion of the megathrust to accumulate slip deficit at a large (50% or greater) percentage of the plate rate. During the earthquake, the rupture was able to propagate into this shallow region, resulting in slip from near the trench to a depth of 45 km.

These models also have important implications for tsunami generation during great megathrust earthquakes, which depends on the displacement of the seafloor independent of whether that motion is accompanied by seismic radiation. The stress cross-sections (Figures 4-4 and 4-7) highlight that the seafloor near the trench experiences large inter-seismic displacements even when the immediately subjacent megathrust is entirely frictionless. If the shallow megathrust near the trench were frictionally locked during the inter-seismic period, the upper plate above it would accumulate elastic strain (effective shear stress values >0.2 MPa), corresponding to large displacement gradients. In this scenario, a rupture to the trench would cause co-seismic seafloor displacements with systematic spatial gradients, and we might expect the seismic waves radiated from the shallow megathrust zone to appear more typical. We may then also expect to see vigorous seismicity, especially aftershocks, located on the shallow megathrust.

In contrast, if the megathrust shallower than 15 km depth is free to slide (as in our models), the inter-seismic displacements of the upper plate that occur within a horizontal distance of ~30 km from the trench occur at very low levels of elastic strain (Figure 4-7) and little seismicity would be expected in this region at any stage of the earthquake cycle. Since there is no
elastic strain to release as fault slip propagates up-dip of the asperity towards the trench, the slip on the shallow megathrust results in the near-trench region of the upper plate moving as a block, i.e., without releasing strain and producing typical seismic radiation. This type of block offset was observed following the Tohoku earthquake, where the region near the trench was shifted ~50 m without significant visible deformation (Fujiwara et al., 2011). We interpret this to indicate that the megathrust up-dip of the Tohoku earthquake asperity was frictionally free to slide prior to the earthquake, which is also be consistent with the low co-seismic heat flow measured near the trench after the Tohoku earthquake (Fulton et al., 2013). Because seismological techniques may be blind to this type of slip near the trench, understanding the motion, slip deficit, and potential tsunami hazard on the shallowest parts of the subduction zone may only be possible through the improved measurements of the displacements of the near-trench ocean floor.

The distribution of slip on the interface accumulated over the inter-seismic stage also has implications for the distribution and amount of afterslip that can occur following an earthquake. If the rupture is confined to the asperity boundaries, then a large area on the interface will still have partial slip deficit after the earthquake. There is no longer a locked region physically restraining the recovery of this region to full slip and we expect this recovery to start immediately after the earthquake. The seismic moment equivalent of afterslip in this scenario is twice that of the initial earthquake. If the rupture propagates outside the asperity into the region of partial slip and the co-seismic slip tapers outside the asperity faster than the partial slip deficit halo, then the total amount of afterslip available is reduced and it will partially overlap with the edges of the earthquake rupture area. In this case, the remaining slip-area product is comparable to or smaller than the earthquake and its observable effects extend to the edge of the slip deficit ring or the slip deficit ring of the next asperity. These results are consistent with recent observations of small amounts of afterslip bounding the 2011 Mw 9.0 Tohoku earthquake rupture zone (e.g., Yamagiwa et al., 2015).

A final consideration in the case that the rupture propagates outside the asperity is the concept of stress drop. Within the asperity, the stress drop has a simple mechanical meaning: the difference in shear stress resolved onto the megathrust before and after the earthquake. However, frictionally unlocked areas have zero shear stress resolved on them, yet they may be able to contribute to the seismic moment if the rupture propagates into this region. The mechanical interpretation of stress drop for this area of the rupture is not as straightforward, and the actual shear stress resolved onto the fault outside the asperity immediately after the rupture may be a complicated function of the rupture details.
5. Conclusion

Finite element models of subduction zones containing a locked region on the megathrust and allowing the rest of the megathrust to slide freely account for the distribution of full and partial coupling in subduction zones to first order. A locked zone restricts nearby regions from sliding, producing a ring of partial slip deficit on the interface around it. Observations of coseismic ruptures indicate that any region that accumulates partial deficit can contribute to slip during the earthquake, irrespective of whether the slip deficit was accumulated through frictional locking or being adjacent to a locked area. This implies that an earthquake can be larger than the size of the asperity might indicate. The partial slip regions of nearby asperities interact and can change the appearance of coupling on the interface as well as the pattern of slip during an earthquake rupture. Finally, the region up-dip of an asperity accumulates significant slip deficit irrespective of its frictional character, implying significant tsunami hazard at most subduction zones globally.

References


The subducting slab is planar and dips at 25°. It is displaced 1 meter at both ends, while the upper plate backstop is held fixed. All points on the interface between upper and subducting plates are allowed to slide freely except for the nodes within the rectangular asperity. At these locations, the plates are restricted from sliding, i.e. are welded together.
Figure 4-2. Interface slip around a 40 km square locked patch.

(A) Upper plate slip on the interface relative to the subducting plate is shown as vectors. The vector magnitudes are contoured with black indicating zero slip and white indicating full plate motion. The rectangular asperity is outlined with a grey box. The value of the fault slip is plotted along the labeled transects. (B) Along-strike transect shows a symmetric recovery to full plate motion outside the asperity boundaries (indicated by the grey bar). (C) Along-dip transect shows a down-dip recovery to full slip similar to the along-strike recovery. Up-dip of the locked patch, the slip at the trench is only 50% of the full plate motion.
Figure 4-3. Surface displacements for a 40 km square asperity. Vectors indicate horizontal motion of the surface, and color contours indicate vertical motion.
Figure 4-4. Stresses on an along-dip cross-section through the middle of a 40 km square asperity.

(A) Effective stresses are highest near the fault where most of the deformation in the model occurs. The maximum effective stress in this model is 0.6 MPa. Plotted above the upper plate surface are the horizontal (blue) and vertical (red) displacements produced by the FEM (solid) and by a back-slip representation of slip only within the asperity (dashed). (B-E) Non-zero components of the stress tensor. (F) Normal stress resolved onto the interface geometry. (G) Shear stress resolved onto the interface geometry.
Figure 4-5. Comparison of fault slip around different sized asperities.

The 80% of plate motion level is highlighted by a bold dashed line. (A) Along-strike transects of slip, normalized to the full plate motion for 10 km (pink), 40 km (blue), and 80 km (green) square asperities. Inset panels show the slip distributions on the interface around these asperities (colored the same as in Figure 4-2). (B) Along-dip transects of the three asperity sizes. The larger the asperity and closer it is to the trench, the lower the amount of shallow slip that occurs. (C) Along-strike transects of recovery to full slip around different sized asperities with respect to the edge of the locked zone. The size of the asperity correlates with the area of the region around it that is restricted from full motion.
Figure 4-6. Comparison of short and long asperity slip distribution.

(A) Slip on the interface around a 40 km by 500 km asperity is plotted as vectors and colored by vector magnitude (as in Figure 4-2). White contours indicate slip around a 40 km square asperity centered at 500 km along-strike position. Blue contours are for the same 40 km square asperity, but plotted so its edge is coincident with the edge of the 40 km by 500 km asperity. The along-strike recovery to full plate motion is very similar for the two asperity sizes. In contrast, the long-asperity generates a much larger along-dip restriction of plate motion. (B) Surface displacements produced by the long asperity are plotted as black vectors, with vertical motions contoured in the background. Superimposed are the horizontal displacements from a 40 km square asperity in red. The dashed line shows the location of the surface displacements plotted in (C) and (D). (C) Horizontal displacements for a short (red) and long (blue) asperity. (D) Vertical displacements for a short (red) and long (blue) asperity.
Figure 4-7. Stresses on an along-dip cross-section through the middle of a 40 km by 500 km asperity, as in Figure 4-4.

(A) Effective stresses are highest near the ends of the fault where most of the deformation in the model occurs. The maximum effective stress in this model is 0.5 MPa. Plotted above the upper plate surface are the horizontal (blue) and vertical (red) displacements produced by the FEM (solid) and by a back-slip representation of slip only within the asperity (dashed). (B-E) Non-zero components of the stress tensor. (F) Normal stress resolved onto the interface geometry. (G) Shear stress resolved onto the interface geometry.
Figure 4-8. Slip deficit around adjacent 40 km square asperities.

(A) Interface slip around two 40 km square asperities separated by 80 km, colored by slip magnitude. (B) When the asperities are very close (20 km apart), there is little slip between them, so they appear like one continuous 100 km long asperity (orange dashed line). (C) As the asperities separate (here shown 40 km apart), more slip occurs between them, but they still interact. (D) Once the asperities are far enough apart (here shown 160 km apart), they begin to behave more like independent locked zones.
Figure 4-9. Surface displacements around adjacent 40 km square asperities.

(A) When asperities are close (20 km separation), surface displacements cannot distinguish them reliably. (B) At a separation distance of 80 km, there is a minimum in displacement magnitude between the asperities that may be resolvable depending on the geodetic network geometry. (C) At separation distances greater than 200 km, the surface displacements show a clear decrease in between the asperities.
Figure 4-10. Effect of rupturing an asperity adjacent to an unruptured locked zone.

If an asperity slips, it can only recover an amount of slip that is not physically restrained by the adjacent section (black bar). If the two asperities are close enough, the amount of slip available on the left side of the ruptured asperity is less (~60% of full plate motion) compared to the amount available on the right side (~80%). After this earthquake ruptures, an amount of residual slip deficit remains on the interface that depends on the earthquake slip and the dimensions of the adjacent asperity (red region).
Figure 4-11. Comparison of our FEM-generated inter-seismic deformation with results from Metois et al. (2016).

(A) Coupling distribution inverted from inter-seismic velocities (black vectors in (C)). (B) Slip distribution around 7 rectangular asperities corresponding to locked patches from inversion in (A). (C) Observed inter-seismic velocities (black) and FEM-produced velocities (red). These show a remarkable agreement in both direction and magnitude.
Chapter 5

Interactions Between Asperities Over Multiple Earthquake Cycles

1. Introduction

Observations of seismicity over multiple earthquake cycles indicate that many subduction zones are seismically segmented, that is, earthquake ruptures are typically confined to occur within similar geographical areas (e.g., in the Chile subduction zone: Kelleher, 1972; Lomnitz, 2004; Bilek, 2010). The boundaries of along-strike segments may be produced by features impinging on the trench from the subducting plate (Carena, 2011; Wang and Bilek, 2011), abrupt along-strike geometrical variations in slab or interface geometry (Hori et al., 2004; Qiu et al., 2016), geological features or topography on the upper plate (Brudzinski and Allen, 2007; Bejar-Pizarro et al., 2013), or variations in frictional characteristics on the interface (Hetland and Simons, 2010).

Sometimes, however, earthquake slip propagates through segment boundaries, resulting in larger, composite “multi-segment” ruptures, even in regions that have historically hosted smaller, single-segment earthquakes. This led to the concept of earthquake “supercycles,” first defined in paleo-seismic studies of coral uplift offshore of Sumatra (Sieh et al., 2008). In this framework, smaller earthquakes (up to Mw ~8) occur with greater frequency than great or giant events (Mw 8+), which have longer recurrence intervals. Earthquake supercycles have also been invoked to account for seismic, geodetic, and paleo-seismic observations in the subduction zones offshore of Japan (Satake, 2015), Cascadia (Goldfinger et al., 2013), Ecuador (Nocquet et al., 2017), and Chile (Melnick et al., 2017). Several explanations have been proposed as the cause of earthquake supercycles, including visco-elastic flow in the ductile deformation region (Lambert and Barbot, 2016), longer down-dip widths of the seismogenic zone (Herrendorfer et al., 2015), and particular frictional characteristics and distribution of friction on the interface (Biemiller and Lavier, 2017).

Another characteristic of supercycles is that the giant earthquakes that rupture the entire length of a subduction zone appear to release essentially all of the slip deficit accumulated since the previous giant event. In contrast, earthquakes that occur in a limited region bounded by other locked (i.e., zero-slip) regions have lower slip than expected based on the amount of slip deficit.
that has accumulated. This slip deficit still remains on the interface; therefore, if a later earthquake ruptures the entire length of the boundary, it can access this residual slip deficit that was not released in the smaller event (Ampuero and Ben-Zion, 2008; Kaneko et al., 2010; Herrendorfer et al., 2015). In the case that the asperity is isolated from surrounding locked regions by sufficiently wide freely sliding regions, then it may not be affected by its neighbors and the earthquake can slip to recover the full accumulated slip deficit (Hetland and Simons, 2010). A basic issue associated with supercycle behavior remains poorly understood: why does an earthquake with a relatively smaller rupture area (i.e., a single-segment rupture) release less than the full locally accumulated slip deficit within its footprint, rather than releasing the entire amount as in a multi-segment event?

Several recent great (Mw 8.0+) earthquakes that have been well constrained by a variety of geophysical analyses may help understand the physical processes responsible for the differences between single- and multi-segment ruptures. For example, the 2014 Mw 8.2 Iquique earthquake was a single-segment event, occurring in the middle of a region inferred to have hosted a multi-segment Mw 8.8 rupture in 1877 (Hayes et al., 2014a; Ruiz et al., 2014; Schurr et al., 2014; Figure 5-1A). In contrast, in the region of the 2011 Mw 9.0 Tohoku event, the record of historical earthquakes in the preceding century includes numerous Mw 7.0-8.0 events (Goldfinger et al., 2013; Satake et al., 2015; Figure 5-1B), but there is no record of a great (Mw 8.0+) event in the region since the 869 Jogan earthquake (Minoura et al., 2001). These earthquakes over the previous millennium indicated the presence of multiple smaller asperities on the megathrust, but a combined rupture of the entire interface was unanticipated.

Understanding the slip release behavior of ruptures, particularly Mw ~8 single-segment events compared to Mw ~9 multi-segment events, on the subduction megathrust is important for evaluating both the remaining seismic hazard in areas that have had recent earthquakes and the hazard in locations without a recent great earthquake, like the Cascadia subduction zone. There has not been a great earthquake in Cascadia since a Mw 9.0 event that ruptured the entire length of the subduction zone in 1700, but there is an extensive geological record of past great earthquakes (Atwater et al., 1995; Goldfinger et al., 2003; Goldfinger et al., 2012). This paleo-seismic evidence suggests that the rupture behavior varies; sometimes slip occurs over the entire length of the subduction zone whereas other times slip only occurs within a smaller region, i.e., there seems to be a supercycle pattern of seismicity (Goldfinger et al., 2013).
2. Model Setup

We simulate an earthquake supercycle using a finite element modeling (FEM) approach to evaluate the slip behavior on a subduction megathrust with small and large rupture segments at each stage of the cycle. The deformation during each time step and at each node in the FEM is determined by solving the static mechanical equilibrium equations using the code GTECTON (Govers and Wortel, 1993; version 2017.1, compiled with PETSc 3.4.2). The models consist of a planar subducting slab dipping at 25° and an upper plate that spans the region from the trench to 400 km arcward of the trench, with the subduction zone extending 1000 km along strike (Figure 5-2). The modeled lithosphere has homogeneous elastic properties (Young’s modulus=1x10^{11} Pa; Poisson’s ratio=0.25). Velocity boundary conditions (10 mm/yr) are applied to the front and back of the subducting plate in the down-dip direction while the back-stop of the upper plate is held fixed.

Discontinuous motion along the megathrust is modeled using the “slippery node” formulation (Melosh and Williams, 1989). The frictional properties of the interface in our model are either fully locked or free to slide with zero resistance; this locking is turned on and off at prescribed times in order to simulate the changing frictional conditions throughout the earthquake cycle. During inter-seismic loading, we lock the seismogenic zones of the interface (10-50 km depth) and allow the rest of the interface outside these locked regions to slide (Figure 5-2). Earthquakes are simulated by unlocking these previously locked segments, which results in reverse slip, facilitating elastic rebound. Regions outside these seismogenic segments are fully locked during the co-seismic stage unless specified otherwise. Although more sophisticated frictional characteristics are used in some models of subduction earthquake cycles (e.g., Hetland and Simons, 2010; Kaneko et al., 2016; Lambert and Barbot, 2016), our relatively simple locked/unlocked frictional state can account for many of the observations (e.g., Chapter 4), with the caveat that we ignore rupture nucleation and elasto-dynamic rupture propagation processes.

The inter-seismic stage of our model progresses in time steps of 5 years, corresponding to 5 cm of relative motion per step. After 20 time steps (100 years, or 1 meter of relative motion), we unlock the middle 150 km long segment (the lightest grey area in Figure 5-2) and allow it to slip while keeping the rest of the megathrust locked. After the earthquake, we relock the fault and unlock the regions of the megathrust up-dip and down-dip of the locked zones, which can then slip in response to co-seismic stress changes. We assume that both co-seismic and post-seismic slip occurs without any time progressing. Estimates of afterslip duration are typically ~2 years or less even for the largest earthquakes (e.g. the 2011 Tohoku earthquake; Diao et al., 2014; Sun et
al., 2014; Yamagiwa et al., 2015), which is sufficiently short relative to the long duration of inter-seismic loading in our models. Following the earthquake and afterslip in the central segment, our model takes 20 more time steps (100 more years for another 1 meter of relative motion since the previous single-segment rupture, or 2 meters of total relative motion). Then, we unlock both the central segment and the longer segments along strike (darker grey regions in Figure 5-2). Again, we allow the fault to slip, then relock the fault and unlock the megathrust up-dip and down-dip to allow afterslip.

We explore models in which the seismogenic zone is continuous along strike; the 150 km long light grey area in Figure 5-2 is locked inter-seismically and unlocks every 100 years (1 meter of relative motion) and both the intervening and dark grey regions are locked inter-seismically and unlock every 200 years (2 meters of relative motion). This geometry, consisting of connected or nearly connected seismogenic zones, likely reflects the state of most subduction megathrusts globally, since there is sparse evidence for long regions (at seismogenic depths) with continuous creep behavior (Chlieh et al., 2008; Loveless and Meade, 2016; Metois et al., 2016). However, if there are areas with relatively long low-coupling segments between locked zones (e.g., the Hikurangi subduction zone: Wallace et al., 2004), we want to know whether those areas will have earthquakes that behave differently. Therefore, we also test the effects of separating the locked segments; in these models the intermediate 50 km long grey areas on either side of the central segment in Figure 5-2 behave like the down-dip region of the megathrust, remaining unlocked during the inter-seismic stage and locking co-seismically.

All of these models start from a zero deformation reference state. As a result, in the first earthquake cycle, the amount of slip on the fault is significantly less than the accumulated slip deficit. Over several cycles, the stress in the system increases and the model reaches a steady state, where co-seismic slip on the fault balances the inter-seismic slip deficit accumulation. The “spin-up” period is ~10 earthquake cycles, so the model requires 2000 years to achieve identical, repeating co-seismic slip events. In discussions of the models, we refer only to these steady state results: time=zero corresponds to the moment immediately following the previous steady state multi-segment earthquake and its corresponding afterslip, when inter-seismic loading has begun again. Time=100 years refers to the rupture of the single-segment event and its subsequent afterslip, and time=200 years is when the multi-segment earthquake occurs. Although these models are geometrically simple with homogeneous elastic material properties, they highlight how multi-asperity interactions may control the rupture characteristics of single segment ruptures and much larger multi-segment ruptures. In particular, they demonstrate how the smaller single-
segment rupture zone has its slip reduced by neighboring locked zones, and how the full slip deficit can only be released in a giant, multi-segment failure.

3. Modeling Results

Since we do not a priori know whether the region up-dip of the seismogenic zone (shallower than ~10 km) will slip during the earthquake (e.g., in the 2014 Mw 8.2 Iquique earthquake, this region did not slip, whereas in the 2015 Mw 8.3 Illapel earthquake, this region appears to have slipped), we test scenarios in which this region remains locked co-seismically, and scenarios in which this region is always free to slide. In the first set of FEMs (Sections 3.1-3.2), we lock the region up-dip of the seismogenic segments during the co-seismic stage, confining any co-seismic slip to a depth range of 10-50 km. We follow that analysis with models that keep the up-dip region unlocked at every stage of the earthquake cycle, including during the earthquake, and allowing slip to occur up to the trench (Section 3.3). Throughout this study, we show fault slip in two ways: (1) slip amplitude contoured and colored on the fault plane (Figure 5-3A, B) and (2) slip vector magnitude plotted versus position on transects taken through the seismogenic zone along strike (Figure 5-3C) and along dip (Figure 5-3D). We take two along-dip transects, one through the center of the model, i.e. through the middle of the smaller asperity (blue and orange lines in Figure 5-3D) and the other through the longer asperity (purple lines in Figure 5-3D).

