Title
Kinematic earthquake source imaging: theory and applications

Permalink
https://escholarship.org/uc/item/50q9p5ps

Author
Fan, Wenyuan

Publication Date
2017

Peer reviewed|Thesis/dissertation
UNIVERSITY OF CALIFORNIA, SAN DIEGO

Kinematic earthquake source imaging: theory and applications

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

in
Earth Sciences

by
Wenyuan Fan

Committee in charge:

Professor Peter M. Shearer, Chair
Professor David Benson
Professor Yuri Fialko
Professor Peter Gerstoft
Professor T. Guy Masters

2017
The dissertation of Wenyuan Fan is approved, and it is acceptable in quality and form for publication on microfilm and electronically:

Chair

University of California, San Diego

2017
DEDICATION

To my mother and my grandparents
We must not forget that when radium was discovered no one knew that it would prove useful in hospitals. The work was one of pure science. And this is a proof that scientific work must not be considered from the point of view of the direct usefulness of it. It must be done for itself, for the beauty of science, and then there is always the chance that a scientific discovery may become like the radium a benefit for humanity.

– Marie Curie
# TABLE OF CONTENTS

Signature Page ......................................................... iii
Dedication ............................................................... iv
Epigraph ................................................................. v
Table of Contents ..................................................... vi
List of Figures ........................................................ ix
List of Tables ......................................................... xiv
Acknowledgements ................................................... xv
Vita ................................................................. xx
Abstract of the Dissertation ........................................ xxii

Chapter 1  Introduction ............................................... 1
    1.1  Kinematic earthquake source imaging .......................... 1
        1.1.1  Multiple moment-tensor inversion ...................... 2
        1.1.2  Finite-fault slip inversion ............................. 3
        1.1.3  Back-projection ........................................ 4
    1.2  Thesis structure ............................................. 5

References .......................................................... 8

Chapter 2  Kinematic earthquake rupture inversion in the frequency domain 13
    2.1  Introduction .................................................. 14
    2.2  Theory ........................................................ 16
        2.2.1  Model setup and forward modeling ...................... 17
        2.2.2  Inverse formulation .................................... 18
        2.2.3  Objective functions .................................... 18
        2.2.4  Negative slip .......................................... 22
        2.2.5  Misfit measure ........................................ 23
    2.3  Synthetic tests ............................................... 24
        2.3.1  Problem description .................................... 24
        2.3.2  Inversion strategies .................................... 25
        2.3.3  No regularization ....................................... 26
        2.3.4  Regularized inversions ................................. 27
    2.4  Discussion ................................................... 29
    2.5  Sensitivity to noise or model assumptions .................. 31
<table>
<thead>
<tr>
<th>Chapter 7</th>
<th>Possible activation of splay faults during the 2006 Java tsunami</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1</td>
<td>Introduction</td>
</tr>
<tr>
<td>7.2</td>
<td>Tectonic Setting and Residual Gravity Anomaly</td>
</tr>
<tr>
<td>7.3</td>
<td>Tsunami Back-propagation and Seismic Back-projection</td>
</tr>
<tr>
<td>7.3.1</td>
<td>Tsunami Tide Gauge Back-propagation</td>
</tr>
<tr>
<td>7.3.2</td>
<td>Seismic P-wave Back-projection</td>
</tr>
<tr>
<td>7.4</td>
<td>Time-dependent P-wave Spectrum Analysis</td>
</tr>
<tr>
<td>7.5</td>
<td>Discussion</td>
</tr>
<tr>
<td>7.6</td>
<td>Conclusions</td>
</tr>
<tr>
<td>7.7</td>
<td>Supplementary Materials</td>
</tr>
</tbody>
</table>

References | 224

Chapter 8 | Investigation of back-projection uncertainties with M6 earthquakes |
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>8.1</td>
<td>Introduction</td>
</tr>
<tr>
<td>8.2</td>
<td>Data and Method</td>
</tr>
<tr>
<td>8.2.1</td>
<td>Data</td>
</tr>
<tr>
<td>8.2.2</td>
<td>Method</td>
</tr>
<tr>
<td>8.3</td>
<td>Results</td>
</tr>
<tr>
<td>8.4</td>
<td>Discussion</td>
</tr>
<tr>
<td>8.5</td>
<td>Conclusions</td>
</tr>
<tr>
<td>8.6</td>
<td>Supplementary Materials</td>
</tr>
</tbody>
</table>

References | 260

Chapter 9 | Conclusions |
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>9.1</td>
<td>Summary</td>
</tr>
<tr>
<td>9.1.1</td>
<td>Theory</td>
</tr>
<tr>
<td>9.1.2</td>
<td>Applications</td>
</tr>
<tr>
<td>9.2</td>
<td>Future research directions</td>
</tr>
<tr>
<td>9.2.1</td>
<td>Kinematic source models of moderate to small earthquakes</td>
</tr>
<tr>
<td>9.2.2</td>
<td>Role of fluids in earthquake triggering</td>
</tr>
</tbody>
</table>

References | 318
LIST OF FIGURES

Figure 2.1: The filtered SIV1 model. ........................................... 39
Figure 2.2: The SIV1 model station distribution. ............................. 40
Figure 2.3: Velocity (km/s) and density (g/cm$^3$) profile for SIV1 ....... 41
Figure 2.4: Inverted source model from least squares without any regular-
izations. ............................................................................ 42
Figure 2.5: Tradeoff ("L") curves of data misfit versus model norm and ex-
ample data fits for station 23. .................................................. 43
Figure 2.6: Preferred ruptured model from combined regularization. ...... 44
Figure 2.7: Synthetic seismograms (0-1Hz) from input and inverted rupture
models. ................................................................................. 45
Figure 2.8: Wavefront snapshots. ................................................... 46
Figure 2.9: Arrival times of the peaks of the slip-rate functions. ............ 47
Figure 2.10: Inverted slip from different subfault sizes. ....................... 48
Figure 2.11: Inverted slip for a sparse rupture model in which only two sub-
faults ruptured. ..................................................................... 49
Figure 2.12: Inverted rupture model by preferred regularization with SIV1
provided data. ....................................................................... 50
Figure 2.13: Regularization strength tradeoff ("L") curves for inversion with
SIV1 provided data. ................................................................. 51
Figure 2.14: Inverted rupture models by different groups. .................... 52
Figure 2.15: Inverted source model from data with Gaussian noise (SNR$_1$). 53
Figure 2.16: Inverted source model from data with Gaussian noise (SNR$_{10}$). 54
Figure 2.17: Inverted source model from data with time shift errors ($\sigma_s = 1$ s)
solved by regularization without iteration. .................................... 55
Figure 2.18: Inverted source model from data with time shift errors ($\sigma_s = 1$ s)
after 15 iterations to correct for time shift errors. ......................... 56
Figure 2.19: Inverted source models with assumed fault dips. .............. 57
Figure 2.20: Inverted source models with assumed rakes. .................... 58

Figure 3.1: P-wave displacements and displacement envelope functions. ... 77
Figure 3.2: Time-integrated images of back-projected P waves. ............... 78
Figure 3.3: Back-projection snapshots. .......................................... 79
Figure 3.4: Rupture evolution of the Mw 7.8 main shock and aftershocks
back-projection images. ......................................................... 80
Figure 3.S1: P wave velocity seismograms. ........................................ 84
Figure 3.S2: Time-integrated image of back-projected P waves from the mid-
dle frequency band 0.1 to 1Hz (MF). ......................................... 85
Figure 3.S3: Snapshots of MF back-projections compared with the stacked-
source time function. ......................................................... 86
Figure 3.S4: Spatiotemporal evolution of LF back-projection ............... 87
Figure 3.S5: Theoretical resolving power of the three frequency bands for the Mw 7.8 mainshock. .................................................. 88
Figure 3.S6: Aftershock back-projection results. ........................................ 89
Figure 3.S7: Images of back-projected P waves for the Mw 7.2 aftershock for three frequency bands. ........................................... 90
Figure 3.S8: Snapshots of the back-projections of the Mw 7.2 aftershock compared with the stacked source-time functions. ....................... 91
Figure 3.S9: Results of the multiple point source synthetic test. ................... 92
Figure 3.S10: Comparison of back-projection images of deconvolved velocity recordings and the original analysis (Figure 3). ..................... 93

Figure 4.1: Tectonic setting of the 2012 Sumatra earthquake sequence. ....... 107
Figure 4.2: Back-projection results from European stations. ..................... 108
Figure 4.3: Stacked BP energy function, aligned velocity seismograms, and dip distribution of possible focal mechanisms. ...................... 109
Figure 4.4: Stacked envelope functions. ................................................ 110
Figure 4.5: Horizontal cross sections of shear stress, normal stress and Coulomb stress. ............................................................... 111
Figure 4.S1: Comparison of the first 15 s of self-scaled P-waves of the Mw 7.2 earthquake and Mw 8.6 earthquake. .............................. 114
Figure 4.S2: P-wave velocity seismogram alignments. ............................... 115
Figure 4.S3: Time-integrated back-projection image from Hi-net stations. .... 116
Figure 4.S4: Comparison between observed very-long-period data and synthetic seismograms. .......................................................... 117
Figure 4.S5: Theoretical resolving powers of the European array and time-integrated back-projection images for a Mw 5.9 earthquake in the same region. ......................................................... 118
Figure 4.S6: Smooth envelope functions of GE.LHMI and GE.GSI. ................ 119
Figure 4.S7: Results of the multiple point source synthetic test. ................... 120
Figure 4.S8: Stacked P-wave velocity spectra of the 10 January 2012 Mw 7.2 Sumatra earthquake. ....................................................... 121
Figure 4.S9: Horizontal cross sections of Coulomb stress changes at 20 km for the three reverse-faulting M5 earthquakes. ......................... 122

Figure 5.1: Tectonic setting of the northern corner of the Tonga subduction zone. .............................................................................. 154
Figure 5.2: Back-projection results. .......................................................... 155
Figure 5.3: Rupture evolution of the 2009 Tonga-Samoa earthquake. .......... 156
Figure 5.4: Aligned velocity seismograms. .............................................. 157
Figure 5.5: Stacked envelope functions. .................................................. 158
Figure 5.6: Comparison of multiple moment tensor solutions with other available source models. ...................................................... 159
Figure 5.7: Comparison of theoretical resolving power and time-integrated back-projection results. ............................................. 160
Figure 5.8: Time-integrated back-projection images of four ~M6 earthquakes in the region. ............................................................. 161
Figure 5.9: Time-integrated back-projection images of four ~M6 earthquakes in the region with waveform alignment derived from the main-shock. ............................................................. 162
Figure 5.10: Back-projection results for a simulated multi-event rupture. . 163
Figure 5.11: One month of aftershock activity from the 2009 Tonga-Samoa earthquake. ............................................................... 164
Figure 5.12: Ensemble of residual bathymetry and swath bathymetry from STRM 15 ................................................................. 165
Figure 5.S1: P-wave velocity seismograms aligned and sorted by similarity to the final waveform stack. ........................................ 169
Figure 5.S2: Multiple moment-tensor inversion grid configuration and station maps. ............................................................... 170
Figure 5.S3: Snapshots of low-frequency and high-frequency back-projection results. ............................................................... 171
Figure 5.S4: Stations used for back-projection and their lower-focal-hemisphere polarities for P, pP and sP. ...................................... 172
Figure 5.S5: Back-projection results of the 2009 Tonga-Samoa earthquake. . 173
Figure 5.S6: Comparison between observed very-long-period data and synthetic seismograms of the GCMT solution for the 2009 Tonga-Samoa earthquake. ......................................................... 174
Figure 5.S7: Comparison between observed very-long-period data and synthetic seismograms of the two-point-source solution for the 2009 Tonga-Samoa earthquake. ......................................................... 175
Figure 5.S8: Data misfit variations with respect to centroid times and durations of the two subevents. ............................................... 176
Figure 5.S9: Theoretical radiation pattern of Rayleigh waves and Love waves for the two-point-source CMT solution. ...................... 177
Figure 5.S10: Comparison of the cumulative distance as a function of time. . 178

Figure 6.1: Twenty-seven early-aftershock-triggering mainshocks and their focal mechanisms. .................................................... 196
Figure 6.2: Back-projection results for three earthquakes with different focal-mechanisms (60% normalized energy contours). ............ 197
Figure 6.3: Time versus distance plot of triggered events. .................. 198
Figure 6.4: Aftershock distribution and relative locations of the triggered early aftershocks. ....................................................... 199
Figure 6.S1: Back-projection results of nine M7 earthquakes from 2004 to 2008 (60% normalized energy contours). ......................... 209
Figure 6.S2: Back-projection results of nine M7 earthquakes from 2009 to 2012 March (60% normalized energy contours). 210
Figure 6.S3: Back-projection results of nine M7 earthquakes from 2012 Oct to 2015 (60% normalized energy contours). 211
Figure 6.S4: Peak power time functions for nine M7 earthquakes from 2004 to 2008. 212
Figure 6.S5: Peak power time functions for nine M7 earthquakes from 2009 to 2012 March. 213
Figure 6.S6: Peak power time functions for nine M7 earthquakes from 2012 Oct to 2015. 214
Figure 6.S7: High-frequency back-projection results of 9 earthquakes with regional arrays. 215
Figure 6.S8: Back-projection results of five earthquakes that triggered early aftershocks. 216
Figure 6.S9: Back-projection results of nine moderate earthquakes from 2004 to 2011 with empirical corrections obtained from the mainshocks 217
Figure 6.S10: Back-projection results of nine moderate earthquakes from 2013 to 2015 with empirical corrections obtained from the mainshocks 218
Figure 6.S11: Time versus distance plot of possible dynamic triggered earthquake pairs from the ISC catalog (1993–2013). 219

Figure 7.1: Residual free-air gravity anomaly, splay faults at Java subduction zone and shallow seismicity near the 2006 Java tsunami earthquake. 243
Figure 7.2: Seismic back-projection and tsunami back-propagation results and finite-fault slip models of the 2006 Java earthquake. 244
Figure 7.3: Back-projection results. 245
Figure 7.4: P-wave spectrum analysis and rupture velocity inferred from cumulative distance as a function of time. 246
Figure 7.S1: Residual bathymetry anomaly, splay faults at Java subduction zone and shallow seismicity near the 2006 Java tsunami earthquake. 251
Figure 7.S2: Ocean surface displacements during the back propagation of tsunami from Christmas, Benoa, and Cocos tide gauges. 252
Figure 7.S3: Ocean surface displacements during the back propagation of tsunami from Broome and Hillarys tide gauges. 253
Figure 7.S4: P-wave velocity seismogram alignments. 254
Figure 7.S5: Stacked envelope function. 255
Figure 7.S6: Globally distributed stations and their P-wave polarities with respect to the GCMT solution of the mainshock. 256
Figure 7.S7: Aligned velocity seismograms (0.02-0.05 Hz) from globally distributed stations. 257
Figure 7.S8: Stacked P-wave spectrum for the 2006 Java earthquake. 258
Figure 7.89: Polar representation (azimuth and takeoff angles) of the falloff rate estimated between 0.3-1Hz.

Figure 8.1: M6 earthquakes within Japan Trench and stations used for investigation.

Figure 8.2: A cartoon illustrating the back-projection method and the biasing effect of 3D structure on travel times.

Figure 8.3: Back-projection results of Event 4 (Figure 8.1) with self-aligned time calibrations (0.05–0.3 Hz, 80% energy contour).

Figure 8.4: Back-projection results for Event 4 (Figure 8.1) with self-aligned time calibrations (0.3–1 Hz, 75% energy contour).

Figure 8.5: Back-projection results (0.05–0.3 Hz, 80% energy contour) for Event 4 with time calibrations from Event 33 (Figure 8.1).

Figure 8.6: Back-projection results (0.3–1 Hz, 75% energy contour) for Event 4 with time calibrations from Event 33 (Figure 8.1).

Figure 8.7: Back-projection results with time calibrations from reference Event 33 applied to 29 other events (Figure 8.1, 75% energy contour).

Figure 8.8: De, Db, and coherence ratios with coherence ratios at 0.05–0.3 Hz.

Figure 8.9: Deviation distance density distribution and coherency ratio versus De with linear and 4th-root stacking (0.05–0.3 Hz).

Figure 8.10: De, Db, and coherence ratios with coherence ratios at 0.3–1 Hz.

Figure 8.11: Deviation distance density distribution and coherency ratio versus De with linear and 4th-root stacking (0.3–1 Hz).

Figure 8.12: Time calibrations for stations.

Figure 8.13: Illustration of relative distances for a given station and relative times versus relative distances.

Figure 8.81: Back-projection results with time calibrations from Event 33 (Figure 1, 75% energy contour).

Figure 8.82: De, Db, and coherence ratios with coherence ratios at 0.05–0.3 Hz without visual inspection.

Figure 8.83: Deviation distance density distribution and coherency ratios versus De with linear and 4th-root stacking without visual inspection (0.05–0.3 Hz).

Figure 8.84: De, Db, and coherence ratios with coherence ratios at 0.3–1 Hz without visual inspection.

Figure 8.85: Deviation distance density distribution and coherency ratio versus De with linear and 4th-root stacking without visual inspection (0.3–1 Hz).
LIST OF TABLES

Table 2.1: Regularization parameters and misfit reduction. . . . . . . . . . 59
Table 5.1: Comparison of available CMT solutions . . . . . . . . . . . . . . 166
Table 6.S1: Triggered early aftershocks. . . . . . . . . . . . . . . . . . . . . . 220
Table 6.S1: Triggered early aftershocks. . . . . . . . . . . . . . . . . . . . . . 221
Table 6.S2: Mainshock-early-aftershock pairs in Fig. S8 . . . . . . . . . . . . 222
Table 6.S3: Moderate intermediate field earthquakes in Fig. S9,10 . . . . . . 223
Table 7.S1: Tide gauges and tsunami arrivals. . . . . . . . . . . . . . . . . . 250
Table 8.S1: M6 earthquakes investigated in the study . . . . . . . . . . . . . 298
Table 8.S1: M6 earthquakes investigated in the study . . . . . . . . . . . . . 299
ACKNOWLEDGEMENTS

Marie Curie said, “I was taught that the way of progress was neither swift nor easy.” That was part of the valuable lessons I learned during my graduate school. The other part was that it was never just about me. I was lucky to get a chance working at Scripps for the past six years, and people around me have been extremely kind and nice.

My advisor, Peter Shearer, is excellent, in my personal opinion, the best! Peter is creative and has the talent to view seismic data differently than anyone else. His unique perspective is magical. He also showed me how to be open-minded while staying rigorous about new findings, and we try to be the harshest reviewers of ourselves. Peter is consistent and always keeps pace with me. He always responds to my emails quickly, and has never let me wait for days to get a manuscript revision. At the same time, he is patient and allows me to make mistakes, which gives me freedom to explore and experiment. Peter is always constructive. Given my erroneous figures or the negative reviews we occasionally get, he always sees the positive parts of them. Peter always makes time for his students. There were times I walked into his office everyday, sometimes multiple times per day (desperately), and he would just put down his work and pay attention to my trivial improvements or complaints. During the meetings, he always tries to understand what I did, and motivate me with encouragements. These interactions boosted my confidence greatly and taught me how to be a professional. Peter puts great effort into promoting me and my work. He has encouraged me in applying for numerous awards and fellowships, and has written I-do-not-even-know-how-many letters for me. Peter is not an expressive person, but for every setback or challenge I encountered, he assured me that he had got my back and we would work it out together. It has been an honor and great pleasure working with Peter, and I will be grateful forever for his training and support.

I thank my thesis committee members for their mentorships and guidance through the years. Peter Gerstoft taught me about inversions, running, and biking. He has keen intuition and is always curious about my work. His approaches for advanced mathematics and numerical algorithms really broadened my horizons.
He showed me how to be a careful, technically solid, open-minded, and skeptical scientist. He also financially supported me for my first two years, only to see me leave noise studies and dive into earthquakes. He has treated me with great patience and has always looked out for me. I thank him for all the lessons he taught me both intellectually and physically. Guy Masters taught me everything I know about long period seismology, moment tensors, and deep Earth structure. He has shown me how to stay focused, perfect my craft, be original, be independent, and solve challenging problems that nobody dares to touch. He always makes time for me, shares his insights generously, and promotes me with great efforts. I thank Guy for all the knowledge, all the guidance, all the protections, and all the kind (occasionally sarcastic) encouragements. I also enjoy his humor a lot and tried to mimic his signature laughter several times without any success. Yuri Fialko taught me everything I know about geodesy and numerical modeling (sounds like a responsibility disclaimer). He has shown me how to appreciate the physics behind earthquake processes and see things from innovative perspectives. Because he works on both observations and modeling, he has shown me that understanding mechanisms helps design methodologies for better observations, while robust observations drive numerical modeling to reveal the not-so-obvious physics. I thank Yuri for always making time for me, all the encouragements, all the insights, and his kind promotions. I would also like to thank David Benson for his timely responses to all my emails and meeting requests.

A lot of the projects would not be done without my coauthors. I thank Chen Ji, Catherine de Groot-Hedlin, Michael Hedlin, Dan Bassett, Marine Denolle, Diego Melgar, and Junle Jiang. Chen taught me to carefully evaluate the assumptions of source imaging and pay close attention to understanding different types of data. He has mentored me through the years with great patience, kindly encouraged and promoted me. Chen is also such a happy person who is lovely to be around. Catherine and Michael gave me an opportunity to work with them with array data innovatively during my last year. They have shown me how to split a large array into subarrays and take advantage of local coherency. They are the nicest and most encouraging people ever despite the fact that I have not done
anything substantial. I appreciate Chen, Catherine and Michael for their kind mentorships. It has been great teaming with Dan, Marine, Diego, and Junle. It has been a lot of fun to work together, and I cannot wait for future collaborations.

I benefited tremendously from mentorships and help from faculty members of IGPP and GRD. I thank Anne Pommier, Adrian Borsa, Emily Chin, Lisa Tauxe, Kerry Key, Dave Stegman, Cathy Constable and Steve Constable, who helped me through my job search and generously offered advice. I particularly appreciate advice from Anne, Adrian, and Emily, who took their valuable time going through my slides and my application materials over and over and over. The interactions with them have been just wonderful.

I thank all the members of Peter’s group with whom I overlapped, including Nick Mancinelli, Xiaowei Chen, Songqiao Wei, Janine Buehler, Robin Matoza, Zhongwen Zhan, Wei Wang, Huajian Yao, Taka Uchide, Daniel Trugman, Qiong Zhang, and Beineng Zhang. They have been always helpful even after some have left IGPP. I am grateful to have had supportive cohorts, including Shi Sim, Qian Yao, Maggie Avery, Peter Kannberg, Kang Wang, Lauren Diperna, and Kyle Withers. I thank them for their friendships, trust, and kindness. I thank my officemates for their company, including Wesley Neely, Adrian Doran, Katia Tymofyeyeva, and Dallas Sherman. I particularly appreciate Wesley for his time reading through my application materials (probably more than ten times). I thank all the students I overlapped with during my days in IGPP, including Soli Garcia, Zhitu Ma, Samer Naif, Xiaopeng Tong, Erica Aaron, Valerie Sahakian, Xiaohua Xu, Andy Barbour, Matt Cook, Dara Goldberg, Jessie Saunders, and Zhao Chen. Soli has been consistently reliable even after he left IGPP. I would also like to thank my non-geophysicist friends, Julia Fiedler, Ning Ma, Megan Barron, Shannon Klotsko, Sean Crosby, and Krystle Chavarria. I thank scientists I met during the years, who offered encouragements and advice, including Jeff McGuire, Shiyong Zhou, Yihe Huang, Martin Mai, and Heidi Houston. Yihe has this mysterious power to light up everything around her. I look forward to working with Jeff, who is scarilly nice and scarilly smart. Martin, Shiyong and Heidi shared their work and life-experiences with me, and have promoted me with lots of effort. I thank the
IGPP Front desk, NetOps, and Scripps administrative staffs, including Maria Rivas, Gilbert Bretado, Wayne Farquharson, Arlene Jacobs, Wayne Chen, Denise Darling, and Maureen McGreevy.

I thank my friends and my family. Their presence in my life made everything enjoyable. I thank Xiaolin Tang, Da Yin, Qinan Luo, Qinghua Wang, Fan Ding, Shelley Das, Xiaojia Zhang, Shanshan Li, and Mengfan Zhu. Da took me to all the fancy restaurants in Beijing and comforted me with his witty jokes. Many years ago, I had a panic attack, and Da told me "if you can not pass it, nobody can!" Xiaolin believes in me and firmly stands by me with no exceptions. She has shown me courage and persistence to move forward no matter how hard life punches into our faces. My friends accepted me as who I am and never asked me to change a thing (which they really should have). My mom financially supported me through my college time, and I would not be writing this if she did not. I spent almost my entire childhood with my grandparents. They fed me everyday and beat me for reasons I could not remember. I love them dearly. I thank my dog Lola, who showed me how be fierce. Last, but definitely not least, I thank my boyfriend Mike for his love and support. I am deeply grateful for his encouragements to pursue an academic career. He has shown me kindness and brought me happiness.

This research was supported by National Science Foundation under Cooperative Agreement EAR1111111, EAR-0710881, EAR-0944109, and EAR-1261681.

Chapter 2, in full, is a reformatted version of the material as it appears in Geophysical Journal International: Fan, W., P. M. Shearer, and P. Gerstoft, Kinematic earthquake rupture inversion in the frequency domain, Geophys. J. Int. 199, 1138–1160, doi:10.1093/gji/ggu319, 2014. I was the primary investigator and author of this paper.

investigator and author of this paper.

Chapter 4, in full, is a reformatted version of the material as it appears in Geophysical Research Letters: Fan, W. and P. M. Shearer, Fault interactions and triggering during the 10 January 2012 Mw 7.2 Sumatra earthquake, Geophys. Res. Lett., 43, 1934–1942, doi:10.1002/2016GL067785, 2016. I was the primary investigator and author of this paper.

Chapter 5, in full, is a reformatted version of the material as it appears in Journal of Geophysical Research–Solid Earth: Fan, W., P. M. Shearer, C. Ji, and D. Bassett, Multiple branching rupture of the 2009 Tonga–Samoa earthquake, J. Geophys. Res. 121, doi:10.1002/2016JB012945, 2016. I was the primary investigator and author of this paper.

Chapter 6, in full, is a reformatted version of the material as it appears in Science: Fan, W., and P. M. Shearer, Local near instantaneously dynamically triggered aftershocks of large earthquakes, Science, 353, 1133-1136, doi: 10.1126/science.aag0013, 2016. I was the primary investigator and author of this paper.

Chapter 7, in full, has been submitted for publication of the material as it may appear in Tectonophysics: Fan, W., D. Bassett, M. A. Denolle, P. M. Shearer, C. Ji, and J. Jiang, Possible activation of splay faults during the 2006 Java tsunami earthquake, Tectonophysics, submitted. I was the primary investigator and author of this paper.

Chapter 8, in full, has been submitted for publication of the material as it may appear in Journal of Geophysical Research–Solid Earth: Fan, W., and P. M. Shearer, Investigation of back-projection uncertainties with M6 earthquakes, J. Geophys. Res., submitted. I am the primary investigator and author of the paper.
VITA

2017  Ph. D. in Earth Sciences, University of California, San Diego
2011  M. S. in Geophysics, Peking University
2008  B. S. in Geophysics, Peking University

PUBLICATIONS


Earthquakes are the primary source of seismic waves in seismology and understanding the earthquake source is essential for predicting ground motions and detailing the physics of rupture. Earthquake kinematic source imaging describes the whole rupture process during an earthquake. It does not directly require the resolved source model to be physically or dynamically plausible, but can help in understanding the conditions of rupture dynamics. Therefore, accurate earthquake source models are highly desirable.

In Chapter 2, we develop a frequency-based approach to earthquake slip inversion that requires no prior information on the rupture velocity or slip-rate functions.
In Chapter 3, we characterize the rupture process of the 25 April 2015 Nepal earthquake with globally recorded teleseismic P waves.

In Chapter 4, we investigate the 10 January 2012 Mw 7.2 Sumatra earthquake in the Wharton basin and detect dynamically-triggered large early aftershocks occurring on or near the subduction interface.

In Chapter 5, we constrain the spatiotemporal evolution of the 2009 Tonga-Samoan earthquake with global P-wave back-projection and a multiple moment-tensor inversion. Our results show that the rupture branches east of the trench axis were controlled by the geometry of bending-related faults on the Pacific plate, and that the rupture branch west of the trench axis may correlate with along-strike fore-arc segmentation.

In Chapter 6, we detect and locate 48 previously unidentified large early aftershocks triggered by $7 \leq M < 8$ earthquakes within a few fault lengths, during times that high-amplitude surface waves arrive from the mainshock. The observations indicate that near-to-intermediate-field dynamic triggering commonly exists and fundamentally promotes aftershock occurrence.

In Chapter 7, we analyze the 2006 Mw 7.8 Java tsunami earthquake with global P-wave back-projection, tsunami tide-gauge back-propagation, and P-wave source spectra modeling. The results suggest that the splay faults may have been reactivated during the Java earthquake.

In Chapter 8, we explore back-projection resolution by imaging M6 earthquakes within the Japan subduction zone. The quantitative analysis of uncertainties can provide guidelines for interpreting back-projection results in the region.

Finally, in Chapter 9, we discuss future research prospects of kinematic earthquake source imaging.
Chapter 1

Introduction

1.1 Kinematic earthquake source imaging

Earthquakes are among the worst natural hazards on Earth and understanding their rupture processes is critical for hazard assessment and mitigation. Earthquakes are transient releases of long-term accumulated tectonic strain energy and provide accessible sources informing Earth structure, tectonics, and rheology, and thus investigating earthquakes can serve as a probe to understand large-scale lithospheric processes.

Seismic waves generally provide the most direct observations of earthquakes, but they contain not only information about earthquakes but also the media that the waves travel through (Earth’s seismic velocity and attenuation structure etc.). This complicates studies of both earthquakes and Earth structure (e.g., Ma and Masters, 2015; Mai et al., 2016). Despite the challenges, considerable progress in earthquake seismology has been made in the past few decades. For example, earthquake locations and centroid moment-tensors (CMT) (Dziewoński et al., 1981; Ekström et al., 2012) are now robustly and routinely obtained, using a point-source approximation.

However, large earthquakes, like the 2004 Sumatra earthquake (e.g., Ishii et al., 2005), rupture for hundreds of kilometers, in which case the point-source approximation is inadequate to describe the kinematics. For such cases, rupture processes need to be modeled with extended fault areas, slip histories, and rupture
propagation directions and speeds. Earthquake kinematic source imaging (e.g., finite-slip inversion, back-projection, and beamforming) describes the whole rupture process spatiotemporally. Such kinematic modeling does not directly require that the resolved source models to be physically or dynamically plausible, but can provide realistic constraints on dynamic rupture simulations.

1.1.1 Multiple moment-tensor inversion

An earthquake moment tensor (MT) solution describes earthquake location, timing, moment, and its double-couple component describes two possible hosting fault planes. It has six independent elements that model an earthquake as a point-source, which are linearly linked by the Green’s function to the ground motions that the earthquake generates (e.g., Gilbert, 1971). This assumption is valid as long as the earthquake dimension and duration are smaller than the wavelengths and period of the seismic waves considered (Dziewoński et al., 1981; Ekström et al., 2012). Moment tensor solutions can be obtained with either linear or non-linear inversions with multiple seismic phases or other geophysical observations (e.g., Ekström et al., 2012; Duputel et al., 2012; Li et al., 2009; Weston et al., 2011; Cambiotti and Sabadini, 2013). The solutions are often termed centroid moment tensors (CMTs) because the location and time are average descriptions for the earthquakes obtained from the long-period and long-wavelength seismic waves. Multiple agencies provide earthquake CMT solutions soon after earthquakes. The most widely used catalog is the Global CMT (GCMT) (Ekström et al., 2012).

The simplified one-point source assumption yields model-induced errors or large compensated linear-vector dipole (CLVD) components, when an earthquake ruptures a very large area or is composed of a series of subevents with different focal mechanisms. For such complex cases, multiple-point sources with independent CMTs can be used to model the observations, providing insights for the complicated rupture process (e.g., Tsai et al., 2005; Li et al., 2009; Nealy and Hayes, 2015; Duputel and Rivera, 2017). In the multiple-point CMT analysis, moment-tensor elements of several sources are solved simultaneously or iteratively. The number of point sources is often pre-assumed.
1.1.2 Finite-fault slip inversion

Kinematic finite-fault slip models describe the spatiotemporal behavior of the rupture process of earthquakes, and often provide real-world constraints on dynamic rupture models and studies of earthquake physics (e.g., Huang et al., 2012; Causse et al., 2014). Various geophysical data can be used individually or in combination to compute finite-fault slip models, for example, seismic, geodetic, and tsunami observations (Yagi and Fukahata, 2011; Tong et al., 2010; Yokota et al., 2011; Koketsu et al., 2004; Melgar et al., 2016; Jiang and Simons, 2016).

Finite-fault slip models are often parameterized as sets of subfaults, and slip or slip-rate functions are solved simultaneously (Trifunac, 1974; Olson and Apsel, 1982; Hartzell and Heaton, 1983). These models can be estimated through linear or nonlinear approaches in either the time or the frequency domains (e.g., Uchide et al., 2009; Fan et al., 2014). For example, Ji et al. (2002a,b) adopted the Meyer-Yamada wavelet transform and simulated-annealing algorithm to determine finite-fault models with its first application to the 1999 Hector Mine earthquake. Recent rapid increases in computational power enables Bayesian inversion implementation for finite-fault slip inversion, which addresses model uncertainties in a more quantitative fashion (e.g., Monelli and Mai, 2008; Monelli et al., 2009; Minson et al., 2013, 2014).

Finite-fault slip inversions are now performed routinely and are often available shortly after major earthquakes (Hayes, 2011; Hayes et al., 2011). However, kinematic rupture models may show substantial variations for one earthquake among different research groups, despite fitting the data equally well (Minson et al., 2013; Fan et al., 2014). The significant variations among the source models are not well understood (e.g., Vallée and Bouchon, 2004; Shao and Ji, 2012; Razafindrakoto and Mai, 2014). The variations prevent detailing rupture physics, and make seismic hazard and induced tsunami hazard assessments challenging. No generally accepted criteria have been established to assess different proposed finite-fault slip models because many factors can cause the model deviations, for example, model parameterization, data type, Green’s functions, and inversion techniques (e.g., Minson et al., 2013; Fan et al., 2014). The Source Inversion Validation
(SIV) project (Mai et al., 2016) provides an online cooperative platform to collect finite-fault slip models and aims to quantify the uncertainty in earthquake source inversion. Current case studies suggest the inverted models tend to converge when multiple geophysical data sets are used (e.g., Yokota et al., 2011; Melgar et al., 2016). However, no theoretical evaluations have been systematically performed, suggesting future research directions.

1.1.3 Back-projection

Back-projection has revolutionized kinematic source imaging since its first application for the 2004 Sumatra earthquake (Ishii et al., 2005). The method is data-driven and the results are robust. It has proven useful in studying complex earthquake ruptures, early aftershock detection, and hazard early warning (Kiser and Ishii, 2012; Fan and Shearer, 2015, 2016b; An and Meng, 2017). The imaging method is widely applied because of its simplicity and the fact that it makes few \textit{a priori} assumptions of the fault geometry or rupture speed. Regional array data, such as Hi-net or USAArray, are often filtered at high-frequency for back-projection imaging because of their highly coherent waveforms (e.g., Kiser and Ishii, 2011; Wang et al., 2012; Meng et al., 2011; Koper et al., 2011; Satriano et al., 2012; Fan and Shearer, 2016a). However, globally distributed networks can provide better array geometries and their generally superior azimuthal coverage leads to better spatial resolution than regional arrays (e.g., Walker et al., 2005; Walker and Shearer, 2009; Yagi et al., 2012; Okuwaki et al., 2014; Fan and Shearer, 2015). The superior spatial resolution of global networks enables utilization of multiple frequency bands for a more complete description of the seismic radiation, providing important information to understand rupture dynamics and local tectonics (e.g., Fan et al., 2016).

Using teleseismic P-waves (30° to 90°) as proxies of earthquake radiations, which retain the phase information while their amplitudes are modulated during the wave propagation, the back-projection method first empirically aligns the records, then shifts the records with respect to the relative time difference of different locations, and finally stacks the seismograms to extract coherent seismic
radiation at hypothetical source locations around the hypocenter. The empirical
time alignments are applied to neutralize 3D velocity structure influences (Houser
et al., 2008), the relative travel times are often calculated with a 1D velocity model,
e.g., IASP91 (Kennett and Engdahl, 1991), and linear or non-linear stacking meth-
ods are applied to suppress noise (Ishii et al., 2005; Rost and Thomas, 2002). For
a set of hypothetical source points, the seismograms will constructively interfere
at true source locations, while destructively interfering at other points. By map-
ing the coherent seismic radiation, we can image the rupture direction, rupture
speed, and rupture extent for a given earthquake. In principle, the source locations
could be three-dimensional, but often only lateral variations can be well resolved
because of the poor depth sensitivity of teleseismic data for shallow earthquakes.
Therefore, the hypothetical source locations are often functions of latitude and lon-
gitude at fixed depth when imaging shallow earthquakes. Rarely, post-smoothing
or post-processing are required for back-projection images.

Despite the popularity of the method, mitigating artifacts and quantifying
image uncertainties are on-going endeavors in the community. They are important
issues because biased back-projected energy-burst locations may lead to erroneous
interpretations of rupture physics. It is challenging to quantify the image uncer-
tainties because back-projection relies on stacking instead of optimizing a misfit
function. Current uncertainty evaluations are based on synthetic tests, resampling
of the data, or back-projection of smaller events near the mainshock of interest.
These approaches do not address the model-induced uncertainties, for example,
for large earthquakes rupturing hundreds of kilometers where it is unclear to what
extent the empirical hypocenter-derived timing corrections fail when the rupture
propagates far away from the hypocenter. The issue is being actively investigated
and should inform future research directions.

1.2 Thesis structure

Chapters 2 to 8 of this thesis were originally written for individual publi-
cation. They can be read in isolation, but the main chapters are closely related to
the theme of kinematic earthquake source imaging.

Chapter 2, in full, is a reformatted version of the material as it appears in Geophysical Journal International: Fan, W., P. M. Shearer, and P. Gerstoft, Kinematic earthquake rupture inversion in the frequency domain, Geophys. J. Int. 199, 1138–1160, doi:10.1093/gji/ggu319, 2014. I was the primary investigator and author of this paper, in which we developed a frequency-based approach to earthquake slip inversion that requires no prior information on the rupture velocity or slip-rate functions.


Chapter 4, in full, is a reformatted version of the material as it appears in Geophysical Research Letters: Fan, W. and P. M. Shearer, Fault interactions and triggering during the 10 January 2012 Mw 7.2 Sumatra earthquake, Geophys. Res. Lett., 43, 1934–1942, doi:10.1002/2016GL067785, 2016. I was the primary investigator and author of this paper, in which we detected dynamically triggered early aftershocks of the 10 January 2012 Mw 7.2 Sumatra earthquake in the Wharton basin.

Chapter 5, in full, is a reformatted version of the material as it appears in Journal of Geophysical Research–Solid Earth: Fan, W., P. M. Shearer, C. Ji, and D. Bassett, Multiple branching rupture of the 2009 Tonga–Samoa earthquake, J. Geophys. Res. 121, doi:10.1002/2016JB012945, 2016. I was the primary investigator and author of this paper, in which we resolved rupture propagation of the 2009 Tonga–Samoa earthquake, and explained the observations with local tectonics.

Chapter 6, in full, is a reformatted version of the material as it appears in Science: Fan, W., and P. M. Shearer, Local near instantaneously dynamically triggered aftershocks of large earthquakes, Science, 353, 1133-1136, doi: 10.1126/science.aag0013, 2016. I was the primary investigator and author of this paper, in
which we detect and locate 48 previously unidentified large early aftershocks, which were buried in the mainshock coda. These aftershocks were near instantaneously dynamically triggered by 27 M7 earthquakes during times that high-amplitude surface waves arrive from the mainshock (< 200 s).

Chapter 7, in full, has been submitted for publication of the material as it may appear in Tectonophysics: Fan, W., D. Bassett, M. A. Denolle, P. M. Shearer, C. Ji, and J. Jiang, Possible activation of splay faults during the 2006 Java tsunami earthquake, *Tectonophysics*, submitted. I was the primary investigator and author of this paper, in which we analyze the 2006 Mw 7.8 Java tsunami earthquake, and suggest that splay faults in the outer wedge of the region have been reactivated during the mainshock.

Chapter 8, in full, has been submitted for publication of the material as it may appear in Journal of Geophysical Research–Solid Earth: Fan, W., and P. M. Shearer, Investigation of back-projection uncertainties with M6 earthquakes, *J. Geophys. Res.*, submitted. I am the primary investigator and author of the paper, in which we explore back-projection resolution limits by imaging M6 earthquakes within the Japan subduction zone. Our quantitative analysis of uncertainties have shown back-projection can serve as a reliable source imaging tool.

Finally, in Chapter 9, we briefly summarize the main findings presented in this thesis and discuss future research plans.
References


Li, X., G. Shao, and C. Ji (2009), Rupture process of Mw 8.1 Samoa earthquake constrained by joint inverting teleseismic body, surface waves and local strong motion, EOS Trans. Am. Geophys. Union, 90(53), U21D–03.


Trifunac, M. D., (1974), A three-dimensional dislocation model for the San Fer-
nando, California, earthquake of February 9, *Bull. seism. Soc. Am.* 64(1), 149–
172.

Tsai, V. C., M. Nettles, G. Ekström, and A. M. Dziewoński (2005), Multiple CMT

Uchide, T., Ide, S., and Beroza, G. C., (2009), Dynamic high-speed rupture from

Vallée, M. and M. Bouchon (2004), Imaging coseismic rupture in far field by slip

Walker, K. T., and P. M. Shearer (2009), Illuminating the near-sonic rupture
velocities of the intracontinental Kokoxili Mw 7.8 and Denali fault Mw 7.9
strike-slip earthquakes with global P wave back projection imaging, *J. geophys.

Walker, K. T., M. Ishii, and P. M. Shearer (2005), Rupture details of the 28 March
2005 Sumatra Mw 8.6 earthquake imaged with teleseismic P waves, *Geophys.

Wang, D., J. Mori, and T. Uchide (2012), Supershear rupture on multiple faults
for the Mw 8.6 Off Northern Sumatra, Indonesia earthquake of April 11, 2012,

Weston, J., A. M. G. Ferreira, and G. J. Funning (2011), Global compilation of
interferometric synthetic aperture radar earthquake source models: 1. Compar-

Yagi, Y. and Y. Fukahata (2011), Introduction of uncertainty of Green’s function
711–720.

Yagi, Y., A. Nakao, and A. Kasahara (2012), Smooth and rapid slip near the
Japan trench during the 2011 Tohoku-Oki earthquake revealed by a hybrid back-
doi.org/10.1016/j.epsl.2012.08.018.