3.1. Continuous Asperity Model

After 100 years, or 1 meter of relative motion, an earthquake occurs on the central, 150 km long zone while the rest of the interface is locked (Figure 5-3A). The co-seismic slip within the fault area has a maximum value of ~0.9 meters in its center, i.e., the earthquake releases most but not all of the slip deficit accumulated over the previous 100 years. One possibility for this lower value is that the system did not cycled through enough earthquakes to build up to steady state slip events; however, the single segment maximum slip is persistently 10-20% smaller than the multi-segment slip during every cycle, suggesting it is a physically realistic phenomenon. Along strike, slip decreases with distance from the center of the earthquake area; within ~25 km of the edges, the amount of slip has fallen below 0.5 meters, less than half of the accumulated slip deficit (Figure 5-3C). In contrast, the amount of slip on a down-dip transect through the center of the fault segment is nearly uniform across the width of the asperity (0.9 m of slip; Figure 5-3D).
Because we lock the regions up-dip and down-dip of the rupture zone and the other fault areas along strike do not unlock during this single segment rupture, these areas experience no co-seismic slip (Figure 5-3A; Figure 5-4A). After the earthquake, when we relock the middle fault segment and unlock the regions up-dip and down-dip (keeping the along-strike segments locked), the co-seismic stress changes drive afterslip on the freely sliding areas (Figure 5-4B). This afterslip is largest immediately adjacent to the up- and down-dip edges of the rupture zone, with a value of ~0.7 meters or 80% of the maximum co-seismic slip. Up-dip of the earthquake, the afterslip has nearly uniform value between the edge of the fault and the trench. In contrast, the after slip decays rapidly with distance from the down-dip edge of the central segment. Nowhere does the afterslip extend significantly beyond the along-strike edges of the rupture zone.

The entire length of the seismogenic zone is unlocked after 2 meters of relative motion since the previous multi-segment event (200 years in the model; Figure 5-4C). Along strike from the previously ruptured central segment, nearly 2 meters of slip occurs uniformly across the rupture area, accounting for most of the slip deficit. In the central segment, 1 meter of relative motion had accumulated since it ruptured in the single-segment earthquake. This 1 meter of slip deficit is in addition to the ~0.1 meters that remained following the previous smaller event. During the multi-segment earthquake, all ~1.1 meters of this slip occurred in the middle of the central region. There is a transition between this low slip and higher slip along strike; slip increases from 1.1 meters in the middle of the 150 km long central segment to 2.0 meters by its edge (Figure 5-3C; Figure 5-4C). Finally, after the rupture of the full length of the seismogenic zone, we lock the fault and unlock the regions up- and down-dip. The post-seismic slip after the multi-segment rupture is geographically extensive, spanning almost the entire down-dip width of the megathrust (Figure 5-4D). Similar to the single-segment event, large, uniform afterslip occurs up-dip of the regions with 2.0 meters of co-seismic slip. Down-dip of these regions, the magnitude of the afterslip decays with distance from the fault and by ~100 km from the fault, afterslip is less than 25% of its maximum value. Adjacent to the central segment, afterslip following the multi-segment event has a larger value than during the single segment earthquake.

We show how fault slip at each stage of the earthquake cycle contributes to the total plate motion slip budget along the fault dip in Figure 5-5 (the locations of these transects are through the central segment, corresponding to the orange line in Figure 5-3B, and farther along strike through the segment that only ruptures every 200 years, corresponding to the purple line in Figure 5-3B). We initialize the slip count immediately after the previous multi-segment earthquake at 0 years. During inter-seismic loading, there is no slip (by definition) within the seismogenic zone.
and the amount of slip on the megathrust increases with distance away from the locked region (e.g., Chapter 4). At 100 years (1 meter of relative motion), right before the single-segment earthquake, the slip conditions are identical at both transects; the locked fault has zero slip, the down-dip edge of the model has the full 1 meter of slip, and the up-dip edge of the model (i.e., the trench) has ~0.2 meters of slip. We unlock the central segment and lock the rest of the interface to simulate the single-segment earthquake. This results in ~0.9 meters of nearly uniform slip within the rupture area (Figure 5-5A). When the central fault is relocked and the regions up- and down-dip are unlocked, the amount of afterslip does not bring the total across the interface all the way to 1 meter. Farther along-strike, the longer segments do not rupture and do not experience significant afterslip, so this region of the model has essentially no deformation during or immediately after the single-segment earthquake (Figure 5-5B).

After another 100 years of loading, the total plate motion has been 2 meters from time zero and 1 meter since the single-segment earthquake. In the transect across the central segment, this slip deficit is accumulated on top of the co-seismic and post-seismic slip, whereas through the segment farther along strike, the slip deficit is simply added to the existing slip deficit. At time 200 years, we unlock the entire length of the subduction zone; this multi-segment rupture recovers the remaining slip budget within the central fault zone, including both the 1 meter accumulated from 100 to 200 years and the amount remaining after the single-segment earthquake (~0.1 meters), for a total of 1.1 meters (Figure 5-5A). Afterslip then fills in the slip deficit over the rest of the interface. In the transect through the section that only slips in multi-segment events, essentially all of the inter-seismic slip deficit is accommodated by co-seismic slip plus afterslip in the large earthquake (Figure 5-5B).

3.2. Isolated Asperities

In the second earthquake cycle FEM, we isolate the central segment from the side segments by 50 km, treating the areas in between (the intermediate grey regions in Figure 5-2) as unlocked during inter-seismic stage and locked during the co-seismic rupture. Changing these interface frictional properties results in significantly different rupture characteristics, particularly in the central region (Figure 5-6; Figure 5-7). After 1 meter of relative motion (100 years), in a single-segment rupture, the central fault region has nearly uniform 1.0 meter of slip over its entire area (Figure 5-6A; Figure 5-7A). There is no reduction in slip magnitude like in the continuous asperity model described in Section 3.1. Following the earthquake, the afterslip pattern up- and
Down-dip of the center region is similar to that of the continuous seismogenic zone model (Figure 5-6B). Because the regions adjacent to the rupture zone along strike are unlocked in the post-seismic stage, these areas also experience afterslip. The maximum afterslip value is ~0.8 meters around the edges of the asperity.

After 2 meters of total relative motion, the multi-segment rupture occurs by unlocking the central and side segments. Note that it is physically unlikely for segments separated by long creeping segments to be able to rupture in the same earthquake (Tse and Rice, 1986; Marone and Scholz, 1988). However, for direct comparison with the connected asperity scenario, we allow the central, smaller segment and longer side segments to slip simultaneously. In the central segment, there is 1.0 meter of uniform slip (Figure 5-6C, Figure 5-7A), whereas in the side regions that have accumulated 2 meters of slip deficit, the slip is ~2.0 meters, essentially the same as in the continuous seismogenic zone model (compare Figure 5-7B with Figure 5-5B). The afterslip following the multi-segment rupture occurs on the interface around the rupture zones, decaying with distance from the fault (Figure 5-6D; Figure 5-7). To account for the remaining slip deficit in the gaps between the seismogenic regions, afterslip is ~1.2 meters in the areas between the central and side segments.

3.3. Up-dip Unlocked

Some earthquakes, like the 2010 Mw 9.2 Sumatra, 2011 Mw 9.0 Tohoku, and 2015 Mw Illapel events, rupture the interface up to near the trench, rather than being confined to depths >10 km. To simulate how such an unlocked up-dip region affects the interface slip patterns in a multiple asperity system throughout the earthquake cycle, we run the same models described in Sections 3.1 and 3.2, except assigning the region up-dip of the locked zone to remain unlocked at every step of the earthquake cycle. This results in the slip deficit generated between the trench and shallow edge of the seismogenic zone being released co-seismically instead of post-seismically (Figure 5-8; Figure 5-9).

In the model with continuous seismogenic segments along strike, slip in the 100 year, single-segment rupture occurs all the way to the trench. The slip decreases from ~0.9 meters at the down-dip edge of the rupture area to ~0.6 meters at the trench (Figure 5-8A; Figure 5-9A), rather than remaining uniform along dip as in the locked up-dip model. The only afterslip that occurs is down-dip of the earthquake area and it has a peak value of ~0.7 meters (Figure 5-8B). After 2 meters of relative motion, the multi-segment earthquake releases the entire 1 meter of slip
deficit plus the amount remaining after the single-segment event (Figure 5-8C; Figure 5-9A). In the regions that did not slip for 200 years, the entire slip deficit is accommodated in this rupture, including the ~1.6 meters of slip at the trench (Figure 5-9B). Afterslip following this large rupture is intense down-dip of the rupture zone for ~100 km along the interface (Figure 5-8D).

When we keep the up-dip region unlocked and isolate the asperities by separating them by 50 km, they exhibit essentially independent behavior. The 100 year rupture of the central segment has nearly uniform slip and slip up to the trench is ~80% of the down-dip slip (Figure 5-8E). Afterslip occurs in the regions down-dip and along strike of the earthquake area and at this point, essentially all the slip deficit is accounted for through the single segment (Figure 5-9C). Finally, the multi-segment rupture results in 1 meter of slip in the central segment and 2 meters of slip in the regions along strike (Figure 5-8G). Afterslip fills in all the remaining slip deficit on the interface before the next cycle begins (Figure 5-8H).

4. Discussion

These simple FEM results demonstrate that a single segment rupture of a smaller asperity immediately adjacent to other large locked segments (a) decreases the maximum slip, and (b) causes slip to be reduced (relative to the accumulated slip deficit) at the edges next to adjacent asperities. The reduction in slip in single-segment earthquakes is essentially due to the “halo effect” discussed in Chapter 4, where a locked asperity prevents the regions around it from sliding the full amount. These results may help understand observations of the 2014 Mw 8.2 Iquique earthquake, which ruptured the central segment of a more broadly locked subduction zone (Figure 5-1A). Maximum slip in the Iquique earthquake was 5-7 meters (e.g., Hayes et al., 2014a; Ruiz et al., 2014; Schurr et al., 2014), significantly less than the 9-10 meters of slip deficit that had accumulated since the 1877 Mw 8.8 multi-segment rupture (assuming 70 mm/yr of convergence; DeMets et al., 2010). The single segment earthquake simulation with a locked up-dip region (Figure 5-4A) is the most relevant model to the Iquique scenario, and suggests that slip in the 2014 Iquique earthquake was restricted by the locked segments to its north and south. This model implies that a subsequent multi-segment rupture (Figure 5-4C) could slip up to ~10 meters outside the Iquique segment and potentially up to the remaining slip deficit (2-5 meters) even within the already ruptured area. If a 200 km long earthquake occurred to the south of the Iquique region, releasing the full 10 meters of slip deficit, the magnitude would be Mw ~8.5. If this earthquake also re-ruptured the 2014 Iquique earthquake area, releasing the remaining ~5 meters
of slip, this could add another Mw ~8.2, increasing the magnitude of the hypothetical multi-segment rupture to Mw 8.6.

This slip reduction effect is also consistent with the seismicity offshore of Honshu prior to the 2011 Mw 9.0 Tohoku earthquake, which consisted of dominantly Mw 8.0 and smaller events (Figure 5-1B). Because large areas of the interface surrounding these asperities remained locked, these events were unable to release all of the slip deficit accumulated within their boundaries. In the Tohoku earthquake, many of these regions slipped simultaneously, along with the interface closer to the trench, releasing a large amount of the slip deficit that had accumulated since the previous multi-segment rupture. The rupture up to the trench in the Tohoku earthquake suggests that our FEM with free slip allowed up-dip of the main asperity is the most representative model for this event. In this model, the near-trench region accommodates 80% of the total slip during the co-seismic stage, with only 20% occurring during inter-seismic loading (Figure 5-9B), despite the fact that it is never frictionally locked.

Finally, the seismic behavior in places like Chile and Japan along with these modeling results provide lessons for how to think about historical earthquakes in Cascadia and the potential multi-segment event that may rupture the entire length of the subduction zone. The history of great earthquakes and the nearly continuous coupling zone along the Cascadia subduction zone suggests it is closer to a continuous asperity system (like Iquique and Tohoku; Section 3.1) than an isolated asperity system (Section 3.2). As a consequence, we interpret that single asperity earthquakes in the Cascadia subduction zone may not release the bulk of the accumulated slip deficit across the length of the subduction zone. Rather, it is dominantly through larger, multi-segment ruptures that the system relieves the majority of the inter-seismic loading.

As a corollary point, there are few subduction zones globally where the isolated asperity model appears to be applicable. There are locations where coupling is inferred to be low from GPS observations over 100s of km, but these regions are often associated with great earthquakes in the past (e.g., in segments of the Aleutian subduction zone; Cross and Freymueller, 2008). In most subduction zones, asperities and locked regions are apparently closer than ~50 km. This means that subduction seismicity behavior will not usually behave as if the segments are independent of each other (Section 3.3), but will typically have segments that interact. As a result, larger, composite, multi-segment ruptures should be considered a possibility at most subduction settings globally.
References


90


Figure 5-1. Examples of subduction zones that ruptured in both single- and multiple-segment earthquakes.

(A) Offshore of northern Chile, the 2014 Mw 8.3 Illapel earthquake (area with greater than 1 meter of slip in grey) occurred in the middle of a larger area inferred to have hosted a Mw 8.8 earthquake in 1877 (dashed grey region). (B) In the Japan subduction zone, earthquakes over the past century (dashed grey regions) ruptured small areas of the subduction interface. When the Mw 9.0 Tohoku earthquake occurred in 2011, it ruptured regions that had already slipped in previous events.
Figure 5-2. Earthquake cycle finite element model setup.

The subducting plate is a planar slab 100 km thick and dipping at 25°. It has velocity boundary conditions applied at its top and bottom of 10 mm/yr. The back of the upper plate is held fixed. On the interface, we lock the area from 10 to 50 km depth (grey regions) in the inter-seismic stage, while allowing the rest of the interface to slide freely (white regions). Every 100 years, or 1 meter of total relative motion, we unlock the middle region (lightest grey region) and hold every other area locked. Every 200 years, or 2 meters of total relative motion, we also unlock the darker grey regions. We test models in which the intervening, 50 km long regions (intermediate grey) are locked during the inter-seismic period and participate in slip during the earthquake as well as models in which these regions are unlocked during the inter-seismic stage and locked co-seismically. We also test models in which the interface shallower than 10 km remains unlocked throughout the entire earthquake cycle.
Figure 5-3. Co-seismic slip on the interface during single- and mult-segment ruptures.

(A) When the single segment is unlocked after 100 years while keeping the along-strike regions locked (grey areas), slip occurs only within this rectangular area, with slip colored by magnitude. (B) When the entire subduction length is unlocked after 200 years, much larger slip occurs in the along-strike regions and significant slip still occurs in the central region. (C) Slip in along-strike transects through the middle of the locked region (transects shown in A and B). The blue line shows slip in the single-segment event and the orange line shows slip in the multi-segment event. (D) Slip in along-dip transects through the model at y=500 km (orange and blue curves) and y=800 km (purple curve), with colors corresponding to the dashed lines shown in A and B.
Figure 5-4. Co- and post-seismic slip for a continuous seismogenic zone with the up-dip region locked during the earthquake.

Slip is colored by magnitude and contour lines are plotted every 0.2 meters. (A) Slip during the earthquake only occurs within the central unlocked segment while the along-strike regions are still locked (grey bars). (B) After the earthquake, when the fault is relocked and the rest of the interface is unlocked, slip occurs up- and down-dip of the rupture area. (C) In the multi-segment rupture, large slip occurs in the segments that have accumulated the most slip deficit, while the central region has lower slip since it ruptures in the previous event. (D) After the multi-segment event, large amounts of afterslip occur down to depths of ~100 km.
Figure 5-5. Total slip at each stage of the earthquake cycle on a down-dip transect through the continuous seismogenic zone model with up-dip region co-seismically locked.

(A) Slip on a transect through the central segment. In the inter-seismic period, the seismogenic zone has zero slip, and slip on the fault grows with distance away from the fault. In the 100-year earthquake, most of the slip is recovered within the rupture limits. After the event, afterslip fills in much (but not all) the remaining slip on the interface. Inter-seismic loading proceeds for another 100 years, then a second earthquake and its afterslip account for the remaining slip deficit. (B) On a transect through the along-strike segments, there is no activity during the single segment rupture, so the system is dominated by the loading signal for 200 years. The earthquake and afterslip then account for the remaining slip deficit.
Figure 5-6. Co- and post-seismic slip for isolated seismogenic zones with the up-dip region locked during the earthquake.

Separating the seismogenic zones (shown as grey regions) by a 50 km long inter-seismically sliding and co-seismically locked region causes the co-seismic slip to be nearly uniform in the central (A, green area) and along-strike (C, red areas) segments. Post-seismically, afterslip occurs up- and down-dip, and also occurs in the regions between the edges of the rupture zones (B, D).
Figure 5-7. Total slip at each stage of the earthquake cycle on a down-dip transect through the isolated seismogenic zone model with up-dip region co-seismically locked.

(A) Slip on a transect through the central segment. In the inter-seismic period, the seismogenic zone has zero slip, and slip on the fault grows with distance away from the fault. In the 100-year earthquake, nearly all the slip is recovered within the rupture limits. After the event, afterslip fills in the rest of the remaining slip on the interface. Inter-seismic loading proceeds for another 100 years, then a second earthquake and its afterslip account for the remaining slip deficit. (B) On a transect through the along-strike segments, there is very little activity during the single segment rupture, so the system is dominated by the loading signal for 200 years. The earthquake and afterslip then account for the remaining slip deficit.
Figure 5-8. Co- and post-seismic slip for models with the up-dip region unlocked throughout the entire earthquake cycle.

The key difference is that the slip deficit up-dip of the seismogenic zone is now accommodated during the co-seismic rupture (A, C, E, and G) instead of post-seismically. (A-D) Co- and post-seismic slip for a continuous seismogenic zone. (E-H) Co- and post-seismic slip for isolated seismogenic zones.
Figure 5-9. Total slip at each stage of the earthquake cycle on down-dip transects through the models with up-dip region always unlocked.

The left column (Panels A and B) show slip when the seismogenic segments are continuous. The right column (Panels C and D) show the case when the seismogenic zones are separated by 50 km.
Chapter 6

Integrated Geophysical Characteristics of the 2015 Illapel, Chile, Earthquake

Abstract

On 16 September 2015, a Mw 8.3 earthquake ruptured the subduction zone offshore of Illapel, Chile, generating an aftershock sequence with 14 Mw 6.0-7.0 events. A double source W-phase moment tensor inversion consists of a Mw 7.2 sub-event and the main Mw 8.2 phase. We determine two slip models for the mainshock, one using teleseismic broadband waveforms and the other using static GPS and InSAR surface displacements, which indicate high slip north of the epicenter and west-northwest of the epicenter near the oceanic trench. These models and slip distributions published in other studies suggest spatial slip uncertainties of ~25 km and have peak slip values that vary by a factor of 2. We relocate aftershock hypocenters using a Bayesian multiple event relocation algorithm, revealing a cluster of aftershocks under the Chilean coast associated with deep (20-45 km depth) mainshock slip. Less vigorous aftershock activity also occurred near the trench and along strike of the main aftershock region. Most aftershocks are thrust faulting events, except for normal faulting events near the trench. Coulomb failure stress change amplitudes and signs are uncertain for aftershocks collocated with deeper mainshock slip; other aftershocks are more clearly associated with loading from the mainshock. These observations reveal a frictionally heterogeneous interface that ruptured in patches at seismogenic depths (associated with many aftershocks) and with homogeneous slip (and few aftershocks) up to the trench. This event likely triggered seismicity separate from the main slip region, including along-strike events on the megathrust and intraplate extensional events.

1. Introduction

1.1. Overview

On 16 September 2015 at 22:54:32 UTC (19:54 local time), a Mw 8.3 earthquake (the Illapel earthquake) ruptured a ~200 km long section of the South America subduction zone in central Chile. The U.S. Geological Survey National Earthquake Information Center (USGS NEIC) located the earthquake hypocenter at (71.67°W, 31.57°S) and a depth of 22 km, ~50 km
west of the city of Illapel and ~230 km north of Santiago (http://earthquake.usgs.gov/earthquakes/eventpage/us20003k7a). As of January 2017, over 500 Mw 4.5+ aftershocks had occurred in and around the source region of the Illapel mainshock. This aftershock zone extends ~250 km along the Chilean coastline and includes numerous Mw 6.0-7.0 earthquakes (Figure 6-1a).