Yokota, Y., K. Koketsu, Y. Fujii, K. Satake, S. Sakai, M. Shinohara, and
T. Kanazawa (2011), Joint inversion of strong motion, teleseismic, geodetic,
and tsunami datasets for the rupture process of the 2011 Tohoku earthquake,
Chapter 2

Kinematic earthquake rupture
inversion in the frequency domain

Abstract

We develop a frequency-based approach to earthquake slip inversion that requires no prior information on the rupture velocity or slip-rate functions. Because the inversion is linear and is performed separately at each frequency, it is computationally efficient and suited to imaging the finest resolvable spatial details of rupture. We demonstrate the approach on synthetic seismograms based on the Source Inversion Validation Exercise 1 (SIV1) of a crustal Mw 6.6 strike-slip earthquake recorded locally. A robust inversion approach is obtained by applying a combination of damping, smoothing, and forcing zero slip at the edge of the fault model. This approach achieves reasonable data fits, overall agreement to the SIV1 model, including slip-rate functions of each subfault, from which its total slip, slip time history and rupture velocity can be extracted. We demonstrate the method’s robustness by exploring the effects of noise, random timing errors, and fault geometry errors. The worst effects on the inversion are seen from errors in the assumed fault geometry.
2.1 Introduction

Kinematic finite-slip inversion resolves the spatiotemporal behavior of the rupture process during an earthquake. It generally does not directly require the resolved slip to be physically or dynamically plausible, although often constraints inferred from earthquake physics are applied to kinematic inversions. Various geophysical data can be used in kinematic studies, including seismic (Yagi and Fukahata, 2011), geodetic (Tong et al., 2010), tsunami (Yokota et al., 2011), and borehole records (Koketsu et al., 2004). Kinematic finite-slip models describe the whole rupture process, including its dimensions and rupture velocity, and help in understanding the conditions for pulse-like ruptures and super-shear ruptures. They provide clues regarding spatial variations in stress drop and fault asperities, while providing real-world constraints on dynamic rupture models and studies of earthquake physics (e.g., Causse et al., 2014). See Ide (2007) for a recent review of finite-slip inversion methods.

The first heterogeneous finite-slip inversion was Trifunac (1974), in which the fault plane was divided into several subfaults and the slip of each subfault was estimated as a Haskell model (Haskell, 1969). The key idea was to construct the fault plane as a set of subfaults and modern finite-slip inversion approaches continue to adopt this parameterization. Important early work includes Olson and Apsel (1982) and Hartzell and Heaton (1983), who developed a linear inversion method with inequality constraints to determine the spatial and temporal slip distribution of the 1979 Imperial Valley earthquake. The method and its extension are referred to as the multi-time-window method, which have been widely applied (e.g., Yagi, 2004; Yue et al., 2012; Uchide et al., 2009). More recently, Ji et al. (2002) adopted the Meyer-Yamada wavelet transform and simulated annealing algorithm to determine the finite-fault model that minimizes an objective function described in terms of wavelet coefficients, and applied it to the 1999 Hector Mine earthquake. Other nonlinear approaches, like Bayesian inversion, have been enabled by rapid improvements of computing power (e.g., Monelli and Mai, 2008; Monelli et al., 2009; Minson et al., 2013, 2014; Dettmer et al., 2014). Without assumptions about slip rate function shape or rupture velocity, solving source time
functions in the time domain is a huge inverse problem, a multidimensional deconvolution problem. It can be simplified and made more computationally tractable by using source time functions with idealized shapes.

Because ruptures do not necessarily have simple slip functions or constant rupture velocities, it is desirable to develop methods that make few assumptions about the model. For computational reasons, this is more likely possible in the frequency domain where the multidimensional deconvolution is replaced with solving a linear set of equations for each frequency (Olson and Anderson, 1988; Mendez and Anderson, 1991; Cotton and Campillo, 1995). Here we describe and further develop a frequency-domain algorithm for finite-slip inversion. For each frequency, the seismic velocity spectrum is a linear superposition of the Green’s functions multiplied by the slip-rate spectra on each subfault. Since this set of equations is solved independently at each frequency, the complexity of the inverse problem is greatly reduced. This computational efficiency enables dense parameterization of the subfaults for good spatial resolution and we avoid assuming a rupture velocity or specific form for the source-time function, which might limit the ability to resolve complex rupture models. To stabilize the inversion, regularization is applied instead.

The frequency domain approach follows Olson and Anderson (1988), but with the following improvements: (1) We derive and apply Bayesian based regularizations which penalize variations in the inverted slip-rate functions, including \( \ell_2 \) norm, roughness, and \( \ell_1 \) norm (i.e., compressive sensing, which produces sparse models), (2) We consider larger problems (e.g., more fault patches), exploiting improved computer resources and recently-developed convex optimization approaches for solving the inverse problem (Boyd and Vandenberghe, 2004).

Finite slip inversions are now performed routinely and are often available within a few days after an earthquake. However, understanding the uniqueness of these solutions remains challenging (e.g., Vallée and Bouchon, 2004; Konca et al., 2013; Shao and Ji, 2012; Razafindrakoto and Mai, 2014). Fault geometry parameterization, assumed Earth structure, seismic data type, Green’s function calculation methods, inversion approaches, and regularization constraints all af-
fect the inversion results. Thus, rupture models often show substantial variations among different groups, even while fitting the data equally well. There are no generally accepted criteria to evaluate the available models and their uncertainties, making interpreting the inverted models inherently subjective. Aiming to quantify the uncertainty in earthquake source inversion, the Source Inversion Validation (SIV) project (Mai et al., 2007) provides an online cooperative platform to test earthquake source inversion approaches through a series of benchmarks.

To test our frequency-domain method, we apply it to SIV Exercise 1 (SIV1, near-vertical strike-slip fault, recorded locally). We do not prescribe the rupture velocity and slip-rate functions, but apply regularization to stabilize the inversion, as we over-parameterize the model to avoid bias from too few unknowns. To provide good spatiotemporal resolution, subfaults are 1km × 1km and the slip-rate spectra are fit up to 1 Hz. We experiment with a variety of physically plausible regularization constraints to stabilize the inversion.

In the following sections, we will describe our method and explore the effects of inversion constraints, regularization strengths, noise and fault geometry. Our results for the SIV1 show that even with perfectly known Green’s functions, noise-free data, and a good station distribution, a wide variety of fault-slip models provide good fits to the data. The finite-slip inversion problem is inherently non-unique and regularization is necessary to obtain a stable result. However, because different regularizations produce different models, choosing the most appropriate regularization is critical and the best regularization strategy will vary depending on the details of each inversion problem.

2.2 Theory

In the frequency domain, the slip-rate spectra and recorded strong-motion spectra are linked by a linear Green’s function relationship. The spatial unknowns for each subfault patch can be inverted independently for each frequency. The problem can be solved efficiently in this fashion without any rupture time restriction (Olson and Anderson, 1988). The increased efficiency allows finer discretization
of subfaults, which gives the inverted rupture model greater resolution to resolve rapid spatial changes. The approach avoids subjective decisions about the slip-rate function shape and rupture velocity and eliminates the trade-off between rupture velocity and rupture front. In this section, we start with a brief review of the forward modeling approach. Using the same fault geometry as in forward modeling, we then construct and perform inversions with various constraints in the frequency domain. Finally, a time-domain misfit function is applied to comprehensively compare data misfit.

2.2.1 Model setup and forward modeling

Using the representation theorem, (Aki and Richards, 2002), the recorded displacement can be expressed as the convolution of the Green’s function of the source-receiver pair and the discontinuity displacement across the fault plane. There is a linear relationship in the frequency domain between the recorded displacement spectrum and the source displacement spectrum. Thus, the recorded velocity spectrum is linearly linked to the slip-rate spectrum by the Green’s function. Using this linear relationship, we modify Equation (15) of Spudich and Archuleta (1987) into a discrete form:

\[ v_i(x, f) = \sum_j M (T^s_i(x, f; \xi^j) T^d_i(x, f; \xi^j)) \left( \begin{bmatrix} v^s(\xi^j, f) \\ v^d(\xi^j, f) \end{bmatrix} \right) \Delta \Sigma(\xi^j) \]  

where the \( i \)th component of ground velocity spectrum \( v_i(x, f) \) at the Earth’s surface \( x \) and frequency \( f \) is expressed as a multiple of the traction and a slip-rate discontinuity across the fault plane. The fault plane is modeled as a set of \( M \) subfaults. Here \( [v^s(\xi^j, f)] \) and \( [v^d(\xi^j, f)] \) are the strike-slip and dip-slip slip-rate spectra at the \( j \)th subfault \( \xi^j \); \( T^s_i(x, f; \xi^j) \) and \( T^d_i(x, f; \xi^j) \) are the strike-slip and dip-slip tractions exerted with respect to the receiver and subfault setting; and \( \Delta \Sigma(\xi^j) \) is the subfault area at \( \xi^j \). The synthetic seismogram \( v_i(x, t) \) is obtained by taking the inverse Fourier transform of the velocity spectrum \( v_i(x, f) \).
2.2.2 Inverse formulation

To set up the inverse problem, we first Fourier transform the observed seismograms and align them into a vector according to station index, obtaining $d^o(f) \in \mathbb{C}^{3N}$, where $N$ is station number and each station has three components ($x$, $y$ and $z$). The problem is thus expressed as

$$d^o(f) = [G_s \ G_d] \begin{bmatrix} m^s(f) \\ m^d(f) \end{bmatrix} + n(f) = Gm + n(f) \quad (2.2)$$

where vectors $m^s(f) \in \mathbb{C}^M$ and $m^d(f) \in \mathbb{C}^M$ are the strike-slip and dip-slip slip-rate spectral components. Matrices $G_s \in \mathbb{C}^{3N \times M}$ and $G_d \in \mathbb{C}^{3N \times M}$ are the Green’s function matrices linking the source spectra and the recording spectra. $G^s_{ij}$ and $G^d_{ij}$ are the media responses of the $i$th recording due to strike-slip and dip-slip slip-rate pulses at subfault $j$, where $G^s_{ij} = T^s_i(x,f;\xi^j)\Delta \Sigma(\xi^j)$ and $G^d_{ij} = T^d_i(x,f;\xi^j)\Delta \Sigma(\xi^j)$. Differences between observations and synthetic seismograms from source models are described as noise $n$, which contains both data noise and any errors due to forward modeling. The inversion is performed at each frequency independently, which greatly reduces the problem size (Olson and Anderson, 1988). Note that the time and frequency parameterizations are physically equivalent, and thus identical solutions should be possible in either domain (Spudich and Archuleta, 1987).

2.2.3 Objective functions

We tackle the problem using a Bayesian approach:

$$p(m|d^o) = \frac{p(d^o|m)p(m)}{p(d^o)} \quad (2.3)$$

where $p(m)$ is the prior model probability density function (PDF), $d^o$ is the observed data, and $p(d^o|m)$ is the conditional PDF. Maximizing $p(m|d^o)$, we obtain the maximum a posteriori (MAP) solution:

$$\hat{m} = \arg \max_m p(m|d^o) \quad (2.4)$$
Least squares

Assume \( \mathbf{n} \) in eq. (2.2) obeys a complex Gaussian distribution, \( \mathcal{CN}(\mathbf{0}, \sigma^2 \mathbf{W}^{-1}) \). Then the likelihood function can be expressed as

\[
p(\mathbf{d}^o|\mathbf{m}) = c_0 \exp \left( -\frac{1}{\sigma^2} (\mathbf{d}^o - \mathbf{Gm})^H \mathbf{W} (\mathbf{d}^o - \mathbf{Gm}) \right)
\]  

(2.5)

The constraints come in via the prior distribution \( p(\mathbf{m}) \), since the \( p(\mathbf{d}^o) \) is just a normalizing constant. In the following sections we derive several regularization approaches and other priors can be easily utilized. The simplest prior model distribution is a non-informative uniform distribution, in which case, eq. (2.4) gives the least squares solution:

\[
\hat{\mathbf{m}} = \arg \max_{\mathbf{m}} c_0 \exp \left( -\frac{1}{\sigma^2} (\mathbf{d}^o - \mathbf{Gm})^H \mathbf{W} (\mathbf{d}^o - \mathbf{Gm}) \right)
\]

\[
= \arg \min_{\mathbf{m}} \| \mathbf{d}^o - \mathbf{Gm} \|_{\mathbf{W},2}^2
\]  

(2.6)

where \( c_0 \) is a constant and the covariance matrix is \( \mathbf{W}^{-1} \), which can be treated as a weighting matrix for the recorded data with respect to site effects and data quality. When \( \mathbf{n} \) is assumed to be identical and independent, \( \mathbf{W} \) is the identity matrix \( \mathbf{I} \).

The objective function (2.6) of the weighted least squares problem is (convex) quadratic, which is a quadratic program (QP) problem (Boyd and Vandenberghe, 2004).

As shown in Section 2.3.3, least squares without regularization often produces rough and unrealistic solutions even with noise-free data. This is because the slip model is typically over-parameterized, such that the data can be fit well with a variety of models, many of which contain rapid spatial or temporal oscillations between positive and negative slip. Regularization can be used to avoid these physically implausible models, and we experiment with both damping and smoothing, as described in the following.
Damped least squares

When the prior model has a complex Gaussian distribution, $\mathcal{C}N(m_0, \sigma^2_m I)$, the posterior PDF (2.3) also obeys a Gaussian distribution:

$$p(m|d^0) = c_0 \exp \left( -\frac{1}{\sigma^2} [(d^0 - Gm)^H (d^0 - Gm) + \frac{\sigma^2}{\sigma^2_m} (m - m_0)^H (m - m_0)] \right)$$

(2.7)

where $m_0$ is the prior model, which can integrate already existing information of the slip-rate into the inversion. When $m_0 = 0$, the MAP is given by

$$\hat{m} = \arg \min_m \|d^0 - Gm\|^2_2 + \alpha^2 \|m\|^2_2$$

(2.8)

where

$$\alpha^2 = \frac{\sigma^2}{\sigma^2_m}$$

(2.9)

is the variance ratio of $m$ and the prior model. It is challenging to find the best value for $\alpha^2$ without information about $\sigma^2_m$. Large $\alpha^2$ enforces the model to be close to the prior model $m_0 = 0$. Small $\alpha^2$ is preferred because it puts more weight on the data. The model maximizing eq. (2.7) is the damped least squares solution of this QP problem and attempts to find the lowest spectral power consistent with the data. The degree of damping controlled by the adjustable parameter $\alpha^2$.

Spatial smoothing

When the complex-valued slip-rate is assumed to be spatially smooth, the prior model can be expressed as:

$$p(m) = c_0 \exp \left( -\frac{1}{\sigma^2_i} m^H L^H L m \right)$$

(2.10)

where $L$ is the Laplacian matrix (Claerbout and Fomel, 2008), a 2D discrete second order finite difference operator and $\sigma^2_i/(L^H L)$ is the covariance matrix. Other smoothing matrices, such as the Gaussian smoothing matrix could also be used. Note that the operator applies to both the magnitude and phase of the spectra of connecting subfaults. The estimation under this assumption has spatial smoothing:

$$\hat{m} = \arg \min_m \|d^0 - Gm\|^2_2 + \lambda^2 \|Lm\|^2_2$$

(2.11)
where the degree of smoothing is controlled by the adjustable parameter

$$\lambda^2 = \frac{\sigma_i^2}{\sigma^2}$$  \hspace{1cm} (2.12)

As in the damped least squares approach, this is also a QP problem.

**Compressive sensing**

If slip is known to be spatially rough, it may be more appropriate to assume a prior that encourages spatially sparse solutions, such as a Laplacian distribution:

$$p(m) = c_0 \exp \left( -\frac{1}{b} \sum_{i} |m_i - \mu_i| \right)$$  \hspace{1cm} (2.13)

where $\mu$ is the model mean, and $b$ is a scale parameter. The MAP solution encourages sparsity and is termed compressive sensing (e.g., Yao et al., 2011; Xenaki et al., 2014), when $\mu = 0$:

$$\hat{m} = \arg \min_{m} \|d^0 - Gm\|_2^2 + \beta^2 \|m\|_1$$  \hspace{1cm} (2.14)

This is a second-order cone problem (SOCP), and can be efficiently solved with convex optimization (Boyd and Vandenberghe, 2004). Similar to smoothing, the degree of roughness is controlled by the adjustable parameter $\beta^2$.

**Combined constraints**

In general, optimal inversion strategies may involve a combination of the constraints listed above and we have experimented with this approach for the SIV1 problem. For example, the model can be required to be damped, spatially smooth, and with fixed boundary patches:

$$\hat{m} = \arg \min_{m} \|d^0 - Gm\|_2^2 + \lambda^2 \|Lm\|_2^2 + \alpha^2 \|m\|_2^2$$

s. t. : $W_0 m = 0$  \hspace{1cm} (2.15)

where $W_0$ is a selecting matrix, where elements corresponding to boundary patches are 1 and the rest are zero. Both the objective function and constraint are convex.
(Boyd and Vandenberghe, 2004), which assures a unique solution for the inverse problem.

The problem is greatly stabilized with the regularization strategy described in eq. (2.15). Fixing the fault boundary patches to zero while asking the slip-rate to be spatially smooth enforces the slip-rate to gradually decrease to zero at boundaries. This makes physical sense and will not introduce errors, since the modeled fault plane is over-parameterized and we can make it larger than the ruptured area.

2.2.4 Negative slip

Equation. (2.15) is in the frequency domain, thus to obtain the time-domain slip function, \( m \) must be Fourier transformed. The method assumes the signal \( d^o \) recorded by the seismic stations is accurately given by the computed Green’s functions, i.e., that there is coherent signal in the frequency-domain to be modeled. However, there are no assumptions of coherence across frequency or time. The slip at each frequency sample is independent and time slip functions are obtained via the inverse Fourier transform. This is in contrast to other slip-inversion schemes where time slip is often assumed to consist of one or more basis rise-time functions. The advantage of basis rise-time functions is that the number of unknowns can be greatly reduced, i.e., typically only a few unknowns are needed per cell (e.g., starting and ending times, amplitude, and a few more to define the function shape). However, this limited parameterization may prevent resolving some of the complexities of real earthquake ruptures, such as multiple slip events on the same subfault. In contrast, eq. (2.15) has one complex-valued unknown per cell per frequency (for the SIV1 inversion, \( 2 \times 50 = 100 \) parameters are used per cell), giving considerably more freedom in obtaining the true slip function.

Negative slip is physically unrealistic and can be eliminated by limiting the search space to only contain positive amplitudes as is often done for global search methods (e.g., Ji et al., 2002) or in classical least squares by using a positivity constraint (e.g., Yagi, 2004). However, implementing a non-negative slip constraint is difficult in frequency-domain methods and we do not impose such a constraint,
but the regularization tends to reduce negative slip. Instead, we use the size of the negative slip patches as a measure of the quality of the inverse solution. Well-resolved models should have only small amounts of negative slip and the positive slip patches have uncertainties similar to the magnitude of the negative slip. Note, however, that the absence of significant negative slip does not guarantee an accurate model, as we will show in Section 2.3.3.

2.2.5 Misfit measure

The inversion is performed in the frequency domain and includes regularization, see eq. (2.15). To quantitatively investigate inverted solutions, we compute the time-domain misfit parameter for the $i$th station:

$$
\epsilon_i = \sum_{j=x,y,z} \| \mathbf{d}_{i,j}(t) - \hat{\mathbf{d}}_{i,j}(t) \|_2^2
$$

(2.16)

where $\mathbf{d}_{i,j}(t)$ is the $j$th component at the $i$th station; $\hat{\mathbf{d}}_{i,j}(t) = \mathcal{F}^{-1}(\mathbf{Gm}(f))_{i,j}$ is the $j$th component synthetic seismogram at the $i$th station from the inverted rupture model. The misfit reduction is defined as

$$
\text{Misfit reduction} = 1 - \sqrt{\frac{E}{E_0}}
$$

(2.17)

where $E = \sum_{i=1}^{N} \epsilon_i$ is the total misfit and we have normalized with the norm of the data $E_0 = \sum_{i=1}^{N} \sum_{j=x,y,z} \| \mathbf{d}_{i,j}(t) \|_2^2$. When noise is present in the recordings, the signal-to-noise ratio (SNR) is defined as signal power over noise power

$$
\text{SNR} = \frac{\sum_{i=1}^{N} \sum_{j=x,y,z} \| \mathbf{d}_{i,j}(t) \|_2^2}{\sum_{i=1}^{N} \sum_{j=x,y,z} \| \mathbf{n}_{i,j}(t) \|_2^2}
$$

(2.18)

where $\mathbf{n}_{i,j}(t)$ is the noise added to the corresponding recording $\mathbf{d}_{i,j}(t)$. In addition to SNR, we construct a normalized noise parameter similar to misfit as

$$
\text{Noise level} = \left( \frac{\sum_{i=1}^{N} \sum_{j=x,y,z} \| \mathbf{n}_{i,j}(t) \|_2^2}{\sum_{i=1}^{N} \sum_{j=x,y,z} \| \mathbf{n}_{i,j}(t) + \mathbf{d}_{i,j}(t) \|_2^2} \right)^{1/2}
$$

(2.19)
2.3 Synthetic tests

2.3.1 Problem description

Exercise 1 of the SIV (http://equake-rc.info/) is used as a synthetic test case (Page et al., 2011; Mai et al., 2007; Mai, 2013). Right-lateral strike-slip motion occurs on a fault with dip 80° and strike 90° (see Fig. 2.1). The rupture remains buried and does not reach the surface. The fault plane is 36 km in length and 18 km in down-dip extent with top and bottom depths of 2.046 km and 19.772 km. The seismic moment $M_0$ is $1.06 \times 10^{19}$ Nm ($M_W$ 6.62). The hypocenter location is (9.2, 2.5, 14 km) in a right-lateral Cartesian system. The earthquake source model and its corresponding synthetic seismograms are generated using a spontaneous dynamic rupture model with heterogeneous initial stress on the fault. While the initial normal stress varies simply with depth, initial shear-stress is parameterized in terms of a von-Karman auto-correlation function (these and other details of the rupture simulation provided by Martin Mai, personal communication, 2014). The dynamic modeling assumes a linear slip-weakening friction law, with static and dynamic friction of 0.6 and 0.55, respectively. The dynamic rupture modeling is performed with a 3D generalized finite-difference method (Ely et al., 2008, 2009), with 100 m spatial discretization and 0.008 s time increments. The resulting slip-rate functions on the fault show Kostrov-type behavior with local variations due to the heterogeneous initial shear-stress. The peak final slip on the fault is smooth with an elliptical shape in the along-strike direction (Fig. 2.1a). Rupture propagation over the fault is fairly smooth, with small variations due to the stress variability (Fig. 2.1b), while the rise time distribution within the rupturing area is quite heterogeneous. Rupture speed averages about 2.8 km/s, but varies over the fault, which causes the rupture time, slip and slip-rate to be somewhat heterogeneous. Total rupture duration is less than 10 s. This assumed rupture model will be the focus of our inversion efforts and will henceforth be termed the SIV1 model. To compute synthetic seismograms, a layered velocity structure and station geometry is assumed (Figs. 2.2 and 2.3). There are a total of 40 three-component receivers, providing 120 recordings that can be used for
inversion. Stations are well-distributed on the surface around the fault. Both the SIV1 model and synthetics are provided as part of the SIV exercise.

We first regenerate the observed data using the provided velocity structure and source-time functions using the COMPSYN program ([Spudich and Archuleta, 1987]). Our forward synthetics are similar to those provided in the SIV1 exercise for frequencies below 1 Hz (misfit reduction 93.4%, SNR=229.5). The differences are likely related to model assumptions and parameter differences between the SIV dynamic model and COMPSYN. Validating synthetic seismograms is an important part of the SIV process, but is not our focus. To examine the performance of the inversion alone, we use our own forward synthetics, and use SIV1 provided data as a sensitivity test. We restrict our analysis of the seismograms to 1 Hz and below, because real slip inversion typically avoid data above 1 Hz where scattering and other effects reduce waveform coherence ([Cormier, 2007]).

2.3.2 Inversion strategies

Assuming that the fault-plane geometry and slip rake are specified, the model unknowns are just \( \mathbf{m}^* \) in eq. (2.2). The fault plane is the same as the true solution, 36 km in length and 18 km in down-dip extent. There are \( M = 72 \times 36 = 2592 \), 0.5 km \( \times \) 0.5 km patches used to generate the observed data. We Fourier transform the first 35 s of observed data into the frequency domain to construct \( \mathbf{d}^o(f) \), including all wave types (i.e. P waves, S waves, multiples and surface waves). For all the station-patch pairs, Green’s functions up to 1 Hz are computed to obtain \( \mathbf{G} \). At each frequency, we invert for the spatial distribution of slip-rate via eq. (2.2). The slip-rate spectra at each point on the fault are then inverse Fourier transformed into the time domain. For the inverse problem, the subfault size is set to be 1 km \( \times \) 1 km, which reduces the spatial unknowns to \( 36 \times 18 = 648 \) for each frequency. When the zero-slip boundary constraint is applied, as in eq. (2.15), the spatial unknowns are \( 34 \times 16 = 544 \) for each frequency.

For all the station-patch pairs, we compute Green’s functions to 1 Hz to obtain \( \mathbf{G} \) (\( \delta f = 0.024 \) Hz, 50 frequency bins are used). At each frequency, we then use convex optimization ([Boyd and Vandenberghe, 2004]) to invert for the spatial
distribution of slip-rate via eq. (2.2) or the equivalent regularized version using the cvx package (Grant and Boyd, 2013, 2008). The slip-rate spectra at each point on the fault are then inverse Fourier transformed into the time domain.

Our applied regularization, regularization strength and obtained misfits are summarized in Table 2.1. Regularization here is used to describe the inversion strategies, such as eqs. (2.8) and (2.15). Regularization strength refers to eqs. (2.9) and (2.12). Retrieved seismic moments ($M_0$) for each case are also listed in the table.

### 2.3.3 No regularization

Fig. 2.4 shows results of a simple least-squares inversion without regularization. The recovered slip model is much rougher than the SIV1 model (compare Figs. 2.1 and 2.4) and contains sharp changes in slip amplitude at the smallest spatial scales of the model. The slip-rate function near the hypocenter has a negative-slip precursor and oscillatory ringing for many seconds after slip has ceased in the starting model. The frequency domain fit to the slip-rate function is also poor, with substantially higher power at 0.3–0.8 Hz than contained in the SIV1 model. Note however that the negative slip component of this model is relatively small, just 0.53% of the moment of the positive slip component, Thus, the absence of substantial negative slip in our inversions does not necessarily imply a high-quality inverted rupture model.

The SIV1 slip-rate functions contain substantial energy above 1 Hz, which cannot be recovered from analysis of seismic data below 1 Hz. Thus, we apply a 1-Hz low-pass cosine filter with passband and stopband 0.5 and 1.2 Hz to the SIV1 slip-rate functions. The filtered SIV1 has seismic moment $M_0$ of $1.05 \times 10^{19}$ Nm ($M_W$ 6.61). This has little effect on fault maps of total slip (e.g., Fig. 2.1), but significantly broadens the slip-rate pulses, as shown in Fig. 2.4b. We will henceforth refer to this 1-Hz low-pass model as the filtered SIV1 model.

More physically realistic models can be obtained by applying regularization. It is important, however, to realize that these models are not required by the data. The model shown in Fig. 2.4 achieves a nearly perfect fit to all 120 components
of the data seismograms. The misfit reduction is 99.2% according to eq. (2.17) (Table 2.1). The model is not unique; there are many models that will produce equivalent fits.

### 2.3.4 Regularized inversions

Regularization can help stabilize the inverse problem and produce more physically plausible models. There is a tradeoff between how well the model fits the data and satisfies the regularization constraints. This tradeoff is controlled by the adjustable regularization parameters that assign relative weights to the regularization constraints and the data fit. This is often illustrated using an L-curve of misfit versus regularization norm. For the SIV1 problem, Fig. 2.5 shows time-domain data misfit reduction (eq. (2.17)) versus model norm for both damping and smoothness constraints. Increasing the regularization strength produces larger misfit and smaller, smoother models. An example of the fit to a seismogram is plotted in Fig. 2.5, both in the time and frequency domains. The Y component of Station 23 is used, which has the largest absolute peak. In principle, the regularization parameters should be frequency dependent, since we are performing the inversions independently at each frequency. However, for simplicity we use the same regularization parameters at all frequencies.

Ideally, choosing regularization strength would be informed by the expected level of data misfit, i.e., to avoid fitting the data better than their noise level. However, data uncertainties can be difficult to quantify, so the process of setting regularization strength can be somewhat subjective. Because our SIV1 synthetics have no noise, we initially apply very weak regularization, enough to provide a more physically realistic solution, but not enough to cause more than a tiny increase in data misfit. For real data sets or synthetic data with errors (Section 2.5), stronger regularization is required.

The inverted source model with the preferred regularization (eq. (2.15)) is plotted in Fig. 2.6. It achieves a 99.2% data misfit reduction and a good overall fit to the filtered SIV1 model, both in terms of total slip and slip-rate functions at selected points on the fault (Fig. 2.6). The total moment of this model is
1.10 \times 10^{19} \text{N} \cdot \text{m} (M_w = 6.62), compared to 1.05 \times 10^{19} \text{N} \cdot \text{m} (M_w = 6.61) for filtered SIV1. Note that these numbers are integrated moments, computed by subtracting the negative slip patches from the positive slip patches. However, the negative slip moment is small compared to the positive slip, just 0.63\% of the moment of the positive slip. Synthetic seismograms from the inverted rupture model and previous cases are shown in Fig. 2.7. Differences with the synthetic seismograms are indistinguishable. These differences between inverted slip models are a clear demonstration of the non-uniqueness of the problem. Even setting the negative slip patches to zero, the resulting model predictions still fit the data reasonably well (91.5\% misfit reduction).

This, and other regularized inversions we have tried, locate the correct hypocenter and resolve the rupture velocity and direction, none of which were prescribed. The colors in Fig. 2.6b give the areas that have slip-rates larger than 0.5 km/s at discrete specified timings, which provides a measure of the rupture front. Because this filtered SIV1 has a wider slip-rate pulse, there are overlapping areas over the fault plane at different times. To better show the rupture progress, the older rupture areas are plotted on top of the newer rupture areas in Fig. 2.6b, which emphasizes the new ruptured area. Fig. 2.8 shows the wavefront snapshots (no time smoothing or averaging) of the original unfiltered SIV1, the filtered SIV1, the least-squares solution, and the inverted source model with our preferred regularization. Smearred rupture fronts are obtained compared to sharp images of SIV1, which is expected given the spatial smoothing regularization. Negative slip patches are scattered over the fault plane in the least-squares solution, see Fig. 2.8c. Regularization tends to reduce negative slip rate as seen in Fig. 2.8d.

Arrival times of the peak of the slip-rate function with respect to distance from the hypocenter are plotted in Fig. 2.9. Rupture velocity can be estimated with wavenumber filtering (Olson and Anderson, 1988). Here, for simplicity, the slope of the curve is inferred as inverse rupture velocity. Inverted versus model rupture velocities along AA’ are compared in Figure 2.9. Spatial changes like the fast rupture velocity near the hypocenter (a result of the assumed 1-km wide slip initiation zone in the SIV1 model) and a nearly constant rupture velocity on the
left fault plane are well resolved. The shaded zone shows the 90%-peak zone of the inverted slip-rate functions. Note that the smoothing constraint introduces some small differences between the inversion result and SIV1, e.g., the inverted source model has earlier rupture at shallower depth (see 1s in Fig. 2.8), earlier initial slip and later peak slip in the slip-rate functions (Fig. 2.6), causing a timing offset between the preferred model and SIV1 (see Fig. 2.9).

2.4 Discussion

These results represent a best-case scenario for rupture inversion from near-field data because they result from a good station distribution, noise-free seismograms, perfectly known Green’s functions, and exact knowledge of the true fault geometry. We will explore the effect of relaxing some of these advantages in Section 2.5, but first we consider the resolution limits. The maximum frequency is 1 Hz, which gives a minimum S-wavelength of about 3.5 km over the fault. This provides an estimate of the spatial resolution, as evidenced by the smearing of the sharp wavefronts in the starting model into broader features in the recovered model (see Fig. 2.8). This agrees with Fukahata et al. (2014), who explored the resolution limits of slip inversions with good data coverage and exact Green’s functions. We also experimented with 2 km × 2 km and 0.5 km × 0.5 km model subfault sizes (see Fig 2.10, Table 2.1). A large patch size intrinsically enforces a simple smoothness constraint (0.5 km × 0.5 km patches within each large patch have the same spectrum). Although this model is coarser, it nonetheless achieves a 99.1% misfit reduction. However, it has abrupt changes in slip across the 2-km cell boundaries. The smaller patch size of 0.5 km is more computationally expensive than our 1-km-patch preferred model, but yields little improvement in model resolution or data fit, due to the spatial resolution limit.

Similarly, the temporal resolution is limited by the 1 Hz maximum frequency, as illustrated by a comparison of the original and filtered source-time function (see Fig. 2.6). For this synthetic example, the spatial and temporal resolution could be improved by including higher frequencies. However, this is difficult
for fault inversions of real earthquakes because of incoherence in the high-frequency waveforms and the lack of Earth models detailed enough to compute accurate synthetic seismograms above 1 Hz (e.g., Cormier, 2007; Ide, 2007).

In addition to resolution limits, the inverted models suffer other problems: they have some negative slip, they underpredict maximum slip amplitudes, and they perform relatively poorly in the low-slip areas of the fault. The low-pass filtering introduces small sidelobes and some negative slip in the time-domain. This negative slip is most visible just before the rupture initiation at each point (see Figure 2.6). The magnitude of the inverted spectra at each sub-fault is smaller than the SIV1 model due to the regularizations. The inversion is able to resolve the source-time function and its spectrum better in the high-slip regions near the hypocenter (#1 to #3 and #5 to #8 of Fig. 2.6) but poorly images the low-amplitude source-time functions at the rupture boundary (#4 of Fig. 2.6). The rupture front can only be partially resolved after 6 s and the slip-rate quickly diminishes to zero after 7 s (Fig. 2.8).

Because the inversion is performed in the frequency domain, constraints on the rupture timing and direction are difficult to apply directly. For example, causality suggests there should be no slip on fault patches at times before the arrival of a P-wave radiated from the hypocenter at the origin time. Slip timing is part of the spectrum phase and direct constraints on the phase has not been developed and requires further investigation. Similarly, physical considerations suggest that negative slip is unlikely, but as mentioned in Section 2.2.4, there is no direct way to implement positivity in our inversion scheme unless we decompose the slip-rate functions into a summation of a group of non-negative functions. This decomposition is intrinsically the same as multi-time-window methods, in which case we lose the flexibilities of the frequency domain approach.

The smoothness regularization worked in large part because the SIV1 model is in fact fairly spatially smooth. Thus, imposing a smoothness constraint helped in recovering the true model, even though rougher models could produce equally good data fits. However, ruptures may not always be spatially smooth. Recent results for several earthquakes have suggested complicated rupture scenarios in which
models of multiple discrete subevents may provide a more realistic description of
the earthquake than smooth, continuous rupture models (e.g., Koketsu et al., 2011;
Maercklin et al., 2012). In these cases, other regularizations will produce better
results, e.g. compressive sensing (Yao et al., 2011), which forces the rupture into
a sparse set of subevents. As an example, Fig. 2.11 shows results for a synthetic
model with just two isolated sub faults, with #1 rupturing once and #2 rupturing
twice at different times.

Compressive sensing (eq. 2.14) can precisely locate the rupturing locations
and times, showing how a frequency domain method can resolve a complicated
multi-subevent earthquake without a well-defined continuous rupture velocity, see
Fig. 2.11a. However, if the same damping and smoothing regularization (eq. 2.15)
is applied as our optimal results for the SIV1 model, the result is an over-damped
and smeared version of the true model (see Fig. 2.11b), which nonetheless fits
the data well. In this case, compressive sensing yields effective point sources.
However, it cannot resolve the slip versus rupture area tradeoff when subfaults
are inaccurately parameterized. Thus, no method is guaranteed to be optimal in
every case. One advantage in applying a smoothness constraint is that any imaged
heterogeneity (e.g., subevents, or multiple slip concentrations) is likely real, as the
smoothness regularization attempts to minimize these features.

2.5 Sensitivity to noise or model assumptions

The inverted source models in Section 2.3 achieve good spatial and tempo-
ral resolution, but are derived from an idealized model with noise-free data, and
perfectly known Green’s functions. These advantages are not present for inver-
sions of real earthquakes, so it is useful to test how more realistic scenarios affect
the resolution. We now explore how the solutions vary with respect to noise and
assumed fault geometry.
2.5.1 Noise and error influence

Noise and errors are inevitable in real slip inversions. Both background microseism noise and signal-generated noise from scattering will contribute to observed seismograms. In addition, Green’s functions for the forward problem will never be known perfectly because of unresolved 3D seismic velocity structure. Thus, it is important to understand the sensitivity of inversion methods to these effects. A method that works for an idealized, noise-free synthetic experiment will be of little practical use if it is overly sensitive to noise and errors. Here we investigate the effects of Gaussian background noise and Green’s function or station timing errors. When noise is added into the data, according to eq. (2.9), the regularization strength should increase to avoid “overfitting” the data beyond its noise level. The slip-rate magnitude and data fit will decrease as a consequence.

SIV1 provided data

Seismic data provided by SIV1 were generated from the dynamic model described in Section 2.3.1. As discussed earlier, our computed synthetic seismograms are a close, but not perfect match, to the synthetics provided in SIV1. If we define noise as the difference between our forward synthetics and those from SIV1 (1 Hz low-pass filtered), the misfit reduction is 93.4%. To test the robustness of our approach to small errors in the Green’s functions, we performed a source inversion with our preferred regularization, using the filtered SIV1-provided data. The inverted source model is plotted in Fig. 2.12. The misfit reduction is 93.1%, which is in good agreement with 93.4% as we do not expect to overfit the data. The inverted source model is nearly identical to that of Fig. 2.6, which is inverted from our synthetic seismograms, and all the key features of the SIV model are recovered, including the slip-rate functions. The total moment of this model is $1.17 \times 10^{19}$N·m ($M_w = 6.64$) and the negative/positive slip moment ratio is 0.53%, even less than the 0.63% for the Fig. 2.6 model. Regularization strength is chosen by an L-curve analysis, see Fig. 2.13.

A comparison of Fig. 2.12 to total slip from other SIV EX1 inversion models from the SIV database is shown in Fig. 2.14. All models used the same data, but
the model parameterizations and inversion techniques were different (Gallović and Zahradník, 2012; Razafindrakoto and Mai, 2014). We define the misfit as the sum of the absolute differences between the inverted source model and SIV1, normalized by the sum of the total slip of SIV1. Our model yields the smallest misfit, but of course we had the advantage of knowing the true model while optimizing our inversion approach.

**Gaussian noise**

We simulate both low-noise and high-noise problems by adding to the data white Gaussian noise with zero mean and a standard deviation set to be 1% and 10%, respectively of the peak amplitude of 32 cm/s of all observed data. Note that 39% (47/120) of the recordings have absolute peak values less than 3.2 cm/s, so the high-noise model severely affects these records. For the 1% peak amplitude case, \( \text{SNR}_{1} = 2.3 \) and \( \text{SNR}_{10} = 0.023 \) for the 10% case.

Inversion results are plotted in Figs. 2.15 and 2.16. Regularization strength and misfit reduction are shown in Table 2.1. For the \( \text{SNR}_{1} \) case, we recover the starting model almost as well as the zero-noise case, while achieving a 81.8% misfit reduction (as much as can be expected given the noise level is 18.8%). We found that an increase in regularization strength of \( 10 \times \) was required to achieve this level of misfit.

In contrast, the \( \text{SNR}_{10} \) noise case introduces considerable error into the inverted source model (see Fig 2.16). Following eqs. (2.15) and (2.12), the strong regularization is 100 times stronger than the \( \text{SNR}_{1} \) case. Although the inverted source model in this case has smeared and small total slip due to the strong regularization (Fig 2.16a), the rupture direction and slip history is recovered reasonably well (Fig 2.16b). The misfit reduction is 12.6%, which is in good agreement with the noise level, 88.6%. The slip-rate functions are recovered with small oscillations (Fig. 2.16c,d). Synthetic seismograms generated by the inverted source model (Fig 2.16e-h) fit the noise-free observations data fairly well. Thus the inversion method is robust with respect to even strong Gaussian noise, at least when 40 stations are available to average out its effects.
However, noise may not be Gaussian and it should be recognized that least-squares methods can be severely damaged by non-Gaussian noise (e.g., spikes, clipped data, or other outliers). In these cases, more robust results should be obtainable by using $\ell_1$, Huber penalty function or other norms, but we have not yet experimented with these approaches.

**Time shift errors**

Another possible source of noise is time shift errors in the Green’s functions caused by unmodeled velocity structure or station timing errors. To simulate this, we randomly time shift the data, using the same shift for all three components of a station. This time shift is assumed Gaussian, $\mathcal{N}(0, \sigma^2_s)$. When the standard deviation $\sigma_s \geq 1$, the inversion scheme (eq. (2.15)) fails to recover a reasonable model (see Fig. 2.17). This is because when the time shifts are too large, the resulting phase lags in the frequency domain are difficult to resolve. In this case, regularization alone does not work to recover the true model.

In real slip inversion problems, static time shifts are often applied to improve the phase alignment between data and synthetics (e.g., Allmann and Shearer, 2007). For our problem with random time shifts, we find that an iterative approach to estimating these time shifts is effective. We invert for a starting model from the data assuming all the time shift terms are zero. We then estimate the time shift for each station by cross-correlating data and synthetics and apply these empirical time shifts to the data before inverting again for the slip model. We repeat this until the process converges to a stable slip model and set of time shifts. For $\sigma_s = 1$, complete convergence is achieved after 15 iterations (see Fig. 2.18), although the bulk of the model improvement occurs in the first few iterations. Initially, the regularization parameters are set to be $\alpha^2 = 6.4 \times 10^{-6}$ and $\lambda^2 = 4 \times 10^{-8}$, and then set to be the same as the preferred regularization (Fig. 2.6) during subsequent iterations.

We find that this iterative approach succeeds for our synthetic problem even for quite large $\sigma_s$ (e.g., $\sigma_s = 5$ s, much longer than most data pulse widths~ 2 s). However, it should be noted that the only source of error is the time shifts;
once these are determined, the slip inversion itself is exactly the same as before. Of course, in reality 3D velocity structure will distort the shape and amplitude of the Green’s functions and may also cause correlated time shifts among nearby stations rather than the independent random time shifts assumed here. However, in some cases timing errors may be caused mainly by unmodeled local velocity structure below each station, in which case the shifts may be largely uncorrelated and the iterative empirical approach described here may be effective in recovering a relatively robust rupture image.

2.5.2 Unknown fault geometry

Unknown dip

The inversions so far have used exactly the same fault geometry as the true model. However, the exact position of the fault is often imprecisely known, so it is important to explore the erroneous effects on the slip model of using an inaccurate fault geometry. One of the least well-constrained fault parameters is the fault dip, so this is the parameter we focus on here. We assume the fault surface trace (i.e., position and strike) is the same as the true model, but vary the dip of the assumed fault plane. Deviations of ±10° from the true dip of 80° are tested, δ₁ = 70° and δ₂ = 90°. In each case, we compute Green’s functions for the erroneous fault location for the inversion, while using data generated using the true 80° dipping fault.

For unchanged regularization, using the wrong fault dip severely affects our solutions, generally having much worse effects than the random Gaussian noise or time shift errors. The models obtained using the same relatively weak regularization as before are quite heterogeneous and bear little resemblance to the true models, particularly for the slip time histories (see Fig 2.19), while achieving relatively poor data fits of 88.7% and 75.1% misfit reduction for the 70° and 90° models, respectively (Table 2.1). For the same σ₂ₘ, taking the unknown fault geometry into account, σ² will increase, which will lead to an increase in α² and λ² (eqs. (2.9) and (2.12)). Increasing the regularization strength can produce more plausible appearing models and slip histories. Of course, for real problems, where
we do not know the true model, it would be difficult to determine the optimal level of regularization.

A likely reason that errors in the assumed fault dip have such a large impact is that they introduce correlated errors, i.e., they cause arrivals to be systematically early on one side of the fault compared to the other. This same kind of error may also occur from unmodeled large-scale velocity variations, e.g., if velocities were faster on one side of the fault and this was not taken into account (e.g., Gallovič et al., 2010). Another more serious problem is the rotation of the radiation pattern from its correct orientation causing the amplitude and even the polarity of some arrivals to be incorrect. To address these difficulties it might make sense to include the fault orientation as part of the inversion. However, this would make the problem non-linear and we not yet experimented with the feasibility of this approach.

**Unknown rake**

Rake can be taken as an unknown parameter in source inversion (e.g., Ji et al., 2002). To stabilize the inversion, it can also be specified for each subfault (e.g., Yue et al., 2012). Until now we have used the true rake for all our inversions. To analyze the errors introduced by possible inaccurately known rake, we perform inversions with two specified rakes ($\pm 10^\circ$ of the true one), while dip is fixed to be the same as SIV1, see Figure 2.20.

For unchanged regularization, the wrong fault rake strongly affects our solutions. The effects are similar to those from using the wrong dip. The models obtained using the same relatively weak regularization as before have large negative slip areas, and no clear pattern can be observed from slip-time histories (see Fig 2.19), while fitting data with 94.8% and 94.0% misfit reduction for the 170° and 190° models, respectively (Table 2.1). Similar to the unknown dip situation, the wrong rake will lead to an increase in $\alpha^2$ and $\lambda^2$ (eqs. (2.9) and (2.12)). Increasing the regularization strength can produce more plausible appearing models and slip histories, but with poor data fit (Table 2.1). It is difficult to decide on the appropriate regularization strength based on the L-curve for the unknown fault
geometry cases, because subjective evaluation is needed to pick the right model besides considering the data misfit and model size.

2.6 Conclusions

We have developed a frequency-based source-inversion method that makes fewer assumptions than most current methods and applied it to SIV Exercise 1. We find that various physically plausible regularizations obtain robust inversion results. We spatially over-parameterize the model (1km × 1km) to avoid errors from too few unknowns. The linear relationship between slip-rate spectra and recorded spectra guarantees finding the global misfit minimum when convex optimization is applied to solve the inversion. Many details of the rupture process can be recovered reasonably well by the method, including rupture velocity and slip-rate functions.

Advantages of the frequency domain approach compared to time domain slip inversion algorithms are: (1) It is fundamentally linear and does not require any limits on the rupture velocity or duration of the slip function. Thus in principle, it should be able to resolve complicated ruptures, including variable rupture speeds, and even reversals of rupture direction, although these complexities were not analyzed. (2) Because it operates on each frequency separately, the complexity is relatively small and the algorithm is computationally efficient enough to permit a very fine spatial sampling of the fault and testing different regularizations.

The main disadvantage of the frequency domain method is that it is less direct than the time domain methods, which makes it more difficult to impose physically plausible constraints on slip direction and timing. In particular, we do not apply a positivity constraint on fault slip or prohibit slip at times earlier than the P-wave arrival from the hypocenter. However, static corrections, as shown in Section 2.5.1, can obtain coherent solutions around the hypocenter near the origin time. For the synthetic examples, regularization reduces negative slip and acausal slip to almost negligible amplitude and physically plausible constrains are not needed. However, physically unrealistic slip may well prove a bigger problem for inversions of real earthquakes.
Aiming to provide an improved understanding of the resolution limits and uncertainties in kinematic source inversions, we explored the robustness of the method with respect to noise, timing errors, and unknown fault geometry. Random Gaussian noise does not have a severe effect on the inverted source model, provided enough stations are available to effectively average it out. Random time shifts caused by station timing errors or poorly known Green’s functions can strongly damage the inversion, but can effectively be removed using an iterative approach that involves cross-correlation of data and synthetics, provided the time shifts are random and uncorrelated. A more severe problem is uncertainties in the assumed fault geometry. Errors in the fault dip and rake angle of 10° requires stronger regularization strength to produce more reasonable models, which are still relatively poorly resolved.

Acknowledgments

We thank Martin Mai for the SIV1 exercise and Paul Spudich for the COMPSYN software package. This work was supported by grants from the National Science Foundation EAR1111111, EAR-0710881 and EAR-0944109. The authors thank two anonymous reviewers and the editor for suggestions that improved the quality of this manuscript.

Chapter 2, in full, is a reformatted version of the material as it appears in Geophysical Journal International: Fan, W., P. M. Shearer, and P. Gerstoft, Kinematic earthquake rupture inversion in the frequency domain, Geophys. J. Int. 199, 1138–1160, doi:10.1093/gji/ggu319, 2014. I was the primary investigator and author of this paper.
Figure 2.1: The filtered SIV1 model, (a) Integrated total slip; (b) Time evolution of slip. For a specific time, only slip-rate functions of the subfaults that are greater than 0.5 m/s. are plotted in colors other than gray. Parts of the gray area do rupture, but the maximum value of the slip-rate functions do not exceed 0.5 m/s. The colored area provides a measure of the rupture front. Because this filtered SIV1 has a wider slip-rate pulse, there are overlapping areas over the fault plane at different times. To better show the rupture progress, the older rupture areas are plotted on top of the newer rupture areas to emphasize the new ruptured areas.
Figure 2.2: The SIV1 model station distribution. There are 40 three-component stations, providing 120 total records. The rupture remains buried and does not reach the surface. Hypocenter location is (9.2, 2.5, 14) km in this (x, y, z) coordinate system.
Figure 2.3: Velocity (km/s) and density (g/cm$^3$) profile for SIV1. A layered isotropic velocity-density structure is provided for the synthetic test. $Q_s$ and $Q_p$ are assumed to be infinite.
Figure 2.4: Inverted source model from least squares without any regularizations. (a) Total slip (m), white star indicates hypocenter; (b) Slip-rate function at hypocenter; (c) Power spectrum of the slip-rate function at hypocenter. The misfit reduction of the least squares model is 99.2%. In (b) and (c), gray lines are from the unfiltered SIV1 model (dynamic simulation); black lines are filtered SIV1 model and red ones are inverted least squares model.
Figure 2.5: Tradeoff (“L”) curves of data misfit versus model norm (top row) and example data fits for station 23 in the time domain (second row), the frequency magnitude (third row) and frequency phase (fourth row). The left column shows results of damped least squares, the right column for spatial smoothing. The L curve is computed in the time domain using the entire model and all of the data fits. The Y-component of station 23 is shown because it has the largest peak value of all the recorded data. The three plotted misfit curves correspond to the three labeled points on the L curves. Horizontal dashed lines in (a) and (e) represent 1% misfit level; red square represents the values we used for the preferred regularization; blue square is the model norm (\(\|m(t)\|_2\)) and zero misfit of the filtered SIV1 model.
Figure 2.6: Preferred ruptured model from combined regularization (see text), (a) Total slip; (b) Slip history; (c) Slip-rate functions at given points; (d) Power spectrum of the slip-rate functions at given points.
Figure 2.7: Synthetic seismograms (0-1Hz) from input and inverted rupture models for least squares, damped least squares, spatial smoothing, preferred regularization combination. The seismograms are indistinguishable.
Figure 2.8: Wavefront snapshots of (a) unfiltered SIV1 model, (b) filtered SIV1 model, (c) least squares (no regularization) and (d) preferred inverted source model with regularization.
Figure 2.9: Arrival times of the peak of the slip-rate function with respect to distance from the hypocenter along AA’ on the fault plane. The star shows the zero location of the horizontal axis. Rupture velocity can be inferred from the inverse slope of the time versus distance line. Shaded zone indicate 90%-peak zone of the inverted slip-rate functions.
Figure 2.10: Inverted slip from different subfault sizes. (a) with subfault size as 2km×2km with misfit reduction 99.1%; (b) 1km×1km, as Fig 2.6 with misfit reduction 99.2%; c, 0.5km×0.5km with misfit reduction 99.2%, the finest subfault size as the Green’s function calculation.
Figure 2.11: Inverted slip for a sparse rupture model in which only two subfaults ruptured. #1 ruptured once and #2 ruptured twice. The rupture locations and times are well resolved with compressive sensing. Black lines show the true model, red lines are inverted slip-rate functions of compressive sensing and blue lines are inverted slip-rate of damping and smoothing. (a) Inverted slip of compressive sensing, i.e., eq.2.14; (b) Inverted slip with damping and smoothing regularization, i.e., eq.2.15. These models yield data variation reductions of 94.98% for compressive sensing and 94.97% for damping and smoothing. The horizontal dashed line indicates the zero axis.
Figure 2.12: Inverted rupture model by preferred regularization with SIV1 provided data (see text), (a) Total slip; (b) Slip history; (c) Slip-rate functions at given points; (d) Power spectrum of the slip-rate functions at given points.
Figure 2.13: Regularization strength tradeoff (“L”) curves for inversion with SIV1 provided data (see text). Panel layout is similar to Fig. 2.5, only the gray dashed line represents 7% misfit level.
Figure 2.14: Inverted rupture models by different groups. Misfit is the sum of absolute differences between the inverted model and SIV1 that are normalized by the sum of total slip of SIV1. (a) parameterization of the fault plane of different groups; (b) filtered SIV1 model; (c) inverted ruptured model by preferred regularization with SIV1 provided data (Fig. 2.12); (d) to (h) inverted slips by other groups (Gallović and Zahradník, 2012; Razafindrakoto and Mai, 2014). (d), (f) and (h) are inverted slips of Razafindrakoto and Mai (2014) using a triangular source-time function, or a Yoffe function with acceleration time Tacc of 0.1 and 0.3 sec, respectively.
Figure 2.15: Inverted source model from data with Gaussian noise (SNR$_1$). (a) Total slip; (b) Slip history; (c) Slip-rate function at hypocenter; (d) Spectrum of the slip-rate function; (e) Seismograms with the largest peak amplitude; (f) Spectra of the seismograms (e); (g) Seismogram with the median peak amplitude; (h) Spectra of the seismograms (g).
Figure 2.16: Inverted source model from data with Gaussian noise (SNR_{10}). (a) Total slip; (b) Slip history; (c) Slip-rate function at hypocenter; (d) Spectrum of the slip-rate function; (e) Seismograms with the largest peak amplitude; (f) Spectra of the seismograms from (e); (g) Seismogram with the median peak amplitude; (h) Spectra of the seismograms from (g).
Figure 2.17: Inverted source model from data with time shift errors ($\sigma_s = 1$ s), solved by regularization without iteration. (a) Total slip; (b) Slip history; (c) Slip-rate function at hypocenter; (d) Spectrum of the slip-rate function; (e) Seismograms with the largest peak amplitude; (f) Spectra of the seismograms from (e); (g) Seismogram with the median peak amplitude; (h) Spectra of the seismograms from (g).
Figure 2.18: Inverted source model from data with time shift errors ($\sigma_s = 1$ s) after 15 iterations to correct for time shift errors. (a) Total slip; (b) Slip history; (c) Slip-rate function at hypocenter; (d) Spectrum of the slip-rate function; (e) Seismograms with the largest peak amplitude; (f) Spectra of the seismograms from (e); (g) Seismogram with the median peak amplitude; (h) Spectra of the seismograms from (g).
Figure 2.19: Inverted source models with assumed fault dips. (a) Assumed fault geometry; (b) Total slip and slip history for $\delta_1 = 70^\circ$ with the same regularization strength as the preferred model. (c) Total slip and slip history for $\delta_1 = 70^\circ$ with strong regularization strength. (d) Total slip and slip history for $\delta_2 = 90^\circ$ with the same regularization strength as the preferred model. (e) Total slip and slip history for $\delta_2 = 90^\circ$ with strong regularization strength.
Figure 2.20: Inverted source models with assumed rakes, true rake $\lambda = 180^\circ$. (a)-(d), assumed rake $\lambda_1 = 170^\circ$, (a), (b) show total slip and slip history with the same regularization strength as the preferred model. (c), (d) show total slip and slip history with strong regularization strength. (e)-(h), assumed rake $\lambda_2 = 190^\circ$, (e), (f) show total slip and slip history with the same regularization strength as the preferred model. (g), (h) show total slip and slip history with strong regularization strength.
Table 2.1: Regularization parameters and misfit reduction.

<table>
<thead>
<tr>
<th>Class</th>
<th>$\alpha^2$</th>
<th>$\lambda^2$</th>
<th>Misfit reduction (%)</th>
<th>Seismic moment $M_0$ (Nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Least Squares</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Preferred DLS*</td>
<td>$9.0 \times 10^{-10}$</td>
<td>99.2</td>
<td></td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>DLS1</td>
<td>$2.1 \times 10^{-5}$</td>
<td>97.0</td>
<td></td>
<td>$8.8 \times 10^{18}$</td>
</tr>
<tr>
<td>DLS2</td>
<td>$2.3 \times 10^{-3}$</td>
<td>65.7</td>
<td></td>
<td>$8.0 \times 10^{17}$</td>
</tr>
<tr>
<td>DLS3</td>
<td>$6.7 \times 10^{-3}$</td>
<td>47.9</td>
<td></td>
<td>$2.8 \times 10^{17}$</td>
</tr>
<tr>
<td>Preferred SS*</td>
<td></td>
<td>$4.0 \times 10^{-8}$</td>
<td>99.2</td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>SS1</td>
<td>$3.6 \times 10^{-5}$</td>
<td>97.8</td>
<td></td>
<td>$1.3 \times 10^{19}$</td>
</tr>
<tr>
<td>SS2</td>
<td>$1.5 \times 10^{-2}$</td>
<td>66.9</td>
<td></td>
<td>$9.7 \times 10^{18}$</td>
</tr>
<tr>
<td>SS3</td>
<td>$5.1 \times 10^{-2}$</td>
<td>47.9</td>
<td></td>
<td>$9.5 \times 10^{18}$</td>
</tr>
<tr>
<td>Preferred Regularization*</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>99.1</td>
<td>$1.2 \times 10^{19}$</td>
</tr>
<tr>
<td>(2km by 2km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Preferred Regularization*</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>99.2</td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>(1km by 1km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Preferred Regularization*</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>99.2</td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>(0.5km by 0.5km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SIV1 DLS1</td>
<td>$2.1 \times 10^{-5}$</td>
<td>95.9</td>
<td></td>
<td>$9.1 \times 10^{18}$</td>
</tr>
<tr>
<td>SIV1 DLS2</td>
<td>$2.3 \times 10^{-3}$</td>
<td>66.2</td>
<td></td>
<td>$8.3 \times 10^{17}$</td>
</tr>
<tr>
<td>SIV1 DLS3</td>
<td>$6.7 \times 10^{-3}$</td>
<td>48.7</td>
<td></td>
<td>$3.0 \times 10^{17}$</td>
</tr>
<tr>
<td>SIV1 SS1</td>
<td>$3.6 \times 10^{-5}$</td>
<td>96.4</td>
<td></td>
<td>$1.4 \times 10^{19}$</td>
</tr>
<tr>
<td>SIV1 SS2</td>
<td>$1.5 \times 10^{-2}$</td>
<td>65.0</td>
<td></td>
<td>$1.0 \times 10^{19}$</td>
</tr>
<tr>
<td>SIV1 SS3</td>
<td>$5.1 \times 10^{-2}$</td>
<td>46.1</td>
<td></td>
<td>$9.9 \times 10^{18}$</td>
</tr>
<tr>
<td>SIV1 provided data</td>
<td>$2.3 \times 10^{-8}$</td>
<td>$1.0 \times 10^{-6}$</td>
<td>93.1</td>
<td>$1.2 \times 10^{19}$</td>
</tr>
<tr>
<td>1% Gaussian noise</td>
<td>$9.0 \times 10^{-9}$</td>
<td>$4.0 \times 10^{-7}$</td>
<td>81.8</td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>10% Gaussian noise</td>
<td>$9.0 \times 10^{-7}$</td>
<td>$4.0 \times 10^{-5}$</td>
<td>12.6</td>
<td>$8.6 \times 10^{18}$</td>
</tr>
<tr>
<td>70° dip¹</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>88.7</td>
<td>$9.0 \times 10^{18}$</td>
</tr>
<tr>
<td>70° dip²</td>
<td>$2.6 \times 10^{-8}$</td>
<td>$1.0 \times 10^{-4}$</td>
<td>75.9</td>
<td>$9.4 \times 10^{18}$</td>
</tr>
<tr>
<td>90° dip¹</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>75.1</td>
<td>$1.1 \times 10^{19}$</td>
</tr>
<tr>
<td>90° dip²</td>
<td>$1.0 \times 10^{-7}$</td>
<td>$1.0 \times 10^{-4}$</td>
<td>71.1</td>
<td>$1.0 \times 10^{19}$</td>
</tr>
<tr>
<td>170° rake¹</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>94.8</td>
<td>$1.0 \times 10^{19}$</td>
</tr>
<tr>
<td>170° rake²</td>
<td>$4.0 \times 10^{-8}$</td>
<td>$4.0 \times 10^{-6}$</td>
<td>92.7</td>
<td>$1.0 \times 10^{19}$</td>
</tr>
<tr>
<td>190° rake¹</td>
<td>$9.0 \times 10^{-10}$</td>
<td>$4.0 \times 10^{-8}$</td>
<td>94.0</td>
<td>$1.0 \times 10^{19}$</td>
</tr>
<tr>
<td>190° rake²</td>
<td>$4.0 \times 10^{-8}$</td>
<td>$4.0 \times 10^{-6}$</td>
<td>91.8</td>
<td>$1.0 \times 10^{19}$</td>
</tr>
</tbody>
</table>
DLS represents damped least squares (eq. (2.8)); SS represents spatial smoothing (eq. (2.11)); Preferred regularization (eq. (2.15)). Misfit reduction is calculated by eq. (2.17). 70° dip\textsuperscript{1,2} represents Fig. 2.19b,c; 90° dip\textsuperscript{1,2} represents Fig. 2.19d,e. 170° rake\textsuperscript{1,2} represents Fig. 2.20a to d; 190° rake\textsuperscript{1,2} represents Fig. 2.19e to h.
References


Chapter 3

Detailed rupture imaging of the 25 April 2015 Nepal earthquake using teleseismic P waves

Abstract

We analyze the rupture process of the 25 April 2015 Nepal earthquake with globally recorded teleseismic P waves. The rupture propagated east-southeast from the hypocenter for about 160 km with a duration of ~55s. Back-projection of both high frequency (HF, 0.2 to 3 Hz) and low frequency (LF, 0.05 to 0.2 Hz) P waves suggest a multi-stage rupture process. From the low frequency images, we resolve an initial slow down-dip (northward) rupture near the nucleation area for the first 20s (Stage 1), followed by two faster up-dip ruptures (20 to 40s for Stage 2 and 40 to 55s for Stage 3), which released most of the radiated energy northeast of Kathmandu. The centroid rupture power from LF back-projection agrees well with the GCMT solution. The spatial resolution of the back-projection images is validated by applying similar analysis to nearby aftershocks. The overall rupture pattern agrees well with the aftershock distribution. A multiple asperity model could explain the observed multi-stage rupture and aftershock distribution.
3.1 Introduction

A Mw 7.8 earthquake struck central Nepal on 25 April 2015 with an epicenter 77 km northwest of Kathmandu. With over 8000 fatalities, it is the largest and most destructive earthquake since the 1934 Bihar-Nepal earthquake in this region (Singh and Gupta, 1980; Bilham, 2004). The Global Centroid Moment Tensor (GCMT) (Ekström et al., 2012) favors a nodal plane with strike 293°, dip 7° to the north, and rake 108°, and the preliminary finite-fault model from the USGS National Earthquake Information Center (NEIC) shares a similar solution (strike 295° and dip 10°). The hypocenter depth is 15 km, indicating the quake occurred on the Main Himalayan Thrust fault (MHT). This fault has been known to host reoccurring large events (e.g., Ambraseys and Douglas, 2004; Bilham, 2004). The GPS-derived convergence rate between India and South Tibet ranges from 17.8±0.5 mm/yr to 20.5±1 mm/yr from central and eastern Nepal to western Nepal (Ader et al., 2012). Because of the large moment deficit accumulation on the MHT within Nepal, it has been suggested that the fault would be able to host a Mw 9.2 earthquake with a return period of the order of 3000 years, if all the moment is released seismically (Ader et al., 2012).

For large earthquakes, the teleseismic P wavetrain contains details of the spatiotemporal slip distribution and rupture propagation. This is illustrated for the Nepal earthquake in Figure 3.1, which shows that the displacement envelope functions of the low-frequency P waves exhibit a clear azimuthal pattern from 40s to 65s, with an average duration of 55s. Assuming that the P-wave signals are mostly from the rupture front, then the rupture direction is about 130°. Following Ni et al. (2005), we can estimate that the rupture length is about 160 km with an average rupture speed of ~2.9km/s (see Supplement). This rupture length agrees well with the aftershock distribution. To resolve details during the rupture propagation, we plot the seismograms versus the directivity parameter defined in Ammon et al. (2005) and Zhan et al. (2014) (see Supplement), assuming a rupture direction of 130°. As shown in Figure 3.1, this plot reveals three distinct subevents, with times and epicentral distances that can be directly estimated from the intercepts and slopes of lines connecting the subevent pulses on each seismogram. These events
occurred at times of 24 s, 36 s, and 47 s, and distances of 54 km, 85 km, and 115 km, with respect to the hypocenter.

Although this approach is useful for providing a quick view of some of the main details of the rupture, it has limited resolution because the records outside of the three largest subevents are too weak to distinguish finer details and the method assumes that the rupture propagates in a one-dimensional fashion only and thus we cannot image possible two-dimensional variations in radiation within the rupture plane. To learn more, we apply P-wave back-projection to image more directly the rupture properties of the Nepal earthquake. Since its introduction by (Ishii et al., 2005) for the 2004 Sumatra earthquake, the back-projection method has been widely applied to large earthquake imaging, most often using regional arrays such as Hi-net or USAArray (e.g., Meng et al., 2011; Wang et al., 2012; Meng et al., 2012; Koper et al., 2011; Kiser and Ishii, 2012). However, in principle, higher resolution can be obtained using globally distributed stations (e.g., (Walker et al., 2005; Yagi et al., 2012; Okuwaki et al., 2014)) and this is the approach we adopt here. We analyze both a high frequency band (0.2 to 3 Hz), similar to that used most often in prior back-projection studies, and a low frequency band (0.05 to 0.2 Hz) to provide a more complete description of the seismic radiation.

Our results indicate that the rupture has three stages, with the first stage rupturing eastward in the down-dip direction and the later two stages involving eastward rupture in the up-dip direction. The second rupture stage radiated most of the energy within the bandwidth of our study, and its location and time agree well with the GCMT solution and preliminary finite-fault models from the NEIC and Geospatial Information Authority of Japan (GSI) (www.gsi.go.jp). Integrating our results with the aftershock distribution, current available finite-fault models and centroid moment solutions, we propose a multiple asperity model. Finally, we compare our results with source models of the 2005 Kashmir earthquake to explore common characteristics and differences among continental earthquakes in the Himalayan region.
3.2 Method and Data

The back-projection method assumes that the initial P-wave arrival comes from the hypocenter, but later parts of the P wave train likely contain overlapping contributions from different parts of the ensuing rupture. We follow the method described in Ishii et al. (2005) and Walker et al. (2005), with \( N \)th root stacking Xu et al. (2009) to suppress noise. A 1D velocity model (IASP91 (Kennett and Engdahl, 1991)) is used to predict theoretical P-wave travel times. We empirically correct the travel-time deviations that are due to 3D velocity structure by aligning the initial P arrival (Reif et al., 2002). The aligned far-field seismograms are then back-projected to a grid of possible sources around the hypocenter to constructively interfere if they are true source locations, or to destructively interfere if they are not. For a given grid point, the start time of coherent energy bursts represents the onset time of the rupture at that position, and the integrated power indicates the relative intensity of P-wave radiation at that point. The source depth is poorly constrained by far-field P waves, therefore, the source depths are fixed to the hypocentral depth and the potential source locations are functions only of latitude and longitude. Non-linear stacking approaches like \( N \)th root stacking were originally designed to reduce false alarms in seismic array detection (Rost and Thomas, 2002). They can effectively suppress noise and enhance the coherent signals when applied to back-projection (Xu et al., 2009), at the cost of losing absolute amplitude information. Detailed discussion about the effects of \( N \) can be found in McFadden et al. (1986); in this study, we use \( N = 4 \) as suggested in Xu et al. (2009).

The investigated area is 400 km by 400 km with a 5 km grid point spacing, within the latitude range 26.34° to 29.92° and longitude range 83.63° to 87.70°. We use seismograms from stations of the broadband Global Seismic Network (GSN) distributed by the Data Management Center (DMC) of IRIS (Figure 3.1). We select 62 stations with high signal-to-noise ratios with epicentral distances ranging from 30° to 90°, thus avoiding the waveform complexities at shorter ranges from the upper-mantle discontinuities and at longer ranges from the lowermost mantle near the core-mantle boundary. The azimuth ranges from 3.9° to 347.7° (Figure 3.1);
the good azimuthal coverage greatly reduces back-projection artifacts and enables a relatively high spatial resolution. With a 40 Hz sample rate, the seismic data are filtered into three frequency bands (0.05–0.2 Hz (low frequency, LF), 0.1–1 Hz (middle frequency, MF), and 0.2–3 Hz (high frequency, HF)) to investigate potential frequency-dependent behavior. All three frequency bands use the same alignment obtained from the cross-correlation of MF band data (Figure S1). To save space, only the LF and HF results are described in the main paper; as might be expected, the MF results generally lie between these two end-member frequency bands (see Supplement).

The amplitude of the P wavetrain is normalized to neutralize the variations caused by site effects, the radiation pattern, and different instrument gains. To avoid biased back-projection results from noisy and/or overrepresented regions, stations are weighted by their average correlation coefficients from the alignment and inversely with the number of contributing stations within 5 degrees. No post-smoothing or post-processing is applied to the images.

### 3.3 Results

The integrated back-projected energy over the ~60s duration of the rupture is shown in Figure 3.2. Both HF and LF back-projection show the rupture zone is mostly east and southeast of the epicenter with a rupture length about 165 km (1.5° in longitude); and both HF and LF back-projection show major energy release north of Kathmandu (Figure 3.2). A Mw 6.7 aftershock occurred half an hour after the mainshock close to the hypocenter; a Mw 6.8 aftershock occurred one day later at the eastern boundary of the mainshock rupture. On 12 May 2015, a Mw 7.2 aftershock occurred 83 km northeast of Kathmandu, which is close to the Mw 6.8 aftershock, and half an hour later, a Mw 6.2 earthquake occurred close to the Mw 7.2 aftershock. Most of the HF energy was released north of Kathmandu, while the LF energy bursts show another energy release peak beneath Kathmandu (Figure 3.2). Locations of the peak energy bursts seen in the LF back-projection with 5s windows are labeled in Figure 3.2, illuminating a multi-stage
rupture process. In the LF integrated energy image, the centroid rupture power agrees well with the GCMT solution (Figure, 3.2).

Snapshots of both the HF and LF radiation distribution show the rupture propagation details (Figure 3.3). From the absolute LF integrated-power images (normalized with the maximum power over the entire 60 s) and stacked LF source time functions, weak rupture propagation during the first 20s can be observed. The normalized LF integrated-power images (normalized with the maximum power of each 5 s window) reveal an initial northeast down-dip rupture (Stage 1). From 20 to 40 s, the LF integrated energy climbed to its maximum and a southeast up-dip rupture propagated toward Kathmandu (Stage 2). From 30 to 35 s, the LF integrated power concentrated around the GCMT centroid location, which is next to Kathmandu. From 40 to 50s, another southeast up-dip rupture broke the northeastern part of the fault and propagated parallel to the 20 to 40s rupture (Stage 3). These three rupture stages are labeled with hexagrams in Figure 3.2 and labeled with yellow arrows in Figure 3.4 (left panel). Compared to the LF images, relatively more HF energy was released during the initiation stage (Figures 3.2, 3.3, and 3.4). From 10 to 25 s, the HF image shows eastward rupture propagation heading into where the Stage 3 rupture starts. The HF energy release reached its maximum from 30 to 35s west of the Stage 3 rupture and ∼50km north of Kathmandu. The HF snapshot at 35 to 40s suggests a rupture around the Stage 3 rupture, and from 40 to 50s, the HF images are similar to the LF images. From 50 to 55s, the normalized LF image indicates new rupture near the hypocenter area. However, bootstrap resampling tests (see next section) suggest this is not a robust feature, so we cannot be confident that it is not some kind of back-projection artifact in the absence of other supporting observations.

3.4 Discussion

The rupture velocity varies among the three stages. The earthquake had a slow start; assuming the Stage 1 rupture followed the path indicated in Figure 3.2, the average rupture velocity is ∼2km/s. In Stage 2, the distance between the
peak energy bursts of 25–30s and 30–35s is \(\sim 46\) km. If the rupture propagated linearly, the apparent rupture velocity is 4.6 km/s, which is faster than the local S-wave velocity (Laske et al., 2013). The rupture velocity during Stage 3 would be \(\sim 2\) km/s if it followed the direct path shown in Figure 3.2. However, we cannot constrain the rupture behavior outside of the times of large radiation bursts. For example, instead of traveling directly from the hypocenter (or other reference point) through an asperity, the rupture could proceed around the asperity before causing it to break from the side and rupture at a misleadingly high apparent velocity away from the hypocenter. In this case the apparent rupture velocity at times could be much higher than the true rupture velocity. By projecting the back-projected energy to certain azimuths, all three stages can display very fast rupture speeds (Figure S4). To better constrain the possible rupture velocities, near-field seismic observations will be needed.

To understand the uncertainties and robustness of the back-projection images, we have performed three types of resolution tests. First, the theoretical resolution can be evaluated by randomly assigning a single recorded P wavetrain to all the stations, then performing back-projection with these traces. This provides a measure of the likely resolution of the station distribution given the frequency content of the data. As seen in Figure S5, as expected the spatial resolution is proportional to the bandwidth used for back-projection, and the theoretical spatial resolution of the LF data is about 50 km in radius.

Another way to test the resolution is to perform back-projection on aftershocks with similar station coverage as the mainshock. In this way, the complexities of the wavefield are taken into account, and due to the fact that the aftershocks have fewer usable stations and may themselves have finite rupture areas, the resolution during the mainshock should be at least as good as that seen in the aftershock images. We performed back-projection on the Mw 6.7, Mw 6.8 and Mw 7.2 aftershocks using 53, 56 and 56 records, respectively. We did not attempt to back-project the 12 May 2015 Mw 6.2 aftershock because of its relatively poor P-wave signal-to-noise ratio. The integrated energies of these three aftershocks are shown in Figures 3.4, S6, and S7. The Mw 6.7 event occurred half
an hour after the mainshock, so the P wavetrain is severely contaminated by the mainshock surface wave, which leads to lower spatial resolution compared to the Mw 6.8 aftershock (Figure S6). Snapshots and stacked source-time functions (see Figure S8) for the Mw 7.2 aftershock indicate a complicated rupture lasting about 20 s, with at least two subevents. The time-integrated LP image for the Mw 7.2 event does not appear larger than the Mw 6.8 rupture image, suggesting the Mw 7.2 rupture may have been relatively compact and likely of higher stress drop than the Mw 6.8 event (Figure S7).

From Figure 3.4, the spatial extent of the Stage 2 and the Stage 3 ruptures imaged in the mainshock is comparable to the spatial extent of the aftershocks, showing that the distinct subevents are clearly resolved. As a final test, we performed bootstrap resampling to verify the stability of our results with respect to random variations in the stations sampling the mainshock. With the exception of the LP results at 50 to 55 s (see above), we found that all of the main features imaged in the back-projection results are robust, i.e., they appear in over 95% of the bootstrap resampled images. Details of the bootstrap analysis are presented in the Supplement.

Artifacts in back-projection images may arise due to complicated waveforms, depth phases, or limited station coverage, which may contribute to misleading or erroneous interpretations of the rupture process. The good azimuthal coverage of the teleseismic data minimizes the “swimming” artifacts (e.g., Xu et al., 2009; Koper et al., 2012) that are troublesome in back-projection images from regional arrays. For large shallow earthquakes, depth phases will be present that cannot easily be separated from the direct phases. Because depth phases occur close in time and at similar slowness to the direct phases, they will back-project to locations near the direct phase image at only slightly later times, and thus depth-phase effects are often ignored in back-projection studies of large earthquakes. However, for the high-resolution imaging at 5-s intervals that we perform here it is important to examine the possible biasing effects of depth phases as well as interference effects from multiple sources. We tested the observed multi-stage rupture propagation with a multiple point source synthetic test (which includes
depth phases computed from the GCMT solution) and a depth-phase deconvolution analysis and found that depth phases and other complexities do not bias our results very much (Figure S9 and Figure S10). All of the main features that we image are robust, although the depth phases extend the duration of the radiation at some source locations by about 5 seconds.

The earthquake rupture propagation is primarily unilateral with possible multiple branches (Figure 3.2, 3.3, 3.4). Combining both LF and HF images, one possible scenario is that the rupture front reached Stage 2 and Stage 3 around the same time after the initiation; but the part of the fault resolved as Stage 3 did not break until the observed Stage 2 rupture passed Kathmandu. The Stage 2 rupture imaged at LF is less obvious in HF results. This suggests that the Stage 2 rupture revealed by the LF images is deficient in high-frequency energy. The cause of this HF deficit is unclear, which will need further investigation. The HF peak energy burst locates at the edge of the Stage 3 rupture and does not collocate with the centroid location. On the other hand, the LF back-projection results agree well with the centroid moment tensor solution both temporally and spatially (Figure 3.2). The aftershocks of the Nepal earthquake are distributed compactly within the back-projection imaged region (Figure 3.2, 3.4), with the Mw 6.7 aftershock near the west edge of the mainshock, and the Mw 6.8 and Mw 7.2 aftershocks on the eastern side. Noticeably, the epicentral distance between the Mw 6.8 and Mw 7.2 aftershocks is within 10 km. In addition, the majority of the aftershocks concentrate at the eastern edge of the mainshock.

One plausible explanation for the observed rupture pattern and the aftershock seismicity is a multiple-asperity model as illustrated in Figure 3.4: the mainshock is dominated by three asperities, indicated as A1, A2 and A3, which correspond to the three rupture stages. The Mw 6.7 aftershock is labeled as A1a in Figure 3.4 and may be an unbroken remnant of A1 from the mainshock rupture. The Mw 6.8 aftershock is labeled as A4 and ruptured one day later at the east boundary of the mainshock. The Mw 7.2 aftershock is labeled as A5 and occurred two weeks later, followed by a Mw 6.2 aftershock denoted as A5a. Because of the spatial clustering of the Mw 6.8, Mw 7.2 and Mw 6.2 events, it is possible that
A4, A5 and A5a are three parts of one large asperity that ruptured sequentially. However, because we can resolve only the main sources of seismic radiation we cannot image the connecting segments that might verify this; thus the three stages of the mainshock are not necessarily directly spatially connected or temporally linked into a sequence.

The GSI quickly released crustal deformation observations obtained with Synthetic Aperture Radar. The interferometric analysis shows a major displacement ($\geq 10$ cm) area extending about 160 km in the east-west direction, which agrees well with our directivity analysis (Section 8.3). The preliminary finite-fault model released by GSI has maximum slip ($>4$m) beneath the area 20 to 30 km northeast of Kathmandu, which is consistent with the observed Stage 2 rupture. In the GSI slip model, there is a significant slip patch north of the hypocenter, which validates the Stage 1 rupture observed in our LF back-projection. The ScanSAR (SAR with a swath coverage) of Advanced Land Observing Satellite 2 (ALOS-2) show two patches of line-of-sight (LOS) deformation outside of the 500mm deformation contour (Lindsey et al., 2015) that are possibly due to the Stage 3 rupture and the aftershocks (Figure 3.4).

The faults defining the India and Eurasia plate boundary have been repeatedly active due to the continental collision. There were four great earthquakes with magnitudes $\sim$Mw 8 along the boundary between 1897 to 1950 (Bilham, 2004). Similar to the 2015 Nepal earthquake, none of the four events ruptured to the surface (Bilham, 2004). The most recent large earthquake occurring on this boundary was the 2005 Kashmir earthquake. Both events appear to be simple shallow crustal events with compact slip distributions and both apparently nucleated at the edge of their main asperities (Avouac et al., 2006). The 2005 Kashmir earthquake initiated at the bottom edge of the main slip patch and the 2015 Nepal earthquake nucleated at the western edge of its rupture zone. Both earthquakes apparently ruptured more than one asperity (Parsons et al., 2006; Pathier et al., 2006). Although the 2005 Kashmir earthquake was smaller than the 2015 Nepal earthquake, it was more destructive, which led to over 80,000 casualties. Besides the difference of population densities (194/km$^2$ for Nepal and 236/km$^2$ for Pakistan, according
to the World Bank), possible explanations lie within the rupture differences: for the 2005 Kashmir earthquake, the dip angle was $\sim 30^\circ$, which caused a very steep thrust-faulting event; the rupture was bilateral and propagated to the surface, which excited more surface waves; the major slip was constrained to be shallower than 10 km; and the short rise time (2–5s) led to severe ground shaking (Avouac et al., 2006; Parsons et al., 2006; Pathier et al., 2006). In contrast, the 2015 Nepal earthquake did not rupture to the surface with unilateral propagation; with a 10° dip, the rupture was concentrated in the depth range of 8 to 20km and the observed Stage 2 is high-frequency deficient, which could be a major reason why there was less ground motion (Figure 3.2).

### 3.5 Conclusion

Teleseismic P waves for the 2015 Nepal earthquake indicate that the rupture propagated for $\sim 160$ km at an azimuth of $\sim 130^\circ$ and an average rupture velocity of 2.9 km/s. We apply a global seismic network back-projection method to both low-frequency (LF, 0.05 to 0.2 Hz) and high-frequency (HF, 0.2 to 3 Hz) data, which provides good spatial resolution. The LF back-projection images suggest a three-stage rupture process: first, down-dip rupture at the nucleation area for the first 20s, then up-dip rupture which released most of the radiated energy from 20 to 40s, and a terminating stage with up-dip rupture northeast of Kathmandu. The total rupture lasted for $\sim 55s$. We observe a relatively compact rupture pattern that agrees well with the aftershock distribution. A multiple asperity model can explain the observed multi-stage rupture and the aftershock distribution. The apparent rupture velocity is significantly higher than S-wave speed during the Stage 2 rupture, but we cannot be sure that this represents the true rupture speed. Given the current plate convergence rate, the Mw 7.8 earthquake is smaller than expected (Ader et al., 2012). Therefore, detailed imaging of the rupture process is important for future hazard assessments.
Acknowledgments

The seismic data were provided by Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS). This work was supported by National Science Foundation grant EAR-1111111. The authors thank the editor, Andrew V. Newman and two anonymous reviewers for suggestions that improved the quality of this manuscript.

Figure 3.1: Upper panel: Station map. Lower left panel: Low-frequency P-wave displacement envelope functions plotted versus azimuth. The envelope functions are smoothed with a moving average window with a 2.5s half-width. The red line shows the expected rupture duration for a fault length of 160 km and an average rupture velocity of 2.9 km/s. Lower right panel: P-wave displacements versus directivity parameter (Ammon et al., 2005), assuming 130° as the rupture direction. The onset of the P wave begins at 0 s. Three different subevents are indicated with the colored lines.
Figure 3.2: Time-integrated images of back-projected P waves. Left panel: high-frequency (HF) back-projection image over 60s with maximum power normalized to 1. Right panel, low-frequency (LF) back-projection image over 60s with maximum power normalized to 1.
Figure 3.3: Snapshots of both low-frequency (LF) and high-frequency (HF) back-projections compared with the stacked source time functions.
Figure 3.4: Left panel: Rupture evolution of the Mw 7.8 main shock imaged with low-frequency (LF) back-projection. The different inferred rupture stages are shown as the yellow arrows, labeled 1 to 3. Right panel: LF back-projection images of Mw 6.7, M 6.8 and Mw 7.2 aftershocks overlaid with the mainshock rupture contour. The integrated energy for three aftershocks are differentiated with different colors: yellow for Mw 6.7, green for Mw 6.8 and blue for Mw 7.2. Left corner: a possible asperity model.
3.6 Supplementary Materials

Text S1: Directivity analysis and subevent analysis

Following Ni et al. (2005), we use the standard directivity formula:

\[ \delta t_i = \frac{L}{V_r} - \frac{L \cos(\theta_i - \theta)}{V_i} \] (3.1)

where \( L \) is the rupture length; \( V_r \) is the rupture speed; \( V_i \) is the apparent velocity for station \( i \); \( \theta_i \) is the \( i \)-station azimuth, and \( \theta \) is the rupture direction; and \( \delta t_i \) is the apparent source duration at station \( i \). For a first order estimation, we take \( V_i \) as the same for all the stations. Then the parameters to be modeled can be simplified as:

\[ \delta t_i = a - b \cos(\theta_i - \theta) \] (3.2)

For the 2015 Nepal earthquake, \( a \) is 55s and \( b \) is 10s when the rupture direction, \( \theta \), is taken as 130°. If we assume \( V_i \) is 16.1km/s, then the rupture length is 161 km with rupture speed 2.9km/s.

If multiple subevents happen during an earthquake, then Equation 3.1 can be modified as

\[ T^n_i = T_n - \frac{L_n \cos(\theta_i - \theta_n)}{V_i} \] (3.3)

where \( T_n \) is the \( n \)-th subevent onset time; \( L_n \) is the distance of the \( n \)-th subevent to the hypocenter; and \( \theta_n \) is the rupture direction of the subevent (Ammon et al., 2005; Zhan et al., 2014). The directivity parameter can be defined as \( x_i = -\frac{\cos(\theta_i - \theta_n)}{V_i} \). Therefore, different subevents can be identified as different straight lines from the aligned teleseismic P wave by the directivity parameter. Figure 1 (right panel) shows this approach applied to long-period P-waves from the Nepal earthquake. There are three well-defined subevents, indicated with the colored lines.

Text S2: Bootstrap analysis

To validate the robustness of the observed features in the back-projection images with respect to random variations in the contributing seismograms, we
perform a bootstrap analysis. For each bootstrap resample we randomly pick 62 seismograms from the original 62 (some are thus picked more than once, some not at all) and perform back-projection with this set of seismograms, retaining their original alignment. The procedure is repeated for 1000 times. In Figure 3 of the main paper, all of the LP peaks observed from 0 to 50s are statistically significant, i.e., they appear in over 95% of the bootstrap resampled images. However, the second rupture at the hypocenter area in the normalized LP image at 50 to 55s is seen in only about half of the bootstrap resamples, so we cannot be sure it is a real feature. Further investigation is warranted to determine why this feature appears and whether it is an artifact or truly indicative of late rupture near the hypocenter.

**Text S3: Depth phase analysis**

For large shallow earthquakes, wave interference and depth phases could potentially introduce bias in the back-projection images. To test whether these effects are important for our results, we designed a synthetic forward test and performed deconvolution for depth phases to confirm the robustness of the main features in our images and evaluate their uncertainties. For the synthetic test, we generate synthetic seismograms with a suite of point sources including depth phases (sP and pP) based on the GCMT solution for the mainshock. To mimic the kind of wave interference that may occur during a complex rupture, we include a suite of point sources that rupture sequentially at 5-s intervals at the ten energy burst peak locations labeled in Figure 2. Each point source generates a displacement pulse with a Hanning window shape and lasts for 10s. All the point sources share the same focal mechanism as the GCMT solution of the mainshock. The first four points and the last point have half of the amplitude of the other points to imitate the weak start and ending phase. Synthetic seismograms with depth phases are then generated for the same global stations, converted to velocity, filtered to the LF band and back-projected following the same procedure described in the paper.

The resulting synthetic seismograms and back-projection images are shown in Figure S9. The synthetic seismograms are aligned by azimuth and normalized by the P-wave radiation factor of the first source. Stations close to the nodal plane
(e.g., DAV and PMG) have abrupt changes and large amplitude variations due to the normalization operation. The red circles in the Input model column show where and how large the energy bursts of that window are. The Normalized column shows the normalized back-projection results with the synthetic seismograms. The results are generally good. No severe bias is seen in the results computed without considering the depth phases. The source duration at each point lasts a little longer than the input model. However, the multi-stage rupture is resolved clearly, which is the main feature discussed in the paper. This test validates our overall conclusions about multi-stage rupture but suggests a note of caution about the uncertainties. Bearing in mind that back-projection is an adjoint operation instead of a formal inversion, we would only expect the back-projected images to be approximations for the rupture process as discussed in the paper.