Reports of shaking intensity, as collected through the online USGS *Did You Feel It?* questionnaire (http://earthquake.usgs.gov/earthquakes/dyfi), indicate Modified Mercalli Intensities (MMI) as large as IX near the epicenter, consistent with peak ground accelerations of ~0.8 g recorded at Centro Sismologico Nacional (CSN) station C110 (70.96°W, 30.70°S; Figure 6-1a). The Illapel earthquake also generated a tsunami with a maximum recorded run-up height of 12 m and run-up heights of over 4 m recorded along ~200 km of coastline, which resulted in widespread damage (Aranguiz et al., 2016; Omira et al., 2016; Melgar et al., 2016).

1.2. Historical Seismicity

The Illapel earthquake occurred on the subduction megathrust interface between the Nazca and South America plates. Here, the oceanic Nazca plate subducts beneath the South America plate at a rate of ~74 mm/yr and an azimuth of 80° (DeMets et al., 2010; Figure 6-1a). The ~3000 km length of this subduction zone offshore Chile (18°-45°S) has hosted ten great (Mw 8.0+) megathrust earthquakes since 1900, as well as many more large and great earthquakes in the historical record over the past several centuries (Kelleher, 1972; Beck et al., 1998; Lomnitz, 2004; Figure 6-1b).

The section of the subduction zone that ruptured during the Illapel earthquake (30°-33°S) has hosted multiple large earthquakes over the past three centuries, including in 1943, 1880, 1822, and 1730. The rupture zones of the 1943 and 1880 earthquakes are in a similar region and appear to be approximately collocated with the 2015 Illapel earthquake rupture zone, whereas the 1822 and 1730 events appear to have been larger earthquakes centered at ~33.5°S with rupture zones that extend as far north as ~30.5°S (Figure 6-1b). The 1943 earthquake had a shallow thrust mechanism consistent with rupture of the subduction plate boundary and estimates of its magnitude range from M 7.9 (Beck et al., 1998) to M 8.3 (Lomnitz, 2004). The 1880 earthquake had a smaller damage region and estimated magnitude than the 1943 earthquake (M 7.5-8.0), whereas the 1822 and 1730 earthquakes had substantially larger damage footprints with estimated magnitudes ranging from M 8.0 to 8.5 and M 8.5 to 9.0, respectively (Kelleher, 1972; Lomnitz, 2004).
The subduction zone south of the 2015 Illapel earthquake has also hosted large earthquakes that did not rupture north of 32°S into the section where the 2015 event occurred. Most recently, the Mw 8.8 Maule earthquake ruptured over 300 km of the subduction megathrust in 2010, with the rupture ending ~100 km south of the 2015 Illapel earthquake. The intermediate section of the subduction zone, south of the 2015 Illapel rupture and north of the 2010 Maule rupture, hosted the large 1971 (Mw 7.8) and 1985 (Mw 8.0) Valparaiso events, which appear to fill the gap between the Maule and Illapel earthquakes (Mendoza et al., 1994; Figure 6-1b).

While the section of the subduction zone to the south of the 2015 Illapel earthquake has a complicated history of overlapping earthquake ruptures, large events have not historically ruptured across 30°S. This has been interpreted to be partly a consequence of the topography of the subducting ocean floor; the Challenger Fracture Zone enters the trench near 30°S (Figure 6-1a) and separates oceanic lithosphere of different age and hence different bathymetry (e.g., Carena, 2011; Metois et al., 2012). There does not seem to be a similar persistent barrier to co-seismic rupture in the subduction zone south of the 2015 Illapel earthquake, although the Juan Fernandez Ridge does enter the subduction trench at ~33°S and has been inferred to act as a partial barrier to propagating slip and afterslip in the 2015 event (e.g. Barnhart et al., 2016; Figure 6-1a).

1.3. Coupling Models

In addition to the historical seismicity that shows that the Illapel segment is capable of hosting large earthquakes, geodetic measurements of surface deformation along the Chile coast over the past 15 years also indicate seismic hazard potential. Over the time period 1993-2009, onshore Global Positioning System (GPS) stations in Chile near the 2015 Illapel rupture zone moved eastward relative to a fixed South America at 20-30 mm/yr, consistent with elastic strain accumulation arising from a locked subduction interface (Vigny et al., 2009; Metois et al., 2012). Inversions of these velocities suggests that the plate boundary was well coupled in the vicinity of the 2015 Illapel earthquake (Metois et al., 2016). The details of the interpreted locking distributions vary due to low resolution of offshore coupling from onshore measurements, but all indicate that the subduction zone south of 31°S was nearly completely coupled prior to 2015, whereas the region north of 31°S had partial to low apparent coupling. The 2015 Illapel rupture and aftershock zone lies at the northern edge of the well coupled region and may straddle the transition in apparent plate boundary coupling behavior around 31°S. Early post-seismic afterslip following the Illapel earthquake was also interpreted to propagate both northward into the zone of
lower coupling as well as southward into regions of high inter-seismic coupling (Huang et al., 2017; Feng et al., 2017). The total accumulated slip deficit since 1943 is 5-6 m (assuming a completely coupled plate boundary), consistent with published estimates of average slip during the Illapel earthquake (see Section 3.1 for a discussion of published models).

1.4. This Study

We evaluate the seismotectonic significance of the Illapel earthquake by identifying robustly constrained features of the Illapel mainshock derived from different geophysical datasets, investigating the relationships between mainshock slip and aftershock activity, and evaluating static stress changes throughout the sequence. These analyses are used to infer the physical properties of the subduction megathrust and processes controlling the seismicity behavior.

2. Methods and Results

We apply a variety of geophysical approaches and synthesize observations from seismological and geodetic analyses of the mainshock rupture, seismological analyses of the aftershock source characteristics, high-quality multiple-event relocations of the aftershock sequence, and modeled stress changes generated by the earthquakes in the sequence.

2.1. Mainshock Moment Tensor

The basic source parameters of the Illapel mainshock are constrained by performing a deviatoric moment tensor inversion of teleseismic W-phase observations (Kanamori, 1993). The USGS single source W-phase moment tensor inversion (see Duputel et al. (2012) and references therein for waveform processing, Green’s function generation, and inversion details) of the Illapel earthquake incorporates data from 44 broadband stations up to 90° from the epicenter, with broad azimuthal coverage of the event (Figure 6-2). The resulting moment tensor solution has a shallowly dipping thrust faulting plane consistent with slip on the plate boundary interface. The seismic moment is $3.19 \times 10^{21}$ Nm, corresponding to Mw 8.3, and the best fitting double couple has a shallowly-east-dipping nodal plane that we interpret as the fault plane (strike=353°, dip=19°, rake=83°; Figure 6-2a). The solution is well represented by a shear dislocation source, i.e. unidirectional slip on a planar fault, with a double couple component of 98%. The event centroid is located at a depth of 26 km and has a time delay (relative to the event origin time) and half-
duration of 49 s. These source parameters are similar to other inversions; the Global Centroid Moment Tensor project solution (GCMT; http://www.globalcmt.org) has a magnitude of Mw 8.3 with a best fitting double couple (strike=7°, dip=19°, rake=109°) and an alternate W-phase moment tensor solution (http://wphase.unistra.fr/events/illapel_2015) has a magnitude of Mw 8.2 with a best fitting double couple (strike=3°, dip=22°, rake=94°).

In large or complex earthquakes, a single source solution may not capture significant variations in the source as the rupture evolves, so Nealy and Hayes (2015) developed a procedure using the W-phase to determine whether an event can be divided into two sub-events. The seismic finite fault analysis of the Illapel earthquake reveals an early peak in moment release (Section 2.3; Figure 6-3), so we apply the double source inversion to the Illapel waveforms to assess whether the mainshock rupture is best described as a double source. This analysis uses the same seismic data as the single source inversion and yields a moment tensor, centroid location, centroid time, and duration for each sub-event. The Illapel mainshock separates into a small (Mw 7.2-7.6) initial sub-event followed by a much larger (Mw 8.2) sub-event accounting for most of the moment release (Figure 6-2b). Because the first sub-event has a seismic moment 8-30x smaller than the second, we have to apply constraints to its timing and mechanism to produce a seismologically and geologically feasible solution, while we are able to leave the larger, second sub-event unconstrained. The first sub-event is constrained to have a time delay and half-duration of 20 s (informed by the small early peak in the source-time function of the mainshock; Figure 6-3) and a dip of 20° (consistent with the megathrust geometry from Slab1.0; Hayes et al., 2012). The source parameters of the first sub-event are sensitive to these constraints while the larger sub-event shows little sensitivity (Figure D-1). In our preferred double source inversion, the first sub-event is equivalent to a Mw 7.2 earthquake with a thrust faulting mechanism (strike=351°, dip=20°, rake=106°) and has a centroid depth of 41 km. The second sub-event begins ~6 s after the event origin time and is Mw 8.2 with an east-dipping thrust mechanism (strike=358°, dip=23°, rake=88°), and has a centroid depth of 24 km and half-duration of 46 s. The double source solution produces synthetic waveforms that fit the observations better than the single source (Figure D-2, D-3); the single source RMS misfit is 0.58, whereas the constrained double source RMS misfit is 0.41. Since the first sub-event is relatively small, there is effectively little difference between the single and double source solutions, as reflected in the similarity between the single source and second sub-event source parameters. However, an Akaike Information Criterion test, which evaluates the improvement of fit relative to the number of added degrees of freedom (Akaike, 1972; Akaike, 1974; Nealy and Hayes, 2015; Appendix D), indicates the Illapel
double source model is a statistically significant improvement over the single source model. Because both sub-events have mechanisms consistent with pure reverse slip on the megathrust, we interpret the Illapel mainshock as a thrust faulting rupture involving only the subduction interface, rather than a more complex rupture of multiple faults.

2.2. Seismic Finite Fault Model

We determine the timing and distribution of mainshock slip with a finite fault model (FFM) analysis of the rupture based on broadband teleseismic observations, following the approach of Ji et al. (2002) (e.g. Hayes et al., 2013; Hayes et al., 2014; Figure 6-3). The waveforms constraining this FFM come from 32 broadband seismometers up to 90º distant from the Illapel epicenter. Waveforms are processed by deconvolving raw time series to velocity in m/s and displacement in m, and rotating to radial, transverse, and vertical components. Two frequency bands are isolated from these waveforms: broadband body waves (P and SH phases, filtered in period range 1-180 s) and long period surface waves (Rayleigh and Love phases, filtered in period range 200-500 s). We use 31 P waveforms, 18 SH waveforms and 49 surface waves. The rupture is constrained to lie on a single fault plane chosen to represent the average megathrust interface geometry from Slab1.0 (strike=6º, dip=17º; Hayes et al., 2012) and the rupture begins at the relocated epicenter (71.673°W, 31.595°S; Section 2.5) at a depth corresponding to the Slab1.0 value at that location (29 km). The fault is divided into 330 rectangular sub-faults (22 along strike, 15 along dip); each sub-fault has an along-strike length of 15 km and a down-dip width of 13.8 km. We use a simulated-annealing algorithm to invert for the combination of rupture velocity, rake angle, slip amplitude, and rise time at each sub-fault that best fits the teleseismic records (Ji et al., 2002). The rupture velocity is allowed to vary from 0.8 to 3.5 km/s and slip amplitude from 0 to 12 m. Both rupture time and slip amplitude are regularized by applying smoothness constraints to the inversion. Sub-fault source time functions are modeled with asymmetric cosine functions. Green’s functions are similar to those used in the W-phase analysis and are computed from a velocity model combining PREM (Dziewonski and Anderson, 1981) with Crust 2.0 for the shallow portion (Bassin et al., 2000); in this study, we add a 1.5 km thick water layer to the top of this velocity model, based on local and regional velocity structure (Contreras-Reyes et al., 2015; Table D-1). We also apply a weighting factor to the total FFM moment so that it is constrained to match the single source W-phase moment (3.19 x 10²¹ Nm).
In this FFM solution, the rupture propagates north and west from the hypocenter, with significant slip both near the coast and the trench (Figure 6-3). The peak slip in our model is 8.6 m and lies west-northwest of the hypocenter at a shallow depth of 5-10 km. Large slip (up to 7.2 m) also occurs at deeper regions of the megathrust interface (25-35 km) north of the hypocenter beneath the coast. The rupture zone with slip greater than 1 m extends ~150 km along-strike and significant moment release lasts ~100 s, consistent with the duration inferred from the W-phase inversion. The FFM explains ~85% of the seismic waveform data, as measured by a least-squares misfit between observed and synthetic waveforms. Although we tested model parameterizations that produced synthetic waveforms matching the observed waveforms better (explaining up to ~88% of the data), the model presented in this study qualitatively balances fitting seismic waveform data with misfit of forward-predicted GPS data (Appendix D; Figure D-4). We did not formally weight the seismic and geodetic datasets, instead selecting models that were deemed to have acceptable fits to both sets of observations.

2.3. Geodetic Slip Model

We also invert nearby geodetic observations for the slip distribution of the Illapel earthquake. Although this approach does not constrain the timing of rupture, it provides an independent estimation of the spatial distribution of slip on the megathrust that is directly inverted from near-field static ground deformation. We jointly invert co-seismic interferometric synthetic aperture radar (InSAR) interferograms from the European Space Agency Sentinel-1a satellite C-band (~6 cm wavelength) radar and Universidad Nacional de Chile continuous GPS observations for the spatial distribution of slip. These geodetic observations span the along-strike length of the rupture and post-event SAR data were acquired ~12 hours following the earthquake, which limits potential signal contamination from early afterslip and post-seismic deformation. The selected datasets, processing steps, model setup, and inversion approach are identical to those used in Barnhart et al. (2016); however, the fault geometry in this study differs so that it matches the seismic FFM geometry (Section 2.2), allowing us to directly compare the solutions.

This geodetic slip model has similar general characteristics to the seismic solution: slip concentrated to the northwest of the hypocenter (peak slip of 6.1 m) and a total moment release of ~2.2 x 10^{21} Nm, corresponding to Mw 8.2 (Figure 6-4). The model produces synthetic co-seismic GPS displacements that fit the observations very well and satisfactory fits to the InSAR observations, although coherent residuals of up to 10 cm persist (Figure 6-4; Figure D-5). Slip is mapped up to the trench, however slip far offshore inferred from onshore static observations may
be smoothed across a broad region and details of the offshore slip distribution at length scales of ~20 km or less cannot be definitively imaged (Barnhart et al., 2016; Melgar et al., 2016). In contrast, the geodetic observations are able to resolve slip beneath or down-dip of the Chilean coast; we find up to 4.0 m of slip occurred immediately west of the coast and a smaller amount of slip (< 2 m) landward of the coast. Although the extent of slip on the interface offshore has low resolution along dip, the along-strike (i.e. latitudinal) location of the slip is better constrained. In addition, since the geodetic observations span the along-strike extent of the rupture (Figure 6-4), it is unlikely that significant along-strike co-seismic slip is missing from the surface deformation record.

2.4. Aftershock Moment Tensors

We perform moment tensor inversions on 234 Mw 4.0-7.0 aftershocks that occurred between 9 September and 4 December 2015 to characterize the kinematics of the Illapel aftershocks and investigate spatial or temporal patterns. Source parameters of events larger than Mw ~5.0 are determined using the single source teleseismic W-phase inversion approach described in Section 2.1. Earthquakes smaller than this magnitude typically generate waveforms with high signal to noise ratio at regional seismic stations and are less reliable for W-phase inversion. For these smaller events, we use the regional waveform inversion approach of Herrmann et al. (2011a), which solves for the source depth, moment magnitude, and strike, dip, and rake of a shear dislocation (i.e. pure double couple) source. For this analysis, broadband regional waveforms are processed similarly to the teleseismic waveforms for W-phase inversion and filtered from 0.02 to 0.06 Hz. Regional Green’s functions are generated from a Western United States velocity model used in NEIC operations (Herrmann et al. 2011a). Previous studies demonstrated that the inverted source parameters are not particularly sensitive to the choice of velocity model; magnitudes vary by up to 0.1 units, depths by up to 4 km, and focal mechanisms remain unchanged (Herrmann et al., 2011b). We have not estimated formal uncertainties for the source parameters, but previous studies indicate magnitude uncertainty is ±0.1 units, source depth is ±6 km, and kinematic parameters, i.e., strike, dip, and rake, are each ±10-15° (e.g., Herrmann et al., 2011b; Herman et al., 2014; Johnson et al., 2016).

The aftershock focal mechanisms are dominated by thrust faulting according to the criteria of Frohlich (1992), with 203 thrust, 19 normal, and 12 intermediate mechanisms (Figure 6-2a). No strike-slip events occurred in this aftershock sequence. Nearly all the aftershocks with thrust mechanisms have a nodal plane dipping shallowly to the east, consistent with rupture on
the subduction interface, although a few small thrust events have nodal planes that are not aligned with the interface. The largest aftershock is Mw 7.0 and the sequence also included 14 Mw 6.0-6.9 aftershocks; all of these largest aftershock events have thrust faulting mechanisms. In contrast, the largest aftershock with a normal faulting mechanism is Mw 5.2. Therefore, thrust faulting accounts for 99.9% of the aftershock moment release.

2.5. Hypocentral Relocation

In order to improve the spatial resolution of the earthquake locations in the Illapel sequence, we relocate 793 event hypocenters initially cataloged by the NEIC in the aftershock sequence from 16 September to 23 November 2015 using bayesloc, a Bayesian multiple-event relocation algorithm (Myers et al., 2007; Myers et al., 2009; Figure 6-5; Figure 6-6). The bayesloc algorithm is a Monte Carlo non-linear multiple-event locater that improves the precision of earthquake locations as well as reducing biases introduced by assumptions of Earth velocity structure. The algorithm incorporates phase arrival times and associated errors, travel times (based on a velocity model) and travel time corrections, prior constraints on the earthquake location and origin, and phase labels. Arrival times used by bayesloc are taken from seismic phases picked from seismograms at the same regional and teleseismic stations used in initial NEIC locations. We use the crustal phases Pg and Sg, the refracted phases Pn and Sn, and the teleseismic P phase in our relocations. Arrival time observations from P and Pn phases dominate our dataset, with ~18,000 observations of each. Our dataset contains only ~1200 Sn observations and ~500 observations of Pg and Sg each. We remove distant stations with few observations from our dataset, as these stations do little to constrain earthquake locations. Source-station travel times are initially derived from the global velocity model AK135 (Kennett et al., 1995) to remain consistent with single-event locations reported by the NEIC. An initial model of the earthquake location and origin time informs our inversion; we place broad constraints on earthquake epicenter and depth locations (10 km for 68% confidence), based on single event locations reported by the NEIC. Because the majority of seismicity occurs offshore, there are limited near-source arrival time observations and therefore hypocenter depths are poorly constrained. In order to best inform our Bayesian inversion, we place tighter limits on the initial depth priors if there are regional or W-phase moment tensor centroid depths. We are also able to constrain the prior distribution for a small subset of onshore earthquakes for which we can obtain calibrated hypocenter locations using the hypocentroidal decomposition algorithm (Jordan and Sverdrup, 1981).
In general, the relocation procedure tends to reduce the spread in the hypocenter locations, i.e. relocated earthquakes have increased spatial clustering (Figure 6-5a). The mean horizontal shift in earthquake locations (from original NEIC epicenter to relocated epicenter), is 7 km, with a maximum shift of 24 km (Figure 6-5b). There is a clear preference for the relocation to move events east-southeast and almost no events relocate northward (Figure 6-5d), suggesting that the relocation procedure is reducing a systematic bias to the north and west in the initial locations. The mean vertical shift (using the absolute values of the vertical shifts) is 5 km, with a maximum shift of 30 km (Figure 6-5c). Although the vertical relocation histogram shows a nearly equal number of events shifted up and down, it is clear that there are some spatial patterns to the vertical relocations; events near the mainshock hypocenter (72.0°W, 31.5°S) dominantly relocate deeper, whereas events north of the mainshock hypocenter near the coast and events near the trench tend to relocate shallower (Figure 6-5a). The Bayesian approach also yields event location uncertainties: the mean location uncertainty is 1.6 km north-south, 3.6 km east-west, and 2.7 km in depth (Figure 6-5e-g). The larger mean east-west uncertainty relative to the north-south uncertainty is a consequence of stations being located onshore, dominantly east of the offshore earthquakes, and the relatively small depth uncertainty is likely due to the tighter prior depth constraints on events with moment tensor solutions.