Another way to examine the influences of depth phases is to deconvolve the synthetic stick seismograms from the observed recordings and then perform back-projection with the deconvolved traces. We first tried this on the synthetic seismograms before applying it to real data. As seen in Figure S9, the rupture process is well resolved. However, due to the instability of the deconvolution, a new artifact is introduced in the 12.5s to 17.5s snapshot. Again, the overall multiple stage rupture is imaged clearly. The depth-phase deconvolution is then implemented on the real data and the results are shown in Figure S10. These images are similar to the results shown in Figure 3 in the paper. All of these tests suggest that depth phases do not bias our results very much, which may be due to the relatively good azimuthal coverage of the stations. However, further research on this topic is warranted.
Figure 3.S1: 0.1–1 Hz (MP) P wave velocity seismograms aligned and sorted by similarity to the final waveform stack. The onset of the P wave begins at 0s. The correlation coefficient of the initial 4 s ranges from 0.45 to 0.89. A polarity shift is allowed to achieve the maximum correlation coefficient. The maximum allowed time shift is ±3s, and the optimal time shifts range from -2.3 to 2.95 s. For back-projection, traces are weighted by their correlation coefficients are thus noisy or less coherent traces contribute less.
Figure 3.S2: Time-integrated image of back-projected P waves from the middle frequency band 0.1 to 1Hz (MF).
Figure 3.S3: Snapshots of MF back-projections compared with the stacked-source time function.
Figure 3.S4: Spatiotemporal evolution of LF back-projection, after projecting power onto a specific azimuth away from a reference point. For Stage 1, the reference point is the hypocenter with projecting azimuth 20°; for Stage 2, the reference point is the peak energy burst location of 20 to 25s with projecting azimuth 130°; for Stage 3, the reference point is the peak energy burst location of 40 to 45s with projecting azimuth 130°.
Figure 3.S5: Theoretical resolving power or array response function of the three frequency bands for the Mw 7.8 mainshock.
Figure 3.S6: HF and MF back-projection images of the Mw 6.7, M 6.8 and Mw 7.2 aftershocks overlaid with the mainshock rupture contours.
Figure 3.S7: Images of back-projected P waves for the Mw 7.2 aftershock for three frequency bands.
Figure 3.S8: Snapshots of the back-projections of the Mw 7.2 aftershock compared with the stacked source-time functions.
Figure 3.S9: Results of the multiple point source synthetic test. Left two panels show aligned synthetic seismograms. Right panels show input model and the snapshots of the back-projected synthetic seismograms and deconvolved synthetic seismograms.
Figure 3.S10: Comparison of back-projection images of deconvolved velocity recordings and the original analysis (Figure 3).
References


Chapter 4

Fault interactions and triggering during the 10 January 2012 Mw 7.2 Sumatra earthquake

Abstract

The 10 January 2012 Mw 7.2 Sumatra earthquake in the Wharton basin occurred three months before the great Mw 8.6 and Mw 8.2 earthquakes in the same region, which had complex ruptures and are the largest strike-slip earthquakes ever recorded. Teleseismic P-wave back-projection of the Mw 7.2 earthquake images a unilateral rupture lasting $\sim 40$ s without observable frequency dependency (low-frequency, 0.05–0.3 Hz, high-frequency, 0.3–1 Hz). In addition to radiation bursts during the Mw 7.2 mainshock, coherent energy releases from 50 to 75 s and from 100 to 125 s are observed about 143 km northeast of the mainshock rupture and landward of the trench. Analysis of globally recorded P-waves, in both 0.02–0.05 Hz velocity records and 1–5 Hz stacked envelope functions, confirms the presence of coherent sources during the time windows. The observed energy bursts are likely to be large early aftershocks occurring on or near the subduction interface. Both dynamic and static triggering could have induced these early aftershocks, as they initiated after the surface wave passed by, and the Coulomb stress perturbations
from the Mw 7.2 mainshock promote earthquakes in the observed locations. The earthquake sequence is a clear example of a seaward-intraplate strike-slip earthquake triggering landward-intraplate earthquakes in the same region, in contrast to previously reported normal-reverse or reverse-normal interactions at subduction zones.

4.1 Introduction

Large earthquakes often involve complex fault interactions, in which ruptures can jump fault boundaries and trigger earthquakes not directly connected to the mainshock. Both static and dynamic stress changes promote triggered earthquakes over a wide range of spatial and temporal scales (e.g., Harris, 1998; Kilb et al., 2000; Kilb, 2003; Lin and Stein, 2004; Toda et al., 2005). For earthquakes occurring in active tectonic regions, multiple fault segments can rupture simultaneously or sequentially (e.g., Wald and Heaton, 1994; Ji et al., 2002a; Wei et al., 2011; Yue et al., 2012; Uchide et al., 2013). At subduction zones, interactions among seaward-intraplate, landward-intraplate, and interplate faults often exhibit complex relationships (e.g., Lay et al., 2009; Li et al., 2009; Ye et al., 2012; Ruiz and Contreras-Reyes, 2015). The northwest Sunda arc subduction zone is of particular current interest because five Mw > 8 earthquakes have struck the region since 2004, beginning with the massive Mw 9.2 2004 Sumatra earthquake. The Indo-Australian plate deforms as multiple rigid units (Minster and Jordan, 1978), which might be responsible for the intricate intraplate and subduction earthquakes of the region (e.g., Robinson et al., 2001; Abercrombie et al., 2003; Antolik et al., 2006). Therefore, understanding interactions among these faults is of great interest as a potential clue to infer seismic properties of the Sunda megathrust.

In 2012, three large strike-slip earthquakes with Global Centroid Moment Tensor (GCMT) magnitude (Ekström et al., 2012) Mw 7.2, Mw 8.6, and Mw 8.2, reactivated old fracture zones and seafloor fabrics west of the Sumatra trench beneath the Wharton basin (Figure 4.1). Source models of the Mw 8.6 earthquake indicate the event ruptured at least four faults over 160 seconds (e.g., Wang et al.,
two hours later, and the Mw 7.2 earthquake occurred three months prior to the Mw 8.6 earthquake. The epicenters of the Mw 7.2 foreshock and Mw 8.6 mainshock are ∼30 km away (Figure 4.1) (International Seismological Centre, 2013). The hypocenter depths are 20.9 km and 26.3 km for the Mw 7.2 foreshock and Mw 8.6 mainshock (International Seismological Centre, 2013), while the GCMT centroid depths vary from 23.7 km to 45.6 km respectively (Ekström et al., 2012). Aside from the location differences, the initial 15 s of the far-field recorded P-waves of these two earthquake bear few similarities (Figure S1). The dissimilar initial P-waves and the location offsets of these two earthquakes suggest they might not have ruptured the same fault, as the fossil Wharton ridges can potentially form a set of strike-parallel fault systems in the region (Singh et al., 2011).

Here, we examine the Mw 7.2 earthquake with far-field global and array seismic records. Using P-wave back-projection, we resolve that the quake ruptured unilaterally for about 40 s, propagating northwest. The results reveal a similar rupture pattern for two frequency bands (low-frequency, 0.05–0.3 Hz, high-frequency, 0.3–1 Hz). In addition, at 50 to 75 s and 100 to 125 s, we image coherent energy radiation originating landward of the trench, about 143 km northeast of the mainshock rupture. The coherent energy radiators indicate triggering of large early aftershocks near the subduction interface. The imaged early aftershocks are also seen in global P-waves in both 0.02–0.05 Hz velocity records and 1–5 Hz stacked envelope functions. The detected earthquake sequence is a rare case of seaward-intraplate strike-slip earthquake triggering landward-intraplate earthquakes, suggesting long-range fault interactions in the region.

4.2 The 10 January 2012 Mw 7.2 Sumatra Earthquake

Great strike-slip earthquakes are rare and often involve perplexing rupture complexities (e.g., Ji et al., 2002a; Bouchon and Vallée, 2003; Dunham and
Archuleta, 2004; Vallée et al., 2008; Walker and Shearer, 2009; Robinson, 2011; Kennett et al., 2014). The 2012 Sumatra strike-slip earthquake sequence is particularly intriguing. First, the Mw 8.6 and Mw 8.2 quakes are the two largest strike-slip earthquakes that have been instrumentally recorded (Meng et al., 2012; Duputel et al., 2012). Second, the sequence happened beneath the Wharton basin, which is considered to be a diffuse deformation zone that is separating the Indian and Australian Plates (Wiens et al., 1985; Delescluse and Chamot-Rooke, 2007; DeMets et al., 2010). Third, the centroid depths, including a Mw 6.2 aftershock (2012/04/15), are all deeper than 20 km (Ekström et al., 2012; Wei et al., 2013), suggesting the earthquakes occurred primarily in the mantle. The Mw 7.2 Sumatra earthquake occurred west of Northern Sumatra and 423 km SW of Banda Aceh, Indonesia (2.43°N, 93.21°E, 20.9 km, International Seismological Centre (2013)).

The GCMT (Ekström et al., 2012) source solution has strike 103°, dip 81° and rake −173°. The finite-fault model from the USGS National Earthquake Information Center (NEIC), based on a W-Phase moment tensor solution, has strike 101.2° and dip 75.1° (Duputel et al., 2011; Hayes, 2011; Hayes et al., 2011; Ji et al., 2002b). The finite-fault model suggests most of the displacement on the fault slipped above the hypocenter, and the moment rate function derived from the finite-fault model shows the rupture lasted for about 40 seconds.

P-wave back-projection (e.g., Ishii et al., 2005; Walker et al., 2005; Walker and Shearer, 2009; Koper et al., 2011; Kiser and Ishii, 2012; Yagi et al., 2012) is useful for resolving the spatiotemporal evolution of complicated earthquakes because it makes few prior assumptions about the fault geometry. The method has been successfully applied in mapping out complex ruptures (e.g., Kiser and Ishii, 2011; Fan and Shearer, 2015), and in detecting subevents and early aftershocks (e.g., Yao et al., 2012; Kiser and Ishii, 2013). Here we apply a variation of the back-projection approach we used to image the 2015 Nepal earthquake (Fan and Shearer, 2015) to the 2012 Mw 7.2 Sumatra earthquake using teleseismic P-wave vertical-component velocity records. Because of the less ideal azimuthal coverage of stations for the Sumatra event, and to avoid potential back projection artifacts due to polarity flips caused by focal mechanism rotations or differences during the
rupture, we restrict our analysis to 89 stations located in Europe (Figure 4.2), and
the Hi-net array located in Japan (Figure S2, S3) (Okada et al., 2004; Obara et al.,
2005). We grid the potential sources at 5-km spacing, within the latitude range
0.6° to 4.2° and longitude range 91.4° to 95.0° (400 km by 400 km). Following Xu
et al. (2009) and Fan and Shearer (2015), we filter the P waves into two frequency
bands, 0.05–0.3 Hz (low frequency, LF) and 0.3–1 Hz (high frequency, HF), to
assess possible frequency dependence in the seismic radiation images. Details about
the back-projection method and data processing are presented in the Supplement.
No post-smoothing or post-processing is applied to the images.

From integrated back-projection images over ∼125 s (Figure 4.2), the earth-
quake does not show frequency-dependent rupture behavior. Most of the radiated
energy was released northwest of the epicenter with a concentration around the
GCMT centroid location. Both LF and HF back-projection snapshots imply a
northwest unilateral rupture (0 to 25 s is denoted as P1, 25 to 50 s is denoted as
P2, Figure 4.2b,d), lasting for ∼40 s (Figure 4.3a). The snapshots show a compact
rupture around the epicenter as suggested by the NEIC finite-fault model. From
50 to 75 s (AE1) and 100 to 125 s (AE2), additional coherent energy bursts are ob-
served landward of the plate boundary (Bird, 2003) and trench-axis (Bassett and
Watts, 2015a,b), well eastward of the mainshock rupture (∼143 km, Figure 4.2,
S3). The clear spatial separation between the first 50 s and the later energy bursts
indicates the landward energy release comes from aftershocks. This landward
region hosted three M5 earthquakes (Figure 4.1,4.2) with reverse faulting focal
mechanisms before and after the Mw 7.2 quake, suggesting the region is capable
of accumulating enough strain for earthquakes and the observed early aftershocks
might share similar focal mechanisms with the M5 earthquakes. Analysis of far-
field very-long-period seismic waves (first 5400 s, 2 mHz to 5 mHz) does not require
a second source to explain the data, which suggests the magnitudes of the early
aftershocks are significantly smaller than the Mw 7.2 mainshock (Figure S4).

Theoretical resolution kernels, back-projection of a Mw 5.9 earthquake in
the same region, and synthetic tests with depth phases indicate that these early
aftershock images are well-resolved and distinct from the mainshock (Figure S5,
S6). To further validate that they are true features, we examine globally recorded far-field P-waves (i.e., not just the European stations used for the back-projection) in two frequency bands beyond the bands used for back-projection (0.02–0.05 Hz, Figure 4.3; and 1–5 Hz, Figure 4.4). We first azimuthally align the 0.02–0.05 Hz globally recorded P-waves with empirical time-shift corrections (Houser et al., 2008), then stack the self-normalized records within each 2-degree azimuthal bin, and finally normalize the stacks with station numbers used for stacking of each bin (Figure 4.3b). The aligned stacks show clear polarity shifts for the first 40s at azimuths of $\sim 12^\circ$, $\sim 103^\circ$, $\sim 192^\circ$, and $\sim 283^\circ$, agreeing well with the moment tensor solutions (Ekström et al., 2012; Duputel et al., 2011). If the observed early aftershocks in Figure 4.2 can be approximated as point sources, globally recorded seismograms should show energy pulses at the predicted arrival times from the 1D velocity model used for back-projection (IASP91 (Kennett and Engdahl, 1991)).

With the peak energy burst location of 50 to 75 s (AE1), we then try to find the best rupture time that can explain the aligned seismograms. The theoretical arrivals from the inferred AE1 location with rupture times 52, 62 and 72 s are shown as colored crosses (Figure 4.3b). The early aftershock AE1 can explain the pulses in the stacks at most azimuths (Figure 4.3b), having an apparent duration of 20 s. The pulses associated with AE1 all share the same positive polarities, suggesting the focal mechanisms of the early aftershocks are likely reverse faulting rather than strike-slip faulting (Figure 4.3b). Assuming the positive polarities are shared by all the stations, we implemented HASH (Hardebeck and Shearer, 2002, 2003) to grid-search 5000 possible focal mechanisms of AE1. Limited by the station distribution, possible focal mechanisms are poorly determined with only polarity information. Resolved dips range from $18^\circ$ to $72^\circ$, with mean and median $50^\circ$ and $52^\circ$ respectively (Figure 4.3c). Strikes cannot be determined, and rakes are consistent with a reverse-faulting focal mechanism. For the focal mechanisms of the three M5 earthquakes (Figure 4.3b, insert), the predicted polarities of all the stations are positive, which further supports the likely reverse-faulting focal mechanism of the observed early aftershocks.

To test for coherent high-frequency energy bursts, we stacked envelope func-
tions from 207 globally recorded P-waves, filtered at 1–5 Hz (Figure 4.4a). To account for time shifts due to differing source locations, the envelope functions are time-shifted using the predicted arrivals of each target energy burst location before stacking (P1 and P2 for the first 50 s; AE1 and AE2 for 50 to 75 s and 100 to 125 s). The stacked function is then normalized with the station number. Although mainshock coda decay dominates the stacks, clear pulses in the stacked envelope functions can be identified in the 50 to 75 s window (Figure 4.4a). To reduce the strength of the mainshock coda, we experimented with restricting the stack to records from the subset of stations that are near the mainshock P-wave nodal planes. When using only stations within 20° of the W-phase CMT nodal planes (36 stations), the 50 to 75 s pulses are more distinct in the stacked envelope functions (Figure 4.4b) and an additional pulse near 110s can be identified (Figure 4.4b), confirming the 100 to 125s energy burst observed in Figure 4.2d. Two nearby GEOFON stations (≤ 5°) have S-wave arrivals in their smoothed envelope functions at the time of the two early aftershocks (Figure S6), preventing their detection of AE1 and AE2.

The observed energy bursts (Figure 4.2) are unlikely to be artifacts or due to water phases. Because of the 25 s stacking window length, the depth phases do not bias the results very much (Figure S7). Water phases would be insignificant for the Mw 7.2 Sumatra earthquake because of the large depth and the shallow water layer of the region (Wiens, 1989). Complex water reverberations usually show strong azimuthal dependence (Chu et al., 2011), which is not observed in this case (Figure 4.4b). Finally, global P-wave spectra of the earthquake do not show resonance-frequency signatures due to a finite water layer (Figure S8) (Zhan et al., 2014).

### 4.3 Faulting interactions

The detected early aftershocks (Figure 4.2, 4.3, and 4.4) could have been triggered by both dynamic and static triggering. Because the spatial separation between the Mw 7.2 mainshock and the early aftershocks is around 143 km, strong
near-field seismic waves can potentially activate the fault systems close to the trench, while the static stress changes could influence the systems as well (e.g., Kilb, 2003; Lin and Stein, 2004). Such correlated seismic activities between fault systems have been observed in other trench areas, for example, the Kuril Islands (Ammon et al., 2008), Japan trench (Lay et al., 2013), Philippine trench (Ye et al., 2012) and northern Tonga (Li et al., 2009).

Rapid-onset dynamic aftershock triggering usually coincides with large-amplitude seismic wave arrivals, for example, surface waves (e.g., Kilb et al., 2000). Assuming the local surface wave speed is $\sim$3 km/s, the distances and the timing of AE1 ($\sim$143 km, $\sim$50s) strongly correlate with the surface wave arrivals. The distances from AE1 to the later parts of the Mw 7.2 rupture range from $\sim$141 to $\sim$145 km (Figure 4.5), extending the shaking in the landward region around AE1 for about 40s.

The static stress changes could either promote or prohibit aftershocks, depending on the local Coulomb stress changes (e.g., Lin and Stein, 2004; Toda et al., 2005). With the NEIC finite-fault source model, we compute Coulomb stress changes induced on shallow reverse-faulting geometries (e.g., the nearby Mw 5.5, 2005/01/29, Mw 5.8 doublet, 2012/04/20, Figure 4.5, S9), as suggested by the polarity analysis (Figure 4.3). In Figure 4.5, shear stress, normal stress and Coulomb stress are calculated with Young’s modulus $8.0 \times 10^5$ Pa, friction coefficient 0.4, and receiving fault geometry $332^\circ/48^\circ/97^\circ$ (strike/dip/rake) of 2012/04/20b, Mw 5.8 (Figure 4.1) at a depth of 20 km. Stress contours of Figure 4.5 show areas where the stress changes exceed 10 kPa. The Coulomb stress increase positively correlates with where we locate the early aftershocks (Figure 4.5, S9), indicating static stress changes might be part of the driving forces triggering these events. However, the static Coulomb stress increase might be insufficient to induce fault failure for certain focal mechanisms (Figure S9a), leaving the importance of static stress triggering undetermined.

The 2012 one-year seismicity of the region (International Seismological Centre, 2013) shows a high density of seismicity where we observe the early aftershocks (Figure 4.1). The GCMT catalog (Ekström et al., 2012) shows a large concentration
of seismicity close to the Sumatra trench (Figure 4.1), showing the trench region is capable of hosting moderate earthquakes frequently. If the early aftershocks we observed in Figure 4.2, 4.3 share similar focal mechanisms of the 2005/01/29 Mw 5.5 earthquake or the 2012/04/20 Mw 5.8 doublet (Figure 4.1), the steep dipping angles of the nodal planes suggest it is unlikely for the early aftershocks to occur at the shallowest portion of the megathrust. More likely, these events are intraplate earthquakes near the subducting slab. The landward seismicities in Figure 4.1 are mostly aftershocks of the Mw 8.6 earthquake, suggesting it might be common for seaward-intraplate earthquakes to trigger landward-intraplate earthquakes in this region.

4.4 Conclusions

As revealed by teleseismic P-wave back-projection, the 10 January 2012 Mw 7.2 Sumatra earthquake unilaterally ruptured northwest for about 40 s. We apply regional seismic network (European array) back-projection to both low-frequency (LF, 0.05–0.3 Hz) and high-frequency (HF, 0.3–1 Hz) data to evaluate the energy radiation. Both frequency bands reveal similar rupture behavior, showing no frequency dependence. Besides the relatively compact rupture pattern around the hypocenter, there are coherent energy bursts at 50 to 75 s and 100 to 125 s observed landward of the trench. Combining global P-wave velocity records (0.02–0.05 Hz) and 1–5 Hz stacked P-wave envelope functions, we infer the observed energy bursts are early aftershocks rupturing within two minutes of the mainshock. Both dynamic triggering and Coulomb stress perturbations over the region may have contributed to the occurrence of the early aftershocks. Our observations indicate that seaward-intraplate strike-slip earthquakes can trigger landward-intraplate reverse-faulting earthquakes in this part of the Sunda arc. Because such early aftershocks may be difficult to detect within the coda of larger earthquakes, it is unclear how commonly such triggering occurs.
Acknowledgments

The seismic data from networks, AF, AU, CH, CZ, DK, ER, G, GE, GR, GT, IC, II, IM, IU, KC, KN, KR, KZ, MN, ND, NL, OE, PL, PS, RO, TW, XV, and ZM were provided by Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS) and National Research Institute for Earth Science Disaster Prevention in Japan (NIED). Coulomb 3 software was developed by S. Toda, R. Stein, J. Lin, and V. Sevilgen. The earthquake catalog was downloaded from the International Seismological Center (ISC). The bathymetry data was processed with the Generic Mapping Tools (GMT) (Wessel and Smith, 1991; Wessel et al., 2013). This work was supported by National Science Foundation grant EAR-1111111. The authors thank the Editor, Andrew V. Newman, reviewer, Claudio Satriano and one anonymous reviewer for suggestions that improved the quality of the manuscript.

Chapter 4, in full, is a reformatted version of the material as it appears in Geophysical Research Letters: Fan, W. and P. M. Shearer, Fault interactions and triggering during the 10 January 2012 Mw 7.2 Sumatra earthquake, Geophys. Res. Lett., 43, 1934–1942, doi:10.1002/2016GL067785, 2016. I was the primary investigator and author of this paper.
Figure 4.1: Tectonic setting of the 2012 Sumatra earthquake sequence. The 2012 Mw7.2, Mw 8.6 and Mw 8.2 strike-slip earthquakes are shown with their epicenters, GCMT locations, and GCMT focal mechanisms. Three reverse-faulting M5 earthquakes are shown with their centroid locations and GCMT focal mechanisms, which are ~100 km away from the Mw 7.2 earthquake. Black dots are earthquakes (EQ) from one-year (2012) ISC catalog of the Sumatra region (International Seismological Centre, 2013). Squares are events from the GCMT catalog of the region since 1979. The plate boundary is from Bird (2003), trench-axis is from Bassett and Watts (2015a,b), and the subduction geometry is from Slab 1.0 with 20 km separation (Hayes et al., 2012). The background bathymetry gradient is from Sandwell et al. (2014) and Garcia et al. (2014).
Figure 4.2: Back-projection results from European stations. (a), Low-frequency (LF, 0.05–0.3 Hz) time-integrated back-projection image. (b), Low-frequency (LF, 0.05–0.3 Hz) rupture evolution. (c), High-frequency (HF, 0.3–1 Hz) time-integrated back-projection image. (d), High-frequency (HF, 0.3–1 Hz) rupture evolution. Two energy bursts from 0 to 50 s are denoted as P1 and P2. The two energy bursts shown at 50 to 75 s and 100 to 125 s are likely from one or more triggered events, shown in the maps northeast of the trench, and not physically connected to the mainshock rupture. They are denoted as AE1 and AE2. The insert in (b) shows the stations used for back-projection of the Mw 7.2 earthquake.
Figure 4.3: (a), Stacked BP energy function at 0.3–1 Hz. The function is a self-normalized stack of all the back-projected seismograms in the source region. (b), Aligned velocity seismograms (low-frequency, 0.02–0.05 Hz) from global stations with predicted travel-time arrivals of AE1 in Figure 4.2. Station map shown as left-upper corner insert, the polarities of the Mw 7.2 strike-slip GCMT focal-mechanism shown as middle insert, and the polarities of the M5 reverse-faulting GCMT focal-mechanisms shown as right insert (Red, positive polarity; Blue, negative polarity). The records are sorted by azimuth. Colored crosses show predicted arrival with AE1 location at three different rupture times. (c), Dip distribution of grid-searched focal mechanisms from the HASH algorithm.
Figure 4.4: (a), Stacked envelope functions of P1, P2, AE1 and AE2 in Figure 4.2, station map plotted as right-upper corner insert. The P-wave seismograms are filtered at 1–5 Hz. The envelope functions are calculated with a standard Hilbert transform without smoothing. (b), Same as (a) with stations within 20° azimuth of the Mw 7.2 mainshock W-phase nodal planes. Peaks related to AE1 are labeled as +. Peaks related to AE2 are labeled as ×.
Figure 4.5: Horizontal cross sections of shear stress (a), normal stress (b) and Coulomb stress (c) changes at 20km. The stress changes are calculated from the mainshock NEIC finite-fault model on a target reverse fault $332^\circ/48^\circ/97^\circ$ (strike/dip/rake) of 2012/04/20b, Mw5.8, using the Coulomb 3 software (Lin and Stein, 2004; Toda et al., 2005). Distances from AE1 to Mw 7.2 epicenter, centroid location, P1, and P2 ranges from $\sim$141 km to $\sim$145 km.
4.5 Supplementary Materials

**Text S1: Back-projection data and method**

Following Fan and Shearer (2015), we first align seismograms using theoretical P-wave travel times calculated with a 1D velocity model (Figure S2, IASP91 (Kennett and Engdahl, 1991)). We align the seismograms by cross-correlating the initial 4 s of the P-waves, allowing ±3 s shifts for European stations and Hi-net separately (Houser et al., 2008). The alignment is applied to neutralize 3D velocity structure influence. No polarity flips are allowed during the alignment, as the stations within each network share the same P-wave polarities (Figure S2). The potential sources are gridded laterally at 5-km spacing, within the latitude range 0.6° to 4.2° and longitude range 100.91.4° to 95.0° (400 km by 400 km) at hypocentral depth (19 km). Non-linear stacking approaches, like Nth root stacking (Rost and Thomas, 2002), can improve spatial resolutions of back-projection images at the cost of losing absolute amplitude information (Xu et al., 2009). In this study, we use \( N = 4 \) as suggested in Xu et al. (2009).

For the European array, 89 vertical-component velocity records from global broadband stations distributed by the Data Management Center (DMC) of IRIS (Figure S2) are used to image the Mw 7.2 earthquake. The epicentral distances of the European stations range from 39° to 87° with azimuth ranging from 293° to 354° (Figure S2). For Hi-net, 703 vertical-component velocity high-sensitivity records from National Research Institute for Earth Science Disaster Prevention in Japan (NIED) are used to image the Mw 7.2 earthquake (Figure S2). The epicentral distances of the Hi-net stations range from 45° to 62° with azimuth ranging from 37° to 49° (Figure S2). The Hi-net network is close to the ∼ 12° nodal plane of the Mw 7.2 earthquake, which limits the resolving power of the array for this special case. The Australian array is divided into two sub-networks by the ∼ 103° nodal plane, leaving only 11 useable stations after removing the ones too close to the nodal plane. Therefore, the European array provides the best array configuration to investigate this earthquake. Back-projection results of Hi-net are shown in Figure S3, and we did not use the Australian array to image this earthquake.
To assess possible frequency dependence in the seismic radiation, we examine two different frequency bands, 0.05–0.3 Hz (low frequency, LF) and 0.3–1 Hz (high frequency, HF). Back-projection for each band uses the same alignment derived from the LF data for the European array and HF data from Hi-net (Figure S2). For each array, the records are normalized, weighted by their average correlation coefficients obtained from the cross-correlation alignment, and also inversely scaled by the number of contributing stations within 3 degrees. This procedure evens out different instrument gains, downweights the noisy records and prevents biases from dense local arrays.

**Text S2: Depth phase analysis**

Following Fan and Shearer (2015), we generate synthetic seismograms for the European stations used for back-projection in the paper, based on source locations from the first two LF peaks in Figure 2. Synthetic seismograms are generated with the sources occurring consecutively at 25 s intervals and include depth phases (sP and pP) based on the GCMT solution for the mainshock (P1 and P2 share the same CMT solution). A 0.3-Hz Ricker wavelet is implemented to generate a displacement pulse for each source, and the synthetic seismogram at each station is a sum of the two sources. The synthetic seismograms with depth phases are then converted to velocity, and back-projected the same way as for the recorded data.

The back-projection results from the synthetic seismograms are shown in Figure S7. The red circles in the Input model row show the locations and amplitudes of the energy bursts in each time window. Since we designed the sources to have the same strength, the input circles all have the same size. The Absolute and Normalized rows show the back-projection results with the synthetic seismograms. The resulting images are centered near the true source locations, without much bias due to the depth phases, although the energy contours are broadened and elliptical due to the array geometry. No artifacts are seen landward of the trench in this synthetic test, supporting our interpretation that the late-arriving energy bursts observed in Figure 2 are due to real events.
Figure 4.S1: Comparison of the first 15 s of selfscaled P-waves (0.05–0.3 Hz) of the Mw 7.2 earthquake and Mw 8.6 earthquake (Figure 1).
Figure 4.S2: P-wave velocity seismogram alignments. 0.05–0.3 Hz (LP) for European array (Left panel) and 0.3–1 Hz (HP) for Hi-net (right panel). The onset of the P wave begins at 0s. The records are sorted by average cross-correlation coefficients for the two arrays respectively. Array locations and the station polarities with respect to the Mw 7.2 GCMT focal-mechanism are shown as inserts (Red, positive polarity; Blue, negative polarity).
Figure 4.S3: Time-integrated back-projection image from Hi-net stations with high-frequency data (HF, 0.3–1 Hz). The landward energy burst is observed ~138 km away from the Mw 7.2 epicenter.
Figure 4.S4: Comparison between observed very-long-period data (200s to 500s) and synthetic seismograms predicted from the GCMT solution for the 10 January 2012 Mw 7.2 Sumatra earthquake. V stands for the vertical component and T stands for the transverse component.
Figure 4.5: (a) and (b), LF and HF theoretical resolving powers or array response functions of the European array for the Mw 7.2 mainshock. The theoretical resolution is evaluated by randomly assigning a single recorded P wavetrain to all the stations, then back-projecting these traces. (c) and (d), Time-integrated back-projection images for a Mw 5.9 earthquake in the same region (2012/04/20, 23:14:30, focal mechanism seen in (d), insert, Red, positive polarity; Blue, negative polarity). Stations ((c), insert) used to image this Mw 5.9 earthquake are similar to those used for the Mw 7.2 earthquake. In both frequency bands, the Mw 5.9 earthquake results spatially span a similar area as the theoretical resolving powers of the Mw 7.2 earthquake. There are no landward artifacts observed in the Mw 5.9 earthquake images, indicating the observed landward energy bursts are real physical processes.
Figure 4.S6: Smooth envelope functions of GE.LHMI and GE.GSI. We first filter the three component records of the two stations at 1–5 Hz, then Hilbert transform the filtered seismograms, and finally calculate the envelope functions. The envelope functions are fitted with cubic smoothing splines (*csaps* of MATLAB, with smoothing factor 0.01). The first P and S arrivals based on IASP91 are shown as vertical lines.
Figure 4.S7: Results of the multiple point source synthetic test. Top panel shows input model. Lower panels show time snapshots of the back-projected image of the synthetic seismograms (Absolute: normalized with the maximum power over the entire 125 s; Normalized: normalized with the maximum power within each 25-s time window).
Figure 4.S8: Stacked P-wave velocity spectra of the 10 January 2012 Mw 7.2 Sumatra earthquake. The Welch’s power spectral density estimate (Oppenheim et al., 1989) is utilized for the first 150s of P-wave velocity records with 75s overlap. The synthetic seismograms are generated with two input point sources (Figure S6) as described in Text S2 with a 0.3-Hz Ricker wavelet as the source time function. Stations in Figure 2b and Figure 3b are used for the calculations. The resonance frequencies are computed assuming a 4.5 km water layer and 1.5 km/s P-wave speed \( f = \frac{V_p}{2H} \), where \( V_p \) is the P-wave velocity and \( H \) is the ocean depth (Zhan et al., 2014). Assuming the local \( V_p \) is 6 km/s, the rupture speed \( V_r \) is 2 km/s, and the rupture length is 50 km/s for the Mw 7.2 mainshock, then the first corner frequency is 0.12 Hz (Eq. 9.28, Shearer (2009)), which can explain the first peak of the stacked spectra. The later peaks are likely due to the depth phases (pP and sP).
Figure 4.S9: Horizontal cross sections of Coulomb stress changes at 20 km for the three reverse-faulting M5 earthquakes in Figure 1. The stress changes are calculated from the mainshock NEIC finite-fault model on target reverse faults with the Coulomb 3 software (Lin and Stein, 2004; Toda et al., 2005). The colorbar and legends are the same as in Figure 5. Event IDs and the receiving fault geometries are listed in the upper left corners of each panel.
References


Chapter 5

Multiple branching rupture of the 2009 Tonga–Samoa earthquake

Abstract

Several source models have been proposed to explain the enigmatic 2009 Tonga-Samoa earthquake. The long-period data require a composite source model and can be fit with a normal-faulting subevent followed by one or more reverse-faulting subevents. The short-period data, in contrast, indicate a more compact rupture pattern around the epicenter. The lack of a unified source model reflects the complexity of the event. We analyze the spatiotemporal evolution of this earthquake with P-wave back-projection from globally distributed stations in different frequency bands (low frequency: 0.05–0.2 Hz, high frequency: 0.2–2 Hz) and a multiple moment-tensor inversion. The rupture propagation revealed by back-projection exhibits frequency-dependent behavior, with two branches of high-frequency-enriched bilateral rupture around the epicenter and a high-frequency-deficient rupture branch at the subduction interface. A composite source model with one Mw 8.0 normal-faulting earthquake east of the trench-axis (seaward) followed by one Mw 8.1 reverse-faulting earthquake along the subduction interface west of the trench-axis (landward) can explain the very-long-period data (200–500 s). Combined with high-resolution swath bathymetry data, the back-projection
images show that the azimuth of rupture branches east of the trench-axis were controlled by the geometry of bending related faults on the Pacific plate, and that the rupture branch west of the trench-axis may correlate with the along-strike forearc segmentation. The rupture along the subduction interface was triggered by the seaward rupture and a partially subducted normal fault which may have played a key role in facilitating the triggering. The apparent normal-reverse faulting interactions pose a higher seismic risk to this region than their individual strands at the northernmost corner of the Tonga subduction zone.

5.1 Introduction

The northern end of the Tonga subduction zone is tectonically complex and a prominent example of tearing plates (Wilson, 1965; Millen and Hamburger, 1998; Govers and Wortel, 2005). Both seismic observations and geodynamic modeling show the Pacific plate lithosphere experiences downwarping and tearing as it subducts into the northernmost segment of the Tonga subduction zone (Millen and Hamburger, 1998; Govers and Wortel, 2005). Both normal-faulting and reverse-faulting events occur in the northern corner that cannot be explained by simple pulling and bending subduction dynamics (Figure 5.1). The termination zone has a linear west-east trend that extends ~ 300 km parallel to the relative Pacific plate motion to the Australian plate. The tearing-induced faults, newly formed outer-rise faults, in conjunction with the pre-existing fabric of the Pacific Plate, produce complicated fault networks in the region. Residual bathymetry gradients clearly reveal the complexities of this system (Figure 5.1) (Bassett and Watts, 2015a,b). East (seaward) of the trench-axis, bending related faults exhibit complex along-strike variations in spacing and strike. Bend faults are approximately trench-parallel (or N-S) with a mean spacing of ~ 20 km south of -16.5°, and are gradually rotated to be trench-perpendicular (E-W) with a mean spacing of ~ 5 km north of -15.3°.

The 29 September 2009 Tonga-Samoa earthquake occurred offshore of the northern termination of the Tonga subduction zone. The Global Centroid Moment
Tensor (GCMT) (Ekström et al., 2012) solution modeled it as a single point source, with strike 346°, dip 62° to the east and rake -63°. The finite-fault model from the USGS National Earthquake Information Center (NEIC) favors a nodal plane with strike 342.5° and dip 57.1°, with bilateral rupture to the surface (Hayes, 2011; Hayes et al., 2011; Ji et al., 2002). The relatively poor waveform fits in the finite-fault modeling indicate a single-fault model is insufficient to explain the observations and several more complex models have been proposed. It was first reported by Li et al. (2009) that the earthquake was composed of both a normal-faulting main shock and an additional “hidden” triggered thrust event, based on analysis of teleseismic body waves, surface waves and local strong motion data. Lay et al. (2010) found evidence for a normal-faulting main shock, closely followed by two triggered reverse-faulting events, based on analysis of R1 source time functions (STFs), array back-projection and surface-wave modeling. Beavan et al. (2010) proposed a two-fault model, with a slow-slip event at the subduction zone before the normal-faulting main shock, based on GPS near-field observations and Deep-ocean Assessment and Reporting of Tsunamis (DART) buoy observations. Kiser and Ishii (2012) argued there were at least two subevents composing the earthquake, which were spatially focused around the epicenter, based on USArray and Hi-Net combined array back-projection. Duputel et al. (2012) and Nealy and Hayes (2015) independently confirmed the normal-reverse doublet with W-phase inversion, agreeing well with the focal mechanisms, while showing variations in the locations and timing of the two events. These studies make clear that the Tonga-Samoa earthquake was complicated, with at least two major subevents, but there is not yet consensus on the subvent locations and other details, which hampers understanding of the rupture dynamics and its relation to structures in the region.

These uncertainties motivate this study. We comprehensively investigate the Tonga-Samoa earthquake rupture process with broadband far-field seismic observations, quantify the resolved source model uncertainties with various tests, and explore possible physical causes of the observed rupture pattern in conjunction with local tectonics. For the high- and low-frequency bands, we apply P-wave back-projection to image the spatiotemporal behavior of the earthquake. For
very-long-period data, we model the earthquake with multiple point sources with different moment tensor solutions to fit the initial 5400 s of the waveforms. To evaluate intrinsic ambiguities in the imaging approaches (e.g., back-projection and moment tensor inversion) and to understand the variations among the available source models, we apply a number of tests to assess the uncertainty and resolution of back-projection and explore the spatiotemporal resolving limits of moment tensor inversion by grid-searching the model space. We show that all the data for the 2009 Tonga–Samoa earthquake can be reasonably explained with a unified multiple branching rupture model and quantify its uncertainty and resolution. Comparisons with seafloor bathymetry suggest that the multiple branching rupture is a natural consequence of the fault-geometry and the fore-arc material segmentation.

First introduced by Ishii et al. (2005) for the 2004 Sumatra earthquake, the back-projection method has proven useful in studying complex ruptures because it makes few a priori assumptions about the fault geometry. High-frequency regional array data, such as Hi-net or USArray, are often used for back-projection imaging because of their good waveform coherence (e.g., Kiser and Ishii, 2011; Wang et al., 2012; Meng et al., 2011; Koper et al., 2011; Satriano et al., 2012; Fan and Shearer, 2016). However, the superior azimuthal coverage of globally distributed stations can provide better resolution than regional arrays (e.g., Walker et al., 2005; Walker and Shearer, 2009; Yagi et al., 2012; Okuwaki et al., 2014; Fan and Shearer, 2015). Following Fan and Shearer (2015), we analyze both a high-frequency band (0.2 to 2 Hz), similar to that used most often in prior back-projection studies, and a low-frequency band (0.05 to 0.2 Hz) to provide a more complete description of the seismic radiation. In addition, we solve for centroid moment tensor (CMT) models (Gilbert, 1971; Dziewoński et al., 1981; Ekström et al., 2012) to constrain the average fault geometry and moment release, based on fits to very-long-period (200s–500s) body and surface waves.

Our results suggest that the rupture had at least three fault branches, with an initial northwestward rupture branch along a curved normal fault, followed by a southeastward rupture branch along the same fault, and a southward final rupture branch at the subduction interface. The first two branches can be charac-
terized as a Mw 8.0 bilateral-rupturing normal-faulting earthquake, which agrees with GCMT, single-point W-phase CMT (Duputel et al., 2012b; Nealy and Hayes, 2015) and NEIC finite-fault solutions. The third rupture branch is depicted by a Mw 8.1 reverse-faulting event with a shallow dip angle, which agrees with other double-CMT models (Li et al., 2009; Duputel et al., 2012; Nealy and Hayes, 2015). Possible early aftershocks are triggered immediately after the mainshock within two minutes. Integrating our results with the aftershock distribution, currently available source models, and local bathymetry data, we conclude the earthquake consisted of two M8 subevents, with at least three fault segments involved. The reverse-faulting rupture branch was triggered by the rupture branches east of the trench. Across the trench, a partially subducted normal fault is revealed by high-resolution swath bathymetry data, which might link faults across the trench axis, and facilitate the initiation of the second subevent at the megathrust. Finally, we discuss the frequency-dependent rupture behavior of this event and compare our results with source models of the 2006–2007 Kuril islands earthquakes doublet to infer general characteristics of this type of multi-fault earthquake sequence.

5.2 Method and Data

5.2.1 Back-projection

Assuming that the P waves are representative of the earthquake rupture process with little distortion during wave propagation, the back-projection method aligns, shifts and stacks seismograms to extract coherent signals which serve as approximations for the radiation from hypothesized source locations relative to the hypocenter. For a grid of possible source points, the aligned seismograms will constructively interfere for true source locations, or destructively interfere for other points. We closely follow the method described in Walker et al. (2005) and Fan and Shearer (2015), with $N$th root stacking (Xu et al., 2009) to suppress noise. Seismograms are first aligned using theoretical P-wave travel times calculated with a 1D velocity model (IASP91 (Kennett and Engdahl, 1991)). We then apply empirical time shifts to account for 3D velocity structure by aligning the initial 5 s of the P ar-
rival (Houser et al., 2008) with allowed time shifts of ±4 s, assuming it comes from
the hypocenter (Figure S1). No polarity flips are allowed during the alignment,
since all the stations used for back-projection share the same P-wave polarities
(Figure 5.2). In principle, the source locations could be three-dimensional, but
because of poor depth sensitivity, we focus only on lateral variations in this study
and grid potential source locations as functions of latitude and longitude with
depth fixed to the hypocentral depth (18 km). Non-linear stacking approaches,
like Nth root stacking (Rost and Thomas, 2002), has been successfully applied to
back-projection to sharpen signals and suppress noise (Xu et al., 2009), at the cost
of absolute amplitude information. In this study, we use $N = 4$ as suggested in
Xu et al. (2009) and discussed in McFadden et al. (1986).

The potential sources are gridded at 5-km spacing, within the latitude range
-17.28° to -13.69° and longitude range -173.96° to -170.23° (400 km by 400 km). We
use vertical-component velocity records from global broadband stations distributed
by the Data Management Center (DMC) of IRIS (Figure 5.2). One hundred and
sixteen stations with high signal-to-noise ratios are hand-selected to perform the
back-projection. The epicentral distances are limited to from 44° to 90° to avoid
waveform complexities introduced by mantle discontinuities. The azimuth ranges
from 10.6° to 353.38° (Figure 5.2), which greatly enhances the back-projection
spatial resolution and suppresses artifacts. To investigate possible frequency de-
pendence in the seismic radiation, we examine three different frequency bands,
0.05–0.2 Hz (low frequency, LF), 0.1–1 Hz (middle frequency, MF), and 0.2–2 Hz
(high frequency, HF). Back-projection for each band uses the same alignment, ob-
tained from cross-correlating the first 5s of the LF data (Figure S1). The LF and
HF results are described in the main paper; the MF results are presented in the
Supplement.

To avoid biasing the back-projection results, the records are normalized
to unit amplitude to remove variations caused by site effects, radiation pattern,
and different instrument gains. They are also weighted by their average correlation
coefficients with the waveform stack during cross-correlation (to down-weight noisy
records) and inversely by the number of contributing stations within 5 degrees (to
reduce the relative influence of large numbers of stations in a single region). No post-smoothing or post-processing is applied to the images.

5.2.2 Multiple Moment Tensor Inversion

The moment tensor (MT) is a point-source representation for earthquakes with six independent elements (e.g., Gilbert, 1971). It is most applicable when the earthquake dimension and duration are smaller than the wavelengths and period of the seismic waves considered (Dziewoński et al., 1981; Ekström et al., 2012). In this parameterization, the ground movement (e.g., velocity) is linearly linked to the six moment-tensor components by Green’s functions. Due to the nature of the long-period and long-wavelength seismic waves used for inversion, the MT solution, including the time and location, are averaged descriptions of earthquakes. This is termed the centroid moment tensor (CMT) and is often generalized to include a source duration as well as the centroid time (Dziewoński et al., 1981; Ekström et al., 2012). When an earthquake ruptures a very large area or is composed of a series of subevents, multiple source points with independent CMTs can be used to model the observations (e.g., Tsai et al., 2005; Li et al., 2009; Duputel et al., 2012; Nealy and Hayes, 2015).

As discussed in Section 7.1, a single CMT is insufficient to represent the Tonga-Samoa earthquake. Consequently, we experiment with using two separate CMTs to model the data, which consist of the first 5400 s of long-period vertical and transverse-component records, sampled at 10 s per sample (Figure S2). Green’s functions are calculated with normal modes (Masters and Widmer, 1995) and the Preliminary Reference Earth Model (PREM) (Dziewoński and Anderson, 1981). We apply a grid-search approach to explore the model space of potential source locations, durations and initiation times. For every source combination, two CMTs are inverted and the waveform misfit is recorded, assuming no isotropic source components. Because shallow source depths are poorly resolved with long-period data, source depths are set to the GCMT depth of 12 km and only lateral locations are grid-searched over 30 km spaced grids within -173.9° and -171.6° in longitude and -14.2° and -16.3° in latitude (Figure S2). Source durations are grid-searched
from 30 to 160 s for the first source and 40 to 160 s for the second one, assuming a symmetric triangular source-time function, and initiation times are grid-searched from 0 to 30 s for the first source and from 40 to 100 s for the second source. Since only a small decrease in the overall data misfit can be achieved with more than two point sources, we restrict our analysis to two CMTs.

5.3 Results

Integrated back-projection images over the $\sim$160 s duration of the earthquake show clear frequency-dependent rupture behavior (Figure 5.2). The LF back-projection shows two distinct rupture zones, one large energy burst locating northwest of the epicenter east of the trench, and another strong energy burst occurring landwardly west of the trench. If the landward rupture happened at the subduction interface, the LF energy burst within the subduction zone does not extend deeper than $\sim$40 km (Hayes et al., 2012). On the other hand, most of the radiated energy appears in the seaward region of the trench-axis for HF back-projection. The bulk of the HF energy was released northwest of the epicenter and there is not much HF energy release imaged west of the trench-axis. Besides the large energy concentration northwest of the epicenter location, two noticeable radiation sources are seen southeast of the epicenter in the integrated HF image (Figure 5.2).

More rupture propagation details can be inferred from the stacked source-time functions (SSTF) and the snapshots (Figure 5.2). Compared to the LF stacked source-time function, two strong HF energy episodes were released from 0 to 20 s and from 60 to 80 s. Two energy episodes, 80 to 100 s and 120 to 140 s, are discernible in the low-frequency SSTF. The normalized LF integrated-power images (normalized with the maximum power within each 20 s window) reveal rupture around the epicenter for the first 20 s, a northwestward rupture followed by an apparent eastward rupture across the trench axis from 20 to 40 s, a southeastward rupture in the seaward region from 40 to 80 s, and a southward rupture in the subduction zone from 40 to 60 s. Bilateral seaward rupture is also
observed in the normalized HF integrated-power images. The apparent westward rupture across the trench axis is also imaged by the HF back-projection from 20 to 40 s. The southward rupture along the subduction interface (40 to 60 s) does not radiate much high-frequency energy. A coherent seismic radiator is imaged by both LF and HF back-projections from 80 to 100 s northwest of the epicenter, which is landward and spatially close to the westward rupture episode crossing the trench axis from 20 to 40 s. Another coherent radiator is located southwest of the epicenter west of the trench-axis from 120 to 160 s, corresponding to the discernible energy episode in the low-frequency SSTF. The radiator is also visible in HF back-projection images, but is less coherent.

The overall rupture process is summarized in Figure 5.3 and characterized by different rupture strands, which we term Branches. Bilateral rupture occurred in the seaward wall of the trench, northwestward for the first 40 s (Branch 1) and southeastward from 40 to 80 s (Branch 2). From \( \sim 27 \) to 60 s (Figure S3), Branch 3 of the earthquake broke the subduction interface and ruptured from north to south no deeper than \( \sim 40 \) km (Slab 1.0 inferred depth \cite{Hayes2012}). The peak locations of the energy bursts are denoted as P1 to P5 for the first 60 s (Figure 5.3). Energy bursts P6 to P8 occurred within two minutes of the earthquake initiation and are possible triggered early aftershocks (Figure 5.3). As part of Branch 2, the energy burst from 60 to 80 s (P6) is identified as a reverse-faulting early aftershock, since the P-wave polarities of pulses associated with P6 are opposite to the other energy bursts east of the trench (P1, P2, and P5, Figure 5.4). The coherent seismic radiators, 80 to 100 s (P7) and 120 to 140 s (P8) may reflect the last stage of Branch 3 rupture, lasting until 160 s. Alternatively, they may also be very early aftershocks triggered by the earlier rupture. In either case, Branch 3 rupture experienced more than one rupture episode. For comparison, the Lay et al. (2010) array back-projection results indicated possible subevents around 52 s and 91 s, which are probably Branch 3 in Figure 5.2 and 5.3. They also observed an early aftershock occurring around 118 s, which might correspond to P8 (Figure 5.3).

Additional evidence for the energy bursts, aftershocks and their focal mechanisms is provided by aligned seismograms and P-wave polarities of the associated
pulses. The LF velocity seismograms are aligned by azimuth in Figure 5.4. Each trace is normalized in two time windows separately (0 to 30 s and 20 to 160 s) to enhance the weak signals in the first 30s. Using the locations of the energy bursts from Figure 5.3, we then try to find the best rupture time to explain the coherent pulses for all the azimuths. The theoretical arrivals (IASP91 (Kennett and Engdahl, 1991)) of the inferred source and aftershock locations are shown as colored bands in Figure 5.4. The eight energy bursts can explain the pulses in the traces at most azimuths. Pulses associated with P1, P2 and P5 share coherent negative polarities, confirming the normal-faulting focal mechanism of the Branch 1 and Branch 2 ruptures. In contrast, pulses associated with P3 and P4 share coherent positive polarities, supporting the reverse-faulting focal mechanism of the Branch 3 rupture. Positive polarities of the pulses associated with P6, P7 and P8 suggest reverse-faulting focal mechanisms. Because of the different focal mechanisms of P5 and P6, the Branch 2 rupture is unlikely to involve continuous rupture from 40 to 80 s and may have had one southward rupture segment (P5, 40 to 60 s), which triggered an aftershock on an adjacent fault (P6, 60 to 80 s). The observed coherent positive polarities for the later event are unlikely to be the sP phase, as a polarity shift is expected in North American stations (azimuth 0° to 60°), if the pulses were sP instead of direct P waves (Figure S4). They are also unlikely to be the pP phase, as there are no identifiable P phases proceeding the observed coherent phases, which should arrive ∼ 6 s ahead of pP phases (Figure S4).

To further verify the existence of P6 to P8, we filtered 935 globally recorded P-waves into two frequency bands (0.5 to 2 Hz and 1 to 5 Hz), and then stacked the envelope functions of the records (Figure 5.5) using the approach of Fan and Shearer (2016). For each target energy burst or possible early aftershock location (P1 to P8), the envelope functions are time-shifted using the predicted arrivals of the burst location before stacking to account for arrival-time differences due to different source locations. The stacked function is then normalized with the station number. From the stacks in both frequency bands, the majority of the quake energy was released before 60s. Clear pulses of P6 to P8 can be identified in the 0.5 to 2 Hz stacked envelope functions (Figure 5.5). P-wave coda decay
dominates the 1 to 5 Hz stacks, yet P6 and P8 can be easily identified.

Figure 5.6 shows results from our double-CMT inversion to model the very-long-period data (200 to 500 s), compared to previously published source models. We performed a grid-search in the seaward region east of the trench for the first subevent location and in the landward region west of the trench for the second subevent location (Figure S2). Our best-fitting solution for the first subevent is located ∼ 50 km north of the epicenter with centroid time and duration of 35 s and 70 s respectively. The first subevent shares similar fault geometries with the GCMT solution with strike 344°, dip 72° to the east and rake -68°. The estimated moment magnitude is M_w 8.0 for the first subevent. The second subevent locates on the shallow megathrust ∼ 80 km southwest of the epicenter with moment magnitude M_w 8.1. The second subevent initiated 80 s after the first subevent and ruptured for about 80 s. The centroid time (∼ 120s) and the focal mechanism of the second subevent are consistent with other two-point source solutions (Li et al., 2009; Duputel et al., 2012; Nealy and Hayes, 2015). With strike 183°, dip 15° to the east and rake 40°, the second subevent fault geometry corresponds well to the subduction interface. Transverse recordings in the North America region require the second source to explain the waveforms (Figure 5.6 and Figure S6), while the vertical recordings are insensitive to the existence of the second source (Figure S6 and Figure S7). For comparison, published CMT solutions are listed in Table 7.S1.

In Lay et al. (2010), the source model of the 2009 Tonga-Samoa earthquake consists of a Mw 8.1 normal-faulting mainshock and two Mw 7.8 reverse-faulting subevents (Figure 5.6, Table 7.S1), rupturing from ∼49–89 s and ∼90–130 s, respectively. Together, the moment of these two Mw 7.8 subevents totals to a Mw 8.0 earthquake. The two triggered reverse-faulting subevents likely correspond to the observed Branch 3 in Figure 5.3. The locations of the subevents in Lay et al. (2010) are offset from our best-fitting CMT solutions, with the initial normal-faulting event located southeast of our corresponding CMT (and south of the GCMT) and the later two events located east and northeast of our second event (Figure 5.6). In Duputel et al. (2012), the source model consists of a Mw 8.03 normal-faulting mainshock and a Mw 8.05 reverse-faulting subevent (Table 7.S1). The time delays
of the two subevents are 32.4 s and 105.4 s, with durations 53.6 s and 40 s respectively. The two subevents of Duputel et al. (2012) are located close to our subevents with depths of 15.5 km. Nealy and Hayes (2015) experimented with fixing the centroid locations of the two subevents, grid-searching the centroids, constraining the second subevent to be west of the trench while grid-searching the other centroids, and matching the time-delays and depths of Lay et al. (2010) while grid-searching the other centroids (Figure 5.6, Table 7.S1). They found the constraints do not notably influence the results, which suggests the spatial and temporal resolution of the very-long-period data are limited, while the resolved focal mechanisms are robust. Their two-point-source solution (constraining the second subevent to be west of the trench) is consistent with a Mw 8.0 normal-faulting subevent and a Mw 8.07 reverse-faulting subevent (Table 7.S1). The time delays of the two subevents are 25 s and 91 s, with durations 50 s and 182 s separately. The first subevent of Nealy and Hayes (2015) is located at the epicenter, and the second subevent \(\sim 50\) km southeast of our solution, with a depth of 17.5 km. The focal mechanisms of these studies (Lay et al., 2010; Duputel et al., 2012; Nealy and Hayes, 2015) are in rough agreement with our results. As discussed below, the differences in the CMT solutions among these studies are likely within the range of uncertainty of multiple-CMT solutions for this complicated earthquake. We prefer our solution because its reverse-faulting event is more clearly westward of the trench and closer to the long-period energy imaged by the LF back-projection.

### 5.4 Resolution and Uncertainties

There are intrinsic ambiguities in back-projection and CMT analysis. The back-projection relies on the reference hypocenter location, which is based on body wave observations. The centroid locations resolved in CMT analysis are mainly based on the surface-wave and W-phase observations. Before drawing conclusions about the rupture process, it is important to understand the resolution and uncertainties of the back-projection images. To do this, we have performed five different types of tests. First, for any given station set, the theoretical resolution
can be computed from the station distribution and the data bandwidth by randomly assigning a single representative recorded P wavetrain to all the stations, then back-projecting these traces. This provides an estimate of the resolution of the station distribution given the frequency content of the data (Figure 5.7). As expected, the spatial resolution is proportional to the bandwidth used for back-projection with the same network. The theoretical spatial resolution of the LF/HF data is roughly circular and about 50 km/25 km in radius, a result of the good azimuthal coverage of the globally located stations.

Another way to assess the robustness of the observed features is to perform back-projection with a spatial subset of the network. Robust features observed with the global network are expected to also appear in regional array data results, albeit at lower resolution. To compare with Lay et al. (2010) and Kiser and Ishii (2012) as well as our own results, we separately used F-net, Hi-net (e.g., Okada et al., 2004; Obara et al., 2005) and USArray (www.usarray.org) to image the Tonga-Samoa earthquake (Figure 5.7), using the same approach we used for the global data. In Figure 5.7, the theoretical resolution for F-net is too limited to distinguish potential multiple subevents. The results from F-net mainly show a concentration of the energy release northwest of the epicenter. Combining multiple arrays can improve the resolution, but distortions are still present. To combine Hi-net high-sensitivity records and USArray (TA) broadband records, we first use a time-domain recursive filter to convert the Hi-net recordings to broadband recordings before the alignment (Maeda et al., 2011). Because 1–5 Hz P waves were used to image the rupture process, Branch 3 rupture was not resolved in Kiser and Ishii (2012). However, as shown in Figure 5.7, the LF energy release within the subduction zone is also imaged with Hi-net and USArray. For the dense arrays, F-net, Hi-net and USArray, each trace is inversely weighted by the number of contributing stations within 1 degree. This weighting scheme is adopted so that we could have a fair comparison of the back-projection results. Other sophisticated weighting strategies might improve the resolution when considering array geometry, epicentral distances and data quality.

To assess potential biases introduced by the complexities of the wave-
field and the path effects, the resolution can be determined by performing back-projection on smaller events within the same region with similar station coverage as the mainshock. The resolution for the mainshock should be at least as good as that seen in the smaller event images, since smaller events may themselves have ruptured extended areas and usually have lower signal-noise-ratios for P waves and fewer usable stations. We performed back-projection on four nearby Mw ≥ 6 earthquakes (Figure 5.8). The images of the smaller earthquakes span a similar spatial extent as that seen in the theoretical resolution tests for LF data (Figure 5.7, 5.8) and are free of obvious artifacts. From Figure 5.8, the back-projection images of the smaller earthquakes are comparable to or smaller than the major energy bursts seen in Figure 5.3, showing that the three mainshock rupture stages and three early aftershocks are clearly resolved.

Finally, we attempt with two more tests to assess the quality of the back-projection when there are multiple subvents in different locations with different mechanisms. In particular, for the 2009 Tonga-Samoa earthquake, the complex structure of the region might cause rapid Green’s function variations near the subduction zone. For an extended rupture, the empirical time corrections based on the initial rupture near the hypocenter might not be appropriate for the later stages of the earthquake, where P polarity flips due to mechanism changes might also be present. Under some conditions, these problems could shift the rupture to erroneous locations/time or cause artifacts in the back-projection images. To address the problem, we performed back-projection of the four M6 earthquakes with time-shifts and weighting factors obtained from mainshock waveform alignment (Figure 5.9). For any of the M6 earthquakes, only stations that recorded both the mainshock and the M6 earthquake are used, which results in fewer useable stations. The empirical corrections are derived from LF waveform cross-correlations of the mainshock records, and they are applied to obtain back-projection images of the M6 earthquakes at both frequency bands. The low-frequency back-projection results of all four M6 earthquakes are well resolved despite their different focal-mechanisms and large spatial extents at all directions (Figure 5.9). The high-frequency back-projections are robustly resolved at the epicenters for all the M6
earthquakes except the 2009/08/30 foreshock, for which the peak energy location shifted ∼25 km northeast of its epicenter (within the claimed uncertainties). The apparent shift could indicate a spatial bias of the HF back-projection when rupture occurs at the trench axis. However, because the other M6 earthquakes are well resolved at both sides of the trench axis, it is possible that the HF spatial drift of the foreshock indicates the termination stage of the earthquake, as it locates close to the edge of the LF image of the earthquake. All of the M6 earthquake back-projection images are well resolved, their extents are within our claimed uncertainties, and the results do not have obvious artifacts. This test strongly supports our observations of the mainshock, and shows the robustness of the imaged multiple branching rupture.

To further test the robustness of the back-projection images for the Tonga-Samoa event, we generate a “synthetic” data set by combining the recorded data from two local ∼M6 events in the region, a normal-faulting seaward wall earthquake (Mw 6.0, 2009/10/19) and a thrust earthquake (Mw 6.3, 2015/03/30) on the subduction interface (Figure 5.10). These two events are close to the observed locations of energy bursts P1 and P3. We sum the first 100 s data from the two events, with a 20 s delay for the thrust event. The summation preserves possible multiples, reflections and noise in the P-waves of the first normal-faulting earthquake, and carries them into the P-waves of the second reverse-faulting earthquake. The realistic noise in conjunction with the rapid Green’s function variation across the trench could introduce biases or artifacts to the images of the second event. We then back-project the “synthetic” data set of the composite events following the same procedure applied to the mainshock, with the empirical time shifts, normalization factors, and weighting factors purely derived from the initiating normal-faulting earthquake (Figure 5.8). The results show that the second earthquake can be well resolved both spatially and temporally, even though the waveform alignment, normalization and weighting factors are based entirely on the initiating normal-faulting earthquake (Figure 5.10). No significant artifacts or biases are observed in the images, such as ghost ruptures offset from the main rupture patches. The results show LF back-projection works better than HF back-projection for this case, indi-
cating the 3D Green’s functions do not bias the LF results very much, the opposite of the expected strong influence due to the complex local velocity variations. This empirical calibration test supports the validity of the observed Branch 3 rupture of the megathrust interface.

By using a globally distributed network, artifacts in back-projection images can be greatly reduced. As shown in Figure 5.7–5.10, the good azimuthal coverage of the global station network minimizes the “swimming” artifacts (e.g., Xu et al., 2009; Koper et al., 2012; Meng et al., 2012) that are troublesome in back-projection images from regional arrays. For large shallow earthquakes, depth phases cannot easily be separated from the direct phases and usually interfere with the ensuing P waves that are associated with later parts of the rupture. However, as shown in Fan and Shearer (2015), the depth phases and other complexities do not bias the back-projection results very much when full azimuthal station coverage is available.

Our best-fitting single moment-tensor solution has 18% RMS data misfit, while the best two-source solution has 9.5% misfit. The significant misfit reduction strongly suggests the necessity of a second reverse-faulting subevent to accompany the initial normal-faulting event. Our preferred reverse-faulting subevent shares similar strikes and dips with other studies (Li et al., 2009; Lay et al., 2010; Duputel et al., 2012; Nealy and Hayes, 2015). But the rake varies from solution to solution for all the models (Table 7.S1). This variation might reflect differences in the data set and/or inversion procedure. Our solution is mainly constrained by the far-field surface waves, while other studies implemented near-field observations. Off-diagonal elements of the moment tensor solutions are poorly resolved for shallow earthquakes. Therefore, different inversion procedures with various approaches may give different values, which will lead to different rake values. Even though all the currently available CMT solutions share similar focal mechanisms, the large variations in their locations and times imply poor spatiotemporal constraints on the two subevents (Figure 5.6, Figure S8). The very-long-period data and Green’s functions calculated from a 1D velocity model are potential causes for some of the variation since the GCMT solution was obtained using 3D velocity corrections. For our two-point-source solution, the misfit contours shown in Figure 5.6 represent
the uncertainties of a subevent location when the other subevent is fixed at the labeled preferred location. When moving the second subevent to the locations in Duputel et al. (2012) and Nealy and Hayes (2015), the overall misfit increases to 10%–11%. The very-long-period data have poor temporal resolution for both subevent durations and centroid times (Figure S8). The second subevent centroid time can vary from 90 to 130 s, with total data misfit less than 10%. Within a misfit of 10%, the second subevent duration can vary from 40 to 100 s, and the first subevent can vary from 50 to 90 s. These small changes in misfit for large shifts in the CMT centroids probably is due to the fact that only stations in North America strongly require the second source, and without good azimuthal coverage its location is not well-constrained. The reverse-faulting subevent only contributes dominantly to the transverse component at certain azimuths, as shown by the theoretical radiation patterns (Figure S9).

The preferred CMT centroid location of the reverse-faulting subevent is \(\sim \)100 km southeast of the landward peak energy burst observed with back projection, and the initiation time is \(\sim \)50 s later (Figure 5.2, 5.3, 5.6, S3). The imaged P8 is spatiotemporally close to the second subevent, but its relative power is much weaker than P3 and P4. Given these differences, it is unlikely that back-projection is imaging the main moment release patch of the reverse-faulting subevent. This is not surprising, given that we filter the data to 0.002–0.005 Hz (200–500s) to determine the CMT solutions while the “low-frequency” back-projection uses P-waves filtered at 0.05–0.2 Hz (5–20 s) and frequency-dependent rupture behavior has been observed in large earthquakes along major plate boundaries (e.g., Lay et al., 2012; Yao et al., 2013; Meng et al., 2014; Wang and Mori, 2011b). Our results are consistent with a scenario in which the reverse-faulting event initiated with a burst of 5–20 s energy seen in the back-projection and then ruptured to the south into a large slip patch that is seen in the CMT solutions, but did not radiate enough short-period energy to coherently appear in the back-projection. Caution is warranted, however, as there are substantial resolution issues for both solutions because of limitations in data coverage and model-dependent uncertainties introduced by unknown velocity structures. The reverse-faulting CMT subevent
is mainly resolved by transverse components of five North America stations (Figure 5.6) within a narrow azimuthal bin. The back-projection resolution depends on the array geometry and locations (Figure 5.7). Strong near-source velocity heterogeneity associated with the subducting slab is not accounted for in our Green’s functions and could bias both the CMT and back-projection solutions. Despite these uncertainties, both preferred models lead to the same conclusion that at least two large subevents are involved in the Tonga–Samoa earthquake, they ruptured sequentially with the reverse-faulting subevent following the normal-faulting subevent, and the reverse-faulting subevent occurred west of the trench axis but no deeper than 40 km.

5.5 Discussion

The majority of aftershocks occurred within the subduction zone (Figure 5.2, 5.11). The first day of aftershocks spans the region where energy release is observed in the LF back-projection image, with a concentration around the peak energy burst (Figure 5.2). One month of aftershocks (International Seismological Centre, 2013) show a similar pattern (Figure 5.10). Combined with the seismicity, a very shallow dipping nodal plane (184°, 16° and 40°) of the reverse-faulting subevent in Figure 5.6 is preferred. A clear aftershock band within the subduction zone concentrates from 20 to 40 km, in contrast to the paucity of seaward-wall aftershocks. The aftershocks of the second subevent are confined from -15.2° to -17° in latitude, indicating the rupture of the reverse-faulting subevent extended about 200 km. All of the CMT solutions locate the second reverse-faulting south of −16°S (Figure 5.6), while there is a large concentration of the aftershocks from -15.2°S to -16°S. This aftershock zone spatially correlates with the P3 and P4 rupture imaged in the low-frequency back-projection, supporting the existence of the northern part of the trench-west rupture.

To constrain the geometry of faults formed seaward of the trench-axis, we calculate residual bathymetry from the SRTM15_PLUS bathymetry compilation (Becker and Sandwell, 2012; Sandwell et al., 2014). Residual bathymetry is calcu-
lated by regionally subtracting a spectral average of the trench-normal topography (Figure 5.11) (Bassett and Watts, 2015a,b). Spectral averaging has been shown to be an effective means of removing steep topographic gradients associated with ocean trenches, and here reveals complex along-strike variations in the spacing and geometry of extensional bend faults. South of -16.5°, bend faults are approximately strike-parallel (or N-S) and have a mean spacing of ~ 20 km. North of -15.3°, in contrast, faults display a strike closer to E-W and have a mean spacing of ~ 5 km. In the intermediary zone, bend faults strike NW-SE and have a mean spacing of ~ 8 km. The transition between approximately trench-parallel to almost trench-normal fault geometries is well correlated with an anti-clockwise curve in the geometry of the trench-axis. The pre-existing seafloor fabric strikes approximately E-W (Billen and Stock, 2000; Downey et al., 2007) and observed along-strike variations in fault spacing and geometry are likely driven by changes in the obliquity of this fabric with respect to the trench-axis and azimuth of plate bending. This hypothesis is consistent with global compilations showing that when the seafloor-fabric is oriented at angles <∼ 30° to the trench, the fabric is reacti-vated during bending and no significant new sets of faults develop (Masson, 1991; Ranero et al., 2005).

Of most significance to this study are clear spatial correlations observed between the location and geometry of energy bursts and fault traces (Figure 5.3,5.12). Branch 1 and 2 of the Tonga-Samoa earthquake are characterized by bilateral rupture NW and SE of the epicenter respectively. Energy bursts P1 and P2 are aligned with the NW-SE trend of bend faults, as are energy bursts P5 and the early af-tershock P6 (Figure 5.3,5.12). Faults in the northern region of the fracture zone strike approximately E-W, which is also mirrored by the azimuth between energy bursts P2 and P3 (Figure 5.3,5.11). It is here that the Tonga-Samoa earthquake appeared to rupture across the trench-axis, linking the seaward components of this earthquake (Branches 1 and 2) to Branch 3, which ruptured the megathrust in a southwesterly direction (Figure 5.3,5.12). We note that the southerly extent of LF back-projection imaged megathrust rupture is approximately coincident with the southern region of a fore-arc block characterized by positive residual bathymetric
anomalies (Figure 5.12). This suggests the fore-arc (and hence megathrust) is segmented along strike, which may have limited the southerly extent of megathrust rupture (Stage 3) (Bassett et al., 2016).

The back-projection images, along with the local bathymetry, provide a simple physical explanation for the rupture propagation and the anomalous nature of the 2009 Tonga-Samoa earthquake. We suggest Branches 1 and 2 are natural consequences of the fault geometries, and the Branch 3 rupture is modulated by the structure and segmentation of the fore-arc (Figure 5.2, 5.3, 5.12). Between P2 and P3, there is an EW trend fault that is partially subducted beneath the Tonga fore-arc, which may link the seaward faults system with the megathrust. The Mw 6.6 foreshock (2009/08/30, Figure 5.8) that occurred one month before the main shock was probably on this or an adjacent linking fault, which is consistent with its normal-faulting mechanism. The change in the geometry of the trench axis enables reactivation of the seafloor fabric. The reactivated normal fault may have played a key role in facilitating the initiation of the Branch 3 rupture. If it holds true, the linking fault may promote both regions to rupture along a single earthquake, as observed for the Tonga-Samoa earthquake.

Although the different rupture branches of the 2009 Tonga-Samoa earthquake are robustly resolved in our back-projection images, the spatial resolution is too coarse to distinguish continuous rupture propagation from a series of triggered subvents for this complicated earthquake. In particular, the apparent trench-crossing rupture from 20 to 40 s observed by both LF and HF back-projection may be more favorably explained as a discontinuous jump across the trench via dynamic triggering (Figure 5.3). Either immediate triggering or continuous rupture through a west-east normal linking fault could possibly initiate the Branch 3 rupture. The observed normal-reverse composite source model of the event connected via the linking fault could also be an early M8 reverse-faulting aftershock triggered simultaneously as the M8 normal-faulting earthquake propagates. With only estimates of average rupture velocities (~2 km/s, Figure S10), there is no clear distinction between multi-subevent rupture and a mainshock-very-early-aftershock rupture sequence. Therefore, we propose the second reverse-faulting subevent is triggered by
the first normal-faulting subevent, with the possibility of continuous westward rupture from the east faults to the west faults of the trench. It makes sense to group subevents by how close they are in time and space, so we suggest that the apparent westward rupture from P2 to P3 as part of Branch 1 is a result of curved fault traces (Figure 5.11), assign the early aftershock P6 as part of Branch 2, as it is spatially and temporally close to P5 (Figure 5.3), and group P7 and P8 to Branch 3.

Frequency-dependent rupture behavior has been observed for large megathrust earthquakes and continental earthquakes (e.g., Lay et al., 2012; Yao et al., 2011; Kiser and Ishii, 2011; Meng et al., 2011; Koper et al., 2011; Wang and Mori, 2011a,b; Fan and Shearer, 2015; Grandin et al., 2015), and variations have been seen both along strike and dip directions (e.g., Kiser and Ishii, 2011; Yao et al., 2013; Meng et al., 2014; Melgar et al., 2016). The observed frequency-dependent variations could be expressions of depth-varying frictional properties (e.g., Yao et al., 2013), multiple segmentation (e.g., Yin and Yao, 2016; Denolle et al., 2015), different rupture properties over the fault plane (e.g., Kiser and Ishii, 2011), or different mechanisms (e.g., Meng et al., 2014). The frequency dependence of the Tonga-Samoa earthquake is different from these cases in the sense that the frequency-dependent rupture behavior is associated with different faults. The high-frequency deficient rupture at the megathrust interface and high-frequency enrichment at the seaward fracture zone is a pattern that is similar to that observed during the 2006-2007 Kuril islands doublet (Ammon et al., 2008; Lay et al., 2009). The significant difference in frequency content likely reflects the special faulting environments of the outer-rise/seaward-wall faults and the subduction interface fault. Moreover, the clear along-strike separation of the LF back-projection energy bursts (P3, P4) and CMT solution centroid of the Branch 3 rupture has not been observed in a single earthquake before, suggesting the complexities of the reverse-faulting earthquake, which itself may rupture more than one episode. The Branch 3 rupture for the Tonga-Samoa earthquake within the subduction zone is critical for the induced tsunami (Zhou et al., 2012) (Figure 5.2, 5.3), as very shallow megathrust ruptures with large vertical displacements can produce strong
tsunamis, which often show little high frequency radiation (Kanamori, 1972).

The Tonga-Samoa earthquake represents an unusual normal faulting / reverse faulting doublet in which the thrust event occurred within 2 minutes of the initiating normal-faulting earthquake. This sequence is in the opposite order of the 2006-2007 Kuril islands doublet, where a reverse-faulting mainshock triggered a normal-faulting intraplate event 60 days later as a slab pulling effect. Other large trench-slope events preceding large interplate ruptures usually involve compressional, rather than extensional faulting (Christensen and Ruff, 1988; Lay et al., 1989). The contrasting observations for the Tonga-Samoa earthquake show another possible scenario for great earthquakes at subduction zones when complex fault systems are present with reactivated pre-existing seafloor fabrics. There is a possible partially subducted normal fault crossing the trench-axis, which may link the seaward-wall of the trench with the megathrust. If the linked fault can effectively transfer rupture from both sides, then this potential linkage would suggest a higher seismic risk for this region as ruptures could efficiently propagate from the seaward fault system to the megathrust and vice versa.

5.6 Conclusions

The 2009 Tonga-Samoa earthquake involved multiple ruptures on a system of faults, initiating as a normal-faulting earthquake located seaward of the trench-axis that curved into the trench, triggering a reverse-faulting subduction thrust event dominated by long-period seismic radiation. As revealed by back-projection in different frequency bands, the spatiotemporal evolution of the earthquake can be characterized as a multi-branch rupture process. Time-integrated back-projection images and snapshots show northwest rupture from 0 to 40 s (Branch 1), southeast rupture from 40 to 80 s (Branch 2), and southward rupture along the megathrust interface from ~30 to 160 s (Branch 3). Very early aftershocks (P6 to P8) are possibly triggered within two minutes located around the boundaries of the three rupture segments. The observed three-branch rupture is a natural consequence of the curved fault geometries associated with reactivation of pre-existing
fabric and fore-arc segmentation, which is illustrated by the collocation of observed energy bursts and bathymetrically inferred fault traces. Combined with a multiple-moment-tensor solution, the first two branches can be modeled as a Mw 8.0 bilateral normal-faulting earthquake and the last stage can be modeled with a Mw 8.1 reverse-faulting earthquake. The second reverse-faulting earthquake is triggered by the first normal-faulting earthquake, with possible involvement of a linking fault that is partially subducted into the megathrust interface, which may facilitate the initiation of the reverse-faulting subevent. One month of aftershocks indicate the reverse-faulting event has a very shallow dip angle and extends about 200 km along strike, and ruptured as north as -15.2°. The back-projection results show the frequency-dependent rupture pattern of the entire sequence, as the Branch 3 rupture of the reverse-faulting subevent is deficient in high frequencies. The rupture behaviors of the three branches reflect different faulting environments. The rupture pattern of the 2009 Tonga-Samoa earthquake reveals an interesting type of fault system involving seaward-wall intraplate normal faults, megathrust reverse faults, and reactivated fabrics, which could possibly link faults at both sides of the trench, and allow rupture sequentially across the trench. The special fault system geometry of the northern Tonga subduction zone poses a high seismic risk for that region as a potential host for great earthquakes and tsunamigenic earthquakes.

Acknowledgments

We thank Guy Masters and Zhitu Ma for providing normal mode data and programs for the Green’s function calculation, Zhongwen Zhan and Soli Garcia for helpful discussions, and Jennifer Nealy and Zacharie Duputel for sharing their W-phase CMT solutions. We thank the editor Yehuda Ben-Zion, the associate editor and the anonymous reviewers for their helpful comments that improved the quality of the manuscript. The seismic data were provided by the Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS) and National Research Institute for Earth Science Disaster Prevention in Japan.
The earthquake catalog was downloaded from the International Seismological Center (ISC) and the Global Centroid Moment Tensor (GCMT) (Ekström et al., 2012). The bathymetry data was processed with the Generic Mapping Tools (GMT) (Wessel and Smith, 1991; Wessel et al., 2013). The data used in the study is publicly available at DMC (http://ds.iris.edu/ds/nodes/dmc/) and the processed data is available from the authors upon request. This work was supported by National Science Foundation grant EAR-1111111.

Chapter 5, in full, is a reformatted version of the material as it appears in Journal of Geophysical Research–Solid Earth: Fan, W., P. M. Shearer, C. Ji, and D. Bassett, Multiple branching rupture of the 2009 Tonga–Samoa earthquake, J. Geophys. Res. 121, doi:10.1002/2016JB012945, 2016. I was the primary investigator and author of this paper.
Figure 5.1: Tectonic setting of the northern corner of the Tonga subduction zone. Seismicity (8274 earthquakes, 1916–2012) is from International Seismological Centre (2013). Focal mechanisms of the M > 6 earthquakes with centroid depths shallower than 60 km within the region are shown as squares; M > 7 earthquake moment magnitudes are listed by their focal mechanisms (Ekström et al., 2012). The plate boundary is from Bird (2003), trench axis is from Bassett and Watts (2015a,b), and the subduction geometry is from Slab 1.0 with 20 km separation (Hayes et al., 2012). The background bathymetry gradient is from Sandwell et al. (2014) and Garcia et al. (2014).
Figure 5.2: Back-projection results. Upper panels: Low-frequency (LF, 0.05–0.2 Hz) time-integrated back-projection image, stacked source-time function and time snapshots; Lower panels: High-frequency (HF, 0.2–2 Hz) time-integrated back-projection image, stacked source-time function and time snapshots. Stations used for the back-projection analysis are mapped in the upper right corner of the integrated LF back-projection image, with their lower-focal-hemisphere polarities plotted in the upper right corner of the integrated HF back-projection image.
Figure 5.3: Rupture evolution of the 2009 Tonga-Samoa earthquake as inferred from back-projection. Left panel: low frequency (LF); right panel, high frequency (HF). Apparent energy burst peaks are labeled as P1 to P8.
Figure 5.4: Aligned velocity seismograms (low-frequency, 0.05–0.2 Hz) with predicted travel-time arrivals of the observed energy busts in Figure 5.3. The records are sorted by azimuth, as listed on the right side every 3rd station with corresponding station names listed on the left side. Inferred rupture times of P1 to P5 are 9 s, 22 s, 31 s, 45 s, 54 s. Inferred rupture times for P6 to P8 are 68 s, 93 s, 130 s.
Figure 5.5: Right panel, stacked envelope functions (0.5 to 2 Hz) of energy bursts P1 to P8, with station map plotted as right-upper corner insert. Left panel, stacked envelope functions (1 to 5 Hz) of energy bursts P1 to P8, with station polarities on the lower hemisphere as right-upper corner insert (Red, negative). 935 stations are used for stacking. The envelope functions are calculated with a standard Hilbert transform without smoothing.
Figure 5.6: Comparison of multiple moment tensor solutions with other available source models. Left panel, multiple-moment-tensor solutions (beachballs), misfit contours for our two-CMT solution (colored lines), and integrated LF back-projection results (gray contours). Focal mechanisms are from Ekström et al. (2012), Lay et al. (2010), Li et al. (2009), Duputel et al. (2012), Nealy and Hayes (2015) and this study. For solutions of Nealy and Hayes (2015), UEv1, 2 are unconstrained solutions from grid-searching the centroids, TEv1, 2 are solutions constraining the second subevent to be landward of the trench while grid-searching the other centroids, CEv1, 2 are solutions matching the time-delays and depths as Lay et al. (2010) while grid-searching the other centroids. Stations used for our multiple-moment-tensor solution are mapped in the lower right corner of the left panel. For blue and teal colored stations, both vertical and transverse recordings are used; for purple stations, only vertical recordings are used. Right panel: the transverse-component data fit for the teal stations. AZ and GC stand for azimuth and great circle distance.
Figure 5.7: Comparison of theoretical resolving power and time-integrated back-projection results from two frequency bands (top: LF; bottom: HF) for the Tonga-Samoa earthquake obtained from F-net, a combined array (Hi-net and TA), and the global stations.
Figure 5.8: Time-integrated back-projection images of four ∼M6 earthquakes in the region. Three normal-faulting earthquakes are in the seaward region, and one reverse-faulting earthquake in the landward region. The Mw 6.6, 2009/08/30 foreshock occurred on a possible linking fault that connects the seaward faults with the megathrust. The Mw 6.0, 2009/10/19 aftershock occurred on the same fault as the mainshock.
Figure 5.9: Time-integrated back-projection images of four ~M6 earthquakes in the region with waveform alignment derived from the mainshock. The symbols are similar to Figure 5.8. The epicentral distances of the M6 earthquakes from the mainshock are shown in the right panel.
Figure 5.10: Back-projection results for a simulated multi-event rupture, created by summing records from the normal-faulting seaward wall aftershock (Mw 6.0, 2009/10/19) and the thrust earthquake (Mw 6.3, 2015/03/30) on the subduction interface. Upper panels: Low-frequency (LF, 0.05–0.2 Hz) time-integrated back-projection image, stacked source-time function (SSTF) and time snapshots; Lower panels: High-frequency (HF, 0.2–2 Hz) time-integrated back-projection image, stacked source-time function and time snapshots. Stations used for the back-projection analysis are mapped in the upper right corner of the integrated LF back-projection image. Because of the magnitude difference and different frequency content, we increase the recording amplitudes four times for the normal-faulting aftershock at LF, and increase the recording amplitudes twice for the reverse-faulting earthquake at HF. Absolute and Normalized represent the power of each window normalized with the maximum power within 160 s and each 20 s window.
Figure 5.11: One month of aftershock activity from the 2009 Tonga-Samoa earthquake (International Seismological Centre, 2013). Seismicity within the trench is denoted with red and the seaward seismicity is shown in blue. Top panel: event number versus time; middle panel: depth distribution along longitude, gray band shows the trench zone; bottom panel: depth distribution along latitude. Depths of events that cannot be vertically located are set to zero.
Figure 5.12: Ensemble of residual bathymetry and swath bathymetry from STRM 15 (Becker and Sandwell, 2012). Left panel: ensemble bathymetry. Right panel: bathymetry gradient. Apparent energy bursts (P1–P8) locations are the same as labeled in Figure 5.3. The integrated LF/HF back-projection results are shown in the left/right panel. Grey solid lines are inferred fault traces in the seaward wall of the trench.
<table>
<thead>
<tr>
<th></th>
<th>Lat(°)/Lon(°)/Depth(km)</th>
<th>Strike(°)/Dip(°)/Rake(°)</th>
<th>Centroid Time (s)</th>
<th>Duration (s)</th>
<th>CLVD</th>
<th>M&lt;sub&gt;W&lt;/sub&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>GCMT</strong></td>
<td>-15.13/-171.97/12.0</td>
<td>346/62/-63</td>
<td>42.6</td>
<td>26.8</td>
<td>0.16</td>
<td>8.1</td>
</tr>
<tr>
<td><strong>Li et al. (2009), Ev1</strong></td>
<td>-15.00/-172.50/12.0</td>
<td>340.7/45.4/-46.3</td>
<td>39.0</td>
<td>46.6</td>
<td></td>
<td>8.0</td>
</tr>
<tr>
<td><strong>Li et al. (2009), Ev2</strong></td>
<td>-16.1/-172.5/10.0</td>
<td>183.7/13.9/69.3</td>
<td>117.0</td>
<td>50.8</td>
<td></td>
<td>8.1</td>
</tr>
<tr>
<td><strong>Lay et al. (2010), Ev1</strong></td>
<td>-15.51/-172.03/18</td>
<td>152/67/-77</td>
<td>33</td>
<td>60</td>
<td></td>
<td>8.1</td>
</tr>
<tr>
<td><strong>Lay et al. (2010), Ev2</strong></td>
<td>-16.01/-172.43/18</td>
<td>185/29/90</td>
<td>69</td>
<td>40</td>
<td></td>
<td>7.8</td>
</tr>
<tr>
<td><strong>Lay et al. (2010), Ev3</strong></td>
<td>-16.01/-172.43/18</td>
<td>185/29/90</td>
<td>110</td>
<td>40</td>
<td></td>
<td>7.8</td>
</tr>
<tr>
<td><strong>Duputel et al. (2012), Ev1</strong></td>
<td>-15.09/-172.16/15.5</td>
<td>157/64/-70</td>
<td>59.2</td>
<td>53.6</td>
<td>-0.07</td>
<td>8.03</td>
</tr>
<tr>
<td><strong>Duputel et al. (2012), Ev2</strong></td>
<td>-16.33/-172.72/15.5</td>
<td>176.9/10.9/79.4</td>
<td>125.4</td>
<td>40</td>
<td>-0.006</td>
<td>8.05</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), FEv1</strong></td>
<td>-15.19/172.20/15.5</td>
<td>349.9/45.9/-84.6</td>
<td>66</td>
<td>66</td>
<td></td>
<td>8.22</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), FEv2</strong></td>
<td>-15.19/172.20/15.5</td>
<td>158.1/39.6/51.7</td>
<td>85</td>
<td>80</td>
<td></td>
<td>8.16</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), UEv1</strong></td>
<td>-15.59/-171.84/30.5</td>
<td>356.0/40.1/-78.5</td>
<td>66</td>
<td>66</td>
<td></td>
<td>8.23</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), UEv2</strong></td>
<td>-15.49/-171.89/23.5</td>
<td>162.4/47.1/57.9</td>
<td>85</td>
<td>80</td>
<td></td>
<td>8.17</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), TEv1</strong></td>
<td>-15.49/-172.10/17.5</td>
<td>330/44/-81</td>
<td>50</td>
<td>50</td>
<td></td>
<td>8.0</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), TEv2</strong></td>
<td>-16.39/-172.94/17.5</td>
<td>178.5/18/59</td>
<td>182</td>
<td>182</td>
<td></td>
<td>8.07</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), CEv1</strong></td>
<td>-15.39/-171.88/17.5</td>
<td>153.6/61.5/-81.6</td>
<td>50</td>
<td>50</td>
<td></td>
<td>8.03</td>
</tr>
<tr>
<td><strong>Nealy and Hayes (2015), CEv2</strong></td>
<td>-15.49/-172.20/17.5</td>
<td>171.3/23.5/51.1</td>
<td>182</td>
<td>182</td>
<td></td>
<td>8.02</td>
</tr>
<tr>
<td><strong>This study Ev1</strong></td>
<td>-15.00/-172.22/12.0</td>
<td>344/72/-68</td>
<td>35</td>
<td>70</td>
<td>-0.05</td>
<td>8.0</td>
</tr>
<tr>
<td><strong>This study Ev2</strong></td>
<td>-16.08/-172.50/12.0</td>
<td>183/15/40</td>
<td>120</td>
<td>80</td>
<td>-0.20</td>
<td>8.1</td>
</tr>
</tbody>
</table>
Compensated linear vector dipole (CLVD) is measured with $\epsilon = \sigma_2/\max(|\sigma_1|, |\sigma_3|)$. For solutions of Nealy and Hayes (2015), $\text{FEv1,2}$ are solutions with one fixed centroid location, $\text{UEv1,2}$ are unconstrained solutions with grid-searching the centroids, $\text{TEv1,2}$ are solutions constraining the second subevent to be landward of the trench while grid-searching the other centroids, $\text{CEv1,2}$ are solutions matching the time-delays and depths as Lay et al. (2010) while grid-searching the other centroids. The first subevent is dipping to the east when the strike is $\sim 344^\circ$, or is dipping to the west when strike is $\sim 153^\circ$. 
5.7 Supplementary Materials
Figure 5.S1: 0.05–0.2 Hz (LP) P-wave velocity seismograms aligned and sorted by similarity to the final waveform stack. The onset of the P wave begins at 0 s. The correlation coefficient of the initial 5 s ranges from 0.88 to 0.99. No polarity shift is allowed during cross-correlation. The maximum allowed time shift is ±4 s, and the optimal time shifts range from -3.3 to 2.7 s. For back-projection, traces are weighted by their correlation coefficients so that noisy or less coherent traces contribute less. The records are sorted by average cross-correlation coefficients, as listed on the right side every 3rd station with corresponding station names listed on the left side.
Figure S2: Multiple moment-tensor inversion grid configuration and stations used for the inversion. Left panel, blue triangles show the grid-searched locations for subevent 1, and red circles are grid-searched locations for subevent 2.
Figure 5.S3: Snapshots of low-frequency and high-frequency back-projection results with a 10s stacking window from 16s to 45s. Blue contours are high-frequency back-projection results, and the filled contours are low-frequency back-projections. The apparent initiation time of the reverse-faulting subevent is $\sim$27s.
Figure 5.S4: Stations used for back-projection and their lower-focal-hemisphere polarities for P, pP and sP. Red, positive polarities; Blue, negative polarities.
Figure 5.S5: Back-projection results of the 2009 Tonga–Samoa earthquake (LF, 0.05–0.2 Hz; MF, 0.1–1 Hz; and HF, 0.2–2 Hz), showing (left) the time-integrated back-projection images, and (right) stacked source-time functions and snapshots of the back-projection results for the three frequency bands. Symbols are the same as Figure 2.
Figure 5. S6: Comparison between observed very-long-period data (200s to 500s) and synthetic seismograms of the GCMT solution for the 2009 Tonga-Samoa earthquake. V stands for the vertical component and T stands for the transverse component. AZ and GC are the azimuth and great circle distance of the station from the 2009 Tonga–Samoa earthquake epicenter.
Table 5.S7: Comparison between observed very-long-period data (200s to 500s) and synthetic seismograms of the two-point-source solution for the 2009 Tonga-Samoa earthquake. Symbols are the same as Figure S6.
Figure 5.S8: Data misfit variations with respect to centroid times and durations of the two subevents, when the two subevent locations are fixed at the preferred locations in Figure 6. (a), waveform misfit variations with respect to changes of delay time, fixing the durations as 70 s for the first subevent and 80 s for the second subevent. (b), waveform misfit variations with respect to changes of durations, fixing the delay time as 0 s for the first subevent and 80 s for the second subevent. White cross shows the optimal grid-searching solution. Centroid time is calculated as the sum of the delay time and half of the subevent duration.
Figure 5.S9: Theoretical radiation pattern of Rayleigh waves and Love waves for the two-point-source CMT solution of Figure 6. Amplitude is proportional to the distance from the center for a given azimuth. Blue stands for the initial CMT and red for the second CMT.
Figure 5.S10: Comparison of the cumulative distance as a function of time, taken as the integrated distance between consecutive peak locations shown in Figure 3 and Figure S4. Due to the complicated rupture pattern, LF cumulative distance shows a more rapid increase compared to MF and HF. As a first-order estimate, the apparent rupture speed of Branches 1 and 2 is \( \sim 2 \) km/s, which is in good agreement with findings in Kiser and Ishii (2012).
References


Chapter 6

Local near-instantaneously dynamically triggered aftershocks of large earthquakes

Abstract

Aftershocks are often triggered by static- and/or dynamic-stress changes caused by mainshocks. The relative importance of the two triggering mechanisms is controversial at near-to-intermediate distances. We detect and locate 48 previously unidentified large early aftershocks triggered by $7 \leq M < 8$ earthquakes within a few fault lengths (~300 km), during times that high-amplitude surface waves arrive from the mainshock (< 200 s). The observations indicate that near-to-intermediate field dynamic triggering commonly exists and fundamentally promotes aftershock occurrence. The mainshocks and their nearby early aftershocks are located at major subduction zones and continental boundaries, and mainshocks with all types of faulting-mechanisms (normal, reverse, and strike-slip) can trigger early aftershocks.
6.1 Content

Earthquake occurrence is modulated by complex fault interactions that often involve static- or dynamic-stress triggering mechanisms (Harris, 1998), which can trigger earthquakes over a variety of spatial and temporal scales (Hill, 1993; King et al., 1994; Gomberg et al., 2001). Aftershock sequences are thought to result from either or both of these mechanisms (Harris, 1998; Freed, 2005). Static stress triggering is most important for near-field aftershocks, whereas dynamic triggering is dominant in the far field. However, it is challenging to quantitatively separate the effects of static and dynamic triggering in the near-to-intermediate field (Voisin et al., 2000; Decrèm et al., 2010), leaving their relative importance controversial (Felzer and Brodsky, 2006; Richards-Dinger et al., 2010).