2.6. Aftershock Catalog

Our catalog of earthquakes contains a total of 844 events, 793 of which are relocated and 235 of which have moment tensor solutions. The largest magnitude and greatest number of aftershocks occurred very soon after the mainshock (Figure 6-7). The decay in aftershock rate follows a standard Omori’s law curve with a decay constant of 1 (Utsu et al., 1995; Figure 6-7b). The magnitude-frequency relationship for the catalog has a break in slope at Mw ~4.3, so we use Mw 4.5 as our magnitude of completeness (Figure 6-7c). We note that there could be a small number of M 4.5+ events that are not in the catalog because their signal is obscured by the waveforms generated by other aftershocks (e.g. Herman et al., 2014). The 844 aftershocks have a total seismic moment of 1.6 x 10^{20} Nm, equivalent to a Mw 7.4 earthquake and comparable to estimates of afterslip moment (Barnhart et al., 2016). Their mechanisms are dominantly thrust faulting with shallowly eastward dipping nodal planes (Figure 6-2a) and in cross-section it is apparent that they occur dominantly on the plate interface (Hayes et al., 2012) over a depth range of 10-50 km, highly concentrated near the rupture zone of the mainshock, especially near the mainshock hypocenter (Figure 6-6). We interpret this to mean most of the aftershocks ruptured
the subduction megathrust. Aftershock activity is conspicuously absent from the shallow regions on the megathrust; at a latitude of 31ºS, no thrust faulting aftershocks are shallower than ~20 km. To the north and south, this shallow megathrust aftershock gap narrows, but no thrust faulting events are located shallower than ~10 km. Few of the relocated hypocenters (and associated focal mechanisms) are indicative of upper plate events. The aftershocks near the trench are mostly shallow (< 15 km) and have normal faulting mechanisms. The events oceanward (west) of the trench clearly occurred within the subducting Nazca plate. However, some normal faulting aftershocks are located landward (east) of the subduction trench, making it difficult to interpret in which plate they occurred.

Aftershock activity began immediately after the mainshock rupture was complete; the first aftershock in our catalog (Mw 6.4) occurred 5 minutes after the origin time of the mainshock and ~6 km southwest of the hypocenter (Figure 6-2a; Figure 6-7). Within an hour, aftershocks spanned the entire geographical footprint of the mainshock rupture area; the largest aftershocks in this timeframe were within 25 km of the mainshock hypocenter, including (in chronological order) Mw 6.1, Mw 6.1, Mw 7.0, Mw 6.5, and Mw 6.7 earthquakes. Almost all of these large aftershocks were located to the west and north of the hypocenter. Only one Mw 6.0+ aftershock (Mw 6.4) in the first 24 hours occurred farther from the mainshock hypocenter and it was located ~50 km to the north. In contrast, 39 of the 45 Mw 5.0-5.9 aftershocks in the first 24 hours were located greater than 25 km from the hypocenter.

Over the week following the mainshock, aftershocks continued to dominantly occur coincident with the mainshock slip zone near the epicenter and to its north and west (including Mw 6.1 and 6.6 events), but the footprint of aftershocks began to expand beyond the immediate mainshock slip area. More distal regions that experienced activity were the portion of the megathrust ~90 km south-southwest of the epicenter (two Mw 6.2 events) and the area near the trench from 30º-31ºS (Mw 5.0 and smaller normal faulting events). A single thrust faulting Mw 6.1 event also occurred on the interface north of 30ºS.

Nine days after the Illapel earthquake, a Mw 6.3 aftershock occurred ~90 km north of the mainshock hypocenter down-dip of the Chile coast. From this event until November 2015 there were numerous events large enough for regional waveform inversion, but no regionally significant aftershocks. Then on 7 November, 51 days after the mainshock, a Mw 6.8 earthquake occurred ~80 km north of the epicenter in the same coastal region as the Mw 6.3 event. Four days later, two Mw 6.9 earthquakes occurred ~230 km north of the hypocenter. Although the rate of seismic activity has remained elevated in the region of the Illapel earthquake sequence even after
December 2015 (relative to rates prior to September 2015), these two Mw 6.9 events are the last significant earthquakes in the sequence as of April 2017, except for a single Mw 6.3 event beneath the coast on 10 February 2016.

2.7. Coulomb Failure Stress Changes

Using the approach described in Herman et al. (2016), we resolve mainshock- and aftershock-generated Coulomb failure stress changes (ΔCFS; a measure of how much closer a fault is brought towards or away from failure; e.g. Reasenberg and Simpson, 1992) on the megathrust plate boundary and on the interpreted fault plane of each aftershock (Figure 6-8). The mainshock source is defined by a slip model. To reduce noise, we set slip less than 15% of the peak value to zero (Ye et al., 2016). As discussed in Section 3.1, there are significant uncertainties in the mainshock slip distribution, so we test 4 different slip models (the seismic and geodetic models from this study; Li et al., 2016; Melgar et al., 2016; Figure D-6). Aftershocks are converted from point sources to rectangular faults with uniform slip centered at the relocated hypocenter using empirical scaling relations (Wells and Coppersmith, 1994) and their fault slip is determined from the seismic moment equation. The material is assumed to be an elastic half-space (Okada, 1992), with a shear modulus of 40 GPa and Poisson’s ratio of 0.25. The ΔCFS from the mainshock is resolved on the plane we defined for our slip models (strike=6º, dip=17º, rake=90º) to compute the loading of the megathrust interface. We also compute cumulative stress changes generated by the mainshock and all preceding aftershocks resolved on the east-dipping nodal planes of thrust faulting aftershocks. For normal faulting events, we test the ΔCFS resolved on both nodal planes because there are no clear constraints on the fault plane, although this does not affect the resulting ΔCFS significantly. We perturb the locations and source parameters of the aftershocks within their uncertainties to assess the robustness of the resulting ΔCFS values (Table D-2). This is a first-order comparison of the relationship between mainshock and aftershocks; we do not attempt to evaluate pre-seismic or local stress conditions and ignore ΔCFS generated by post-seismic processes such as afterslip or viscous relaxation, focusing only on the stress changes caused by seismicity.

Independent of the slip model used, within the high-slip zone of the mainshock rupture, ΔCFS is generally reduced on the megathrust, consistent with the Illapel earthquake accommodating a large amount of slip deficit (Figure 6-8; Figure D-6). Inside the mainshock rupture zone there are also regions of positive ΔCFS associated with areas of relatively low slip. Outside the rupture area, ΔCFS is increased by >0.01 MPa (an empirical threshold for triggering
seismicity; Stein et al., 1992) over a broad region of the interface, with positive loading extending ~100 km along-strike to the north and south of the rupture limits. Although the modeled stress effects of this earthquake also extend far down-dip, the plate boundary deeper than ~50 km is at a high enough temperature that inter-seismic shear stresses can relax relatively rapidly and the interface frictional characteristics may be unfavorable for nucleating earthquakes (Scholz, 1998; Hayes et al., 2012). Therefore, we do not consider co-seismic $\Delta CFS$ as a potential megathrust earthquake triggering mechanism deeper than 50 km, which is consistent with the scarcity of seismicity both before and after the Illapel earthquake in this part of the plate boundary.

The $\Delta CFS$ pattern resolved onto the megathrust interface is similarly reflected in the cumulative $\Delta CFS$ resolved onto the aftershock fault planes, consistent with the interpretation that most of the aftershocks occurred on or near the megathrust interface. Using our seismic model, the relocated hypocenters and source parameters of the aftershocks, and the average elastic properties of the half-space, of the 234 relocated aftershocks with moment tensor solutions, 128 (55%) are positively loaded by the preceding seismicity, 96 (41%) are negatively loaded, and 10 (4%) are negligibly loaded. As a test of whether the mainshock actually increased the likelihood of aftershocks, we compare the percentage of positively loaded aftershocks from the Illapel sequence to a control set of events in the same geographical region (71.0°-73.0°W, 29.5°-32.5°S) from 1980 to just before the 2015 Illapel rupture (following the approach of Hardebeck et al., 1998). Resolving the $\Delta CFS$ from the seismic FFM onto 123 historical events from the GCMT catalog near the rupture area, we find that 61 (50%) have positive $\Delta CFS$ and 62 (50%) have negative $\Delta CFS$. The increase in percentage of positively loaded events from the control set is similar to or slightly smaller than other earthquake sequences along the subduction zone, e.g., during the 2010 Maule aftershock sequence, where 57-66% of aftershocks occurred in areas of increased $\Delta CFS$ (from 55-60% in the control set; Hayes et al., 2013), or the 2014 Iquique foreshock sequence, where 70% of the foreshocks occurred in areas of increased $\Delta CFS$ (from 46% in the control set; Herman et al., 2016). As Hayes et al. (2013) recognized, the relatively small increase in positively loaded events is likely due to the large number of aftershocks superimposed on the down-dip region of high mainshock slip, where the $\Delta CFS$ varies significantly over short spatial scales and is particularly sensitive to the mainshock slip distribution and aftershock locations. Out of the 4 slip models tested, 3 produce similar numbers of aftershocks with positive loading (the seismic and geodetic models from this study and the Melgar et al. (2016) model), whereas the Li et al. (2016) model generates a significantly lower number of positively loaded aftershocks (46%) (Table D-2).
3. Discussion

3.1. Mainshock Slip Models

In addition to the seismic and geodetic slip models presented in this study, numerous other slip distributions have been published for the Illapel earthquake. These include models inverted from a variety of datasets: teleseismic broadband waveforms (Ye et al., 2015; Lee et al., 2016), teleseismic broadband data augmented by tsunami observations (Heidarzadeh et al., 2015; Li et al., 2016), joint teleseismic broadband, regional strong motion, high-rate GPS, static GPS, and InSAR (Tilmann et al., 2016), joint high-rate GPS, strong motion, InSAR, and tsunami (Melgar et al., 2016), joint tsunami and static GPS and InSAR (An and Meng, 2017; Williamson et al., 2017), and static geodetic observations alone (Barnhart et al., 2016; Ruiz et al., 2016; Feng et al., 2017). Because the various geophysical observations are sensitive to different aspects of the deforming subduction system and the model setups and inversion approaches differ, there are differences in the precise timing (in kinematic models), distribution, and amount of slip. However, in general these models have broadly similar characteristics. Here we explore the features of the Illapel rupture that are robustly constrained by observations, i.e. that are insensitive to modeling approaches, and are common across multiple models incorporating different datasets.

Kinematic models of the Illapel mainshock all start at a hypocenter near the coast in the vicinity of the initial USGS-determined location (71.67°W, 31.57°S). Although the east-west position of the hypocenter varies from study to study by 15-20 km and the north-south position varies by up to 10 km, the effects of shifting the hypocentral position within this range on the resulting slip model appear to be much smaller than the effects of fault geometry and rupture constraints. The timing of moment release throughout the duration of the rupture is well constrained in any model making use of datasets that capture seismic wave propagation (i.e. teleseismic broadband, regional strong-motion, and near-field high-rate geodetic observations). In sensitivity tests of our seismic FFM geometry and rupture constraints, we find that every model source-time function peaks at ~50 s and is similar in shape to the one shown in Figure 6-3. This timing is consistent with the moment release indicated by our W-phase moment tensor inversions (Section 2.1), as well as most other published FFM solutions. All of these analyses indicate a rupture lasting 100-120 s, although later, relatively low moment release has been inferred by some to continue until ~250 s after the origin time (Lee et al., 2016).
While the timing of co-seismic slip varies little with model parameterization in kinematic models of the Illapel mainshock, the precise location and amplitude of slip are more model dependent, particularly in models based on far field observations. In sensitivity tests of our seismic FFM to fault geometry, velocity model, and rupture velocity, the peak slip location shifts by up to 50 km in the along-strike (north-south) direction and up to 30 km in the horizontal along-dip (east-west) direction (Figure D-4). Even inversions based on near-field observations (e.g. GPS, InSAR, and regional strong motion) show model-dependent slip location variability. For example, the geodetic slip inversion in this study is identical to the model from Barnhart et al. (2016), except for the fault geometry, and the two solutions are qualitatively similar. However, despite good theoretical resolution of the slip latitude provided by onshore geodetic data, the center of shallow slip determined in this study is 25 km north of the center of shallow slip in Barnhart et al. (2016). The variability in slip location from other published slip models is similar, suggesting an average uncertainty in slip position of ~25 km in both along-strike and along-dip directions. This uncertainty is significantly larger than our relocated aftershock hypocenter location uncertainty (less than 6 km), making direct comparison between slip and aftershocks difficult, especially within the rupture footprint.

Although the precise slip location is uncertain, there are several common features of all models irrespective of the data used and the model setup. All models have dominantly reverse slip north of the hypocenter (along-strike) and northwest of hypocenter at a shallower depth (i.e. along-strike and up-dip), with peak slip estimates ranging from 6-12 m. Most models have an along-strike length of 100-150 km spanning a latitude range of ~30.5°-31.5°S and slip extending from near the trench to near the coast. On average, the deep and shallow slip have similar amplitudes and do not differ in any model by more than a factor of 2.

Some features of the slip distribution do not appear in every model and interpreting their significance is more speculative. In models with less smoothing, the deep and shallow high slip patches are separated by a local minimum in slip (e.g. our seismic model). This distinction between slip regions may also be reflected in the radiated frequency content from these two regions; the deeper slip radiates higher frequency seismic waves and the shallower sections radiate lower frequency seismic waves (Melgar et al., 2016; Tilmann et al., 2016). The amount of rupture to the trench seems to be one of the most poorly resolved features and even tsunami-based models (sensitive to ocean floor displacements and thus shallow slip) disagree as to whether large slip occurs shallower than ~10 km depth (i.e., within a horizontal distance of ~25 km from the trench). For example, Li et al. (2016) and Melgar et al. (2016) have large slip essentially up to the
trench, whereas the Heidarzadeh et al. (2015), An and Meng (2017), and Williamson et al. (2017) models do not have significant shallow slip at depths less than 10 km. Finally, we do not interpret differences in peak slip amplitude, since there are multiple possible explanations for the variability, including the resolution inherent in the observations, smoothing or other regularization constraints, model elastic properties, and sub-fault dimensions.

3.2. Mainshock-Aftershock Relationship

Previous studies of the Illapel sequence have identified that aftershock seismicity is generally anticorrelated with the mainshock slip and that aftershocks are conspicuously absent in the area of shallow slip (Heidarzadeh et al., 2015; Melgar et al., 2016; Tilmann et al., 2016). Our catalog of precise aftershock locations and kinematics may provide some additional insights into the rupture process of the mainshock and the physical properties of the subduction system. Specifically, we explore (1) the high intensity of seismic activity near the hypocenter early in the sequence, (2) the shallow gap in subduction interface aftershocks, (3) the relationship of deeper megathrust aftershock activity with co-seismic slip or post-seismic deformation, (4) the concentration of normal faulting adjacent to the rupture near the trench, and (5) the aftershocks offset along-strike from the edges of the rupture zone. In addition, we evaluate whether it is possible to interpret the aftershocks in context of static stress changes that occur with the mainshock and then further evolve throughout the sequence.

The region near the Illapel mainshock hypocenter had high seismic activity following the event, hosting several relatively large aftershocks in the first few days. These aftershocks are difficult to interpret through ΔCFS analysis because they are very close to the co-seismic rupture zone and as a consequence the ΔCFS resolved onto their shallow-dipping nodal planes is very sensitive to the mainshock slip distribution, event locations, and fault scaling. Qualitatively, this part of the megathrust appears to have relatively low slip in many models, consistent with having positive ΔCFS and triggering thrust faulting activity. Intriguingly, the Omori decay constant of events within 40 km of the epicenter is 1.4, higher than the rest of the sequence and indicative of unusually high early aftershock activity and/or more rapid decay than the sequence as a whole (Figure 6-9). Chan and Stein (2009) observed a similar positive anomaly in Omori decay constant in the 1999 Chi-chi, Taiwan, aftershock sequence and inferred the anomaly to be associated with static stress loading caused by afterslip and post-seismic relaxation. This interpretation may also apply to the Illapel sequence; since most of the co-seismic slip occurred north and northwest of the hypocenter, the hypocentral region may have been co-seismically positively loaded, causing
unreconciled slip deficit in the region to be accommodated immediately after the rupture through a combination of aftershocks and aseismic afterslip. Following this immediate slip activity (over 2 days), the seismicity decreased back to the aftershock rates associated with the rest of the sequence.

The shallow gap in aftershock activity can in part be explained by negative ΔCFS caused by co-seismic slip. In particular, large amounts of co-seismic slip between the trench and the updip limit of the aftershocks produces negative ΔCFS with amplitude >0.01 MPa on the interface (i.e., the mainshock accounts for a large amount of the slip deficit; Figure 6-8), which appears to inhibit thrust faulting aftershocks. The scarcity of aftershocks within this shallow co-seismic rupture region suggests that slip in this area is relatively homogeneous, with few low-slip areas for aftershocks to fill in. We interpret that this homogeneous shallow slip extends from near the trench to a depth of 20-25 km at 31.0°S, corresponding to the shallowest aftershocks at this latitude. Along-strike to the north and south, the shallowest aftershock depths continue to trace the down-dip edge of the co-seismic slip region, so that where the mainshock slip width narrows, the minimum aftershock depth decreases. However, even at latitudes where there is little to no co-seismic slip, aftershocks do not occur shallower than ~10 km depth, indicating an alternative (but not mutually exclusive) explanation for this remaining gap in aftershock activity; the shallowest region of the megathrust (depth <10 km) has frictional and material properties that inhibit aftershock nucleation (but may allow large ruptures to propagate through). This interpretation is consistent with sparse shallow thrust faulting seismicity along the entire length of the South America subduction zone (e.g., Figure 6-1a), as well as the observations of the Illapel mainshock frequency content; seismic radiation emitted from the shallow subduction interface is lower in frequency than radiation from the deeper portions of the rupture and is interpreted to reflect differences in frictional characteristics of the shallow megathrust (Melgar et al., 2016; Tilmann et al., 2016).

In contrast to the apparently straightforward behavior of the shallow parts of the megathrust in the Illapel sequence, deeper aftershocks do not clearly reflect a static stress change triggering mechanism. Although aftershock hypocenters in the down-dip region of the subduction zone are clustered into coherent strands surrounding relative gaps in seismicity, these gaps do not correlate with particular features of mainshock slip models. This mismatch may be due in part to the uncertainties in co-seismic slip location. One possibility is that the aftershocks bound the high-slip areas of the mainshock rupture, consistent with being triggered by positive ΔCFS around the edges of slip regions (e.g., Briggs et al., 2006). This interpretation implies that the co-
seismic slip deeper than ~25 km is significantly more heterogeneous than the shallow slip, according to the geometry of the gaps between aftershock strands. Several published slip models, including our seismic model, have proposed such heterogeneous slip in the depth range 25-45 km, but with the current inability to resolve rupture details at the same scale as aftershock locations, this interpretation remains speculative.

Normal faulting aftershocks are found only near the trench, dominantly between 30°S and 31°S, although some near-trench seismicity extends as far south as 32°S. These events have positive $\Delta CFS$ from the mainshock if shallow co-seismic slip occurs immediately adjacent to them, as in our geodetic slip model (Figure 6-10). If mainshock slip is displaced along-strike or occurs farther down-dip, the normal faulting events move into a negative or low $\Delta CFS$ zone. Therefore, if the location and amplitude of shallow mainshock slip are interpreted to exert control through static stress transfer on the location and intensity of normal faulting seismicity, then shallow slip must occur within ~15 km of the trench (measured horizontally), corresponding to a depth of ~5 km, in the section of the subduction zone from 30°S to 31°S. Such a relationship may also imply a connection between the stress state within the subducting plate and the stage of the seismic cycle, from inter-seismic compression (indicated by thrust faulting mechanisms) to co-seismic tension (e.g. Lay et al. 2009). However, historical seismicity in this segment does not clearly reveal any temporal variations in earthquake focal mechanisms in this part of the plate boundary.

Finally, $\Delta CFS$ very likely accounts for the two clusters of aftershocks offset along strike from the edges of the rupture zone (Figure 6-6; Figure 6-8), because irrespective of the model used, these events are in a dominantly positive $\Delta CFS$ region. These earthquakes are slightly delayed from the immediate aftershock activity; to the south, aftershocks started the day after the mainshock rupture, whereas in the north seismic activity did not occur until 55 days after the Illapel event (Movie S1). This may reflect a static triggering process for remote events that differs from the process that induces seismicity near the hypocenter and rupture footprint. We propose that these aftershock clusters are asperities that have been pushed over the frictional failure threshold by elastic mainshock deformation (plus any afterslip or post-seismic deformation) acting at a distance. This along-strike loading of the interface may also trigger future earthquakes whose ruptures are able to propagate over a larger area on the megathrust, producing larger magnitude events.
3.3. Inter-seismic Coupling and Mainshock Rupture

One important seismotectonic implication of this analysis is the relationship between inferred coupling on the megathrust interface and the mainshock rupture. A typical expectation for large megathrust earthquake rupture zones is that they will dominantly correspond with regions of high coupling on the interface. In the Illapel segment of the South America subduction zone, pre-earthquake GPS observations indicated a transition from high coupling south of 31ºS, to intermediate/low apparent coupling north of 31ºS (Metois et al., 2016; Figure 6-11). Metois et al. (2016) argue that the 2015 Illapel earthquake rupture halted at this transition with minimal overlap. We revisit this interpretation with an understanding that the slip models for the Illapel earthquake vary significantly in their along-strike extent. Similarly, inter-seismic coupling distributions derived from the motion of onshore stations have low resolution (and thus comparably large spatial uncertainties) in the shallow, offshore regions of the megathrust near the trench (e.g., Loveless and Meade, 2010). Our seismic model has the northern edge of the rupture at 31.0ºS, whereas several other rupture models have significant slip north of 31.0ºS (our geodetic model; Li et al., 2016; Melgar et al., 2016; Figure 6-11). These latter models suggest there may be 20-40 km of overlap between the northern rupture area and the apparent low-coupling zone defined by Metois et al. (2016). This does not preclude the interpretation that the rupture slowed and halted as it entered this zone to the north, but may suggest that large megathrust ruptures can propagate into zones interpreted to be low-coupling rather than remaining constrained to the high-coupling areas.

In addition, the interpretation that the South America subduction zone from 28º-31ºS has relatively low coupling is inconsistent with the historical record of great earthquakes in the segment (1819 M ~8.5; 1922 M ~8.4; Figure 6-1b). This implies that coupling inferred over a relatively short interval of time may not always be useful for anticipating the likely locations of large megathrust ruptures (e.g., Witter et al., 2016). A low level of apparent coupling may reflect either (a) that the degree of coupling can vary over time throughout the earthquake cycle (e.g., Meltzner et al., 2015; Tsang et al., 2015), or (b) the plate boundary is in fact coupled north of 31ºS, but some other process generates surface deformation overprinting the locking signal (e.g., Furlong et al., 2016). If inter-seismic coupling models are to be effectively applied as a tool for anticipating megathrust earthquakes, more work needs to be done to understand why some earthquakes rupture into or occur entirely within segments that today appear to have intermediate to low coupling values.
3.4. Subduction Zone Seismic Behavior

Through these combined geophysical datasets, we place the observations from the Illapel earthquake sequence into a synoptic model of subduction zone earthquake behavior based on the model of Lay et al. (2012) (Figure 6-12). The mainshock slip distribution, relocated hypocenters, and ∆CFS analysis are consistent with a segmented megathrust (Lay et al., 2012): a seismogenic zone at depths of 20-40 km characterized by heterogeneous slip and vigorous aftershock activity; a shallow megathrust that clearly accumulated inter-seismic slip deficit (as evidenced by large amounts of co-seismic slip) but appears to have physical characteristics that result in comparatively homogeneous slip and few aftershocks; and a down-dip transition to a state that is unfavorable for either rupture propagation or aftershock nucleation at depths greater than 50 km.

One important issue is how the shallow section of the megathrust can have high co-seismic slip amplitudes when its inferred frictional properties are incompatible with accumulating significant amounts of elastic strain in between large earthquakes. This is particularly critical for constraining tsunami hazards in subduction zones. Models of elastic strain accumulation around a locked asperity indicate that the regions on the megathrust adjacent to the frictionally locked components of the system should accumulate slip deficit irrespective of their frictional properties (e.g. Burgmann et al., 2005; Herman et al., 2016). If the up-dip edge of the frictionally locked zone is at a depth of 20 km, these models suggest that the slip deficit accumulation rate at the trench should still be ~50% of the full plate motion. Thus, a rupture that is able to propagate into the shallow region, as apparently happened during the Illapel earthquake, would result in substantial slip occurring at very shallow depths even in regions considered to be frictionally unsuitable for earthquake nucleation. A similar situation occurs at the down-dip limit of the seismogenic zone at depths of 40-50 km or greater. However, the higher temperatures and corresponding visco-elastic rheology of the surrounding rocks complicate the loading process in this part of the system and thus ruptures may not be able to propagate much farther down-dip.

In the Illapel sequence, the aftershocks can be separated into three main categories (Figure 6-12): (1) thrust-faulting events within the zone of or immediately adjacent to the mainshock rupture; (2) thrust-faulting events distinctly separated from the mainshock rupture; and (3) normal-faulting events near the trench. A fourth category for which we observe few events in this sequence is intraplate thrust or strike-slip aftershocks. In general, we interpret that aftershocks in all three of these categories were positively loaded prior to their rupture, although because of the complexity of mainshock slip and other uncertainties, ∆CFS remains a poor predictive tool for precise aftershock locations. One question that arises from such a
categorization is whether these earthquakes reflect different deformation processes. It is clear that
the normal-faulting earthquakes are reflecting intraplate deformation (perhaps in both the
subducting and upper plates) and they may occur in response to stress changes caused by shallow
slip in the mainshock. In contrast, it is more difficult to distinguish thrust-faulting earthquakes
that are all interpreted to have occurred on the plate boundary. However, since we interpret
Category 1 events to be filling in gaps between mainshock asperities, whereas it is possible to
interpret Category 2 events as rupturing asperities separated from the mainshock zone, these
aftershocks may have different seismological characteristics that could help distinguish their role
in the seismotectonic setting of the subduction zone. Specifically, we might expect seismic
analyses of the Category 2 events to reflect a “typical” earthquake rupture, whereas Category 1
events may be associated with afterslip, post-seismic relaxation, or temporal changes in friction
that generate seismic signals with different spectral characteristics.

4. Conclusion

Through multiple geophysical approaches, we constrained the deformation processes
acting during the 2015 Illapel mainshock and its aftershock sequence. The mainshock was a
simple thrust faulting earthquake occurring on the subduction interface. It ruptured a large area on
the megathrust north and northwest of the hypocenter, from near the trench (< 10 km deep) to
beneath the coast of Chile (~50 km deep). There is some uncertainty in the amount and
distribution of slip, but the deep and shallow slip patches appear to have similar amplitudes. On a
regional scale, slip occurs in the region with the lowest number of precisely located aftershocks,
but comparing slip and aftershocks at a resolution of < 25 km is difficult because of the inherent
uncertainties in mainshock slip location and amplitude. Nevertheless, some aftershocks appear to
be associated with static stress changes generated by the Illapel rupture, particularly the events
that occur at the down-dip edge of the shallow rupture region, the events that are offset along
strike from the northern and southern ends of the rupture, and the normal faulting events. Relating
aftershocks in the depth range 25-45 km with mainshock slip through static stress change
mechanisms is difficult given the current inability to precisely constrain the spatial distribution of
mainshock slip, but the clustering of these aftershocks suggests they might bound high-slip areas.
The observations of mainshock slip from coast to trench, aftershocks surrounding the deeper slip,
and a dearth of aftershocks associated with shallow slip help understand the frictional and
deformational state of the megathrust throughout inter-seismic loading, co-seismic rupture, and
post-seismic aftershock and aseismic activity. They reveal a frictionally heterogeneous
megathrust that accumulates slip deficit from near the trench to depths of ~50 km, which can all be released in ruptures of the interface, as apparently occurred during the Illapel earthquake.

**Acknowledgements**

This work was supported by NASA Earth and Space Science Fellowship 16-EARTH16R-72. Many of the figures in this manuscript were created using the Generic Mapping Tools (Wessel and Smith, 1991). Data, techniques, and results presented are available in Appendix D and references. Seismic waveform data and associated station metadata are available from the Incorporated Research Institutions for Seismology Data Management Center (http://www.iris.edu). We appreciate helpful comments from R. Briggs, M. Guy, D. McNamara, two anonymous reviewers, and the Associate Editor. Any use of trade, firm, or product names is for descriptive purposes only and does not imply endorsement by the U.S. Government.
References


Figure 6-1. Tectonic setting and historical seismicity near Illapel, Chile.

(A) Tectonic setting and seismicity in the Illapel section of the South America subduction zone (defined as the region between where the Challenger Fracture Zone and Juan Fernandez Ridge enter the trench). The relative velocity between the subducting Nazca plate and the upper South America plate is 74 mm/yr, at an azimuth of ~80° (black arrow). The depth of the megathrust plate boundary (Slab1.0; Hayes et al., 2012) is indicated by the light grey dot-dash contour lines, drawn every 10 km. The 2015 Illapel earthquake hypocenter is shown as a yellow star and the rupture area with slip larger than 1 m from the seismic slip model determined in this study is the shaded grey region. The seismic station C110, where the largest ground acceleration measurements were obtained, is shown as an orange triangle. Earthquakes M 4.5+ from 1900-2016 are plotted as circles scaled by magnitude. Historical earthquakes prior to the 2010 Maule earthquake are colored transparent grey, seismicity from the Maule earthquake up to the 2015 Illapel earthquake are light green and the dark blue to light red circles are the aftershocks of the Illapel earthquake. The left offset panel shows earthquake epicenter latitude versus number of days after the Illapel mainshock, with events colored by time corresponding to the map. (B) Historical large and great earthquakes in the South America subduction zone (Kelleher, 1972; Beck et al., 1998; Lomnitz, 2004). The Illapel earthquake is highlighted in red for reference and the grey bar indicates the Illapel section. Bold lines are events measured or inferred from shaking effects to be M 8.0+ and smaller significant events are in grey.
Figure 6-2. Moment tensor solutions for Illapel earthquakes.

(A) Single source moment tensor solutions for the Illapel mainshock and 234 aftershocks M 4.0-7.0. Earthquake focal mechanisms are plotted at their relocated epicenters, scaled by magnitude, and colored by time (as in Figure 6-1). The single source mainshock moment tensor is shown at the bottom right corner of the map. Events M 6.0+ have their magnitudes indicated above the focal mechanism. Note the dominance of thrust faulting mechanisms east of the trench and normal faulting mechanisms near the trench. In the bottom left is a ternary diagram to categorize the focal mechanisms (Frohlich, 1992). Thrust mechanisms dominate the aftershock sequence, with a secondary population of normal faulting events and no strike-slip events.

(B) Double source moment tensor solution for the Illapel mainshock (sub-event kinematics indicated by strike/dip/rake). The event separates into two sub-events: a smaller Mw ~7.2 initial sub-event (left mechanism) and the dominant Mw 8.2 sub-event (right mechanism). Waveform fits from observations recorded at 3 broadband stations are shown below the sub-event focal mechanisms. The locations of the stations are indicated as red dots on the map and other stations used for the W-phase inversion are shown as green dots. Black traces are observed waveforms and red dashed traces are synthetic. The amplitudes of the traces are given in mm. The grey box indicates the W-phase window.
Figure 6.3. Map of the seismic finite fault model solution determined in this study.

Each rectangular sub-fault is colored by its slip amplitude and slip vectors are shown as scaled arrows in each sub-fault. Rupture front timing contours are plotted on top of the slip distribution and labeled in seconds. The earthquake epicenter is shown as a green star. The inset plot shows the relative moment release rate over time. Waveform fits for displacement and velocity traces recorded at 3 broadband stations are shown at right, including fits to P, S, and surface waves. Black traces are observed waveforms and red dashed traces are synthetic. The maximum absolute amplitude of the trace is given at the left of each trace in um or um/s. Locations of seismic stations where the plotted waveforms were observed are indicated on the map as red dots and the locations of other stations whose waveforms were used in the FFM inversion are the green dots. Black concentric circles centered on the earthquake epicenter (yellow star) indicate teleseismic distances of 30º and 90º from the earthquake.
Figure 6-4. Map of the geodetic slip model solution determined in this study.

Each triangular sub-fault is colored by its slip amplitude using the same color scale as in the seismic FFM in Figure 6-3. The earthquake epicenter is shown as a green star for reference. Fits to static surface displacements at nearby GPS stations are shown as arrows; black arrows are observed and red arrows are synthetic. The scale of the vectors is shown in the bottom right. Note the different scale for horizontal and vertical displacements. Fits to InSAR interferograms are provided in Figure D-4.
Figure 6-5. Results of hypocenter relocation.

(A) Map of relocated earthquake epicenters. The grey dots are the original NEIC locations and these are connected to the relocated epicenters. The relocated epicenters are colored by whether the event was moved up (red) or down (blue). (B) Horizontal relocation histogram, showing events shifted by up to 25 km, but dominantly shifted by 1-10 km. (C) Vertical relocation histogram, colored the same as the circles on the map. The relocation moved an approximately equal number of events up and down, with an average vertical shift of ~5 km. (D) Polar histogram of horizontal shift azimuths. The relocation moves a large fraction of events to the east-southeast and very few events to the north. (E-G) Hypocenter location uncertainty determined from the relocation procedure. The east-west uncertainty is much larger than the north-south and even vertical uncertainties, reflecting the distribution of stations relative to the events and prior constraints.
Figure 6-6. Map of relocated aftershocks, scaled by magnitude and colored by time.

Cross-sections of earthquake depth labeled on the map as dashed lines are shown at right and are superimposed on Slab1.0 (dashed line; Hayes et al., 2012) and ETOPO1 bathymetry/topography (solid line; Amante and Eakins, 2009). Each cross-section is separated by 120 km (perpendicular to the orientation of the cross-section) and earthquake hypocenters within 60 km of the cross-section are plotted on each panel. Note the clusters of events offset from the central region of aftershocks both to the north of cross-section A and south of cross-section C. Events with moment tensor solutions are shown with their focal mechanisms projected onto the cross-section. These cross-sections show that the majority of the seismicity can be interpreted to be ruptures on the megathrust, with minor intraplate activity near the trench and little significant upper plate seismicity.
Figure 6-7. Illapel earthquake sequence catalog statistics, using the same set of events as shown in Figures 6-1 and 6-6.

(A) Earthquake magnitude versus number of days after the mainshock. There is a clear decrease in aftershock magnitudes in the 50 days following the mainshock, but two of the largest aftershocks in the entire sequence occur after this period. (B) Number of events per day versus number of days after the mainshock (black line). The aftershock rates decrease following a standard Omori decay law with a decay constant of 1 (red line; Utsu et al., 1995). Activity is extremely vigorous for the first 15-20 days before decaying to lower (but still elevated relative to pre-2015 levels) seismicity rates. (C) Cumulative number of events in our catalog larger than magnitude versus moment magnitude.
Figure 6-8. Coulomb failure stress change analysis.

The background colors show the $\Delta$CFS of the Illapel mainshock (using the seismic slip model, with 1 m slip contours shown in white) resolved onto the megathrust fault plane. Areas of high mainshock slip tend to have reduced $\Delta$CFS (blue) and areas of low mainshock slip have increased $\Delta$CFS (red). The coarseness of the $\Delta$CFS distribution reflects the sub-fault dimensions from the slip model. Slab1.0 contours are shown for reference and the plate boundary deeper than 50 km is shaded because this region does not tend to produce seismic activity. Aftershocks are colored by the cumulative $\Delta$CFS from all preceding seismicity resolved at their relocated hypocenter, on the eastward dipping nodal plane for thrust events. Normal faulting event fault planes are more ambiguous, but the choice of fault plane makes little difference in the $\Delta$CFS value for these aftershocks.
Figure 6-9. Results of a grid search over the geographical area of the Illapel aftershock sequence for Omori decay constant within a circular area of radius 40 km.

When the center of the selected region is near the Illapel mainshock hypocenter (yellow star), the Omori decay constant for the enclosed events is above 1.3, which is significantly higher than the value corresponding to the rest of the sequence (0.8-1.2). This suggests relatively higher early activity in this region than in the rest of the sequence.
The thrust faulting geometry is (strike=0°, dip=20°, rake=90°) and the fault lies on a planar subduction interface dipping at 20° to the east. The target normal faulting geometry is (strike=0°, dip=45°, rake=-90°). ∆CFS is resolved at a target fault depth of 10km. (A) When the up-dip edge of the thrust fault is 15 km from the trench at a depth of 5 km, normal faulting is promoted at a depth of 10 km in the along-strike footprint of the rupture. (B) In contrast, if the up-dip limit of the rupture is 40 km from the trench at 15 km depth, then normal faulting is not promoted near the trench at all. The result is not significantly changed by using a normal faulting plane that dips to the west instead of the east.
Figure 6-11. Relationship between inter-seismic coupling and Illapel mainshock slip. The 50% or greater apparent coupling region from Metois et al. (2016) is shown shaded in red. We overlay the 1 m slip contour from four slip models: the seismic model from this study (green), the geodetic model from this study (blue), the model from Li et al. (2016) (orange), and the model from Melgar et al. (2016) (purple). Although in all the models slip dominantly occurred coincident with the wide coupling zone south of 30.5ºS, three of the models have significant slip north of this latitude.
Figure 6-12. Schematic interpretation of Illapel sequence seismotectonics.

The main shock slip patches are represented by grey areas and reflect two different rupture characteristics: deeper slip (20-40 km) is more heterogeneous, with aftershock seismicity (blue dots) interpreted to surround the slip patches, filling in slip deficit immediately adjacent to the mainshock rupture zones. At shallower depths (0-20 km), the rupture is smoother and is associated with very low levels of megathrust aftershock activity. However, this shallow slip may correspond spatially to the normal faulting aftershocks interpreted to occur within the subducting plate (purple dots). Finally, there are earthquakes separated from the mainshock rupture area and its immediate aftershock footprint (green dots). These events are unambiguously positively loaded by the mainshock (loaded area shown as a red background on the interface) and are interpreted to occur in response to these stress changes.
Chapter 7

Patterns of Loading and Unloading in the Upper Plate Throughout the Earthquake Cycle in Japan

1. Introduction

Following the elastic rebound of the upper plate that occurs large subduction zone earthquakes, several post-seismic processes act to facilitate continued trenchward motion of the upper plate (Cohen, 1999; Wang et al., 2012): aftershocks and aseismic slip on the interface (Marone, 1998), bulk visco-elastic relaxation of the lower lithosphere (Melosh and Raefsky, 1983), and plate motion across a re-locked subduction interface (Savage, 1983). Determining the relative contribution of each of these processes to the observed post-seismic surface displacement field constrains how strain and stress are transferred to and from the rupture zone and the process by which the interface frictionally re-locks after an earthquake (Pollitz et al., 2006; Hu et al., 2016; Klein et al., 2016), which are critical aspects of subduction zone seismotectonics and the associated buildup of seismic hazard after a large event.

In this study, we focus primarily on the bulk viscous relaxation component of the post-seismic deformation signal; viscous flow after an earthquake is commonly inferred to respond directly to co-seismic stress changes (Nur and Mavko, 1974; Pollitz, 1997; Wang et al., 2001). However, analyses of post-seismic viscous relaxation often implicitly assume that the pre-seismic state of stress is zero; in fact, the pattern of post-seismic viscous relaxation depends on the residual distribution of stresses that remains following the earthquake (Kato, 2002; Johnson and Segall, 2004). Therefore, models of post-seismic viscous relaxation should consider the integrated deformation history of the subduction zone throughout previous earthquake cycles (Lambert and Barbot, 2016).