Dynamic triggering is most clearly seen at large distances from earthquakes where earthquakes and/or non-volcanic tremor sometimes occur during the passage of surface waves, which are generally the highest-amplitude wave arrivals from shallow sources (Kilb et al., 2000; Obara et al., 2002; Velasco et al., 2008). However, observing possible dynamic triggering close to earthquakes is hampered by the mainshock coda, leaving existing catalogs incomplete and the local dynamic triggering effects uncertain (Kagan, 2004; Peng et al., 2006; Peng and Zhao, 2009). More complete catalogs would help in understanding local tectonics, constraining fault strength, and forecasting potential host faults for large earthquakes (Wesnousky, 1994). Local dynamic triggering can facilitate multi-fault rupture for a single earthquake, which can pose a much higher seismic risk than single fault ruptures (Harris et al., 1991, 2002). Recently, local near-instantaneous dynamic triggering has been observed at both subduction zones (Fan and Shearer, 2016) and continental plate boundaries (Nissen et al., 2016; Wang et al., 2016). But it is unclear how commonly this type of triggering occurs. Here we perform a comprehensive global search for early aftershocks of $7 \leq M < 8$ earthquakes and find that local near-instantaneous dynamic triggering is common and that multiple fault systems often dynamically interact with each other within a few fault lengths of the mainshocks within the first $\sim 200$ s.

We analyze teleseismic P waves from 88 large earthquakes ($7.0 \leq M_w <$
8.0) from January 2004 to September 2015, with Global Centroid Moment Tensor (GCMT) centroid depths shallower than 40 km (Fig. 1) (Ekström et al., 2012). We do not examine 12 larger earthquakes (Mw ≥ 8.0) in the same period because of their duration and complexity. We apply back-projection to detect and locate early aftershocks. Back-projection has proven to be effective to resolve complex spatiotemporal evolution of large earthquakes because the method requires few prior assumptions (Ishii et al., 2005), and it has also been successfully implemented to detect and locate both subevents (Allmann and Shearer, 2007; Kiser and Ishii, 2011; Yue et al., 2012; Meng et al., 2012) and early aftershocks (D’Amico et al., 2010; Kiser and Ishii, 2013; Fan and Shearer, 2016; Nissen et al., 2016; Wang et al., 2016).

Our data are from global stations distributed by the Data Management Center (DMC) of the Incorporated Research Institutions for Seismology (IRIS). The P waves are filtered between 0.05 to 0.5 Hz for back-projection, and are aligned for each event to reduce the effect of 3D velocity structure (Supp, 2016). No post-smoothing or post-processing is applied to the back-projection images. We search for potential early aftershocks that occurred within 200 s and between 50 and ∼300 km from the target earthquakes with a three-step screening criteria (Supp, 2016). We have validated our detection algorithm with three tests (Supp, 2016), which include confirming that our detected early aftershocks can be seen in high-frequency regional array data (Fig. S7), detecting and locating five cataloged mainshock-early-aftershock pairs within 100 km (Fig. S8), and performing back-projection on 15 local Mw 5.5 to 6.5 earthquakes located 200 km to 400 km away from mainshocks, using the same corrections as we used on the mainshocks (Fig. S9,10).

Twenty-seven of the 88 target earthquakes clearly triggered early aftershocks, and include events at most of Earth’s subduction zones and continental boundaries (Fig. 1, Fig. S1–3). None of the 88 target earthquakes have cataloged aftershocks in the time/distance window that we examined for this study (International Seismological Centre, 2013). Earthquakes with all types of faulting-mechanisms are capable of triggering early aftershocks (16 reverse-faulting, 4 strike-
slip, and 7 normal-faulting) (Fig. 1,2, Table S1). Normal-faulting earthquakes have the highest triggering rate (50%), while triggering rates of reverse-faulting (28.1%) and strike-slip (23.5%) earthquakes are similar. For robustness, our back-projection approach focuses on the phase of the P-wave arrivals, at the cost of losing absolute P-wave amplitudes (Supp, 2016), which makes estimating the magnitudes of the very early aftershocks challenging. Nevertheless, by comparing to historical nearby earthquakes, the triggered early aftershocks are likely to be M 5 to 6.5 earthquakes (Supp, 2016).

Within the Sunda arc subduction zone, a 2005/07/24 earthquake (Mw 7.2) occurred near the northwestern boundary of the great 2004 Sumatra-Andaman earthquake (Fig. 2A) (Ishii et al., 2005). The earthquake is a strike-slip event, which likely ruptured a different fault than the megathrust. The earthquake triggered two early aftershocks ∼177 and ∼221 km away from the epicenter, and ∼68 and 120 s after its initiation at the landward region of the subduction zone. The triggered events strongly correlate with the surface-wave arrivals from the mainshock. The exact focal mechanisms, magnitudes and depths of the triggered events are difficult to determine with only teleseismic P waves (Fan and Shearer, 2016), but the triggered events must be at least M5 to be observed in the far-field (Fan and Shearer, 2016). Within 100 km of the 2005/07/24 earthquake, a 2004/12/26 reverse-faulting earthquake (Mw 7.2) triggered an event in the seaward region (Fig. S1E). Both M7 earthquakes may be aftershocks of the 2004 Sumatra-Andaman earthquake, yet have very different focal mechanisms. Their triggered events are located in both the seaward and landward regions of the trench, indicating the region is potentially critically stressed at both sides of the trench.

Within the Japan subduction zone, a 2013/10/25 normal-faulting earthquake (Mw 7.1) broke the shallow part of the Pacific plate mantle two years after the 2011 Tohoku earthquake (Fig. 2B). The earthquake is seaward of the trench and triggered an early aftershock landward of the trench axis ∼40s later. The triggered event is ∼133 km away from the epicenter, close to the 2011 Tohoku earthquake centroid location, and can be either an interplate or intraplate earthquake (Nakamura et al., 2016). This triggered event underlines that stress can be
near-instantaneously dynamically transferred within complex multiple fault systems, in a region with long-term plate bending and converging deformation.

At the New Britain trench, the 2015/03/29 and 2015/05/05 Mw 7.5 doublet occurred on or near the subduction interface 130 km away from Kokopo, Papua New Guinea (Fig. 2C). The doublet events share similar focal-mechanisms (Ekström et al., 2012), and the 2015/05/05 event triggered two early aftershocks within the first two minutes after its initiation. The rupture propagated north-eastward, toward the two triggered events (Fig. 2C). The first early aftershock is triggered ∼40s after the mainshock origin time, located ∼120 km northeastern of the mainshock epicenter, and is close to the 2015/03/29 Mw 7.5 earthquake epicenter. Even though the first early aftershock struck within the trench-parallel region, its clear spatiotemporal separation from the mainshock and spatial correlation with the 2015/03/29 earthquake suggest it is an early aftershock rather than part of the mainshock. The second triggered early aftershock is located further north of the mainshock, which we refer to as the horizontal down-dip direction of the mainshock. The second triggered event is located by the Manus trench, where the Australian and Pacific plates converge at more than 70 mm yr\(^{-1}\) (DeMets et al., 1994). The convergence dominates the local tectonic evolution (Johnson and Molnar, 1972) and the triggered early aftershock implies that the interplate fault of the two plates might be critically stressed.

In total the 27 large earthquakes triggered 48 early aftershocks with epicentral distances ranging from 54 km to 334 km (Fig. 3). For each triggered event, the triggering time is taken as the delay from the origin time to the peak amplitude time within a 20s stacking window at the triggered location, ranging from 29.7s to 193.3s, from which a triggering velocity can be derived from the epicentral distance divided by the triggering time. To assess standard errors in the epicentral distances and triggering times, we implement jackknife resampling for each earthquake with the records used for the back-projection (Supp, 2016). Within one standard error, all the triggered early aftershocks occur after the surface wave passed through (at 3∼4 km/s) (Fig. 3). This strong correlation shows the 48 early aftershocks are triggered by the mainshocks, and suggests that dynamic stress is the physical
process that drives the observed triggering. Assuming the M7 earthquakes ruptured for about $\sim 40$ s, then 26 triggered early aftershocks coincide with the passing surface waves (Fig. 3). The rapid-onset dynamic triggering events with small delay time indicate frictional failures caused by dynamic stress changes induced by the transient surface waves (Kilb et al., 2000). In addition, 22 triggered events are delayed for seconds to minutes, which might reflect nonlinear friction behavior or a hydraulic response of the receiver faults (Gomberg et al., 1997; Parsons, 2005; Nur and Booker, 1972). The diversity of the triggered responses suggests the heterogeneity of the stress field and the variability of the frictional strength at each given fault (Rivera et al., 2002). These observations imply dynamic triggering modulates near-to-intermediate field seismicity, and commonly promotes large early aftershocks in a near-instantaneous fashion.

There are 11 triggered early aftershocks that occurred within 50 s of the origin time, with the remainder occurring within 200 s. The short temporal delay has two implications. First, a large portion of the early aftershocks are missing from global catalogs, which do not have these events, despite the fact they are large enough to be detected in teleseismic records. Second, the transition time from mainshock to aftershocks is near-instantaneous at most of the subduction zones and plate boundaries via dynamic triggering. If the early aftershock sequence follows Omori’s law, then the relative aftershock deficit, related to parameter $c$, will be pushed to as short as tens of seconds (Supp, 2016). If the aftershock activities are dominated by rate-and-state friction, then the derived $c$ can be used to probe frictional properties of the local fault systems (Peng et al., 2007).

Most of the observed early aftershocks are unlikely to be on or near the mainshock slip surface. The early aftershocks have a distinctly different epicentral distance distribution than the aftershocks cataloged by the International Seismological Centre (ISC) (International Seismological Centre, 2013) catalog or U.S. Geological Survey National Earthquake Information Center Preliminary Determination of Epicenters Bulletin (PDE) (Fig. 4A). The aftershock distribution of the ISC and PDE events can be used to estimate mainshock rupture areas. The majority of the catalog aftershocks are within $\sim 90$ km, while the majority of the
dynamically triggered early aftershocks are over ~90 km away from the epicenters (Fig. 4A) (Supp, 2016).

Fifteen of the 27 large earthquakes are reverse-faulting earthquakes at subduction zones, seven are normal-faulting earthquakes at subduction zones, four are strike-slip, and one is the 2008 China Wenchuan continental earthquake (reverse-faulting). All the subduction zone reverse-faulting earthquake centroid locations are within the landward-region, while all the normal-faulting earthquakes are in the seaward-region of the subduction zone (Fig. S1–3). To better understand the triggering mechanisms, we horizontally divide the triggered locations as down-dip-region, trench-parallel-region, up-dip-region, and seaward-region for reverse-faulting earthquakes with respect to the mainshock and trench axis; we divide the triggered locations as landward- and seaward-regions for normal-faulting earthquakes with respect to the trench axis (Fig. 4B, 4C). For the reverse-faulting earthquakes, the dynamically triggered early aftershocks tend to occur in the down-dip-region (44.5%), rather than the trench-parallel-region (33.3%) or the seaward-region (22.2%). We did not observe any triggered early aftershocks in the up-dip region for reverse-faulting earthquakes. For the normal-faulting earthquakes, occurrence of the triggered early aftershocks in the landward region is three times higher than in the seaward region (landward-region, 76.9%, and seaward-region, 23.1%). In total, the faults in the landward region, either on or near the megathrust, are more susceptible to near-field dynamic triggering. Although tsunami earthquakes often rupture the shallowest portion of the megathrust (Kanamori, 1972), the material or the faults in that region may be too weak to accumulate enough strain to be dynamically triggered. Generally, extensional regions are more easily triggered than compressional regions (Prejean and Hill, 2009; Harrington and Brodsky, 2006). If this holds true, some of the triggered earthquakes in the landward region are likely normal-faulting instead of reverse-faulting.

The detected early aftershocks all radiated less energy than the mainshocks and are likely moderate in size (M 5 to 6.5) (Supp, 2016). Although observed remotely triggered earthquakes to date have been relatively small (Parsons et al., 2012), it is possible on rare occasions that remote dynamic triggering could cause
damaging earthquakes (Pollitz et al., 2012). Near-to-intermediate field triggering has been observed before (Eberhart-Phillips et al., 2003; Allmann and Shearer, 2007; Oglesby et al., 2004; Kiser and Ishii, 2011; Yue et al., 2012; Meng et al., 2012; Delouis et al., 2009; Peyrat et al., 2010), and occurs in physics-based rupture models showing multi-fault systems can rupture together in a single earthquake (Harris et al., 1991, 2002; Anderson et al., 2003; Oglesby, 2005; Lozos, 2016), but is often considered to be part of the mainshock rupture process, which may involve multiple subevents, some triggered in response to both static and dynamic stress changes. Since the majority of the triggered early aftershocks are seen in the landward region of the trenches (Fig. 4), which could be on or near the megathrust, it is possible that this type of early dynamic triggering could lead to a great earthquake (M ≥ 8.0), with contributions from both static and dynamic triggering (Anderson et al., 2003; Lozos, 2016; Agnew and Jones, 1991). However, in the absence of known fault geometries for the triggered earthquakes we observe, it is difficult to perform stress calculations to explore the triggering mechanisms in more detail.

Early dynamically triggered aftershocks can also be seen in standard earthquake catalogs in favorable circumstances. To compare with our back-projection-detected events, we systematically searched through the ISC catalog from 1993 to 2013 to find earthquakes occurring in the same space/time window that we searched using back-projection, which follow target events of varying sizes (Supp, 2016). The local near-instantaneous triggering rate drawn from the ISC catalog for M7 earthquakes (4/198) is much lower than what we observe with back-projection (27/88), highlighting the difficulty in detecting these early aftershocks with standard methods (Fig. S11). The catalog results are most likely to be complete in cases where the triggered event is larger than the target event, i.e., the target event is a foreshock to the later event. This occurs globally in our space/time window for 3 of 1532 (0.2%) of M > 6 earthquakes. Assuming the dynamically triggered earthquake magnitudes are drawn randomly from a $b = 1$ Gutenberg-Richter distribution and aftershock triggering rates are self-similar with magnitude (Felzer et al., 2004), a 0.2% rate of triggering a larger event implies a 20% rate of $M 5-7$
early aftershocks following $M_7$ mainshocks in the same space/time window. This is comparable to our observed rate of 30%.

Our analysis represents a lower limit on the number of near-source dynamically triggered earthquakes that are large enough to be seen teleseismically, as we likely missed many events owing to our conservative selection criteria and the poor station coverage for some mainshocks. Thus fault interactions and triggering may be a relatively common feature for large earthquakes near subduction zones and continental boundaries. The near-zero to short delay time of the observed dynamic triggering suggests that in a large complex fault system, such as exists in most subduction zones, a few faults will always be critically stressed and close to failure. By studying where triggered events are most common, it may be possible to infer properties of the interacting faults in specific regions. Finally, early aftershocks can potentially illuminate unknown faults, and the observed fault interactions of earthquake sequences can inform future hazard assessment.

**Acknowledgements**

The facilities of IRIS Data Services, and specifically the IRIS Data Management Center (DMC), were used for access to waveforms used in this study (National Science Foundation EAR-1261681). Hi-net data were obtained from the National Research Institute for Earth Science Disaster Prevention in Japan (NIED). The earthquake catalogs were downloaded from the Global Centroid Moment Tensor project (GCMT) (Ekström et al., 2012), U.S. Geological Survey National Earthquake Information Center Preliminary Determination of Epicenters Bulletin (PDE), and the International Seismological Center (ISC) (International Seismological Centre, 2013). The data used in the study are publicly available at DMC and NIED, and the processed data are available from the authors upon request. This work was supported by National Science Foundation grant EAR-1111111.

Chapter 6, in full, is a reformatted version of the material as it appears in Science: Fan, W., and P. M. Shearer, Local near instantaneously dynamically trig-
gered aftershocks of large earthquakes, *Science*, 353, 1133-1136, doi: 10.1126/science.aag0013, 2016. I was the primary investigator and author of this paper.
Figure 6.1: Twenty-seven early-aftershock-triggering mainshocks and their focal mechanisms. The 27 triggering mainshocks are color coded at their GCMT centroid locations (colored circles). The white circles show the rest of the 88 large earthquakes ($7 \leq M_w < 8$) that we investigated with back-projection. Strike-slip earthquakes are defined when the rake of both nodal planes are within $45^\circ$ deviation of $0^\circ$ or $180^\circ$. Normal-/Reverse-faulting earthquakes have rakes within $45^\circ$ deviation of $-90^\circ/90^\circ$. Triggering rates for the three types of earthquakes are shown in the lower left insert.
Figure 6.2: Back-projection results for three earthquakes with different focal-mechanisms (60% normalized energy contours). (A) Rupture evolution of the 2005/07/24, Mw 7.2 strike-slip earthquake in the Sunda arc. Stations used for back-projection and their P-wave polarity with the GCMT focal-mechanism are shown as inserts. Negative polarities are red, and positive polarities are blue. (B) Rupture evolution of the 2013/10/25, Mw 7.1 normal-faulting earthquake in the Japan subduction zone. The inserts are the same as (A). (C) Rupture evolution of the 2015/05/05, Mw 7.5 reverse-faulting earthquake in the New Britain trench. The inserts are the same as (A).
Figure 6.3: Time versus distance plot of triggered events. Forty-eight triggered early aftershocks are shown as stars with the same color of their triggering large earthquakes. The colored bars for each triggered early aftershock show one standard error of the epicentral distances and triggering times (Table S1). The shaded region shows the likely influence of passing surface waves of a $\sim 40s$ duration M7 earthquake. The insert shows the distribution of the triggering velocity for the 48 triggered events.
Figure 6.4: Aftershock distribution and relative locations of the triggered early aftershocks. (A) The averaged ISC aftershock distribution of 18 mainshocks (Supp, 2016) is shown as the black line. Average PDE aftershock distribution from 5 mainshocks is shown as the grey line. The dynamically triggered early aftershock distribution is shown as the red line. The distribution is obtained from partitioning the triggered early aftershocks into 30-km-wide bins in epicentral distance. The probabilities are placed in the middle of the center of each epicentral bin. (B) Triggering rate and distribution for reverse-faulting mainshocks. (C) Triggering rate and distribution for normal-faulting mainshocks. The divisions of the relative locations of triggered early aftershocks are shown in the inserts.
6.2 Supplementary Materials

Materials and Methods

Back-projection

Due to their simple ray paths, P waves recorded at 30° to 90° provide a relatively unbiased record of seismic radiation from earthquakes with little distortion during wave propagation. The back-projection method then relies on the coherent signals recorded by regional or global arrays as approximations for the relative radiation at given locations close to the hypocenter (Ishii et al., 2005; Walker and Shearer, 2009; Xu et al., 2009; Meng et al., 2012; Koper et al., 2012; Yagi et al., 2012; Wang et al., 2012; Satriano et al., 2012; Okuwaki et al., 2014; Fan and Shearer, 2015). It is computationally efficient and stable as the method uses stacking instead of inversion. It also does not assume any particular fault geometry, which enables the method to detect abnormal or complex rupture behaviors. Following (Xu et al., 2009; Fan and Shearer, 2015), we first align the seismograms by cross-correlating the initial P-waves to neutralize 3D velocity structure influence (filtered in 0.05 to 0.5 Hz or 0.5 to 2 Hz for cross-correlation, cross-correlation window lengths vary from 5 to 8s, and allowed time shifts vary from 4 to 5s). We do not allow polarity shifts during the alignment for back-projection (Houser et al., 2008). Then, the aligned seismograms are shifted at given hypothesized source locations with theoretical P-wave velocity travel times from the 1D IASP91 velocity model (Kennett and Engdahl, 1991). Finally, the shifted seismograms are stacked at each location with a 20s window using Nth root ($N = 4$) stacking to extract coherent signals, while suppressing artifacts from unmodeled amplitude variations (McFadden et al., 1986; Rost and Thomas, 2002; Xu et al., 2009).

The potential sources are gridded laterally at the hypocentral depths with 10 km spacing within 600 km by 600 km of each earthquake, where the epicenters are at the center of the grids. Only records that share the same polarities from GCMT focal-mechanism best double-couple solutions are used. The seismograms are hand-selected to have high signal-to-noise ratios, filtered at 0.05 to 0.5 Hz with a second-order Butterworth filter for all the earthquakes. For each earthquake, the records used for back-projection are normalized, weighted by their average
correlation coefficients obtained from the cross-correlation alignment, and also inversely scaled by the number of contributing stations within 5 degrees, which serves to equalize the contributions from different instruments, to down-weight noisy records, and to prevent biases from dense local arrays. Results are shown in Fig. S1-S3. No post-smoothing or post-processing is applied to the images. Only the first 200 s after the earthquake origin time is evaluated to avoid complicated later phases. The small spatial extent (∼300 km radius) minimizes biases from changes in the 3D structure away from the hypocenter. Because of these time and distance limits in our method, we cannot exclude the possibility that additional aftershocks are present at greater distances and times. However, such events would be much more easily detected with standard methods than the time/space region we examine here, as their seismic arrivals would be more separated from the main-shock coda.

The jackknife

To understand the uncertainties of the early aftershock locations and triggering times, we perform jackknife resampling over the records used for back-projection for each earthquake (Efron and Tibshirani, 1994). The jackknife is very similar to the bootstrap. But because the station distribution and usable stations are different for each earthquake, the jackknife is more appropriate for this problem.

Suppose we have $n$ stations to estimate an early aftershock epicentral distance (location) and triggering time:

$$\hat{L} = BP_l(n)$$

(6.1)

and

$$\hat{T} = BP_t(n)$$

(6.2)

where $\hat{L}$ and $\hat{T}$ are estimations for the epicentral distance and triggering time for an early aftershock, $BP_l$ and $BP_t$ are the estimators. Denote the estimations as $\hat{L}_i = BP_l^i(i)$ and $\hat{T}_i = BP_t^i(i)$, when the $i$th station is excluded. Then the jackknife
estimate of standard errors (SE) are

\[
\hat{SE}_L = \left( \frac{n - 1}{n} \sum_{i=1}^{n} (\hat{L}_i - \bar{L}) \right)^{1/2}
\]  

(6.3)

\[
\hat{SE}_T = \left( \frac{n - 1}{n} \sum_{i=1}^{n} (\hat{T}_i - \bar{T}) \right)^{1/2}
\]  

(6.4)

for the epicentral distance and the triggering time, where \(\bar{L}\) and \(\bar{T}\) are the averages of \(\hat{L}_i\) and \(\hat{T}_i\). Due to the grid size and frequency band used for back-projection, \(\hat{SE}_L\) is set to be 10 km when \(\hat{SE}_L < 10\), and \(\hat{SE}_T\) is set to be 2 s when \(\hat{SE}_T < 2\) (Table. S1).

**Detection criteria**

We detect very early aftershocks with a three-step screening criteria:

(1) For each event, we first stack the back-projected time-series at each grid point with a non-overlapping 4 s window. At each time step, we then identify the grid point (greater than 50 km from the epicenter) with the maximum back-projected power and generate a time series of these maximum powers:

\[
SP(t) = \max_{d_i > 50km} |s_i(t)|
\]  

(6.5)

where \(i\) is the grid point for the \(i\)th potential source location, \(d_i\) is the epicentral distance of the grid point, and \(s_i(t)\) is the back-projection stack at the \(i\)th grid point at time \(t\). To avoid complexities associated with the mainshock rupture, we set the first 20 s of the peak power time functions to zero. This generates a time series for each event, in which the power typically decreases with time (as expected from coda decay from the mainshock), but which contains local peaks that are possible early aftershocks (Figs. S4–6). Note that each peak generally represents a different possible source location. We flag for further analysis every peak with amplitude at least twice the power in the lower of the two adjacent troughs. This resulted in 169 candidate early aftershocks for our 27 mainshocks, shown as the circles in Figs. S4–6.

(2) Next, we used a 20 s stacking window to image each M7 earthquake. The beam power of each time window is normalized by the maximum power within
each window. The flagged time windows from the first step are visually inspected
to examine the peak energy locations for possible early aftershocks. An energy
peak will pass this screening step and continue to be labeled as an early aftershock
candidate only if it satisfies three criteria: (a) its spatial peak is 50 km away
from the epicenter, (b) there are no side lobes and only one dominant spatial
peak location is present (no other energy peak within the source grid during the
20 s stacking window can exceed 75% of the amplitude of the candidate energy
peak), and (c) it is at least 10 km away from the edges of the back-projection
grid. If multiple episodes occur at the same location sequentially, only the first
one is labeled as an early aftershock candidate. These step reduced the number of
candidate early aftershocks to 67.

(3) Finally, we performed jackknife resampling to examine the standard
errors of the early aftershock candidates. We reject any events with spatial stan-
dard errors greater than 40 km and temporal standard errors greater than 15 s
(Table. S1). This reduced the number of early aftershocks to our final list of 48
events.

Magnitude estimation of the very early aftershocks

The amplitude information of the detected very early aftershocks is either
relative or distorted (owing to the non-linear Nth root stacking used in the study),
which makes it hard to determine the absolute magnitude. Therefore, we empir-
ically estimate the likely magnitudes of the observed very early aftershocks (Ta-
ble S1). We examine the historical earthquake magnitude distribution within 50
km of the very early aftershocks, and take the median as characteristic magnitudes
of the very early aftershocks. The catalog used to estimate each very early after-
shock is obtained from the ISC catalog (International Seismological Centre, 2013),
listing earthquakes from 1993/01/01 to 2013/01/01, with hypocentral depths shal-
lower than 60 km and magnitude larger than 5 (prime authors). Because of the
magnitude (M > 5) cutoff and estimation procedure, characteristic magnitudes of
18 of the 48 very early aftershocks are not resolved. Characteristic magnitudes
of the remaining 30 of the 48 very early aftershocks range from 5.0 to 6.5 with
average magnitude 5.3 (Table S1).

**Validation I: High frequency back-projection with regional array data**

We experimented with back-projection of regional array data at relatively high frequency (0.8 to 2 Hz) to see if we could confirm the very early aftershocks (Fig. S7). This frequency band was demonstrated to be effective to detect hidden aftershocks of the 2011 Tohoku earthquake by coherency functions with USAArray (*Kiser and Ishii*, 2013). In principle, higher frequencies should be better for detecting small events buried in mainshock codas when the waveforms are coherent, as the amplitude ratio between the aftershock and the mainshock coda at high frequency is larger. Shown in (*Kiser and Ishii*, 2013), using different choices for the frequency range of the filtered data (e.g., 0.2 to 2 Hz) may lead to fewer detected events, but the overall results are similar (Figure S1 of (*Kiser and Ishii*, 2013)). In our main study, we used a relatively low-frequency band (0.05 to 0.5 Hz) because we have found this works better for global back-projection, for which the waveforms are less coherent than for more localized arrays but the global coverage increases spatial resolution and reduces artifacts.

We performed back-projection of the 27 mainshocks with globally detected aftershocks using high-frequency regional North American, European array or Hi-net data (Fig. S7). Eleven very early aftershocks of 9 mainshocks are well resolved with high-frequency array data, and their locations/time agree well with the global array detected ones (Fig. S1–3,7). Even though not all the 48 very early aftershocks were detected with high-frequency data, the test supports our findings with the very early aftershocks.

It appears that regional array data are often not as effective as global data for back-projection detection of early aftershocks. Small aperture arrays (e.g., HiNet) have limited spatial resolution for moderate to small earthquakes, and often suffer from imaging artifacts, which hamper detection of small earthquakes. In addition, even for regional arrays, the timing corrections derived from the mainshock may not be accurate enough for coherent stacking of high-frequency energy from early aftershocks occurring far away from the mainshock epicenters, owing
to the effects of 3D velocity structure. However, further experimentation with regional array data is warranted to see whether in some circumstances they might detect early aftershocks that are missed by other techniques, including global back-projection.

**Validation II: Imaging of cataloged early aftershocks**

Both synthetic tests and aftershock imaging of smaller events show that the back-projection procedure we use has a resolution of \( \sim 50 \) km (at 0.05 to 0.3 Hz) and is not biased very much by depth phases because of the 20s long stacking window (Fan and Shearer, 2015; Fan et al., 2016). To further validate the resolving capability of the back-projection procedure, we imaged five earthquakes having ISC cataloged early aftershocks following the same back-projection method (Fig. S8, Table. S2). The aftershocks occurred less than 240 s and 350 km away from the mainshocks, and are chosen to be at least \( \sim 0.5 \) magnitude smaller than the mainshock to test our detection capability. All five mainshock-aftershock pairs are well resolved with aftershock location errors within our claimed uncertainties, while no additional aftershocks are seen at other locations and times.

In Fig. S8B, the cataloged aftershock location is shifted 27.6 km away from the peak energy location of the back-projected image. The same shift is also observed in Fig. S7D, which was imaged with that European array in the high-frequency band (0.8 to 2 Hz). The location uncertainty for this cataloged aftershock ranges from 4.0 to 14.3 km (6 other location solutions, with average deviation 9.8 km from the ISC preferred solution) (International Seismological Centre, 2013). Systematic comparisons of earthquake source locations using seismic data and interferometric synthetic aperture radar (InSAR) show the median epicentral location deviation is \( \sim 20 \) km (Weston et al., 2011, 2012), suggesting the possible location uncertainty of the cataloged aftershock can be as large as \( \sim 20 \) km. The uncertainty of the back-projection imaged aftershocks for the mainshock is \( \sim 20 \) km (Table S1). The shifted location could be explained by a combination of these two uncertainties.
Validation III: Imaging of moderate intermediate field earthquakes

In principle, our back-projection approach will be less effective for aftershocks further away from the mainshocks (because our timing corrections will be less accurate) and for events with different mechanisms than the mainshock (because of polarity flips). Since many of the detected early aftershocks in this study are across trenches, they might have different faulting mechanisms than the mainshocks or be subject to strong 3D velocity variations. These problems could shift the detected very early aftershocks to erroneous locations/times or cause artifacts in the back-projection images. To test for this, we performed back-projection of 15 moderate earthquakes (Mw 5 to Mw 6.5, Fig. S9,10, Table S3) with time-shifts and weighting factors obtained from 12 mainshock waveform alignments following the same procedure described above (Fig. S9,10, Table S3). These 15 moderate earthquakes are located ∼200 to ∼450 km away from the mainshocks, and occurred within 60 days before or after the mainshocks. Only common stations that recorded both the mainshock and the moderate earthquake are used, which results in fewer usable stations. Normalized power of the first 180 s are shown as black lines in Figure S9,10. All the 15 moderate earthquakes are robustly resolved at their epicenters with maximum deviation 35.9 km and average deviation 13.9 km without obvious artifacts or biases (Table S3). This test strongly supports the reliability of our detections of early aftershocks, and shows the robustness of the imaging algorithm.

Supplementary Text

Omori’s law

Omori’s law (Omori, 1895) describes how aftershocks decay with time following a power law after the mainshock, and was later generalized to

\[ R(t) = \frac{K}{(t + c)^p} \]  \hspace{1cm} (6.6)

where \( R(t) \) is the number of aftershocks per unit time, \( K \) is the aftershock productivity, \( p \) is the decay rate, and \( c \) is relevant to a relative deficit of aftershocks immediately following the mainshock, which is introduced to prevent singularities.
when the time goes to zero.

**Aftershock catalogs**

Both the International Seismological Centre (ISC) (*International Seismological Centre*, 2013) catalog and U.S. Geological Survey National Earthquake Information Center Preliminary Determination of Epicenters Bulletin (PDE) are used to investigate the conventional aftershock distributions for the mainshocks. We downloaded the first month of aftershock catalogs within 300 km for each earthquake. Two doublets, 2004/9/5 (Fig. S1B, S1C) and 2014/04/12 and 2014/04/13 (Fig. S3F, S3G) occurred sequentially within two days at nearby locations. Therefore, only the first event aftershock catalog is used to prevent duplicity. The 2004/12/26 (Fig. S1E) earthquake occurred within one day of the 2004 Sumatra-Andaman earthquake, and the 2010/2/27 event (Fig. S2C) is an aftershock of the 2010 Maule earthquake and occurred within one day. Therefore, they are excluded because the local seismicity does not necessarily represent its own aftershock sequence.

For each mainshock, the aftershock distribution is the normalized probability calculated after partitioning the aftershocks into 30-km-wide bins in epicentral distance. The probability distributions are averaged separately for both catalogs to show the aftershock distribution (Fig. 4A).

**ISC earthquake triggering search**

We systematically searched through the ISC (*International Seismological Centre*, 2013) catalog from 1993/01/01 to 2013/01/01 to find earthquakes potentially dynamically triggered by target events of varying sizes with hypocentral depths shallower than 60 km. Only earthquakes reported by reliable magnitude authors (prime author) are included in the analysis. M8 earthquakes and their aftershocks within 350 km and 15 days are removed, leaving 1725 M6–M7 earthquakes in the catalog. The potentially dynamically triggered early aftershocks are searched within 50–350 km away from the target event epicenters, and occurred 20–200 s later after the target event origin times. There are 1532 $6 < M < 7$ and
198 $M > 7$ earthquakes occurring in the time period. Four M7 triggered early aftershocks are cataloged, 3 $M > 6$ foreshocks triggered larger mainshocks, and there are 20 $M > 6$ target earthquakes triggered early aftershocks in total fitting our search criteria (Fig. S11).
Figure 6.S1: Back-projection results of nine M7 earthquakes from 2004 to 2008 (60% normalized energy contours). Stations used for back-projection and their P-wave polarity with the GCMT focal-mechanism are shown as inserts. Negative polarities are red, and positive polarities are blue. Plate boundaries are from (Bird, 2003), slab contours with 20 km separation are from (Hayes et al., 2012), and trench axes are from (Bassett and Watts, 2015a,b). Stars and squares are the epicenters and GCMT centroid locations of the M7 mainshocks. Crosses are the detected early aftershocks.
Figure 6.S2: Back-projection results of nine M7 earthquakes from 2009 to 2012 March (60% normalized energy contours).
Figure 6.S3: Back-projection results of nine M7 earthquakes from 2012 Oct to 2015 (60% normalized energy contours).
Figure 6.S4: Peak power time functions for nine M7 earthquakes from 2004 to 2008. Grey circles show possible early aftershocks (EA), where the peak powers reach local maxima that are twice larger than the powers of the adjacent lower troughs. Colored lines show those candidates that passed our three-step screening procedure.
Figure 6.S5: Peak power time functions for nine M7 earthquakes from 2009 to 2012 March. See Fig. S4 caption for details.
Figure 6.S6: Peak power time functions for nine M7 earthquakes from 2012 Oct to 2015. See Fig. S4 caption for details.
Figure 6.S7: High-frequency back-projection results of 9 earthquakes with regional arrays (95% normalized energy contours of Fig. S7I, 75% for the rest). A cataloged early aftershock (180 to 200 s) of the 2006/4/20 earthquake was also imaged with the high-frequency band in (D).
Figure 6.S8: Back-projection results of five earthquakes that triggered early aftershocks within 240s after their origin times (80% normalized energy contours of 180 to 200 s in Fig. S8B, 60% normalized energy contours for the rest). The aftershocks are chosen to be at least \( \sim 0.5 \) magnitude smaller than the mainshocks. Triggering time is denoted as \( dT \). Black/Red stars are the ISC mainshock/aftershock epicenters (Table S2). Station maps and their polarities are the same as Fig. 2.
Figure 6.S9: Back-projection results of nine moderate earthquakes from 2004 to 2011 with empirical corrections obtained from the mainshocks (75% normalized energy contours). Stations used for back-projection and their P-wave polarity with the GCMT focal-mechanism are shown as inserts. Negative polarities are red, and positive polarities are blue. Grey focal-mechanisms are for moderate earthquakes, colored ones are for the mainshocks. M stands for the mainshocks, and S stands for the moderate earthquakes. Stars and squares are the epicenters and GCMT centroid locations of the mainshocks (colored) and moderate events (red).
Figure 6.S10: Back-projection results of nine moderate earthquakes from 2013 to 2015 with empirical corrections obtained from the mainshocks (75% normalized energy contours).
Figure 6.S11: Time versus distance plot of possible dynamic triggered earthquake pairs from the ISC catalog (1993–2013). Green dots show early aftershocks dynamically triggered by M7 earthquakes, red dots show $M > 6$ foreshocks triggering larger mainshocks, and blue dots show early aftershocks dynamically triggered by M6 earthquakes. Inserts show the three foreshock-mainshock pairs, red stars are the triggering earthquakes, and grey stars are the triggered earthquakes.
Table 6.S1: Triggered early aftershocks.

<table>
<thead>
<tr>
<th>Triggering Event</th>
<th>Epicentral distance (km)</th>
<th>Triggering time (s)</th>
<th>Station number (Jackknife (n))</th>
<th>Characteristic magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>2004/1/3a</td>
<td>149.53 ± 10.00</td>
<td>72.85 ± 7.09</td>
<td>123</td>
<td>5.2</td>
</tr>
<tr>
<td>2004/1/3b</td>
<td>60.51 ± 10.00</td>
<td>117.85 ± 2.00</td>
<td>123</td>
<td>5.2</td>
</tr>
<tr>
<td>2004/9/5A</td>
<td>90.19 ± 10.00</td>
<td>53.45 ± 2.00</td>
<td>206</td>
<td>5.0</td>
</tr>
<tr>
<td>2004/9/5B</td>
<td>105.94 ± 19.66</td>
<td>102.92 ± 2.00</td>
<td>129</td>
<td></td>
</tr>
<tr>
<td>2004/11/15a</td>
<td>98.08 ± 35.38</td>
<td>103.42 ± 2.00</td>
<td>82</td>
<td></td>
</tr>
<tr>
<td>2004/11/15b</td>
<td>213.05 ± 19.05</td>
<td>137.82 ± 2.00</td>
<td>82</td>
<td>5.8</td>
</tr>
<tr>
<td>2004/12/26</td>
<td>158.71 ± 27.31</td>
<td>133.97 ± 2.48</td>
<td>100</td>
<td>5.2</td>
</tr>
<tr>
<td>2005/7/24a</td>
<td>177.24 ± 10.00</td>
<td>68.85 ± 2.00</td>
<td>98</td>
<td>5.1</td>
</tr>
<tr>
<td>2005/7/24b</td>
<td>221.00 ± 10.00</td>
<td>120.57 ± 2.00</td>
<td>98</td>
<td>5.2</td>
</tr>
<tr>
<td>2006/4/20a</td>
<td>66.73 ± 10.00</td>
<td>33.17 ± 2.00</td>
<td>204</td>
<td>5.2</td>
</tr>
<tr>
<td>2006/4/20b</td>
<td>91.71 ± 19.00</td>
<td>43.90 ± 2.00</td>
<td>204</td>
<td>5.2</td>
</tr>
<tr>
<td>2006/4/20c</td>
<td>208.84 ± 13.95</td>
<td>82.03 ± 2.00</td>
<td>204</td>
<td></td>
</tr>
<tr>
<td>2008/5/12a</td>
<td>147.93 ± 27.46</td>
<td>64.97 ± 2.00</td>
<td>111</td>
<td>5.1</td>
</tr>
<tr>
<td>2008/5/12b</td>
<td>260.34 ± 39.45</td>
<td>95.10 ± 2.00</td>
<td>111</td>
<td>5.2</td>
</tr>
<tr>
<td>2008/11/16a</td>
<td>129.53 ± 13.93</td>
<td>55.85 ± 2.00</td>
<td>57</td>
<td></td>
</tr>
<tr>
<td>2008/11/16b</td>
<td>200.87 ± 10.00</td>
<td>60.12 ± 2.00</td>
<td>57</td>
<td>5.2</td>
</tr>
<tr>
<td>2009/1/3</td>
<td>120.99 ± 10.00</td>
<td>105.20 ± 2.00</td>
<td>123</td>
<td>5.9</td>
</tr>
<tr>
<td>2010/1/12a</td>
<td>84.53 ± 10.00</td>
<td>29.70 ± 8.70</td>
<td>62</td>
<td></td>
</tr>
<tr>
<td>2010/1/12b</td>
<td>75.87 ± 10.00</td>
<td>40.10 ± 2.00</td>
<td>62</td>
<td></td>
</tr>
<tr>
<td>2010/1/12c</td>
<td>160.60 ± 31.76</td>
<td>131.55 ± 5.40</td>
<td>62</td>
<td></td>
</tr>
<tr>
<td>2010/2/27a</td>
<td>260.18 ± 23.22</td>
<td>65.42 ± 2.00</td>
<td>125</td>
<td></td>
</tr>
<tr>
<td>2010/2/27b</td>
<td>168.27 ± 18.02</td>
<td>100.00 ± 2.00</td>
<td>125</td>
<td>5.3</td>
</tr>
<tr>
<td>2010/12/21a</td>
<td>111.35 ± 10.00</td>
<td>43.65 ± 2.00</td>
<td>191</td>
<td>5.4</td>
</tr>
<tr>
<td>2010/12/21b</td>
<td>99.60 ± 20.03</td>
<td>151.80 ± 2.00</td>
<td>191</td>
<td>5.4</td>
</tr>
<tr>
<td>2010/12/21c</td>
<td>334.37 ± 39.98</td>
<td>193.25 ± 2.00</td>
<td>191</td>
<td>5.1</td>
</tr>
</tbody>
</table>
**Table S6.1:** Triggered early aftershocks.

<table>
<thead>
<tr>
<th>Triggering Event</th>
<th>Epicentral distance (km)</th>
<th>Triggering time (s)</th>
<th>Station number (Jackknife ( n ))</th>
<th>Characteristic magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>2011/7/6a</td>
<td>112.63 ± 10.00</td>
<td>82.20 ± 2.00</td>
<td>152</td>
<td>5.4</td>
</tr>
<tr>
<td>2011/7/6b</td>
<td>286.46 ± 13.02</td>
<td>148.80 ± 2.00</td>
<td>152</td>
<td>5.1</td>
</tr>
<tr>
<td>2011/8/20</td>
<td>228.49 ± 10.00</td>
<td>163.18 ± 2.00</td>
<td>189</td>
<td></td>
</tr>
<tr>
<td>2012/1/10</td>
<td>168.35 ± 30.32</td>
<td>74.00 ± 5.73</td>
<td>89</td>
<td>5.1</td>
</tr>
<tr>
<td>2012/2/2a</td>
<td>82.02 ± 10.00</td>
<td>44.65 ± 2.00</td>
<td>99</td>
<td>5.2</td>
</tr>
<tr>
<td>2012/2/2b</td>
<td>197.37 ± 10.00</td>
<td>174.15 ± 2.00</td>
<td>99</td>
<td></td>
</tr>
<tr>
<td>2012/3/25a</td>
<td>204.42 ± 26.71</td>
<td>122.68 ± 2.00</td>
<td>90</td>
<td>5.2</td>
</tr>
<tr>
<td>2012/3/25b</td>
<td>119.43 ± 20.18</td>
<td>144.45 ± 2.00</td>
<td>90</td>
<td>5.2</td>
</tr>
<tr>
<td>2012/10/28a</td>
<td>53.64 ± 10.00</td>
<td>93.12 ± 2.00</td>
<td>177</td>
<td></td>
</tr>
<tr>
<td>2012/10/28b</td>
<td>135.46 ± 13.47</td>
<td>108.65 ± 2.00</td>
<td>177</td>
<td></td>
</tr>
<tr>
<td>2012/11/7a</td>
<td>109.38 ± 16.29</td>
<td>45.12 ± 11.09</td>
<td>133</td>
<td>6.5</td>
</tr>
<tr>
<td>2012/11/7b</td>
<td>258.05 ± 17.91</td>
<td>88.55 ± 2.00</td>
<td>133</td>
<td></td>
</tr>
<tr>
<td>2013/2/6</td>
<td>94.02 ± 16.13</td>
<td>110.57 ± 2.00</td>
<td>86</td>
<td>5.1</td>
</tr>
<tr>
<td>2013/10/25</td>
<td>133.18 ± 10.00</td>
<td>50.35 ± 2.00</td>
<td>149</td>
<td>5.2</td>
</tr>
<tr>
<td>2014/4/3</td>
<td>219.51 ± 10.00</td>
<td>49.50 ± 14.24</td>
<td>154</td>
<td></td>
</tr>
<tr>
<td>2014/4/12</td>
<td>189.00 ± 14.02</td>
<td>184.30 ± 2.00</td>
<td>109</td>
<td>5.4</td>
</tr>
<tr>
<td>2014/4/13a</td>
<td>80.04 ± 10.00</td>
<td>65.28 ± 2.00</td>
<td>137</td>
<td></td>
</tr>
<tr>
<td>2014/4/13b</td>
<td>212.39 ± 10.00</td>
<td>160.03 ± 2.00</td>
<td>137</td>
<td></td>
</tr>
<tr>
<td>2014/4/19a</td>
<td>162.90 ± 10.00</td>
<td>44.83 ± 2.00</td>
<td>93</td>
<td>5.5</td>
</tr>
<tr>
<td>2014/4/19b</td>
<td>142.21 ± 15.09</td>
<td>114.30 ± 2.00</td>
<td>93</td>
<td></td>
</tr>
<tr>
<td>2014/4/19c</td>
<td>333.49 ± 10.00</td>
<td>135.70 ± 2.00</td>
<td>93</td>
<td>5.3</td>
</tr>
<tr>
<td>2015/5/5a</td>
<td>119.40 ± 12.28</td>
<td>42.88 ± 2.00</td>
<td>61</td>
<td>5.2</td>
</tr>
<tr>
<td>2015/5/5b</td>
<td>315.38 ± 10.00</td>
<td>104.62 ± 2.00</td>
<td>61</td>
<td>5.1</td>
</tr>
</tbody>
</table>

Epicentral distances (± standard error), triggering times (± standard error), station numbers used for back-projection and jackknife resampling (\( n \)), and characteristic magnitudes of dynamically triggered early aftershocks.
Table 6.S2: Mainshock-early-aftershock pairs in Fig. S8

<table>
<thead>
<tr>
<th>Event</th>
<th>ISC ID</th>
<th>Time</th>
<th>Location</th>
<th>Magnitude</th>
<th>Magnitude type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fig. S8A mainshock</td>
<td>7421058</td>
<td>2004-10-23, 08:55:58.63</td>
<td>37.3032/138.7706/8.5</td>
<td>Ms</td>
<td>6.4</td>
</tr>
<tr>
<td>Fig. S8A aftershock</td>
<td>7421061</td>
<td>2004-10-23, 08:59:37.28</td>
<td>37.3158/138.8497/20.4</td>
<td>mb</td>
<td>5.3</td>
</tr>
<tr>
<td>Fig. S8B mainshock</td>
<td>10697894</td>
<td>2006-04-20, 23:25:02.96</td>
<td>61.0398/167.0973/26.6</td>
<td>Ms</td>
<td>7.6</td>
</tr>
<tr>
<td>Fig. S8B aftershock</td>
<td>10697895</td>
<td>2006-04-20, 23:28:04.11</td>
<td>60.8687/167.2054/10.0</td>
<td>mb</td>
<td>5.8</td>
</tr>
<tr>
<td>Fig. S8C mainshock</td>
<td>13227629</td>
<td>2008-05-07, 16:02:04.92</td>
<td>36.2246/141.5949/32.6</td>
<td>Ms</td>
<td>6.6</td>
</tr>
<tr>
<td>Fig. S8C aftershock</td>
<td>10943871</td>
<td>2008-05-07, 16:03:40.60</td>
<td>36.2265/141.7973/36.6</td>
<td>Ms</td>
<td>5.9</td>
</tr>
<tr>
<td>Fig. S8D mainshock</td>
<td>15155483</td>
<td>2010-09-03, 16:35:46.48</td>
<td>-43.3608/171.9023/4.0</td>
<td>Ms</td>
<td>7.3</td>
</tr>
<tr>
<td>Fig. S8D aftershock</td>
<td>601459571</td>
<td>2010-09-03, 16:37:02.62</td>
<td>-43.5543/172.3237/10.0</td>
<td>ML</td>
<td>5.9</td>
</tr>
<tr>
<td>Fig. S8E mainshock</td>
<td>15265938</td>
<td>2010-10-25, 19:37:30.47</td>
<td>-3.0065/100.3700/22.6</td>
<td>Ms</td>
<td>6.1</td>
</tr>
<tr>
<td>Fig. S8E aftershock</td>
<td>15609469</td>
<td>2010-10-25, 19:39:26.42</td>
<td>-3.0194/100.4880/28.0</td>
<td>Ms</td>
<td>5.7</td>
</tr>
<tr>
<td>Event</td>
<td>Time</td>
<td>Location (Lat°/Lon°/Depth(km))</td>
<td>Magnitude (Mw)</td>
<td>Distance from mainshock (km)</td>
<td>Back-projection deviation from ISC epicenter (km)</td>
</tr>
<tr>
<td>-----------</td>
<td>------------------</td>
<td>-------------------------------</td>
<td>---------------</td>
<td>-----------------------------</td>
<td>---------------------------------</td>
</tr>
<tr>
<td>Fig. S9A</td>
<td>2003/11/6, 10:38:4.3</td>
<td>-19.26/168.89/113.7</td>
<td>6.6</td>
<td>341.87</td>
<td>13.87</td>
</tr>
<tr>
<td>Fig. S9B</td>
<td>2004/10/6, 14:40:39.9</td>
<td>35.95/139.92/64</td>
<td>5.7</td>
<td>439.60</td>
<td>9.88</td>
</tr>
<tr>
<td>Fig. S9C</td>
<td>2004/12/28, 11:17:43.9</td>
<td>4.73/95.21/36</td>
<td>5.7</td>
<td>346.81</td>
<td>9.93</td>
</tr>
<tr>
<td>Fig. S9D</td>
<td>2004/12/29, 1:50:52.6</td>
<td>9.11/93.76/8</td>
<td>6.0</td>
<td>259.55</td>
<td>14.10</td>
</tr>
<tr>
<td>Fig. S9E</td>
<td>2005/5/26, 10:8:26.8</td>
<td>5.69/93.21/30</td>
<td>5.7</td>
<td>271.86</td>
<td>22.22</td>
</tr>
<tr>
<td>Fig. S9F</td>
<td>2005/7/30, 15:13:20.1</td>
<td>5.18/94.48/38</td>
<td>5.8</td>
<td>395.30</td>
<td>14.03</td>
</tr>
<tr>
<td>Fig. S9G</td>
<td>2008/5/25, 8:21:48.7</td>
<td>32.57/105.42/10</td>
<td>6.1</td>
<td>263.86</td>
<td>35.94</td>
</tr>
<tr>
<td>Fig. S9H</td>
<td>2010/3/21, 18:31:4.1</td>
<td>-36.35/-73.16/36.3</td>
<td>5.5</td>
<td>229.93</td>
<td>9.93</td>
</tr>
<tr>
<td>Fig. S9I</td>
<td>2011/7/31, 14:34:47.3</td>
<td>-17.02/171.58/10</td>
<td>6.1</td>
<td>383.12</td>
<td>22.29</td>
</tr>
<tr>
<td>Fig. S10A</td>
<td>2012/2/11, 2:58:17</td>
<td>-37.46/-73.88/20.2</td>
<td>5.6</td>
<td>291.48</td>
<td>22.25</td>
</tr>
<tr>
<td>Fig. S10B</td>
<td>2013/2/20, 0:9:18.1</td>
<td>-10.78/166.9/6</td>
<td>5.6</td>
<td>224.55</td>
<td>9.97</td>
</tr>
<tr>
<td>Fig. S10C</td>
<td>2014/4/18, 4:13:12.2</td>
<td>-11.15/164.81/10</td>
<td>6.1</td>
<td>283.63</td>
<td>14.10</td>
</tr>
<tr>
<td>Fig. S10D</td>
<td>2014/3/20, 21:15:11.4</td>
<td>-5.17/152.73/25.9</td>
<td>5.7</td>
<td>297.64</td>
<td>0.00</td>
</tr>
<tr>
<td>Fig. S10E</td>
<td>2015/3/20, 15:42:52.5</td>
<td>-4.77/154.84/25.7</td>
<td>5.7</td>
<td>335.27</td>
<td>9.94</td>
</tr>
<tr>
<td>Fig. S10F</td>
<td>2015/5/18, 17:4:52.3</td>
<td>-7.16/154.4/8.1</td>
<td>5.7</td>
<td>334.47</td>
<td>0.00</td>
</tr>
</tbody>
</table>
References


Hill, D., et al. (1993), Seismicity remotely triggered by the magnitude 7.3 Landers, California, earthquake, Science 260, 1617–1623.


Supplementary materials are available on Science Online.


Nakamura, W., N. Uchida, T. Matsuzawa (2016), Spatial distribution of the faulting types of small earthquakes around the 2011 Tohoku-oki earthquake: A comprehensive search using template events, J. Geophys. Res. 121, 2591.


Walker, K. T., P. M. Shearer (2009), Illuminating the near-sonic rupture velocities of the intracontinental Kokoxili Mw 7.8 and Denali fault Mw 7.9 strike-slip earthquakes with global P wave back projection imaging, *J. Geophys. Res.* 114.


Omori, F., On the after-shocks of earthquakes (1895), *College of Science, Imperial University*.


Chapter 7

Possible activation of splay faults during the 2006 Java tsunami earthquake

Abstract

The 2006 Mw 7.8 Java earthquake was a tsunami earthquake, exhibiting frequency-dependent seismic radiation along strike. High-frequency global back-projection results suggest two distinct rupture stages. The first stage lasted $\sim65$ s with a rupture speed of $\sim1.2$ km/s, while the second stage lasted from $\sim65$ to 150 s with a rupture speed of $\sim2.7$ km/s. In addition, P-wave high-frequency radiated energy and fall-off rates indicate a rupture transition at $\sim60$ s. High-frequency radiators resolved with back-projection during the second stage spatially correlate with splay fault traces mapped from residual free-air gravity anomalies. These splay faults also collocate with a major tsunami source associated with the earthquake inferred from tsunami first-crest back-propagation simulation. These correlations suggest that the splay faults may have been reactivated during the Java earthquake, as has been proposed for other tsunamigenic earthquakes, such as the 1944 Mw 8.1 Tonankai earthquake in the Nankai Trough.
7.1 Introduction

Tsunami earthquakes generate disproportionately large tsunamis and are often characterized by having anomalously large magnitudes at low seismic frequencies compared to magnitudes at higher frequency (Kanamori, 1972; Kanamori and Kikuchi, 1993). The July 17, 2006 Java earthquake was a classic tsunami earthquake with body-wave magnitude mb = 6.1, surface-wave magnitude Ms = 7.1, and moment magnitude Mw = 7.7 (International Seismological Centre, 2013; Ekström et al., 2012). Such a large variation in magnitude estimates exceeds that seen for most earthquakes and indicates a likely deficiency in high-frequency radiation compared to low-frequency radiation (Newman and Okal, 1998; Ammon et al., 2006). The 2006 Java earthquake initiated at shallow depth (20 km, ISC; Figure 7.1) and ruptured eastward along the trench axis for ~200 km (Ammon et al., 2006; Bilek and Engdahl, 2007). Given the source dimension, the unusually long source duration (~185 s) indicates anomalously slow rupture propagation for the event (Ammon et al., 2006; Bilek and Engdahl, 2007). The earthquake generated a large tsunami (~8 m) resulting in over 800 fatalities (Fritz et al., 2007; Fujii and Satake, 2006; Mori et al., 2007). This was the second tsunami earthquake that struck the Java region since instrumental records began, and a Mw 7.8 earthquake in June 1994 produced an even larger tsunami (~13 m), resulting in 250 fatalities (Abercrombie et al., 2001; Mori et al., 2007). These two earthquakes are only 600 km apart, highlighting the major tsunami hazard along the south coast of Indonesia (Mori et al., 2007). Is the Java trench prone to more tsunami earthquakes and if so, what properties of the margin promote this type of rupture?

Finite-fault slip models of the 2006 Java earthquake suggest a smooth slip distribution with an unusually slow (~1 km/s) rupture propagation (Figure 7.2b). Finite-fault slip models obtained from body waves (P and SH waves, ~0.001–0.2 Hz) have similar slip distributions, with the largest slip concentrated near the hypocenter (Figure 7.2b) (Ammon et al., 2006; Bilek and Engdahl, 2007; Yagi and Fukahata, 2011; Ye et al., 2016a,b). In contrast, finite-fault slip models obtained from both body and surface waves (both Rayleigh and Love waves) suggest the largest slip is close to the trench and is ~50 km east up-dip from the hypocenter.
Surface waves have been shown to be effective at resolving near-trench slip distributions, which are difficult to resolve just with body waves (Shao et al., 2011).

The 2006 Java earthquake was one of the best-recorded tsunami earthquakes with modern instruments. Combining the wealth of data with new observational approaches enables us to investigate the earthquake in great detail. We first analyze bathymetry and gravity anomalies in conjunction with active-source seismic profiles to constrain margin structure and the location of splay faults. We then back-propagate first-crest arrivals in tsunami waveforms of five nearby tide gauges at various azimuths to locate tsunami sources. Finally, we build on published kinematic slip models of the 2006 Java earthquake source by performing global P-wave back-projection using two different frequency bands to examine the earthquake kinematics and analyze the P-wave source spectrum to investigate the rupture dynamics. Our high-frequency back-projection results suggest a unilateral rupture extending \( \sim 200 \) km with a slow first-stage rupture (\( \sim 1.2 \) km/s) from west to east until \( \sim 65 \) s and a fast second-stage rupture (\( \sim 2.7 \) km/s) from \( \sim 65 \) to 150 s. This is supported by time-dependent P-wave radiated energy estimates and high-frequency fall-off rates, which also suggest a transition around \( \sim 60 \) s. The spatial correlation observed between the stage-two rupture imaged by back-projection and splay fault traces delineated by gravity data suggests that splay faults may have been reactivated during the 2006 Java earthquake and possibly contributed to tsunamigenesis, which is further supported by the tsunami back-propagation analysis.

7.2 Tectonic Setting and Residual Gravity Anomaly

The Java subduction zone accommodates underthrusting of the Indo Australian plate beneath Eurasia at approximately 67 mm/yr (Tregoning et al., 1994). The incoming plate offshore western Java and is structurally complex, hosting a dense population of seamounts and the Roo Rise oceanic plateau (Shulgin et al.,
2011). The forearc is characterized by an outer-arc high, which typically extends 100 km from the trench-axis with water-depths of 2-3 km (Kopp et al., 2002; Plan- ert et al., 2010). Landward of the outer-arc high, the Lombok forearc basin extends along the coastline of Java for over 400 km.

Short wavelength topographic and gravimetric anomalies can illuminate detailed structure of the overthrusting and subducting plates. These short wavelength features can be effectively extracted using spectral averaging methods designed specifically to suppress steep topographic and gravimetric gradients across subduction zones (Bassett and Watts, 2015a,b). Application of these methods to the Java subduction zone reveals a long array of lineations in the residual gravity field, encompassing the full ∼100 km trench-normal width of the outer-arc high and the full ∼800 km along-strike extent of the Java margin (Arrows, Figure 7.1). Where 2D seismic reflection and refraction profiles traverse the forearc (Red line, Figure 7.1), the gravity lineations are consistent with the locations of splay faults imaged in the overthrusting plate (Kopp et al., 2009). The residual gravity field allows us to extend this interpretation along strike, which indicates that the outer-arc high is pervasively faulted along strike and that splay faults are almost certainly present within the source region of the 1994 and 2006 tsunami earthquakes.

7.3 Tsunami Back-propagation and Seismic Back-projection

7.3.1 Tsunami Tide Gauge Back-propagation

To constrain tsunami source locations, we perform tsunami back propagation with five nearby tide gauges recording the tsunami of the 2006 Java earthquake (Figure 7.2c). We first estimate initial and first-crest arrivals in tsunami waveforms. With the first-crest arrivals, tsunami back-propagation is simulated from the tide gauges to possible source locations of sea surface displacements (Figure S2,3). We consider Gaussian-shaped seafloor uplifts as tsunami sources centered at the gauge locations with half-widths of 2 km. Tsunami propagation is computed with nonlin-
ear shallow water-wave equations (Liu et al., 1995) and General Bathymetric Chart of the Oceans (GEBCO) 30 arc-second bathymetry (Weatherall et al., 2015). To account for the long-wave dispersion that is missing in our numerical simulations, the observed first-crest arrivals are shifted 1% earlier for all stations. This shift is derived from comparisons between tsunami models and observations from recent great earthquakes (Tsai et al., 2013; Watada et al., 2014).

Regions bounded by multiple arcs of the back-propagated tsunami first-crest wavefronts are the possible tsunami source areas (Figure 7.2c). These tsunami sources were excited by local large seafloor displacements. To account for the uncertainty in tsunami modeling, we identify source regions using the first-wave bands instead of crest lines, which are contours with tsunami amplitude greater than 50% of the crest (Figure 7.2c). The back-propagation results suggest two possible main sources for the observed tsunami (Table S1, Figure 7.2c, Figure S2.3). The first source is bounded by two arcs close to the epicenter (from Christmas and Hillarys), and the second source is bounded by four arcs close to the second stage high-frequency seismic radiation (from Benoa, Cocos, Broome, and Hillarys). The second source is more than 100 km eastward of the epicenter. Tsunami back-propagation of the Cocos gauge, west of the 2006 Java earthquake, only tracks the eastern tsunami source, suggesting that the western source may be a weaker source compared to the source located to the east. The tsunami sources we resolve are generally consistent with Fujii and Satake (2006), which suggested a major tsunami source ∼150 km east of the epicenter. The eastward source is within the zones of inferred splay faults and correlates with high-frequency radiation from ∼60 to 150 s (Figure 7.2c).

7.3.2 Seismic P-wave Back-projection

We perform P-wave back-projection using the procedure described in Fan and Shearer (2015), detailed in the supplementary materials. We use vertical-component velocity records from the International Federation of Digital Seismograph Networks (FDSN) seismic stations that are available and distributed by the Data Management Center (DMC) of the Incorporated Research Institutions for
Seismology (IRIS). Because back-projection techniques do not make assumptions about fault geometry or rupture velocity, they often can resolve complex earthquake behavior, such as variable rupture velocity, multiple events, and very early aftershocks (Ishii et al., 2005; Xu et al., 2009; Koper et al., 2011; Kiser and Ishii, 2011; Meng et al., 2012; Satriano et al., 2012; Nissen et al., 2016; Wang et al., 2016). Global back-projection is particularly effective in detecting frequency-dependent radiation because of its superior spatial resolution (e.g., Walker et al., 2005; Yagi et al., 2012; Okuwaki et al., 2014). In this study, P-wave velocity seismograms are filtered into two frequency bands, a high-frequency (HF) band (0.3-1 Hz) and a low-frequency (LF) band (0.05-0.3 Hz), to examine potential frequency-dependent seismic radiation. We obtain a peak-power time function with a non-overlapping 2 s window that is the maximum back-projected power of the potential sources (location of high-frequency bursts) (Kiser and Ishii, 2013; Fan and Shearer, 2016). The back-projection snapshots are computed with 20-s stacking windows and are normalized by the maximum power within each window (Figure 7.2, 7.3). The robustness of the resolved snapshots is assessed by jackknife resampling (Efron and Tibshirani, 1994; Fan and Shearer, 2016) and we reject snapshots with peak-power spatial standard errors greater than 0.5° for either latitude or longitude (≈ 50 km). No post-processing is applied to the final images.

Our back-projection peak-power time functions agree with prior studies that indicate the 2006 Java earthquake had an abnormally long duration. The LF peak-power time function suggests it lasted ≈180 s, which is consistent with long-period finite-fault modeling (e.g., Ammon et al., 2006), while the HF peak-power time function indicates at least ≈150 s of continuous seismic radiation (Figure 7.2a). Stacked envelope functions (1-5 Hz) with globally distributed station also suggest a very long rupture duration lasting ≈150 s (Figure S5).

In the first 60 s, both LF and HF back-projection results show similar seismic radiation (Figure 7.3). The time-integrated back-projection image (Figure 7.3a,c) suggests that the bulk of seismic radiation was excited around the epicenter during the early phase of the earthquake (Figure 7.3b,d). After 60 s, the back-projection snapshots indicate frequency-dependent seismic radiation (Fig-
Figure 7.3). HF back-projection snapshots show west-to-east linear rupture propagation from 60 to 160 s. The 100 – 120 s LF back-projection snapshot seems to correspond to the overall rupture propagation (Figure 7.2b,7.3a) because its location and average rupture speed agree with the expected rupture propagation (∼1 km/s), while the 120 – 180 s LF back-projection snapshots are significantly down-dip of the mainshock epicenter (∼70 to ∼170 km), suggesting possible nearby triggered early aftershocks (Fan and Shearer, 2016). Finite-fault models have limited resolution for the later stage of the mainshock rupture, suggested by their discrepancies, showing minor to negligible slip after 140 s (Ammon et al., 2006; Bilek and Engdahl, 2007; Yagi and Fukahata, 2011; Ye et al., 2016a,b). As shown by globally recorded 0.02–0.05 Hz P-wave, identifiable phases are present from 120 s to 200 s (Figure S7). These phases are coherent in the azimuthal range of the stations used for back-projection, which is likely why back-projection detected those coherent energy bursts. The polarity patterns of these phases are different than those of the mainshock, and the amplitudes vary gradually with azimuth (Figure S7). Based on the observations, these phases are unlikely to be water phases nor part of the mainshock because of the azimuthally dependent radiation pattern. However, the noisy records leave the focal mechanisms of these possible early aftershocks yet to be determined with future analysis.

A plot of cumulative distance as a function of time (Figure 7.4c) provides average estimates of rupture speeds. The cumulative distance was computed from HF back-projection peak-power locations (20 s stacking window with 1 s temporal increment, Figure 7.2b). The results suggest an increase in rupture velocity around 65 s (Figure 7.4c). Similar to the rupture speed resolved from finite-fault inversions (e.g., Ammon et al., 2006; Yagi and Fukahata, 2011), the first stage ruptured slowly (∼1.2 km/s) for about 65 s, while the second stage ruptured much faster (∼2.7 km/s) from ∼65 to 150 s. Intriguingly, the HF back-projection snapshots indicate west to east migration in seismic radiation for the second stage of the event (Figure 7.2b,7.3d), which spatially correlates with the location of slay fault traces inferred from residual gravity anomalies (Figure 7.1).
7.4 Time-dependent P-wave Spectrum Analysis

Spectral analysis of the far-field P-wave pulse provides constraints on rupture dynamics through measures of radiated energy and estimates of stress drop, radiation efficiency, and rupture velocity (e.g., Houston and Kanamori, 1986; Houston, 1990; Newman and Okal, 1998; Baltay et al., 2010; Conyers and Newman, 2011; Baltay et al., 2014; Denolle et al., 2015; Denolle and Shearer, 2016). By studying high-frequency fall-off rates and radiated energy during earthquakes, we can potentially gain insight into dynamic rupture processes (e.g., Denolle et al., 2015). High-frequency fall-off rates provide a direct constraint on the type of seismic radiation at the source and complement back-projection observations. Radiated energy enters directly into the earthquake energy budget and thus constrains the earthquake dynamics. We select stations between 30° and 90° epicentral distance, remove the instrumental response to retrieve displacement seismograms (see Figure S8 and supplementary materials for data processing details). We obtain the P-wave moment-rate source spectra through a two-step empirical Green’s function approach (Denolle and Shearer, 2016) and use one local event to remove the path effects. Because the Green’s function is assumed stationary, we perform a time-dependent analysis, similar to using spectrograms (Denolle et al., 2015), to estimate the radiated energy (in both LF and HF) and high-frequency falloff rates throughout the 180 s rupture.

HF radiated energy (normalized by the moment $M_0 = 4.1 \times 10^{20}$ Nm) and HF fall-off rates (0.3-1 Hz, Figure 7.4a,b) suggest a possible change in the character of seismic radiation. Both measures show that the initiation of the seismic radiation occurs over 20 s, suggesting a slow nucleation. From that point on, the temporal evolution of both HF scaled radiated energy and HF fall-off rates support the proposed multi-stage rupture (Figure 7.4a,b). On one hand, the HF scaled radiated energy slowly reaches its maximum at $\sim 60$ s and plateaus after. On the other hand, the LF scaled energy indicates at least three sub-events, one between initiation and 50 s, one between 60 s and 100 s, and finally starting from 110 s until the end (Figure 7.4a). The HF fall-off rate initiates at 3 (strong depletion of high frequencies) and stabilizes to 2 after $\sim 60$ s with a mean
value of 2.4 (Figure 7.4a,b). Both time-dependent scaled energy and falloff rates support the back-projection analysis and imply that the 2006 Java earthquake had a possible transition around 60 s in high-frequency seismic radiation.

7.5 Discussion

Tsunami waveform inversion suggests that the tsunami source of the 2006 Java earthquake was about 200 km long with the largest slip (~2.5 m) stably located about 150 km east of the epicenter, regardless of the assumed earthquake rupture velocity (Fujii and Satake, 2006). This tsunami-derived slip model is significantly different from the seismic slip models (Fujii and Satake, 2006; Ammon et al., 2006; Bilek and Engdahl, 2007; Yagi and Fukahata, 2011; Ye et al., 2016a), which suggest the largest slip occurred within 50 km of the epicenter. The slip model discrepancies may be attributed to two possibilities: (1), tsunami data and seismic data have different spatial sensitivities over the slip distribution (e.g., Melgar et al., 2016; Jiang and Simons, 2016). (2), slip at the plate interface is not the only source responsible for the observed tsunami. As tsunami waves are most sensitive to seafloor displacements rather than the earthquake slip distribution (Jiang and Simons, 2016), the resolved tsunami sources (Figure 7.2c,S2,3) may suggest that more than one fault was involved for the seafloor displacements.

Rupture velocity, radiated energy, and high-frequency fall-off rates suggest that the 2006 Java earthquake radiated high-frequency energy in a two-stage fashion, with a transition around 60 s (Figure 7.4). Stage one was characterized by a rupture velocity of ~1.2 km/s and high-frequency fall-off rate decays from 3 to 2 after 55 s (Figure 7.4b), suggesting an initial deficiency in high-frequency radiation. In contrast, stage two was enriched in high-frequency radiation and the high-frequency radiators migrated from west to east at more than twice the rupture speed observed during stage one (~2.7 km/s). This atypical abrupt two-stage HF energy release may indicate that more than one source generated the high-frequency radiation.

The observations cannot distinguish whether the rupture transition oc-
curred sharply or gradually. The possible HF radiation transition timing ranges from 55 s to 65 s, leaving the exact initiation time of the second stage unclear. It is also possible that the second stage started around 90 s, after which the average rupture velocity shows a larger increase (Figure 7.4c). Nonetheless, the differences between the two stages are robust. The observed two-stage rupture may reflect solely the rupture complexities along strike, which has been reported for large earthquakes (e.g., *Kiser and Ishii*, 2011; *Wei et al.*, 2011). Fault geometry, heterogeneous initial stress at the plate interface, or heterogeneous friction properties could all produce along-strike variations of high-frequency radiation (e.g., *Madariaga*, 1977; *Bernard and Madariaga*, 1984; *Spudich and Frazer*, 1984; *Fukahata et al.*, 2014; *Denolle et al.*, 2015; *Bassett et al.*, 2016). Sea floor dense seismic and geodetic networks would improve understanding of rupture propagation and provide more conclusive evidence to determine the details of such complex rupture, suggesting future research directions.

Alternatively, the collocations of the second-stage high-frequency radiators with splay faults and the eastward tsunami source likely locating near the splay faults may indicate splay-fault reactivation during the 2006 Java earthquake (Figure 7.2,7.3). Active-source seismic profiles 100 km west of the 2006 Java earthquake epicenter resolve steep north-dipping splay faults that originate from the plate interface and rise to the sea floor (*Kopp et al.*, 2009). The surface traces of the splay faults are well correlated with their locations inferred by residual gravity anomalies. These splay faults extend along the forearc and are present in the vicinity of both the 2006 and 1994 Java tsunami earthquakes (Figure 7.1). The ISC catalog (1993-2013) locates shallow seismicity close to the 2006 Java earthquake (59 earthquakes shallower than 10 km, 122 earthquakes at 10 to 20 km), suggesting splay faults being seismically active (Figure 7.1) (*International Seismological Centre*, 2013). Although back-projection method does not have the depth resolution to discriminate between radiation from the splay faults and from the plate interface, the transition in HF seismic radiation and the strong spatial correlation between the stage-two rupture and the splay fault traces suggest that reactivated splay faults may have been a key source of HF seismic radiation and seafloor displacement.
during the second-stage rupture. (Figure 7.2, 7.3).

Splay fault activation during the mainshock rupture has been reported for earthquakes in the Japan, Kuril, Alaska, and Sumatra subduction zones (Plafker, 1969, 1972; Fukao, 1979; Moore et al., 2007; DeDontney and Rice, 2012; Waldhauser et al., 2012). Numerical models have also validated the reactivation feasibilities (Wang and He, 1999; Kame et al., 2003; Wang and Hu, 2006; Wendt et al., 2009; Tamura and Ide, 2011; DeDontney and Hubbard, 2012). In Nankai, the presence of megasplay fault in conjunction with evidence of large-scale sediment slumping suggests that the splay faults were likely activated during the 1944 Mw 8.1 Tonankai earthquake (Moore et al., 2007). Similarly, the 1964 Mw 9.2 Alaska earthquake also activated splay faults (Plafker, 1969, 1972). Tsunami-wave analysis with space geodesy also suggests possible involvement of splay faults during the 2004 Mw 9.3 Sumatra-Andaman earthquake (DeDontney and Rice, 2012). The steep geometry of these faults may substantially increase vertical seafloor uplift thus enhancing tsunamigenesis (Moore et al., 2007). We suggest a similar scenario may have occurred along the Java trench, with coseismic splay fault reactivation providing one viable mechanism to explain our observations.

### 7.6 Conclusions

The 2006 Mw 7.8 Java earthquake ruptured more than 200 km from west to east, lasting for more than ~180 s. Finite-fault slip models suggest a smooth and slow rupture with the largest slip patch within ~50 km away from the hypocenter (Ammon et al., 2006; Bilek and Engdahl, 2007; Yagi and Fukahata, 2011; Ye et al., 2016a,b), which is supported by our low-frequency back-projection results. In contrast, high-frequency global back-projection results, time-dependent estimate of radiated energy, and fall-off rates suggest a unilateral two-stage rupture. The first stage ruptured with an unusually low rupture speed of ~1.2 km/s, agreeing well with that of finite-slip fault models (e.g., Ammon et al., 2006), while the second stage rupture was much faster with a rupture speed of ~ 2.7 km/s. Although back-projection cannot resolve the depth of the radiators, the abrupt transition
of both kinematic and dynamic rupture characteristics and the strong spatial correlation between the second stage rupture and splay fault traces delineated from residual gravity anomalies may indicate that the splay faults were possibly reactivated during the earthquake and acted as the primary source of high-frequency radiation during the second stage rupture. The hypothesis is further supported by the tsunami first crest arrival back-propagation, which shows at least two sources were involved for the observed tsunami. The two sources were separated more than 100 km with the first source close to the epicenter and the second source spatially collocating with the inferred splay fault traces. The residual gravity anomalies delineate multiple trench parallel splay faults near both the 1994 and 2006 Java tsunami earthquakes, raising concerns of enhanced tsunami hazard of the region. Similar tsunamigenic earthquakes, such as the 1944 Mw 8.1 Tonankai earthquake and the 2004 Mw 9.3 Sumatra-Andaman earthquake, have been proposed to also activate splay faults during the coseismic rupture propagation, and splay fault networks may play a critical role in enhancing tsunamigenesis during large megathrust earthquakes.

Acknowledgments

We thank Lingling Ye for sharing her finite-fault slip model. Finite-fault slip model of Yagi and Fukahata (2011) is downloaded from the Source Inversion Validation (SIV) database (SRCMOD, http://equake-rc.info/) (Mai et al., 2016). Finite-fault slip model obtained with both body and surface waves is downloaded from the U.S. Geological Survey National Earthquake Information Center (Dупетел et al., 2011; Hayes, 2011; Hayes et al., 2011). The facilities of IRIS Data Services, and specifically the IRIS Data Management Center, were used for access to waveforms, related metadata, and/or derived products used in this study. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience and EarthScope (SAGE) Proposal of the National Science Foundation under Cooperative Agreement EAR-1261681. Tsunami arrival times are estimated for waveforms compiled from University of Hawaii Sea Level Center
and Bureau of Meteorology, Research Centre, Australian Government (Fujii and Satake, 2006). The earthquake catalog was downloaded from the International Seismological Center (ISC). The bathymetry and gravity data were processed with the Generic Mapping Tools (GMT) (Wessel and Smith, 1991; Wessel et al., 2013). This work was supported by National Science Foundation grant EAR-1620251 at Scripps Institution of Oceanography, UC San Diego.

Chapter 7, in full, has been submitted for publication of the material as it may appear in Tectonophysics: Fan, W., D. Bassett, M. A. Denolle, P. M. Shearer, C. Ji, and J. Jiang, Possible activation of splay faults during the 2006 Java tsunami earthquake, Tectonophysics, submitted. I was the primary investigator and author of this paper.
Figure 7.1: Residual free-air gravity anomaly, splay faults at Java subduction zone and shallow seismicity near the 2006 Java tsunami earthquake. Black arrows show splay faults revealed by residual gravity. Insert: black circles are earthquakes (EQ) from 1993-2013 ISC catalog with $M > 4$ and depth shallower than 10 km, gray circles are earthquakes (EQ) from 1993-2013 ISC catalog with $M > 4$ and depth in between of 10 and 20 km (International Seismological Centre, 2013). Black lines are the interpreted fault traces. Red line is coincident seismic reflection and refraction profile SO137-03/SO138-05. Trench-axis is from Bassett and Watts (2015a,b).
Figure 7.2: Seismic back-projection and tsunami back-propagation results and finite-fault slip models of the 2006 Java earthquake. (a), Peak-power time functions of two frequency bands. Peak-power time functions are self-normalized. (b), Finite-slip model obtained with both body and surface waves are the filled contours from USGS, NEIC. Finite-slip model from Ye et al. (2016a,b) is contoured from 0.5 to 4.5 m with 2 m separation, finite-slip model from Yagi and Fukahata (2011) is contoured from 0.5 to 2.5 m with 1 m separation. Diamonds show the peak-energy locations of high-frequency back-projection with 20 s averaging window and 1 s time increment. Stations used for back-projection and their P-wave polarity with the GCMT focal-mechanism are shown as inserts. The subduction geometry is from Slab 1.0 with 20 km separation (Hayes et al., 2012). (c), Tsunami wave first peak back-propagation results of five tide gauges. The solid lines show the first crest back-propagations with the shades regions of 50% crest amplitude. Stations are shown in the insert and listed in Table S1. Colored contours are high-frequency (HF, 0.3-1 Hz) 20 s snapshots.
Figure 7.3: Back-projection results. (a), (b), Low-frequency (0.05-0.3 Hz) back-projection time-integrated energy release and snapshots. (c), (d), High-frequency (0.3-1 Hz) time-integrated energy release and snapshots. The background bathymetry gradient is from Sandwell et al. (2014) and Garcia et al. (2014). Low-frequency back-projection is contoured above 50% normalized energy contours, high-frequency back-projection is contoured above 20% normalized energy contours.
Figure 7.4: Radiated energy of high-frequency (HF, 0.3-1 Hz) and low-frequency (LF, 0.05-0.3 Hz), high-frequency fall-off rates from P-wave spectrum analysis, and rupture velocity inferred from cumulative distance as a function of time. (a), Radiated energy in two frequency bands normalized by the moment ($M_0 = 4.1 \times 10^{20}$ Nm). (b), High-frequency fall-off rates evolution with mean fall-off rates as 2.4. (c), Cumulative distance as a function of time obtained from HF back-projection with 20 s averaging window and 1 s time increment.
7.7 Supplementary Materials

Text S1: Back-projection data and method Data Processing

1. Globally recorded P-waves are filtered in two frequency bands, 0.05–0.3 Hz (low-frequency, LF) and 0.3–1 Hz (high-frequency, HF).

2. Seismograms are visually inspected, and only traces with clear P-wave initial arrivals are kept.

3. Based on the GCMT solution (Ekström et al., 2012), stations with theoretical negative lower-hemisphere polarities are removed.

4. Centered at the epicenter, the Earth surface is divided into 1° by 1° azimuthal-epicentral-distance grids. Within each cell, traces are aligned with cross-correlation, and only the station with the highest cross-correlation coefficient is kept per grid.

5. The remaining 68 traces are aligned with cross-correlation with the initial few seconds of P-waves separately in two frequency bands (Houser et al., 2008). For LF, the cross-correlating window is from -3 s to 4 s based on the IASP91 (Kennett and Engdahl, 1991) predicted arrival with allowed maximum time shifts as 5 s. For HF, the cross-correlating window is from -1 s to 4 s based on the IASP91 (Kennett and Engdahl, 1991) predicted arrival with allowed maximum time shifts as 4 s. The alignment is applied to neutralize 3D velocity structure influence. No polarity flips are allowed during the alignment. (Figure S2).

6. The potential sources are gridded at 10-km horizontal spacing, and fixed at the hypocentral depth (20 km). The grid latitudes range from $-12^\circ$ to $-6.6^\circ$, and grid longitude range from 105.5° to 111.1° (600 km by 600 km).