An additional complication to determining this strain/stress history in the subduction system is its rheological variability, particularly the temporal variations in deformation mechanisms that depend on the time scales and strain rates acting during the particular stage of the earthquake cycle (Ranalli, 1987). Over the short time scales of the co-seismic rupture and seismic wave propagation (10^1-10^4 seconds), strain rates are high enough that the lithosphere and asthenosphere deform dominantly through elastic and brittle mechanisms (Figure 7-1A).
Deformation in the subduction system over longer, inter-seismic and post-seismic time scales (10⁵-10⁹ seconds, i.e., days to years) is more appropriately seen as visco-elastic. Loading of the subduction system involves deformation of the lithosphere at low strain rates (ranging from 1x10⁻¹⁶ s⁻¹ far from the locked fault to 5x10⁻¹⁴ s⁻¹ adjacent to the fault in these models, assuming plate motion of 80 mm/yr). At these strain rates, cooler rocks (<600ºC; Turcotte and Schubert, 2002) deform through brittle or elastic processes, whereas hotter rocks relax elastic stresses through viscous flow. The thermal structure of a subduction zone consists of a cold slab and an upper plate that increases in temperature with depth; the depth-varying temperature in the upper plate implies that its inter-seismic rheology also varies with depth (e.g., McKenzie, 1969; Andrews and Sleep, 1974; Anderson et al., 1978; Furlong et al., 1982; Honda, 1985; Peacock, 1996; Syracuse et al., 2010; Figure 7-1b). The top ~50 km of the upper plate is cooler and will be dominantly elastic, accumulating elastic strain and stress while the subjacent warmer section will relax such stresses on the time scale of the earthquake cycle. Thus, the deformational state of the upper plate immediately before an earthquake is that of an elastically compressed upper section and a deformed but elastically relaxed lower section.

Because of these differences in inter-seismic and co-seismic rheology, the stress distribution (and corresponding surface displacements) in the rheologically heterogeneous upper plate immediately before an earthquake is not identical to the pattern of stress change in the co-seismic, elastic-only Earth. As a result, there may be locations where elastic stresses remain unreleased by the earthquake or are in excess of their expected values; these stresses may relax through additional deformation after the earthquake.

Observations of surface displacements throughout the earthquake cycle corroborate how the spatial pattern of deformation in subduction zones varies substantially between the inter-, co-, and post-seismic stages, demonstrating that the co-seismic strain is not simply the reverse of the pre-seismic strain accumulation. The rheological variations at each stage of the earthquake cycle are apparent in the surface displacement record in Japan, where there is a dense network of Global Positioning System (GPS) stations that recorded inter-seismic motion prior to 2011, displacements associated with the 2011 Mw 9.0 Tohoku earthquake, and the post-seismic deformation following the event (Figure 7-2; Nishimura, 2014). Prior to the earthquake, the island of Honshu was shortening and moving westward as a consequence of plate coupling on the interface with the converging Pacific plate (Figure 7-2A). The surface velocity (GPS station position change over time) vector magnitudes decrease approximately linearly from east to west, indicating that the upper plate was essentially being uniformly shortened. During the earthquake,
the co-seismic surface displacements are not simply the inverse of the loading velocities; rather, the eastward co-seismic displacement magnitudes increase non-linearly from western Honshu to the eastern coast (Figure 7-2B). Finally, the post-seismic velocities indicate continuing eastward motion of Honshu that appears to be occurring at a relatively constant rate, i.e. it appears to have a lower spatial gradient than either the pre-earthquake or co-seismic pattern (Figure 7-2C).

In this study, we simulate the distribution of inter-seismic and co-seismic stresses and surface displacements to evaluate whether the post-seismic deformation seen in Japan is consistent with relaxation of the residual stresses remaining after the earthquake. We model elastic strain accumulation using boundary conditions and material properties that represent heterogeneous lithospheric rheology. We compare the deformation accumulated during the inter-seismic stage to the detailed time evolution of displacements during the Tohoku earthquake rupture and the corresponding co-seismic stress changes. The co-seismic rebound process is well matched by elastic theory. We then compute the difference between the rheologically more complex inter-seismic deformation state and the co-seismic elastic deformation and compare the post-seismic deformation predicted by this modeling to the observed post-seismic surface displacements from Japan.

2. Inter-seismic Deformation

We use a finite element model (FEM) to simulate the deformation produced by inter-seismic loading of a rheologically layered upper plate (Figure 7-3). The subducting plate has a uniform dip of 25° and a thickness of 100 km. It is displaced by 1 meter at its top and bottom, parallel to the dip direction. The upper plate is coupled to the subducting plate from 10 to 50 km depth, which corresponds to a typical seismogenic depth range (e.g. Hayes et al., 2012; Lay et al., 2012). To approximate the deformation of a rheologically layered (elastic over viscous) upper plate, the upper plate in this model is only 50 km thick (Turcotte and Schubert, 2002); along its bottom boundary, we allow free horizontal motion to simulate the stress-relaxation behavior of the subjacent viscous layer. To avoid the upper plate flexing freely into the empty region below it, we assign Winkler (spring-like) boundary conditions along the base of the upper plate, with a spring constant of 1x10^2 N/m. This spring constant is chosen based on testing a range of FEMs with known analytical solutions. The top of the subducting plate at the base of the upper plate is constrained to move along the plate interface. The deformation in this layered model is compared to that of a model with a homogeneous elastic upper plate, i.e., the upper plate is elastic from the
surface to the megathrust (e.g., Chapter 4). In this study, we treat the slab as entirely elastic, although its base is likely warm enough to deform through viscous flow. We do not expect the relaxation of stresses at the base of the slab to have a major effect on the deformation field, because models with varying slab thickness showed little sensitivity in the resulting displacement field for slabs thicker than ~20 km.

A cross-sectional view through the layered model (Figure 7-4) shows large displacements in the upper plate that decrease in magnitude and rotate to vertical towards the backstop. The largest effective shear stresses (the second invariant of the deviatoric stress tensor) in this model are ~0.15 MPa (for 1 meter of displacement) and these are distributed almost uniformly throughout the top elastic layer of the upper plate (Figure 7-4A). Examination of the stress components indicates that the dominant contribution to the deformation of the upper plate is from $\sigma_{xx}$, i.e., horizontal shortening of the upper plate (Figure 7-4B). This uniform compression of the upper plate differs from the pattern of stresses that are generated in a model with a completely elastic upper plate that extends from the surface to the megathrust (Chapter 4; Figure 7-5). When the entire upper plate volume is elastic, stress is produced throughout the upper plate, generating horizontal surface displacements that decrease non-linearly with distance from the trench (Figure 7-4; Figure 7-5). The visco-elastic representative model also generates smaller near-fault stress values because these are relieved through transference to the subjacent visco-elastic region. The deformation state generated in this representation of an elastic over visco-elastic layered subduction zone is that of a compressed, elastically stressed upper section and a deformed but elastically relaxed lower section (Figure 7-1). To assess how this elastic stress is released in the co-seismic stage, we use elastic theory and examine high-rate GPS observations from the 2011 Tohoku earthquake.

3. Co-seismic Observations

During the earthquake rupture, a variety of deformational processes affect the displacement evolution of the upper plate. As the upper plate is rebounding to its post-rupture state, transient seismic waves radiate from the slipping fault zone through the region of heterogeneous loading (Aki and Richards, 2002). We explore the nature of these transient waves and how they interact with the rebounding upper plate in Appendix E. The synthetic final displacements generated by placing the teleseismic rupture model in an elastic half-space agree with the observed displacements reasonably well, although there are large systematic misfits
throughout northern Honshu (Figure 7-6). If the synthetic final displacements exceeded the observed displacements throughout Honshu, we might interpret the signal as elastic rebound restricted due to the effect of coupling an elastically strained top of the upper plate with an elastically relaxed base of the upper plate. However, because the misfit is spatially isolated, we prefer the interpretation that slip in the rupture model beneath the coast at ~39ºN is mislocated and is contributing excess displacements at nearby GPS sites. Such slip mislocations (up to 50 km) can occur in teleseismic-based inversions, since the rupture is not anchored in space by regional observations (Hayes et al., 2013). We treat the co-seismic deformation as a purely elastic process. Therefore, to evaluate how co-seismic rebound changes the stresses in the subduction system and compute the remaining post-seismic stresses, we apply analytical solutions for faulting in an elastic half-space (Okada, 1992).

For direct comparison with the deformation produced in the inter-seismic layered rheology FEM, we calculate the distribution of co-seismic stress changes for the same subduction and faulting geometry, but placed in an elastic half-space. We assign 1 meter of reverse slip to occur everywhere within the region that was locked in the inter-seismic loading model (Figure 7-3; Figure 7-4) and compute the stress changes for a homogeneous elastic half-space using the equations of Okada (1992). The half-space has identical mechanical properties as the elastic components of the inter-seismic model (Young’s modulus=1x10¹¹ Pa; Poisson’s ratio=0.25).

We sample the components of stress on a dip-parallel cross-section through the center of the fault zone. The effective shear stress (Figure 7-7A) is very large near the fault (>0.5 MPa) and all of the non-zero components of the stress tensor (Figure 7-7B-E) have significant contributions to these high effective shear stresses. These produce large shear (Figure 7-7F) and normal (Figure 7-7G) stresses resolved onto the megathrust geometry; in particular, large shear stresses extend more than 200 km down-dip from the earthquake. The co-seismic, elastic-only stress changes have distinctly different footprints than stresses generated in the layered FEM representing inter-seismic loading. In the following section, we quantitatively compare these co-seismic and inter-seismic stresses in order to understand the spatial pattern of post-seismic deformation.

4. Post-seismic Stress

We estimate the distribution of post-seismic stresses in the subduction system by adding the co-seismic stresses generated by fault slip in an elastic half-space (Figure 7-7) to the inter-seismic stresses in the layered FEM (Figure 7-4), resulting in a representation of the stresses after
the earthquake (Figure 7-8). The stresses within ~150 km horizontal distance from the trench are significantly reduced by the fault slip, but the top layer of the upper plate farther from the trench remains compressed, as indicated by large negative $\sigma_{xx}$ values ($< -0.10$ MPa; Figure 7-8B). Another effect of co-seismic slip is to load the subjacent region of the upper plate, which is apparent from $\sigma_{xz}$ and $\sigma_{zz}$ components and shear stresses resolved onto the megathrust geometry $>0.01$ MPa that extend nearly 300 km down-dip from the edge of the fault. Therefore, the state of the subduction system following a large megathrust earthquake is that of (a) an elastically relaxed near-trench region, (b) an elastically compressed top of the upper plate far from the trench, (c) an elastically loaded upper plate down-dip of the megathrust, and (d) an elastically relaxed upper plate immediately beneath the compressed region (Figure 7-9). Some of the loading down-dip of the rupture is released by slip on the plate boundary immediately following the earthquake, but this distribution of stresses likely persists and drives longer term post-seismic deformation.

On the time scales of the years following the earthquake, the subduction rheology reverts to the layered elastic on top of visco-elastic rheology. Under these material properties, the compressed part of the upper plate transfers its compressive elastic stresses to the underlying viscous material through continued extension and the loaded region of the upper plate relaxes these stresses through ductile creep, resulting in quasi-linear trenchward motion. This interpretation is consistent with the post-seismic displacements seen after the Tohoku earthquake (Figure 7-2C). At distances $>300$ km to the north and south of the rupture zone (along strike), the co-seismic stress changes are small and the motion after the earthquake reverts rapidly back to the inter-seismic pattern. Between these regions is a transition zone where surface velocities are intermediate between the trenchward post-seismic relaxation and arcward inter-seismic loading. Observations of displacements near the Tohoku trench show westward motion within a year of the earthquake, indicating that the trench relocks quickly after the earthquake (Sun et al., 2014). Therefore, as the relaxation of the upper plate far from the fault is occurring, the region near the locked zone is already accumulating slip deficit and elastic stresses. As the relaxation process decays and slip deficit accumulates, the system transitions from post-seismic back to the inter-seismic pattern of loading.

5. Conclusion

Elastic strain accumulation during the inter-seismic stage of the earthquake cycle and strain release during the co-seismic stage develop under different rheological conditions. In the
inter-seismic stage, the upper plate behaves like a layered elastic on visco-elastic material, whereas the entire system is elastic in the co-seismic stage. As a result, the spatial distribution of stress differs; inter-seismic loading is that of a compressed and stressed beam overlying an elastically relaxed layer, whereas co-seismic stresses are concentrated near the fault and distributed throughout the entire volume. These stresses do not balance each other, leaving residual elastic deformation that is resolved through post-seismic processes that occur under the rheologically layered boundary conditions.

Acknowledgements

Daily GPS position data were provided by the Geospatial Information Authority of Japan (http://www.gsi.go.jp/ENGLISH). High-rate GPS data were provided by the Geospatial Information Authority of Japan through Nippon GPS Data Services Company (http://www.gpsdata.co.jp). VERIPOS (http://www.veripos.com) provide the GPS satellite clock and orbit corrections. Processing was performed by RTNet software developed by GPS Solutions (http://www.gps-solutions.com) and Hitachi Zosen Co., Japan (http://ww.hitachizosen.co.jp/english/index.html)

References


Figure 7-1. Schematic representation of subduction zone material properties.
(A) At the rapid strain rates of the co-seismic stage, the entire system behaves elastically. In an elastic system, deformation is linearly proportional to stress. (B) During slow inter-seismic loading, the cooler top regions of the lithosphere and the subducting plate deform elastically. The warmer regions below deform through creep mechanisms, relaxing stresses. As a result, the shallow regions are deformed and stressed while the deeper regions are deformed but unstressed.
Figure 7-2. Surface motion of Japan before, during, and after the 2011 Mw 9.0 Tohoku earthquake.

(A) From 2005 to 2011, the island of Honshu was shortening and moving westward at 2-6 cm/yr as a result of being coupled to the subducting Pacific plate, which moves at 8.3 cm/yr towards Japan. A transect across the island shows that the magnitude of westward motion increases approximately linearly from west to east. (B) Co-seismic displacements immediately following the Tohoku earthquake. The spatial pattern of strain release is not the inverse of the loading pattern; the displacements increase non-linearly from west to east. (C) Post-seismic velocities 2 years after the earthquake show continuing eastward motion at ~8 cm/yr that are quasi-uniform across the island.
Figure 7-3. Finite element model setup of inter-seismic loading in a subduction zone.

The upper plate is 50 km thick, and its bottom boundary conditions allow free sliding with spring-like restrictions on the vertical motions to simulate the stress relaxation effects of a viscous material. The subducting plate moves down-dip and is coupled to the upper plate from 10 to 50 km depth.
Figure 7-4. Cross-sectional stresses through the center of the rheologically layered finite element model.

(A) Effective shear stress is nearly uniform through the upper plate, with an value of ~0.15 MPa. The horizontal displacements are linear throughout the upper plate (blue solid line) and it also shows some vertical flexure (red solid line). In contrast, a model with a fully elastic upper plate has distinctly different surface displacements (dashed lines). (B-E) Non-zero components of the stress tensor. The most significant of these is $\sigma_{xx}$ (B), which shows the upper plate being uniformly shortened. (F) Normal and (G) shear stresses resolved onto the megathrust geometry.
Figure 7-5. Cross-sectional stresses through the center of a finite element model with a fully elastic upper plate.

Panels correspond to the same stress components as in Figure 7-4. The primary difference is that stresses can extend through the base of the upper plate instead of being relaxed through viscous flow.
Figure 7-6. Fit between observed and synthetic static displacements.

Finite fault model sub-faults are colored by slip value. (A) The black arrows are observed static displacements. Red arrows are displacements predicted by putting the rupture model in an elastic half-space. (B) Residual displacements at each GPS station. There are systematic misfits northern Honshu, but the fits are reasonably good elsewhere.
Figure 7-7. Cross-sectional stresses through the center of a fault with 1 meter of slip in an elastic half-space.

Panels correspond to the same stress components as in Figures 7-4 and 7-5. These stresses are concentrated dominantly near the region of the rupturing fault.
Figure 7-8. Sum of co-seismic and inter-seismic stresses.

Panels correspond to the same stress components as in Figures 7-4, 7-5, and 7-12. The key stress components are (B) $\sigma_{xx}$, which shows that the top of the upper plate remains compressed after the earthquake, (C) $\sigma_{xz}$, and (E) $\sigma_{zz}$, which highlight the loading down-dip of the fault into the base of the upper plate.
Figure 7-9. Schematic representation of stresses immediately after the earthquake.

Dark colors indicate high levels of elastic stress, and light colors indicate lower levels of elastic stress. The top of the upper plate is relaxed in the area near the fault while the regions farther away are still compressed. The subjacent layer is loaded by fault slip down-dip of the fault while the region farther away remains relaxed.
Chapter 8

Related Work

1. Introduction

As part of my doctoral research, I have been involved as a co-author on seismo-tectonic studies related to the overarching themes of my dissertation in which I provide analytical, modeling, and interpretation expertise. These studies reveal a variety of deformation processes acting before, during, and after earthquakes that are closely tied to the results discussed in previous Chapters. In particular, these events and studies contribute to the breadth of observations that constrain (a) the detailed rupture processes of earthquakes, (b) interactions between faults via stress transfer, and (c) the role aseismic slip has leading up to and immediately following the main event. In this Chapter, I highlight my specific contributions to the science, the key interpretations from these studies, and how they are related to the goals of my dissertation.

In Section 2, I describe the analysis of the 2013 Santa Cruz Islands earthquake sequence, in which stress transfer calculations were used to infer slow slip occurring simultaneously with the earthquake. This work applies the elastic half-space stress transfer tools and methodology also utilized in Chapters 2, 3, and 6. The same tools were applied to analyze the 2014 Iquique earthquake sequence described in Section 3. Observations from this study of a sequence of earthquakes leading up to the Iquique mainshock led to the more detailed analysis of how these events triggered the larger rupture in Chapter 2. Finally, in Section 4, I describe how stress transfer can be used to infer rupture complexity and estimate future seismic hazards in a non-plate boundary setting, similar to Chapter 3.

2. 2013 Santa Cruz Islands Earthquake Sequence

The role of aseismic slip in the loading and unloading of fault systems is poorly understood primarily because it may produce subtle variations in the observable deformation. However, as part of the slip deficit budget, knowing the amount of aseismic slip is important in evaluating the stress and strain state of the subduction system, balancing the slip budget on the interface, and mapping the distribution of slip deficit available in future earthquakes (e.g., Chapters 4 and 5). In the absence of direct evidence for aseismic slip (e.g., surface displacements
that are not explained by earthquakes; Chapter 2), slow slip may be identifiable from the effect it has on the surrounding stress field. One example of interpreting slow slip based on the effect it produced on the nearby stress field comes from the 2014 Santa Cruz Islands earthquake sequence (Hayes et al., 2014a). On 6 February 2014, a Mw 8.0 earthquake occurred at the northern end of the Vanuatu subduction zone, where the Australia plate subducts eastward under the Pacific plate (Figure 8-1). Although the mainshock characteristics were consistent with it rupturing the subduction megathrust, its aftershock sequence had very few megathrust events. In contrast, the aftershock sequence contained numerous shallow strike-slip and normal aftershocks.

As a component of the analysis of the deformation throughout this seismic sequence, we computed Coulomb failure stress changes (ΔCFS) generated by the mainshock and resolved on the aftershocks. A common expectation for aftershocks is that the majority of them will occur in response to positive ΔCFS and negative ΔCFS will inhibit aftershock activity (e.g., Chapters 2, 3, and 6); however, a spatially coherent, large number of the aftershocks in the Santa Cruz sequence were in regions that were negatively loaded by the mainshock (Figure 8-2A). In particular, the Santa Cruz earthquake had 4 large strike-slip aftershocks (M6.7, M6.8, M7.0, M7.1). The first aftershock (northeast of the mainshock rupture zone) had positive ΔCFS resolved onto its nodal planes, consistent with being triggered by the mainshock rupture. The other three, south of this first aftershock and east of the mainshock rupture zone, have negative ΔCFS resolved onto their nodal planes.

This stress transfer calculation implies that the mainshock stress changes alone were not responsible for triggering these strike-slip aftershocks, and in fact we would expect left lateral strike-slip motion in these locations. One mechanism for positively loading these aftershocks is by adding a region of slip on the megathrust to the south of the main rupture zone. Because such a slip patch is not apparent from seismic observations, this proposed deformation would need to occur relatively slowly; Lay et al. (2013) suggest that at least a portion of it may have occurred rapidly enough (in the minutes following the mainshock) to generate a tsunami. In order to positively load the crustal strike-slip aftershocks, the slow slip patch has to include most of the seismogenic width, spanning a horizontal distance of nearly 100 km. Its along-strike length is constrained to fit between the large-slip extent of the rupture to the north and the 1966 earthquake to the south. Without constraints on the amount of deformation, the average slip/area ratio is assumed to be similar to the seismic region to the north. Thus, this rupture patch was computed to have a magnitude equivalent of Mw 7.6.
By adding slow slip to the megathrust south of the seismic rupture patch, nearly all of the aftershocks are now in regions that are positively loaded (Figure 8-2B). The slip from both the seismic and aseismic regions appears to be relatively homogeneous, which accounts for the lack of megathrust aftershock activity. Although lack of direct evidence, e.g., local GPS displacements, makes it difficult to confirm the existence of such a slow slip patch as the source of the aftershock loading, it is difficult to account for the distribution of aftershocks without the stress added by the slow slip.

3. 2014 Iquique Earthquake Sequence

After a large earthquake, it is usually assumed that the regional seismic hazard has decreased because the fault has released a significant amount of its accumulated strain energy. Although this is largely true immediately adjacent to the section that slipped, the ruptured region may be only a fraction of the total locked area (e.g., Chapter 5). In these cases, it is important to assess the remaining level of residual seismic hazard and how the earthquake has affected the potential for additional events in the region. The 1 April 2014 Mw 8.2 earthquake that occurred on the subduction megathrust offshore of Iquique, Chile, appears to be such an event (Hayes et al., 2014b). In 1877, a much larger Mw 8.8 earthquake occurred in the same region of the megathrust (Figure 8-3A). The 2014 event appears to have only ruptured a small part of the subduction interface that slipped in 1877. Therefore, slip deficit and corresponding elastic strain still remain in the regions of the megathrust around the 2014 Iquique rupture zone, posing a future seismic hazard. Slip in the 2014 event also caused additional loading on the surrounding regions, which may promote the rupture of these regions.

One way to test whether stress changes from an earthquake promote future seismicity is to compute Coulomb failure stress changes (ΔCFS) resolved on the subsequent events. If the later events are generally positively loaded, they can be interpreted as being triggered by the preceding earthquakes. I computed the ΔCFS throughout the 2014 Iquique earthquake sequence, which not only demonstrated that a large earthquake triggers aftershocks in a ΔCFS sense, but also that smaller earthquakes (foreshocks) can help trigger the mainshock (Chapter 2; Figure 8-3B). Following the 2014 Iquique earthquake, there were 138 aftershocks from April to August 2014, including the largest aftershock, a Mw 7.7 event on 3 April 2014. These events appear to be responding dominantly to stress changes from the mainshock (Figure 8-3C):

The hypocentral region of the M 7.7 aftershock was loaded 0.25 MPa by the mainshock and the first 27 hours of aftershocks. Aftershocks have generally
nucleated in areas of increased ΔCFS, surrounding the main slip patches of the largest events. Overall, the hypocentres of ~70% of relocated aftershocks (94 of 138 events) occurred in areas of positive ΔCFS. If uncertainties in relocated hypocentres (±2–3 km) are considered, more than 80% of aftershocks occurred in regions of positive ΔCFS (Hayes et al., 2014b).

The positive stress loading region generated by the mainshock and aftershocks extends ~100 km farther along-strike than the edge of the rupture and aftershock zone, into regions that have not slipped since 1877. The Nazca-South America relative plate velocity at this location is 70 mm/yr (DeMets et al., 2010), implying nearly 10 m of slip deficit have accumulated since the last great earthquake. No large earthquakes have occurred in these regions since 1877, indicating that most of this slip deficit potentially still exists to the north and south of the Iquique rupture zone. The ΔCFS computations demonstrate that slip from the Iquique earthquake has brought these regions closer to failure. Although we cannot use these results to estimate the timing of potential great earthquakes in the remaining Iquique gap, these areas should be considered major seismic hazards still capable of rupturing in up to Mw 8.5 earthquakes.

4. 2014 Chiang Rai Earthquake Sequence

Although moderate magnitude (Mw ~6.0) earthquakes only account for a small fraction of the global seismic moment budget, they can generate severe local and regional shaking and secondary effects such as liquefaction and landslides. In addition, because these smaller magnitude earthquakes rupture relatively short faults (<25 km long), they can occur on structures other than the larger and better known hazardous faults corresponding to plate boundaries. Despite their small dimensions, moderate sized earthquakes may exhibit complex spatial rupture and aftershock distributions more typically associated with larger events. Quantifying the rupture complexity through analyses of the mainshock and aftershock source properties and understanding the sources of such rupture complexity are important components of constraining the hazards related to the immediate aftershock sequence and triggered events that may rupture long after the immediate seismicity has decayed (e.g. Herman et al., 2014; Herman et al., 2016; Chapter 3). The 5 May 2014 Mw 6.2 Mae Lao, Thailand, earthquake and its aftershock sequence in the Chiang Rai province is an example of a complex rupture in a moderate magnitude earthquake (Pananont et al., 2017). I provided many of the analyses and corresponding figures for this study.
We constrain the spatial pattern of deformation throughout the sequence by relocating the aftershocks in the sequence and I determined source parameters using regional seismic data for the largest events (Mw ~4.3+; Figure 8-4). The aftershocks relocate onto two vertically dipping planes: a north-striking plane whose northern end abuts against an east-striking plane. The geometries of the aftershock regions match the nodal planes of the mainshock and most aftershocks; these events are strike-slip with either right lateral slip on the north-striking plane or left lateral slip on the east-striking plane. Because the mainshock relocates into the center of the north-striking cluster of aftershocks, we interpret the right lateral kinematics as the mainshock fault plane. The majority of aftershocks and the largest events are dominantly located in the east-striking cluster of events, suggesting they are left lateral strike-slip.

I also computed Coulomb failure stress changes (ΔCFS) generated by the mainshock and resolved onto the location and left lateral geometry of the east-striking plane, which puts many of the aftershocks there into a negatively loaded state that should inhibit seismicity (Figure 8-5a). However, significant seismic activity on the east-striking left lateral plane began within an hour after the mainshock, suggesting some triggering event. To account for these aftershocks, we hypothesize that the mainshock rupture “leaked” onto the east-striking plane, generating co-seismic left lateral offset. If there is a relatively small amount of left lateral slip (Mw ~5.7), the center of the east-striking fault becomes a region of positive ΔCFS (Figure 8-5b). Such a sub-event would account for the immediate left lateral activity northwest of the mainshock epicenter.

The ΔCFS extends ~25 km beyond the source region of the earthquake. There are no mapped active faults in the ΔCFS footprint, however, faults in this region are typically inferred by their surface expression; if there are faults oriented correctly for slip in the subsurface, these may be able to host future earthquakes in the region and may have been brought closer to failure by the Chiang Rai sequence.

References


Figure 8-1. Tectonic setting of the 6 February 2013 Mw 8.0 Santa Cruz Islands earthquake.

The location of the map is highlighted by a red box on the inset global map. The direction of relative motion between Australia and Pacific plates is indicated by the vector. The earthquake has a thrust faulting mechanism (U.S. Geological Survey W-phase solution), consistent with rupture of the northern end of the Vanuatu subduction zone.
Figure 8-2. Coulomb failure stress changes ($\Delta$CFS) resolved on aftershock focal mechanisms, modified from Hayes et al. (2014a).

The color shown corresponds to the most positive value resolved on on the two nodal planes, with warm colors indicating stresses that promote slip and blue indicating stresses that inhibit slip. (A) When the mainshock alone generates the stress changes, a large number of aftershocks to the east have negative $\Delta$CFS. (B) When a slow slip patch is added to the southeast of the mainshock, these events become positively loaded.
Figure 8-3. Regional tectonic setting and stress changes for the 2014 Iquique earthquake.

(A) The 2014 Mw 8.3 Iquique earthquake (solid line shows the 1 meter slip contour) occurred in the middle of the region inferred to have hosted the 1877 Mw 8.8 earthquake (dashed line). Coulomb failure stress changes (ΔCFS) generated by (B) foreshocks and (C) the mainshock and largest aftershock resolved onto the subduction interface. Yellow to red colors indicate positive ΔCFS and blue colors indicate negative ΔCFS, with both saturating at 0.1 MPa. Dark foreshocks and aftershocks have positive ΔCFS resolved onto their east-dipping nodal plane and light events have negative ΔCFS.
Figure 8-4. Regional moment tensor solutions for events in the Chiang Rai sequence.

Earthquakes are plotted at their relocated positions and scaled by magnitude, indicated above each event. The aftershocks form two vertical planes, highlighted in red dashed lines. The mechanisms are dominantly strike-slip, as indicated by the ternary diagram in the bottom right.
Figure 8-5. Coulomb stress change caused by the mainshock resolved onto east-striking left lateral strike-slip faults.

(A) If the mainshock is interpreted as just a right lateral strike-slip rupture on a north-striking plane, most of the aftershocks are in a region of negative $\Delta CS$. (B) In contrast, if mainshock slip also occurs on the left lateral plane (equivalent to Mw 5.7) then the aftershocks become positively loaded.
Chapter 9
Conclusion

The studies presented in this dissertation demonstrate the significant variability in deformation observed throughout the earthquake cycle, particularly in subduction zones. They also reveal underlying processes that are common to any seismo-tectonic setting. The transfer of stress from a slipping fault to surrounding region is shown to trigger subsequent earthquakes on nearby structures. Although these stress changes cannot be used to predict the precise timing events, they can be applied to anticipate which faults are more likely to slip or identify the regions that may be more likely to host future earthquakes. In the inter-seismic stage, stress and deformation around a locked region on a plate boundary structure also affect how adjacent parts of the interface move and accumulate elastic strain. In particular, the amount of slip that is observed in an earthquake may be reduced by adjacent locked structures, resulting in a larger remaining slip deficit and associated seismic hazard. These results highlight the importance of treating plate boundary zones as interacting fault systems rather than individual structures with characteristic dimensions, slip, and recurrence intervals.

These studies also identify several important rheological properties of subduction zones that are essential for understanding the patterns of strain accumulation and release. A relatively simple description of the interface as locked or unlocked helps understand multiple geophysical observations, including partial slip deficit accumulation on the megathrust, slip reduction in earthquakes, and low seismic radiation in tsunamigenic regions of the megathrust. The rheological variations throughout the lithospheric volume also have a critical role in the observed deformation. At low strain rates, the pattern of deformation in a visco-elastic upper plate are significantly different than the elastic-only deformation associated with the high strain rates of the co-seismic rupture. Slow post-seismic observations help constrain the remaining deformation in the system and are consistent with these rheological changes.

Future work related to this thesis involves continued exploration of rheological and frictional complexity in subduction zones through 3-D finite element modeling. This includes models with more complex visco-elastic rheology that may more accurately reflect the thermal state of the upper plate, mechanical heterogeneity in the upper plate reflecting along-strike
geological variations, and how seismogenic zones with different recurrence intervals interact throughout the earthquake cycle. From an observational perspective, this research also demonstrates the need for improved monitoring networks. In particular, deploying sensors offshore near subduction trenches will be crucial for further constraining how the shallow regions of the megathrust deform. Developing more widespread networks of instruments capable of recording from short periods to infinite period, like the high-rate GPS network in Japan, will also be essential for recording deformation at each stage of the earthquake cycle. Through these observational and modeling developments, we will be able to better constrain the seismotectonic processes that occur in subduction zones and better anticipate the hazards associated with great megathrust earthquakes.
Appendix A

Supporting Information for “Foreshock Triggering of the 1 April 2014 Mw 8.2 Iquique, Chile, Earthquake”

Figure A-1. Evolution of stress changes throughout foreshock sequence.

Evolution through time of the Coulomb failure stress change (ΔCFS) resolved onto the megathrust interface (assuming a rake of 90°) and onto the fault plane of each event (with the rake given by the focal mechanism). Focal mechanisms are colored by the sign of the ΔCFS resolved onto the fault plane, with positive colored red and negative blue. This figure is identical to Figure 2-3 in the main text, except that all of the events with thrust faulting mechanisms have their depths shifted so that they lie on the megathrust interface. Note that this changes the amplitude of the ΔCFS, but not the general pattern.
Appendix B

Supporting Information for “Revisiting the Canterbury Earthquake Sequence After the 14 February 2016 Mw 5.7 Event”

Canterbury Plains Seismicity

One definition of the end of the Canterbury sequence is when the rate of seismicity has returned to background levels. To visually establish this background rate and compare it with the recent activity, we plot seismicity in the Canterbury Plains (within the mapped region shown in Figure 3-1a) from January 2000 through April 2016 (Figure B-1). The seismicity data come from the New Zealand GeoNet catalog, and can be accessed through their Quake Search tool (http://quakesearch.geonet.org.nz).

Figure B-1. Seismicity in the Canterbury Plains over time.

(A) Local magnitude versus time (similar to Figure 3-1b). There was some seismicity prior to September 2010, but the rate was low (~0.2 ML 2.0+ events per day), and there were only two events larger than ML 4.5. Seismicity from 2013 to 2015 was similar in that there were few large earthquakes, but the rate was still significantly higher than the pre-Darfield rate (~1.8 ML 2.0+ events per day), consistent with the continuation of the aftershock sequence. (B) Seismicity in the Canterbury Plains from 2010 to present. The seismicity in the Canterbury Plains from 2010 to the present is dominated by the aftershock activity immediately following the main events in the Canterbury sequence (the 2010 Mw 7.0 Darfield and 2011 Mw ~6.0 Christchurch earthquakes). The vigorous activity in the immediate aftermath of the main events, the relatively diminished
seismic activity from 2013 to 2015, and the pre-Darfield background seismicity rate span several orders of magnitude, so here we plot the number of events per day on a logarithmic scale. On this scale, the differences in activity during these time periods are clearly distinguishable. We also show the total seismic moment release in the Canterbury Plains from 2000 onward, including the Mw 7.0 Darfield mainshock.

**Coulomb Stress Change Sensitivity Analysis**

The Coulomb stress change (ΔCS) is a measure of whether a fault has been brought closer to or farther from shear failure, based on the Mohr-Coulomb failure criterion (Reasenberg and Simpson, 1992). It is a function of several parameters, whose uncertainties propagate to uncertainties in the ΔCS distribution. We perform a sensitivity test to assess this ΔCS uncertainty, varying the parameters that have the largest impact on the ΔCS distribution. The input parameters for the ΔCS computation are listed below, and the specific parameters tested are indicated in parentheses.

**ΔCS Input Parameters**

- Earthquake locations and source parameters. As discussed in the manuscript, these are derived using seismological analyses, and the events are converted to rectangular faults with uniform slip through empirical relations between magnitude and fault area. (The dominant source uncertainty is the 2010 Darfield mainshock slip distribution; we test the Hayes (2010) and Beavan et al. (2012) models).
- Material properties of the host medium. We model the material as an elastic half-space with shear modulus 40 GPa and Poisson’s ratio 0.25. Variations in these values have a second-order effect on the ΔCS distribution, and are not considered in the sensitivity analysis.
- Target fault locations and kinematics. (There is significant uncertainty in the target fault kinematics, but we have some constraints on the three general fault types from the earthquake seismicity and seismic reflection surveys. We show ΔCS distributions for E-W right lateral strike-slip faults, N-S left lateral strike-slip faults, and NE-SW reverse faults. The depth of the target faults also has a significant impact on the ΔCS analysis, and we test depths of 5, 10, and 15 km).
- Effective coefficient of friction on the fault. This varies due to rock type and pore fluid pressure. (We test coefficient of friction values 0.2, 0.5, and 0.8).
Figure B-2. Results of sensitivity test on the ∆CS distribution.

To summarize, we explore variations in friction value (0.2, 0.5, and 0.8; the columns of each figure), ∆CS computation depth (5, 10, and 15 km; rows), mainshock slip distribution (Hayes, 2010; Beavan et al., 2012; pages), and target fault kinematics (A: right lateral strike-slip; B: left lateral strike-slip; C: reverse). Although substantial variability in the ∆CS distribution is evident from these plots, the general pattern of positive and negative lobes is fairly similar for each target fault type. It is also apparent that the ∆CS distribution resolved on the two strike-slip orientations (i.e. conjugate strike-slip faults) is essentially the same.
Darfield FFM: Beavan et al. (2012)

Target Faults

Strike: 80°
Dip: 90°
Rake: 180°
Darfield FFM: Hayes (2010)

Target Faults
- Strike: 80°
- Dip: 90°
- Rake: 180°
Darfield FFM: Beavan et al. (2012)

**Target Faults**
- Strike: 150°
- Dip: 90°
- Rake: 0°
Darfield FFM: Beavan et al. (2012)

Target Faults

Strike: 40°
Dip: 45°
Rake: 90°
Darfield FFM: Hayes (2010)

Target Faults

Strike: 40°
Dip: 45°
Rake: 90°
Appendix C

Supporting Information for “Modeling Slip Deficit Accumulation Around Locked Regions on the Subduction Megathrust”

Derivation of Effective Shear Stress

Consider a three-dimensional stress tensor, $\sigma$:

$$
\sigma = \begin{bmatrix}
\sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\
\sigma_{xy} & \sigma_{yy} & \sigma_{yz} \\
\sigma_{xz} & \sigma_{yz} & \sigma_{zz}
\end{bmatrix}
$$

We are primarily interested in the components of the stress tensor that drive geological deformation, so we subtract the isotropic component from the total stress tensor, resulting in the deviatoric stress tensor, $\sigma'$:

$$
\sigma' = \begin{bmatrix}
\sigma_{xx} - I & \sigma_{xy} & \sigma_{xz} \\
\sigma_{xy} & \sigma_{yy} - I & \sigma_{yz} \\
\sigma_{xz} & \sigma_{yz} & \sigma_{zz} - I
\end{bmatrix}
$$

where

$$
I = \frac{1}{3} (\sigma_{xx} + \sigma_{yy} + \sigma_{zz})
$$

The invariant quantities of the deviatoric stress tensor reflect a total amount of shearing deformation. These invariants can be computed through eigenanalysis of the deviatoric stress tensor:

$$
0 = \det \begin{bmatrix}
\sigma'_{xx} - \lambda & \sigma_{xy} & \sigma_{xz} \\
\sigma_{xy} & \sigma'_{yy} - \lambda & \sigma_{yz} \\
\sigma_{xz} & \sigma_{yz} & \sigma'_{zz} - \lambda
\end{bmatrix}
$$

and

$$
0 = (\sigma'_{xx} - \lambda) \left( (\sigma'_{yy} - \lambda) (\sigma'_{zz} - \lambda) - \sigma'_{yz}^2 \right)
$$

Rearranging the terms to group powers of $\lambda$:

$$
0 = \lambda^3
$$

This characteristic equation is unique for this stress tensor, irrespective of the coordinate system used. Therefore, the coefficients of $\lambda$ must also be identical in any coordinate system, i.e., they are invariant. By construction of the deviatoric stress tensor, the coefficient of $\lambda^3$ (the first invariant) is zero. The coefficient of $\lambda^1$ is the second invariant, and we take the square root of the absolute value to yield a quantity with units of stress. This is the effective shear stress:

$$
\sigma_{eff} = \sqrt{\sigma'_{xx} \sigma'_{yy} + \sigma'_{xx} \sigma'_{zz} + \sigma'_{yy} \sigma'_{zz} - \sigma'_{xy}^2 - \sigma'_{yz}^2 - \sigma'_{zx}^2}
$$

In terms of the non-deviatoric stress tensor, and rearranging the terms, we get the equation in Chapter 4:

$$
\sigma_{eff} = \sqrt{\frac{1}{3} \left( \sigma_{xx}^2 + \sigma_{yy}^2 + \sigma_{zz}^2 - \sigma_{xx} \sigma_{yy} - \sigma_{xx} \sigma_{zz} - \sigma_{yy} \sigma_{zz} \right) + \sigma_{xx}^2 + \sigma_{zz}^2 + \sigma_{yy}^2}
$$

181
Appendix D

Supporting Information for “Integrated Geophysical Characteristics of the 2015 Illapel, Chile, Earthquake”

Introduction

In this Supporting Information, we provide details and results for the geophysical analyses performed in this study. The double source moment tensor inversion constraints are described and we show the resulting solutions in Figure C-1. Our final seismic finite fault model was chosen from a suite of model parameterizations, and we show how several representative models fit GPS displacements (Figure C-2). In Figure C-3, we provide observed, modeled, and misfits for InSAR interferograms and GPS displacements generated by our geodetic finite fault model. Finally, we show the results of using different finite fault model solutions on the Coulomb failure stress change distribution (Figure C-4).

Double Source Moment Tensor Inversion

We test the sensitivity of the double source moment tensor inversion to the choice of constraints on the source parameters of the first sub-event. These constraints are informed by the seismological finite fault model and the slab geometry. The unconstrained solution results in an initial Mw 7.6 thrust faulting sub-event (strike=227°, dip=3°, rake=123°) at a depth of 40.5 km followed by a Mw 8.2 thrust faulting sub-event (strike=358°, dip=22°, rake=91°) at a depth of 23.5 km (Figure C-1a). In order to better match the subduction interface geometry, we constrain the first event to have a dip of 20.3°; this produces a double source solution with an initial Mw 7.3 thrust faulting sub-event (strike=170.1°, dip=20.3°, rake=65.2°) at a depth of 40.5 km and a second Mw 8.2 thrust faulting sub-event (strike=358.0°, dip=22.3°, rake=91.2°) at a depth of 23.5 km (Figure C-1b). Finally, the source-time function determined in the finite fault solution has a small peak at ~20 s, so we fix the time delay and half-duration of the smaller sub-event to 20 s, in addition to fixing the dip at 20.3°, for our final constrained inversion. This results in a Mw 7.2 first sub-event (strike=350.6°, dip=20.3°, rake=106.2°) at 40.5 km and a Mw 8.2 second sub-event (strike=357.6°, dip=22.9°, rake=87.5°) at 23.5 km depth (Figure C-1c).

In order to determine whether a single source W-phase inversion or a double source W-phase inversion produces a more realistic model, an Akaike Information Criterion (AIC) test is performed. The AIC test evaluates the relative fit of each inversion by considering the model residuals and the degrees of freedom. For both the single and double source W-phase inversion, the degrees of freedom for the model are based on the moment tensor being used. In the case of the Illapel earthquake, deviatoric moment tensors are used, so the single source solution has five degrees of freedom, whereas the double source solution has 10 degrees of freedom. Since only two models are being compared, the difference in each model’s AIC value can be used to assess the most appropriate model for the Illapel earthquake. The difference in AIC values is given by

\[ \Delta \text{AIC} = N \ln(\text{SS}_2/\text{SS}_1) + 2 \Delta df \]
where $N$ is the number of data points, $\Delta df$ is the difference in degrees of freedom between the two models, and $SS_i$ is the sum-of-squares error between synthetic and observed waveforms, with $i=1$ representing the single source model and $i=2$ representing the double source model. A negative $\Delta AIC$ value indicates that the more complex model (the double source inversion) is preferred and a positive $\Delta AIC$ value indicates that the simpler model (the single source inversion) is still the preferred solution; in this case, the $\Delta AIC$ value is negative.

**Seismic Finite Fault Model**

We generated numerous seismic finite fault models using different Green’s functions, fault parameterizations, and rupture constraints. Specifically, we determined a suite of FFM inversions with the following variations:

- Green’s functions from Crust 2.0 (Bassin et al., 2000) on top of PREM (Dziewonski and Anderson, 1981) versus Green’s functions generated from multiple 1-D profiles through a regional velocity model (Contreras-Reyes et al., 2015)
- Green’s functions from velocity models with and without top water layer
- Including velocity (in addition to displacement) waveforms in the inversion
- Fault dips ranging from 15º to 19º
- Minimum rupture velocities ranging from 0.4 km/s to 0.8 km/s
- Unconstrained total moment versus total moment weighted to match W-phase solution
- Unaligned versus aligned observed and synthetic waveforms

Although the FFM results generated using different Green’s functions, fault parameterizations, and rupture constraints produce fairly similar fits to waveforms, they vary significantly in the distribution and magnitude of slip. Six models representative of this variability are shown in Figure C-4. This variability in rupture characteristics maps into differences in forward-modeled static displacements at onshore GPS stations. We compare synthetic and observed displacements to select a preferred result from the suite of FFMs, considering a balance of fitting observed teleseismic waveforms (through the inversion procedure described in Section 2.2), fitting both horizontal and vertical GPS displacements (through visual inspection of the observed and synthetic displacement vectors; Figure C-4).

**Geodetic Finite Fault Modeling**

Our geodetic finite fault model is derived by inverting co-seismic GPS displacements and InSAR interferograms. The data sources and details of the analytical techniques can be found in Barnhart et al. (2016) and references therein. In Figure C-5, we show the interferogram and GPS fits for the model determined in this study, which has a slightly different fault geometry from the model in Barnhart et al. (2016) (this study: strike=6º, dip=17º).

**Coulomb Stress Dependence on FFM**

The four mainshock slip models discussed in this study produce broadly similar Coulomb stress change distributions on the plate boundary; however, the number and location of aftershocks positively loaded by the models varies significantly (Table D-1; Figure D-6). We graphically
present the results from Table D-1 in Figure D-6 to demonstrate the geographical variability in ΔCS distribution between the four models.
Figure D-1. Double source moment tensor inversion solutions applying different constraints (A) unconstrained solution; (B) first sub-event dip constrained to 20.3°; (C) first sub-event dip constrained to 20.3°, centroid time and half duration constrained to 20 s. The solution shown in (C) is preferred since the first sub-event is consistent with rupturing the megathrust. The source parameters of the second, larger sub-event are similar in each inversion, suggesting they are robust with respect to the choice of constraints.
Figure D-2. Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b.

The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-2 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b. The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-2 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b. The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-2 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-b. The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-2 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b. The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-2 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by a single source W-phase solution, similar to Figure 6-2b. The station location is shown in the map adjacent to the time series plot and the component is indicated in the station name. The W-phase fit is based on the part of the observed signal between the two red dots in each time series.
Figure D-3. Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b.

Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-3 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b. Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-3 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b. Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-3 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b. Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-3 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b. Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-3 (continued). Comparison of observed (black) and synthetic (red) seismograms produced by the preferred constrained double source W-phase solution, similar to Figure 6-2b. Notice the improvement in fit, especially time alignment, as compared to the waveforms shown in Figure D-2.
Figure D-4. Comparison of predicted and observed GPS displacements from a selection of six representative seismic finite fault model inversions.

Observed displacements are black and predicted displacements are red. At the top of each panel is the root-mean-squared misfit between observed and synthetic displacements for the component shown in the panel. (A) Preferred model used in this study. (B) Increasing the fault dip to 19º leads to shallow slip located ~20 km east of the trench. (C) Best model using PREM + Crust 2.0 velocity structure (no water layer) (D) When the magnitude of the model is unconstrained, the FFM-derived magnitude (Mw 8.4-8.5) tends to be significantly larger than the moment tensor magnitudes (Mw 8.2-8.3). (E) Best model using PREM + Crust 2.0 + water layer velocity structure (F) Lowering the minimum rupture velocity to 0.4 km/s brings slip closer to the epicenter.
Figure D-5. Fits to co-seismic InSAR interferograms and GPS displacements generated by the geodetic slip model from this study.

The two rows of interferograms represent results from two different image pairs. The image dates are shown to the left of the row. The left column shows the unwrapped interferograms, the center column is the modeled displacements, and the right column is the observed-predicted residual. At the bottom is the observed (black) and predicted (red) horizontal and vertical GPS displacements, also shown on Figure 6-4.
Figure D-6. Coulomb failure stress changes from four Illapel earthquake slip models.

Aftershocks are colored by the mean $\Delta$CFS value resolved on their interpreted fault planes determined by perturbing the aftershock locations and source parameters. (A) Seismic slip model from this study (identical to Figure 6-8): 54% positively loaded. (B) Geodetic model from this study: 54% positively loaded. (C) Li et al. (2016) model: 46% positively loaded. (D) Melgar et al. (2016) model: 55% positively loaded.
<table>
<thead>
<tr>
<th>Vp (km/s)</th>
<th>Vs (km/s)</th>
<th>Density (g/m$^3$)</th>
<th>Thickness (km)</th>
<th>P Attenuation</th>
<th>S Attenuation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.50</td>
<td>0.00</td>
<td>1.02</td>
<td>2.7</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>3.81</td>
<td>1.94</td>
<td>0.92</td>
<td>0.00</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>2.50</td>
<td>1.07</td>
<td>2.11</td>
<td>0.10</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>6.00</td>
<td>3.50</td>
<td>2.72</td>
<td>10.44</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>6.60</td>
<td>3.80</td>
<td>2.86</td>
<td>10.44</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>7.20</td>
<td>4.10</td>
<td>3.03</td>
<td>10.76</td>
<td>1000</td>
<td>500</td>
</tr>
<tr>
<td>8.08</td>
<td>4.47</td>
<td>3.38</td>
<td>196.00</td>
<td>1200</td>
<td>500</td>
</tr>
<tr>
<td>8.59</td>
<td>4.66</td>
<td>3.45</td>
<td>36.00</td>
<td>360</td>
<td>140</td>
</tr>
<tr>
<td>8.73</td>
<td>4.71</td>
<td>3.49</td>
<td>108.00</td>
<td>370</td>
<td>140</td>
</tr>
</tbody>
</table>

Table D-1. 1-dimensional velocity model used in finite fault model inversion (Section 2.2).

<table>
<thead>
<tr>
<th>Mainshock FFM</th>
<th>Seismic FFM (this study)</th>
<th>Geodetic FFM (this study)</th>
<th>Li et al. (2016)</th>
<th>Melgar et al. (2016)</th>
</tr>
</thead>
<tbody>
<tr>
<td>∆CFS &gt; 0.01 MPa, stdev &lt; ∆CFS</td>
<td>75 (32%)</td>
<td>78 (33%)</td>
<td>70 (30%)</td>
<td>79 (34%)</td>
</tr>
<tr>
<td>∆CFS &gt; 0.01 MPa, stdev ≥ ∆CFS</td>
<td>50 (22%)</td>
<td>48 (21%)</td>
<td>37 (16%)</td>
<td>50 (21%)</td>
</tr>
<tr>
<td>∆CFS &lt; -0.01 MPa, stdev &lt;</td>
<td>58 (25%)</td>
<td>62 (27%)</td>
<td>84 (36%)</td>
<td>56 (24%)</td>
</tr>
<tr>
<td></td>
<td>∆CFS &lt; -0.01 MPa, stdev ≥</td>
<td>41 (17%)</td>
<td>34 (14%)</td>
<td>23 (10%)</td>
</tr>
<tr>
<td>∆CFS</td>
<td>≤ 0.01 MPa, stdev &lt; 1 x 10$^4$ Pa</td>
<td>5 (2%)</td>
<td>9 (4%)</td>
<td>13 (6%)</td>
</tr>
<tr>
<td>∆CFS</td>
<td>≤ 0.01 MPa, stdev ≥ 1 x 10$^4$ Pa</td>
<td>4 (2%)</td>
<td>2 (1%)</td>
<td>6 (2%)</td>
</tr>
</tbody>
</table>

Table D-2. Coulomb failure stress statistics resulting from randomly perturbing aftershock locations and source parameters within their uncertainty bounds.

We performed 1000 aftershock perturbations for each of four mainshock slip models. Events are categorized as positively/negatively loaded by their mean ∆CFS value, and further subdivided by the relative size of the standard deviation as compared to the mean ∆CFS value.
Appendix E

Exploring the Nature of Transient Waves Observed at High-Rate GPS Stations During the Tohoku Earthquake

Transient seismic waves are related to and may help facilitate the rebound of the upper plate as the finite period components of the complete elastic wavefield. However, there may be additional complications in the spatial and temporal patterns of rebound as these transient surface waves extend to different depths and thus interact with differently stressed volumes of the upper plate. Depending on the depth extent and particle motion associated with these waves, they may promote or inhibit the rebound. We characterize the elastic rebound deformation, including its velocity and static value, and explore whether the spatial pattern of rebound is affected by interactions of the seismic wavefield with the heterogeneously stressed upper plate.

D.1. High-rate GPS Dataset

GPS observations directly measure ground displacements and record both static (zero-frequency) and transient displacements. In contrast, standard seismometers measure velocity or acceleration and therefore reliably constrain only transient displacements without the zero-frequency component. In Japan, the GPS Earth Observation Network (GEONET) consists of over 1200 stations with an average spacing of ~20 km (Figure 7-2; Figure E-1A). Receiver position data are recorded at each station with a sampling rate of 1 second. With this high sampling rate, the GEONET stations can capture both the transient seismic waves and the final total offset from the earthquake. The surface displacements generated by the 2011 Mw 9.0 Tohoku earthquake were recorded by GEONET (Nishimura et al., 2011) and processed to retain a displacement time series period range of ~10 s to static (infinite) (Wright et al., 2012; http://www.gps-solutions.com/data_2011_tohoku_eq).

These time series reveal a spatially coherent westward propagation of the signal from the offshore slip region evolving towards the final offset (Figure E-1B). At stations closest to the Tohoku earthquake on the east coast of Japan, ground displacement amplitude first exceeds 10 cm ~50 seconds after the origin time (e.g., at station 0918, represented by the yellow curve in Figure E-1B). The peak displacement for station 0918 is ~5.0 meters eastward and ~1.8 meters to the south, which occurs ~120 seconds after the rupture begins. The final displacements are smaller than the peak values and are reached 30-60 seconds after the peak displacement occurs.
The seismic waves propagate westward; at stations on the west coast of Japan, 200 km from the east coast, the first significant displacements occur at ~80 seconds after origin time, the peak displacements occur at ~140 seconds, and the displacement amplitudes are significantly lower (e.g., station 0555, green curve in Figure E-1B). Next, we quantify the rate at which these displacement signals propagate westward across Honshu.

D.2. Modeling High-Rate GPS Records

Previous studies have mapped these high-rate GPS data to fault slip using full waveform inversion approaches that incorporate a zero-frequency component (e.g. Yue and Lay, 2011). In order to understand how the growth of static displacement is related to the deformation produced by fault slip, we first constrain the rate at which the signal propagates from the rupturing fault. We use the teleseismic-based (seismic stations 30°-90° away from the source) finite fault model (FFM) from Hayes (2011) as our input, deliberately selecting an independently derived rupture model that does not incorporate GPS data into the fit.

The approach we take to determining the propagation (“moveout”) velocity of the static displacement signal is simple; we assume that the displacements generated by a slipping sub-fault from the finite fault model propagate away from the sub-fault at a uniform velocity. At the time the signal reaches the position of the receiver, the corresponding static displacement appears as a step function in the time series (Figure E-2). These step function contributions to the static displacement time series are computed for the sub-fault station pair using the Okada (1992) solutions for fault slip in an elastic half-space. The moveout velocity is identical for all sub-faults and at every point in the model, and we run forward models with a range of velocities to find the propagation speed that best accounts for the observed evolution of the static offset.

The best average fits between synthetic and observed time series occur with a moveout velocity of 3.0 km/s (Figure E-3; Movies 1 and 2, available on Penn State ScholarSphere; https://scholarsphere.psu.edu). The best fit at some stations suggest a slightly faster moveout velocity of ~3.5 km/s, indicating either a change in velocity structure between the rupture area and the island of Honshu or a misfit in the precise location of slip in the rupture model. In none of the time series do we see significant static offset associated with faster P-wave velocities (~6.5 km/s). Rather, the static offsets evolve with a moveout velocity consistent with the group velocity of surface waves, particularly 10-50 s period Rayleigh waves (Oliver, 1962). We also determine the particle motion (relative to the earthquake epicenter) at each station by rotating the horizontal
components of the GPS signal from east-north to radial-transverse, where radial points away from the Tohoku epicenter and transverse is perpendicular to both the radial and vertical components (Figure E-4). The motions are generally retrograde elliptical, also consistent with Rayleigh waves, but the particle motion is complicated by the arrival of a large packet of seismic energy and rapid growth of total displacement generated by the high value of slip on the fault that occurred up-dip of the hypocenter. This package of seismic energy passes through Honshu ~30 s after the first static signal starts growing and is responsible for the bulk of the total displacement. After the large amplitude transient waves pass through, which takes 100-150 seconds for most stations on Honshu, all of the stations have an eastward (negative in the radial component) offset from their previous position. In other words, the upper plate has rebounded towards the trench during the time the deformation we interpret as Rayleigh waves passed through.

Assuming that the waves associated with the growth of the total displacement are Rayleigh waves, we assess which properties of these waves facilitate the observed deformation. As they propagate, Rayleigh waves displace a thickness of the Earth that is a function of the wavelength. The longer the period, the deeper in the Earth the particle motion extends (Figure E-5). For Rayleigh waves in the 10-50 s period range, the principal depth extent of the wave is 0-20 km for a 10 s period wave, 10-50 km for a 30 s period wave, and 20-90 km for a 50 s period wave. Therefore, the shorter period (10-30 s) waves that are propagating through the rebounding lithosphere only perturb the top, elastically strained volume, while the longer period (30-50 s) waves extend through both the top (stressed) section and the lower (relaxed) section.
Figure E-1. GEONET network and observed displacement time series.

(A) GEONET consists of stations spaced on average ~20 km apart throughout Japan recording at 1 Hz. (B) Displacement time series recording the 2011 Tohoku earthquake. Colors correspond to the station locations on Honshu. These show the first and largest signals arriving at the eastern coastal station (0918) and the propagation and amplitude decay westward across the island.
Figure E-2. Moveout velocity modeling scheme.

Each sub-fault contributes a step fault that propagates away from the fault. When the signal arrives at the station, the step function is added to the time series. The teleseismic based finite fault model of the Tohoku earthquake has 300 sub-faults, which are added together to make a synthetic static displacement time series.
Figure E-3. Moveout velocity fits to displacement time series.

Black curves are the observed time series at the stations labeled on the map. Colored curves are the synthetic static displacement time series for 2.0 (orange), 3.0 (red), and 4.0 (purple) km/s. A moveout velocity of 3.0-4.0 km/s fits the observed data best.
Figure E-4. Particle motions during the growth of static offset.

The observed displacements at the stations indicated on the map are rotated to radial (away from the epicenter) and plotted versus the vertical component, with color indicating time from the origin of the earthquake. These reveal generally reverse elliptical transient motions around the growing total displacement (solid colors). We also plot the difference between the observed particle motion and the particle motion predicted by the synthetic static time series (see Figure 7-8) as transparent curves.
Figure E-5. Surface wave depth sensitivity kernels for the western United States velocity model used in U.S. Geological Survey analysis (Herrmann et al., 2014).

The kinks in the curves represent abrupt transitions in the velocity model. Surface waves with different periods extend to different depth ranges; the 10 s period wave extends to ~20 km, the 30 s period wave extends to ~50 km, and the 50 s period wave extends to ~100 km. As a result, they move parts of the subduction system that have different levels of elastic stress, i.e. the elastic portion above 50 km and the visco-elastic region below 50 km.
I arrived at Penn State in Fall 2010 to begin a Master’s degree working with Kevin Furlong after taking a year off of school. I liked doing the Master’s so much that I stayed at Penn State to do a Doctorate degree. Kevin has sent me around the country and the world to work with a variety of excellent scientists and present my research. I have also worked closer to home, helping develop coursework, the weekly Geodynamics Seminar, the research group website, and much more. Basically, I have been the lab fixer over the past several years, which has given me a particularly broad set of skills for my career in science.

**Education**

- **Ph.D. Geosciences**, Penn State  
  Summer 2017
- **M.S. Geosciences**, Penn State  
  Summer 2012
- **B.A. Geology and Physics**, Amherst College, Magna cum laude  
  Spring 2009

**Awards**

- PSU Geosciences Graduate Student Colloquium 1st place Ph.D. (Post-Comps) Talk (Spring 2017)
- Outstanding Student Paper Award, Tectonophysics Section, 2012 AGU Fall Meeting
- Charles Knopf, Sr. Memorial Scholarship for Outstanding First-Year Graduate Students (2010)

**Teaching and Research Experience**

- 2016 Plate Tectonics Elective, Department of Geosciences, Penn State
- 2015, 2016 Deformation Modeling Short Course, Kasetsart University, Bangkok, Thailand
- 2012-2016 Geodynamics Seminar Coordinator, Department of Geosciences, Penn State
- 2010-2015 Summer Intern, National Earthquake Information Center, USGS, Golden, CO

**Research Papers**