7. Back-projection is performed with $N$th root stacking ($N = 4$), which can improve spatial resolutions of back-projection images at the cost of losing absolute amplitude information (Rost and Thomas, 2002; Xu et al., 2009) (Figure S3). When performing back-projection, the records are normalized,
weighted by their average correlation coefficients obtained from the cross-correlation alignment, and inversely scaled by the number of contributing stations within 5 degrees, which downweights the noisy records and prevents dominance of dense local arrays.

**Uncertainty analysis** Following *Fan and Shearer* (2016), we perform jackknife resampling over the records used for back-projection *Efron and Tibshirani* (1994). For the *i*th resampling, we suppose we have *n* stations to estimate peak energy location (latitude and longitude) of each 20 s seismic radiation:

\[
\hat{\text{Lat}} = BP^{\text{lat}}(n) \quad \hat{\text{Lon}} = BP^{\text{lon}}(n)
\]

(7.1)

where \( \hat{\text{Lat}} \) and \( \hat{\text{Lon}} \) are location estimations. When the *i*th station is excluded, the estimations can be written as \( \text{Lat}_i = BP^{\text{lat}}_i(i) \) and \( \text{Lon}_i = BP^{\text{lon}}_i(i) \). Then the jackknife estimate of standard errors (SE) are

\[
\hat{SE}_{\text{Lat}} = \left( \frac{n-1}{n} \sum_{i=1}^{n} (\hat{\text{Lat}}_i - \bar{\text{Lat}})^2 \right)^{1/2}
\]

(7.2)

\[
\hat{SE}_{\text{Lon}} = \left( \frac{n-1}{n} \sum_{i=1}^{n} (\hat{\text{Lon}}_i - \bar{\text{Lon}})^2 \right)^{1/2}
\]

(7.3)

where \( \bar{\text{Lat}} \) and \( \bar{\text{Lon}} \) are the averages locations.

**Text S2: P-wave spectral analysis**

We use P-wave displacement seismograms (instrumental response removed) from the Java 2006 earthquake and from a nearby event M 7.0 (seismic moment \( M_0 = 4.4 \times 10^{19} \) Nm) that occurred on 2009/09/02 at 07:55:01 UTC around 40 km depth. *Denolle and Shearer* (2016) found the nearby event to have a simple single-corner frequency source model with an omega-2 decay with an approximate corner frequency of 0.13 Hz.

For each station, we calculate the P and S travel time using TauP in the IASP91 velocity model (*Kennett and Engdahl*, 1991). We also assign each station to an azimuth and takeoff angle bin. We ignore the stations where the difference in travel time between S and P is less than the source duration (180 s). We define a
noise window between 300 s before the P arrival time and 25 s before the P arrival time. We demean, detrend, and taper the noise window with a Tukey window, allowing the tapering to occur on 25 s on each edge of the window. We compute the noise amplitude Fourier transform and interpolate the discrete spectrum onto a log-spaced frequency vector. We window the P wave between 25 s before the P arrival and 25 s before the S arrival. We demean, detrend, taper, and compute the amplitude Fourier spectrum similarly to the noise window. Our signal-to-noise ratio criterion imposes that the mean amplitude spectrum of the P wave between three frequency bands (10-100 s, 2-10 s, 0.5-1.5 Hz) must exceed 2.5 times that of the noise level.

We proceed as such for the small event that serves us as an empirical Green’s function. The Green’s function is assumed stationary through time during the the Java rupture. For the time-domain analysis, we proceed as following for each azimuth and takeoff angle bin:

1. For each station within this bin, we construct a spectrogram using 20 s running windows, tapered using a Hanning taper, that each have 0.5 s of overlap.
2. We perform the same SNR selection using the noise spectrum of that station as done for the full spectrum.
3. We remove the path effects using the station records of the small event and multiplying by the source spectrum of the small event found by Denolle and Shearer (2016), obtaining a spectrogram between 0.1 and 1.5 Hz at each time increment during the 180 s of the rupture.
4. For each time increment, we stack the spectrograms and estimate:
   - Radiated energy within 0.1–1.5 Hz or other frequency bands.
   - High-frequency falloff rate.
5. We obtain time series of radiated energy and high-frequency falloff rates for each azimuth and takeoff bins that we can average for an ensemble view on the earthquake.
Table 7.S1: Tide gauges and tsunami arrivals.

<table>
<thead>
<tr>
<th>No.</th>
<th>Station name</th>
<th>Longitude (°)</th>
<th>Latitude (°)</th>
<th>$T_o$ (min)</th>
<th>$T_p$ (min)</th>
<th>$T_p'$ (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Benoa</td>
<td>115.20</td>
<td>-8.77</td>
<td>82</td>
<td>91</td>
<td>90</td>
</tr>
<tr>
<td>2</td>
<td>Christmas</td>
<td>105.67</td>
<td>-10.53</td>
<td>16</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>3</td>
<td>Cocos</td>
<td>96.87</td>
<td>-12.13</td>
<td>96</td>
<td>102</td>
<td>101</td>
</tr>
<tr>
<td>4</td>
<td>Broome</td>
<td>122.23</td>
<td>-18.00</td>
<td>280</td>
<td>290</td>
<td>287</td>
</tr>
<tr>
<td>5</td>
<td>Hillarys</td>
<td>115.70</td>
<td>-31.83</td>
<td>249</td>
<td>260</td>
<td>257</td>
</tr>
</tbody>
</table>

$T_o$: Time of arrivals in the waveform. $T_p$: Time of the first peak in the waveform. $T_p'$: Adjusted time used in the tsunami modeling. Tsunami arrival times are estimated for waveforms compiled from University of Hawaii Sea Level Center and Bureau of Meteorology, Research Centre, Australian Government.
Figure 7.S1: Residual bathymetry anomaly, splay faults at Java subduction zone and shallow seismicity near the 2006 Java tsunami earthquake. Insert: black circles are earthquakes (EQ) from 1993-2013 ISC catalog with $M > 4$ and depth shallower than 10 km, gray circles are earthquakes (EQ) from 1993-2013 ISC catalog with $M > 4$ and depth in between of 10 and 20 km (International Seismological Centre, 2013). Black lines are the interpreted fault traces. Trench-axis is from Bassett and Watts (2015a,b).
Figure 7.S2: Ocean surface displacements during the back propagation of tsunami from Christmas, Benoa, and Cocos tide gauges (red triangles). Water displacements (in color) are scaled so that the first wave front is clearly seen. The crest and outer contour (defined by 50% of the closest peak amplitude) of the first wave-front are marked by black solid and dotted lines, respectively. Event hypocenter is denoted by the red star and back-projection radiators are represented by black circles. $T_p'$: Adjusted time used in the tsunami modeling.
Figure 7.S3: Ocean surface displacements during the back propagation of tsunami from Broome and Hillarys tide gauges (red triangles). Legends are the same as Figure S2.
Figure 7.54: P-wave velocity seismogram alignments. 0.05–0.3 Hz (LF, Left panel) and 0.3–1 Hz (HF, right panel). The onset of the P wave begins at 0s. The records (station name left axis) are sorted by average cross-correlation coefficients (right axis).
Figure 7.55: Stacked envelope function. Stations used for the calculation are shown in the insert. The P-wave seismograms are filtered at 1–5 Hz. The envelope functions are calculated with a standard Hilbert transform without smoothing. Each envelope function is self-normalized before stacking. The stacked envelope function is divided by the number of stations.
Figure 7.S6: Globally distributed stations and their P-wave polarities with respect to the GCMT solution of the mainshock.
Figure 7.S7: Aligned velocity seismograms (0.02-0.05 Hz) from globally distributed stations (Figure S4). Blue lines shows the nodal plane strikes of GCMT. Predicted travel time shows as crosses with color corresponding to Figure 3.
Figure 7.S8: Stacked P-wave spectrum for the 2006 Java earthquake. The gray spectra are estimated at each station and the red spectrum is their log-average.
Figure 7.S9: Polar representation (azimuth and takeoff angles) of the falloff rate estimated between 0.3-1Hz. The mean high-frequency falloff rate is 2.4.
References


Chapter 8

Investigation of back-projection uncertainties with M6 earthquakes

Abstract

We investigate possible biasing effects of inaccurate timing corrections on teleseismic P-wave back-projection imaging of large earthquake ruptures. These errors occur because empirically-estimated time shifts based on aligning P-wave first arrivals are exact only at the hypocenter and provide approximate corrections for other parts of the rupture. Using the Japan subduction zone as a test region, we analyze 46 M6–7 earthquakes over a ten-year period, including many aftershocks of the 2011 M9 Tohoku earthquake, performing waveform cross-correlation of their initial P-wave arrivals to obtain hypocenter timing corrections to global seismic stations. We then compare back-projection images for each earthquake using its own timing corrections with those obtained using the time corrections for other earthquakes. This provides a measure of how well sub-events can be resolved with back-projection of a large rupture as a function of distance from the hypocenter. Our results show that back-projection is generally very robust and that sub-event location errors average about 20 km across the entire study region (∼700 km). The
back-projection coherence loss and location errors do not noticeably converge to zero even when the event pairs are very close (<20 km). This indicates that most of the timing differences are due to 3D structure close to each of the hypocenter regions, which limits the effectiveness of attempts to refine back-projection images using aftershock calibration, at least in this region.

\section{Introduction}

Back-projection is one of the primary tools to investigate large earthquake rupture propagation without requiring many prior assumptions about fault geometry or rupture speed. Ever since its first application to the 2004 Sumatra–Andaman earthquake (Ishii et al., 2005), back-projection has proven useful in studying complex earthquake ruptures, early aftershock detection, and hazard early warning (e.g., Walker et al., 2005; Allmann and Shearer, 2007; Koper et al., 2011; Meng et al., 2011; Kiser and Ishii, 2011, 2012; Yagi et al., 2012; Okuwaki et al., 2014; Fan and Shearer, 2016a; An and Meng, 2017). Back-projection images, in conjunction with rupture dynamics simulations and knowledge of local tectonics, have led to improved understanding of earthquake physics and subduction zone stress transfer patterns (e.g., Huang et al., 2012; Fan et al., 2016).

Back-projection takes advantage of source-receiver reciprocity. It assumes that teleseismic P-waves provide relatively undistorted records of seismic radiation, such that differences in source locations on the fault can be approximated as time shifts in the records. This permits stacking methods to be used to extract coherent signals to map earthquake rupture propagation, an adjoint approximation that is often more robust than formal inversion (Claerbout and Fomel, 2008). Back-projection works even with high-frequency waveforms, which can provide higher spatial and temporal resolution, but are often challenging to model deterministically with current forward calculation methods and Earth structure models (e.g., Komatitsch et al., 2004; Mancinelli et al., 2016). These advantages, together with the need for few prior assumptions, make back-projection suited for resolving even very complex earthquake ruptures (e.g., Walker and Shearer, 2009; Satriano
The simplicity of back-projection helps to provide stability in its results, such that different groups often get very similar back-projection source models even when different datasets and stacking approaches are used (e.g., Fan and Shearer, 2015; Yagi and Okuwaki, 2015; Grandin et al., 2015; Zhang et al., 2016; Wang and Mori, 2016). However, the method can suffer from imaging artifacts or uncertainties because it heavily relies on the station network geometry and data quality. For example, swimming artifacts commonly appear when only using regional arrays, in which back-projection energy migrates toward the arrays because of tradeoffs between the radiation origin time and source-receiver distance (e.g., Xu et al., 2009; Koper et al., 2012; Meng et al., 2012). Understanding back-projection resolution and uncertainty is important because biased back-projection source models may cause erroneous interpretations of rupture physics (e.g., Ishii et al., 2007; Walker and Shearer, 2009; Koper et al., 2012; Meng et al., 2016). However, assessing the uncertainties is challenging because back-projection relies on waveform stacking instead of optimizing a misfit function. Examples of methods used to measure back-projection resolution and uncertainty include: (1) using synthetic tests to estimate resolution kernels and the effect of depth phases, (2) seismogram resampling tests to assess the statistical significance of features, and (3) back-projection of smaller events near the mainshock of interest to provide empirical measures of image resolution (e.g., Fan and Shearer, 2016b; Fan et al., 2016). However, none of these approaches addresses model-related uncertainties, i.e., how much bias in back-projection images is caused by inaccuracies in the assumed source-receiver Green’s functions. Here we focus on the specific issue as to whether the empirical timing corrections obtained from the initial P-wave alignment for the epicenter are applicable to other parts of the fault.

At the relatively high-frequencies used in back-projection, time shifts due to 3D structure along the ray paths will prevent coherent stacking of the records if only a 1D model is assumed. Because 3D tomographic models do not yet have the resolution to predict these time shifts, the standard approach for estimating them is to align the first-arriving waves from the hypocenter using waveform cross-
correlation. These empirical time shifts are then used to correct the times for all
the hypothetical source locations in the back-projection calculation. However, the
timing corrections are only exact at the hypocenter and presumably can become
increasingly inaccurate at more distant locations on the fault, where the rays will
traverse different parts of the underlying 3D velocity structure. This will cause the
back-projection stacks to be less coherent at these locations, reducing the apparent
amplitudes of features. Perhaps more seriously, inaccurate time corrections might
bias the locations of sub-events away from their true locations.

One way to address the timing correction issue is to use records from other
earthquakes within the mainshock rupture region (often aftershocks of the main-
shock) as calibration events to obtain more accurate time shifts at varying lo-
cations on or near the fault (Ishii et al., 2007). This idea has been explored
by applying waveform cross-correlation of these aftershocks to extract empirical
time corrections for the regions close to the aftershocks (e.g., Ishii et al., 2007;
Meng et al., 2016). For the 2004 Sumatra–Andaman earthquake, the aftershock-
calibrated back-projection image shows more heterogeneous small-scale features
compared to the smoother hypocenter-calibrated image, although the main fea-
tures are very similar (Ishii et al., 2007). However, aftershocks that can be used
for calibration (M ≥ 5.5) generally do not span the entire rupture region and it
is unclear how close they need to be to the back-projection source locations in or-
der to improve the image and whether interpolation of timing corrections between
aftershock locations is a valid approach.

Here we explore these issues by studying P-wave timing corrections for 46
M6 earthquakes within the Japan subduction zone, including how they vary as a
function of event location and how these differences will affect back-projection im-
ages, in particular to what extent the empirical timing corrections fail when large
earthquakes rupture hundreds of kilometers. By back-projecting events using the
timing corrections derived from different events, we show that a single set of em-
pirical time corrections can produce reasonably unbiased images across our entire
study region (∼700 km). There is loss of coherence, particularly at higher frequen-
cies, but errors in imaged source locations are mostly less than 40 km (averaging
about 20 km) and the errors are only a weak function of distance from the calibration event. Our results suggest that it is challenging to improve back-projection images with aftershock calibration because the region of improvement is confined to the close vicinity of each aftershock, limiting the effectiveness of interpolation of timing corrections between events, given that the aftershock coverage is typically sparse and nonuniform. Our results provide quantitative guidelines for interpreting back-projection results in the Japan trench, and show that current back-projection source models are reasonably well resolved and robust.

8.2 Data and Method

8.2.1 Data

To explore back-projection uncertainties, we study moderate earthquakes (6 ≤ Mw ≤ 7) within the Japan subduction zone, which can be approximated as point sources in the far field. We select this region because of its extensive M6 earthquakes, particularly after the 2011 Tohoku earthquake, covering the whole area without large spatial gaps along the trench axis. The station coverage is good, with regional arrays in the United States, Europe, and Australia, in addition to the stations of the Global Seismic Network (GSN), within epicentral distances of 30° to 90° from the earthquakes, providing good azimuthal coverage for back-projection (Figure 8.1).

We examine 46 M6 earthquakes occurring during the time period from 2004 October to 2015 February with Global Centroid Moment Tensor (GCMT) centroid depths shallower than 40 km (Figure 8.1, Table S1) (Ekström et al., 2012). The median and average moment magnitudes of the M6 earthquakes are 6.2 and 6.3, and the median and average hypocenter depths are 15.5 and 17.8 km. Twenty five of the events are reverse-faulting earthquakes, most of which occurred at the plate interface, fifteen are normal-faulting earthquakes, and six are strike-slip earthquakes (Figure 8.1). The seismograms used for analysis are from 2749 globally distributed stations with 431 stations recording more than 15 earthquakes with high-quality data (Figure 8.1). These stations are in the 30° to 90° epicentral dis-
tance range that complex P waveforms introduced by mantle discontinuities and the core-mantle boundary are unlikely to be present. The data were downloaded from the Data Management Center (DMC) of Incorporated Research Institutions for Seismology (IRIS) (Figure 8.1), and filtered into two frequency bands (low frequency 0.05–0.3 Hz and high-frequency 0.3–1 Hz). Some events only had usable data in one frequency band—in total, 39 and 41 earthquakes were analyzed in the low- and high-frequency bands, respectively. For example, several aftershocks of the 2011 M9 Tohoku earthquake have poor quality data in the low-frequency band, but have relatively clear signals in the high-frequency band. For a given earthquake, stations close to the GCMT nodal planes are removed, and only stations sharing the same GCMT P-wave polarities are used for back-projection analysis after self-normalization. All the traces are visually inspected to ensure good data quality. For our study region, we loosely define stations within 0° to 90° azimuthal range as USArray, and will refer to all the usable stations, including North American stations, as the global array.

8.2.2 Method

We follow closely the back-projection procedures described in Ishii et al. (2005); Xu et al. (2009); Fan and Shearer (2015) to image the 46 M6 earthquakes. We apply a standard time-domain back-projection approach, in which we empirically align the initial P waves, shift the records using the 1D-model-predicted relative time offsets for different locations, and stack the shifted records to map coherent seismic radiators. Strong signals will be observed when the seismograms constructively interfere, as should occur for true sources of seismic radiation, whereas weak signals are due to destructive interference, which should occur in regions without seismic radiation. We experiment with both linear and non-linear stacking approaches. Linear stacking retains the relative radiation strength but is more likely to suffer from artifacts due to anomalously strong signals on individual seismograms. Non-linear $N$th-root stacking can sharpen signals and suppress noise at the cost of absolute amplitude information (McFadden et al., 1986; Rost and Thomas, 2002; Xu et al., 2009). In this study, we focus on exploring variations in
the back-projection images caused by (1) different frequency bands (low-frequency 0.05–0.3 Hz and high-frequency 0.3–1 Hz), (2) different stacking approaches (linear and $N$th-root stacking), and (3) different array configurations (global and regional arrays).

In practice, we first grid potential sources at 5 km spacing across a box centered on the epicenter, spanning 200 km in latitude by 200 km in longitude. In principle, the source locations could be three-dimensional, but teleseismic P-wave data cannot resolve vertical rupture propagation for shallow earthquakes because of its poor depth sensitivity. Therefore, we consider only horizontal source grids fixed at the hypocentral depths. We compute empirical timing corrections for each event by cross-correlating the first few seconds of the records ($\sim 8$ s for 0.05–0.3 Hz and $\sim 5$ s for 0.3–1 Hz) to empirically align the seismograms to neutralize 3D velocity structure influences (Figure 8.2a) (Houser et al., 2008). Polarity flips are not allowed during cross-correlation because only stations sharing the same P-wave polarities are selected. We then shift the seismograms to stack at all the hypothetical source grids with respect to the relative time differences between the grids and the epicenter (Figure 8.2b). The time differences are calculated with the IASP91 1D velocity model (Kennett and Engdahl, 1991). Both linear and non-linear stacking are employed to explore their effects on the imaging results with $N = 4$ for $N$th-root stacking. During stacking, the seismograms are inversely weighted by the number of contributing stations within 5 degrees to avoid biasing the results by a single densely instrumented region. No post-smoothing or post-processing is applied to the back-projection images.

As an example, the 2005/12/02 Mw 6.5 reverse faulting earthquake (Event 4, Figure 8.1) is imaged in two frequency bands with both global stations and US-Array (Figure 8.3, 8.4). The waveforms are highly coherent for the stations in the northern hemisphere, whereas the Australian array data did not pass the quality control criteria for this particular event. Both global and array data can clearly resolve the event, as shown by the first 20 s integrated images with either linear or non-linear stacking, although the global array has higher spatial resolution and $N$th-root stacking leads to more compact energy burst images (Figure 8.3d–
High-frequency back-projection has higher spatial resolution with more compact images compared to low-frequency back-projection (Figure 8.3d–g, 8.4d–g). The peak energy locations of the integrated images may deviate from the epicenters, suggesting finite rupture extent and finite duration of the Mw 6.5 earthquake (Figure 8.2d, 8.3d–g, 8.4d–g). When integrating just the first 10 s at 0.05–0.3 Hz and the first 5 s at 0.3–1 Hz for both global and regional arrays, the peak energy locations are centered on the epicenter, showing the effectiveness of the cross-correlation approach in neutralizing 3D velocity effects.

For giant earthquakes like the 2004 Sumatra–Andaman earthquake, rupture lengths can exceed a thousand kilometers. This large spatial extent may cause the initial time calibrations at the epicenter region to fail at later/further rupture stages when the rays are crossing different parts of the underlying heterogeneous 3D velocity structure (Figure 8.2c) (Ishii et al., 2005, 2007). The time corrections may also be flawed if the earthquake contains subevents with different focal mechanisms, such that data polarity flips are occurring (Wang et al., 2012; Nissen et al., 2016). For these cases, the imaged rupture may be shifted to erroneous locations/time, or artifacts may be present, which could be difficult to distinguish from the true rupture features. To assess the robustness and uncertainties of hypocenter-calibrated back-projection across extended regions, we perform back-projection of the M6 earthquakes with time calibrations obtained from waveform alignments from other earthquakes (Figure 8.2c,d). The empirical time calibrations are derived from waveform cross-correlation for each earthquake individually and separately at two frequency bands, then the time calibrations are applied to other earthquakes to obtain back-projection images. For a given earthquake, P-wave polarity-flip artifacts are unlikely to be present due to our strict data selection criteria. For any of the M6 earthquakes, only stations that recorded both the imaged and the calibration earthquakes are used, which often results in fewer useable stations. For example, USAArray moved eastward during its deployment, which reduces the shared stations for events occurring at different times even when they are spatially close.

As described above, we apply the time calibration obtained from the 2011
276

/ 08 / 17 Mw 6.2 normal-faulting earthquake (Event 33) to the shared stations to image the 2005/12/02 Mw 6.5 reverse-faulting earthquake (Event 4, Figure 8.1). These two events occurred six years apart, reducing the number of common stations (Figure 8.5a,8.6a). The waveforms of the reverse-faulting earthquake (imaged earthquake) are less well aligned when using the reference time calibrations (Figure 8.5a,8.6a). Interestingly, the misalignments in the first few seconds do not defocus the images, but shift the energy bursts instead (Figure 8.5d–g,8.6d–g). These spatial shifts can be measured from either the epicenter or the self-aligned back-projection energy locations imaged from only the stations shared in common (Figure 8.2d), which can be defined as the deviation distance from the epicenter (De) and the deviation distance from the peak location of the self-aligned back-projection image (Db). In addition, the misalignment will reduce the strength of the stacks, causing incoherence. Because different events have different magnitudes, different focal mechanisms, and may have been recorded with different stations, we use coherence ratios to quantify the incoherence introduced by the misalignment. Coherence ratios are calculated from the 20-s integrated energy images, as the ratio between the peak amplitude in the reference-aligned and self-aligned back-projection images, using identical traces but different alignments.

8.3 Results

For every M6 earthquake, we apply its time calibration to back-project all the other events that were recorded by more than 20 common stations with high quality data. As an example, Figure 8.7a shows 11 of 29 events that were imaged with 0.05–0.3 Hz P waves with time calibrations obtained from Event 33 (Nth-root stacking). Despite different focal mechanisms between Event 33 and the imaged events, earthquakes with all types of focal mechanisms (reverse, normal, and strike-slip) were imaged clearly at both sides of the trench, hosted by faults at the plate interface, the outer rise, and the upper plate, extending across the whole study region (Figure 8.7a). Most of the reference-alignment-imaged events are close to their epicenters with maximum De of 63 km (Figure 8.7b) with no preferred
direction or clear trend in the deviations (Figure 8.7d). The imaged events retain high coherence ratios (∼80%) over the entire region without noticeable spatial decay when they move away from Event 33 (Figure 8.7c).

In most back-projection studies, defocused images lacking a clear peak are considered likely artifacts and are not interpreted due to their large uncertainties. Although ruptures occurring simultaneously at multiple parts of the fault have been reported in both dynamic simulations and some rare cases (e.g. Oglesby et al., 2004; Dunham and Archuleta, 2004; Melgar et al., 2016), we conservatively restrict our analysis to well-focused back-projection images, assuming the M6 earthquakes have simple rupture kinematics. We visually inspect the back-projection images and remove the defocused ones (more than one peak or a peak too close to the grid edges) to avoid potential artifacts, with an example shown in Figure S1. The inspections are performed for Nth-root stacking results at two frequency bands respectively, and the same selections are applied to linear stacking for comparison. Global and USArray results are inspected separately. In total, 257 and 300 event pairs are removed for global-array back-projection images at the two frequency bands (0.05–0.3 Hz and 0.3–1 Hz), and 60 and 150 for USArray. We then examine the spatial patterns of De, Db, and the coherence ratio with all the event pairs that passed visual inspections, the measurements include 1007 and 974 event pairs for the global array at the two frequency bands (0.05–0.3 Hz and 0.3–1 Hz), and 1016 and 996 pairs for USArray. Summaries of the results are shown in Figure 8.8 to 8.11.

The deviation distances from the epicenters (De) show similar spatial patterns for both the global array and USArray in the two frequency bands (Figure 8.8–8.11). De does not significantly increase with epicentral distance between the calibration events and the imaged events, while De also does not converge to zero when the separation distance approaches zero (Figure 8.8a,b,8.10a,b). For example, 78.4% of De are below 40 km for the global array with 4th-root stacking at 0.05–0.3 Hz (Figure 8.8a). The median of De remains at ∼20 km even when the event pairs are 700 km apart. De medians for USArray are slightly larger than those from the global array for both frequency bands, but they do not exceed
40 km. Different stacking approaches produce similar spatial patterns for De at both frequency bands respectively, although the imaged energy bursts often have larger spatial extents for linear stacking (Figure 8.5, 8.6, 8.8a,b, 8.10a,b). De medians share very similar spatial patterns for both the low and high frequency bands (Figure 8.8a,b, 8.10a,b). No preferred deviation directions can be identified in both USArray and the global array at the two frequency bands (Figure 8.9a–d, 8.11a–d). The deviation distances from the peak locations of self-aligned back-projection images (Db) are similar to those of De for all the event pairs (Figure 8.8–8.11).

Coherence ratios decay significantly just kilometers away from the calibration M6 epicenters and then remain stable up to ～700 km away without any obvious drop (Figure 8.8e,f, 8.10e,f). For example, the coherence ratio drops to ～80% 20 km away from the epicenters and remains at ～80% up to ～700 km with 88.1% event pairs above 60% for the global array with 4th-root stacking at 0.05–0.3 Hz (Figure 8.8e). USArray has higher coherence ratios compared to the global array at both frequency bands, and linear stacking has higher coherence ratios compared to Nth root stacking at both frequency bands. As might be expected, in all cases the low-frequency band (0.05–0.3 Hz) has higher coherence ratios compared to the high-frequency band (0.3–1 Hz). The coherence ratio is a weak function of De, showing no obvious decay when De increases (Figure 8.9i–l, 8.11i–l).

8.4 Discussion

Back-projection imaging has similarities to the earthquake location problem. One of the leading sources of errors in global earthquake locations is the biasing effect of unknown or incorrect 3D velocity structure (e.g., Thurber, 1983, 1985). In the absence of improvements to 3D velocity models, empirical approaches can be used to improve relative earthquake locations within localized regions by accounting for the correlated travel time residuals from the region to more distance stations (Douglas, 1967; Jordan and Sverdrup, 1981; Waldhauser and Ellsworth, 2000; Hauksson and Shearer, 2005; Shearer et al., 2005; Waldhauser and Schaff, 2008; Trugman and Shearer, 2017). A particularly simple approach is the master-
event technique, in which travel-time residuals for a single reference event are used as time corrections to relocate other nearby events. This is analogous to back-projection imaging using the time corrections derived from the hypocenter to better locate energy bursts at other parts of the rupture. A key question for both problems is quantifying how close the reference event needs to be to the target event in order to improve the location, which is related to the spatial coherence of travel-time residuals due to 3D structure as a function of event separation distance.

One way to assess uncertainties in relative earthquake locations due to influences of near-source 3D velocity structure is to “relocate” stations, which takes advantage of source-receiver reciprocity and the fact that the true station locations are precisely known (Shearer, 2001; Buehler and Shearer, 2016). The station relocation results show that the station mislocation vectors change rapidly over short distances (Buehler and Shearer, 2016). For example, relocations of the USArray stations show that mislocation vectors do not converge to zero even for very close station pairs (Buehler and Shearer, 2016). Such biases are very similar to what we observe for De (Figure 8.8–8.11). On average, De does not converge to zero even when event pairs are within 20 km (Figure 8.8–8.11), suggesting that strong near-source 3D velocity variations are a leading source of location error and are approximately random in their effects. This makes it difficult to use calibration events to improve back-projection images because the events would need to span the entire region at a spacing of 10 km or less (we have not resolved exactly how close they need to be). However, the good news is that the travel-time perturbations are largely random and uncorrelated at different locations, such that the average deviations (De) remain stable around 20 km over the whole study region (∼700 km) without obvious fluctuations. The relatively small De implies that the standard approach of producing hypocenter-calibrated back-projection images is fairly robust and can be used to map very large ruptures.

The observed spatial patterns of De also suggest that low-frequency back-projection can provide sub-event location accuracy comparable to high-frequency back-projection despite the broader imaging kernels at low frequencies (Figure 8.8, 8.10). Deviations in the peak energy locations are mostly introduced by local
heterogeneities in the source regions, which randomize the time calibrations for both frequency bands in similar ways. In general, the global array has higher resolution than USArray because of its superior azimuthal coverage. However, the mislocation vectors, De, from USArray observations do not have a preferred azimuthal distribution, suggesting the biases caused by 3D structure do not relate to array geometry or local tectonics in simple ways (Figure 8.9,8.11). Nonlinear stacking yields more compact back-projection images (Figure 8.3–8.6), but does not reduce the mislocation errors. However, it may still improve back-projection images for large earthquakes, because non-linear stacking equalizes amplitudes and enhances the later signals (Xu et al., 2009).

Another consequence of heterogeneous near-source 3D velocity structure is the sharp loss of coherence when event pairs are just kilometers apart (Figure 8.8e,f,8.10e,f) due to differences in the time calibrations. Coherence ratios remain stable for the whole study region, up to ~700 km between event pairs, indicating the time calibrations do not vary smoothly and instead have sharp changes caused by random shifts at each event location. This stability suggests that the relative strength between different subevents in back-projection images is likely not biased very much by coherence differences from inaccurate time calibrations. However, low-frequency back-projection is less sensitive to the inaccurate time calibrations, while high-frequency back-projection is more sensitive (Figure 8.8e,f,8.10e,f). This hampers efforts to quantitatively evaluate radiation strength between different frequency bands for a given earthquake by back-projection, including attempts to relate or scale integrated back-projection images to earthquake slip distributions. It is especially true when non-linear stacking approaches are employed, which reduces the coherence ratios by more than 50% at 0.3–1 Hz during our experiments (Figure 8.10e). The linear stacking approach preserves the coherence ratio strengths much better, for example, the median coherence ratios are ~90% at 0.05–0.3Hz for both USArray and global stations (Figure 8.8f). Coherence ratios obtained with USArray are higher than those with global data, as a result of highly coherent waveforms from regional arrays. Moreover, the coherence ratios do not scale with De (Figure 8.9i–l,8.11i–l), suggesting that the cos(azimuth) part of
the time calibrations that causes location differences is weakly correlated to total variance in the time calibrations. An example is shown in Figure 8.7c, in which the coherence ratios are above \( \sim 80\% \) and vary widely from region to region, and do not appear to correlate with local tectonics. Some nearby events have quite different coherence ratios, suggesting near-source shallow 3D structure controls the process.

Our observations of coherence ratios, \( \text{De} \) and \( \text{Db} \), imply it is difficult to improve back-projection images with aftershock calibration. Large aftershocks \((M \geq 5.5)\) typically sample the mainshock fault plane very unevenly, so that for each station, time calibrations derived from aftershocks must be either interpolated or extrapolated onto all of the hypothetical source locations for the back-projection image. In practice, weighting can also be used to enhance contributions from the higher-quality aftershock records \((\text{Ishii et al.}, 2007)\). An underlying assumption behind this approach is that the near-source velocity structure is relatively homogeneous or varying smoothly. For fine-scale structures, which vary rapidly at shallow depth, such assumptions often cannot be met \((\text{e.g. Wang and Zhao}, 2005; \text{Zhao et al.}, 2011)\), as is suggested by the spatial pattern of \( \text{De} \) and coherence ratios observed in this study \((\text{Figure 8.8–8.11})\).

A related approach to accounting for inaccurate time calibrations in back-projection involves an extra grid-station based time correction \((\text{Meng et al.}, 2016)\). For a given source grid, this extra time calibration term is \( \delta s \cdot (x_i - x_0) \), the product of the distance between the grid point \( (x_i) \) and the epicenter \( (x_0) \) and a slowness correction term \( \delta s \), which is designed to correct to first-order for 3D velocity variation effects between the source grid and the station \((\text{Meng et al.}, 2016)\). This approach computes a station-specific slowness correction and assumes the correction is valid for the whole rupture region, implying a smooth variation in the time corrections across the source grid. If this is the case, we would expect the aftershock-derived time calibrations to have smooth and continuous spatial distributions for a given station, but this is not what we have found for the Japan subduction zone. As shown for an example station in Figure 8.12a,b, the spatial distributions of our M6 derived calibrations do not vary smoothly or continuously. Another way to test the hypothesis is to examine the relationship between the
relative time calibrations and the relative distances to the stations. For a given station, the relative distances are defined as the distances between the closest M6 earthquake (minimum epicentral distance) and the other earthquakes, and the relative time calibrations are defined as the differences between the time calibrations of the minimum epicentral distance M6 earthquake and other earthquakes (Figure 8.13a). If the slowness correction hypothesis is correct, then a scaling relationship is expected between the relative distances and relative time calibrations. For four stations at different azimuths, we do not find clear scaling relationships between relative distances and the time calibrations for both frequency bands (Figure 8.12c–f). A more comprehensive analysis of all stations recording at least 15 events indicates that the station calibration times generally do not vary smoothly with event distance. This is consistent with the fact that the event mislocation vectors do not vary smoothly with position (e.g., Fig. 7d), as gradual location changes would be expected if the calibration times varied smoothly. These observations suggest that the slowness correction approach would not be effective in improving back-projection quality in our study region.

8.5 Conclusions

We systematically evaluate uncertainties in P-wave back-projection from velocity-model errors by imaging 46 M6 earthquakes within the Japan subduction zone. We derive empirical time corrections by cross-correlating waveforms of each event, and then apply the corrections to other events for back-projection imaging. The analysis quantifies the location and coherence uncertainties introduced by time shifts caused by unknown 3D velocity variations near the source. Our results show that the standard back-projection method in this region is robust and should be able to location subevents with reasonable accuracy, even for ruptures extending ~700 km away from the epicenter used for time calibration. Errors in imaged subevent locations, De, are mostly controlled by near-source fine-scale structures, and do not converge to zero even when the event pairs are very close. However, because these near-source variations are largely random and uncorrelated between
sources, the biases in the locations do not vary systematically and the location errors average about 20 km from very close to the calibration event to 700 km away. The spatial patterns of $D_e$ are similar in both frequency bands (0.05–0.3 Hz and 0.3–1 Hz), but suggest that the global array has better spatial resolution than single regional arrays. Similar spatial patterns are observed for coherence ratios, which decay rapidly kilometers away from the calibration M6 epicenters, while remaining stable up to $\sim$700 km away without a clear decay. On average, regional arrays have higher coherence than the global array, lower-frequency back-projection has higher coherence than higher-frequency back-projection, and linear stacking has higher coherence than non-linear stacking. These observations for the Japan subduction zone reveal no systematic biases that can be correlated to local tectonics or array geometries, suggesting that although it is difficult to improve back-projection images with aftershock calibration, standard back-projection images in the region are reasonably robust with respect to biases from unmodeled 3D velocity structure. Similar analyses could be used to evaluate back-projection uncertainties in other parts of the world.

Acknowledgments

The facilities of IRIS Data Services, and specifically the IRIS Data Management Center, were used for access to waveforms used in this study. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience and EarthScope (SAGE) Proposal of the National Science Foundation under Cooperative Agreement EAR-1261681. The earthquake catalog was downloaded from the Global Centroid Moment Tensor (GCMT) project (Ekström et al., 2012). Trench-axis data was from Bassett and Watts (2015a,b). The background bathymetry data was from Sandwell et al. (2014) and García et al. (2014), and were processed with the Generic Mapping Tools (GMT) (Wessel and Smith, 1991; Wessel et al., 2013). This work was supported by National Science Foundation grant EAR-1620251 at Scripps Institution of Oceanography, UC San Diego.

Chapter 8, in full, has been submitted for publication of the material as
it may appear in Journal of Geophysical Research–Solid Earth: Fan, W., and P. M. Shearer, Investigation of back-projection uncertainties with M6 earthquakes, *J. Geophys. Res.*, submitted. I was the primary investigator and author of this paper.
Figure 8.1: M6 earthquakes within the Japan Trench and the seismic stations used. Forty six earthquakes (of these, 39 and 41 were investigated in the low- and high-frequency bands, respectively) with centroid depths shallower than 40 km recorded globally from 2004 to 2015 are shown with GCMT focal mechanisms (Ekström et al., 2012). The upper-left corner insert shows the stations used in this study. The lower-right corner insert shows the distribution of focal mechanisms (Strike-slip earthquakes are defined when the rake of both nodal planes are within 45° deviation of 0° or 180°. Normal-/Reverse-faulting earthquakes have rakes within 45° deviation of −90°/90°).
Figure 8.2: A cartoon illustrating the back-projection method and the biasing effect of 3D structure on travel times. (a) Travel times from the hypocenter (star) are perturbed by 3D structure, such that the wavefront at the stations (wiggly solid line) is distorted compared to that predicted by a 1D model (dashed line). However, by cross-correlating and aligning the first-arriving P waves from the hypocenter, time correction factors can be applied to bring the arrival times into agreement with theoretical predictions. (b) Back-projection imaging stacks the seismograms for each source grid point using theoretical differential times relative to a source at the epicenter. (c) However, the time corrections computed for the hypocenter are not necessarily correct for other source points because the rays travel through different 3D structure and thus the back-projection images may be biased. (d) Using cross-correlation results from pairs of earthquakes, we test to see how much this bias will shift the back-projection images of target earthquakes, as measured by the distance of the image peak from the epicenter (De) or the back-projection image with self-aligned time calibrations (Db).
Figure 8.3: Back-projection results for Event 4 (Figure 8.1) with self-aligned time calibrations (0.05–0.3 Hz, 80% energy contour). (a), Velocity records sorted with azimuth. (b), Station map, station color represents time calibrations obtained with cross-correlation. (c), Lower-hemisphere P-wave polarities of the stations in (b). (d)/(e) Integrated back-projection result of the first 20 s with linear/4th-root stacking. (f)/(g), Similar to (d)/(e), but only using stations within the azimuthal range from 0° to 90°.
Figure 8.4: Back-projection results for Event 4 (Figure 8.1) with self-aligned time calibrations (0.3–1 Hz, 75% energy contour). (a), Velocity records sorted with azimuth. (b), Station map, station color represents time calibrations obtained with cross-correlation. (c), Lower-hemisphere P-wave polarities of the stations in (b). (d)/(e) Integrated back-projection result of the first 20 s with linear/4th-root stacking. (f)/(g), Similar to (d)/(e), but only using stations within the azimuthal range from 0° to 90°.
Figure 8.5: Back-projection results (0.05–0.3 Hz, 80% energy contour) for Event 4 with time calibrations from Event 33 (Figure 8.1). (a), Velocity records sorted with azimuth. (b), Station map, station color represents time calibrations obtained from the reference event. (c), Lower-hemisphere P-wave polarities of the stations in (b), the light green lines are nodal planes of the target event. (d)/(e) Integrated back-projection result of the first 20 s with linear/4th-root stacking. (f)/(g), Similar to (d)/(e), but only using stations within the azimuthal range from 0° to 90°.
Figure 8.6: Back-projection results (0.3–1 Hz, 75% energy contour) for Event 4 with time calibrations from Event 33 (Figure 8.1). (a), Velocity records sorted with azimuth. (b), Station map, station color represents time calibrations obtained from the reference event. (c), Lower-hemisphere P-wave polarities of the stations in (b), the light green lines are nodal planes of the target event. (d)/(e) Integrated back-projection result of the first 20 s with linear/4th-root stacking. (f)/(g), Similar to (d)/(e), but only using stations within the azimuthal range from 0° to 90°.
Figure 8.7: Back-projection results with time calibrations from reference Event 33 applied to 29 other events (Figure 8.1, 75% energy contour). (a), Examples of integrated back-projection images of the first 20 s. For clarity, only 11 events are shown here, the rest can be seen in Figure S1. (b), Deviation distances from the epicenters (De) for the 29 events. (c) Coherency ratio for the 29 events. (d) Deviation distance (De) and azimuth for the 29 events. For visibility, the length of the arrows are exaggerated compared to the map scale.
**Figure 8.8:** Deviation distances from epicenters (De) and back-projection peak energy loci (Db), and coherence ratios with linear and 4th-root stacking (0.05–0.3 Hz), as a function of epicentral distance.
Figure 8.9: Deviation distance density distribution and coherency ratio versus De with linear and 4th-root stacking (0.05–0.3 Hz). (a)–(d), De density distribution, normalized with total event-pair number, with epicenters located in the center for global stations and USArray with linear and 4th-root stacking. (e)–(h), Db density distribution, normalized with total event-pair number, with epicenters located in the center for global stations and USArray with linear and 4th-root stacking. (i)–(l), Coherency ratio versus De obtained for global stations and USArray with linear and 4th-root stacking. USArray is loosely defined as stations within the azimuthal range from 0° to 90°.
Figure 8.10: Deviation distances from epicenters (De) and back-projection peak energy loci (Db), and coherency ratios with linear and 4th-root stacking (0.3–1 Hz), as a function of epicentral distance.
Figure 8.11: Deviation distance density distribution and coherency ratios versus De with linear and 4th-root stacking (0.3–1 Hz). Legends are similar to Figure 8.9.
Figure 8.12: Time calibrations for station LAO in two frequency bands (a,b) and relative time versus relative distance for station LAO (c), SDCO (d), KBK (e), and HGN (f).
Figure 8.13: Illustration of relative distances for a given station (a), and relative time versus relative distance for all the stations recording more than 15 events in two frequency bands (b,c).
### Supplementary Materials

#### Table 8.S1: M6 earthquakes investigated in the study

<table>
<thead>
<tr>
<th>No.</th>
<th>Time</th>
<th>Lat(°)/Lon(°)/Depth(km)</th>
<th>Mw</th>
<th>LF</th>
<th>HF</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2004/10/23, 8:56:0.9</td>
<td>37.23/138.78/16</td>
<td>6.6</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>2</td>
<td>2005/10/19, 11:44:42.8</td>
<td>36.4/140.84/32</td>
<td>6.3</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>3</td>
<td>2005/11/14, 21:38:51.4</td>
<td>38.11/144.9/11</td>
<td>7.0</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>4</td>
<td>2005/12/2, 13:13:9.5</td>
<td>38.09/142.12/29</td>
<td>6.5</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>5</td>
<td>2007/7/16, 1:13:22.4</td>
<td>37.53/138.45/12</td>
<td>6.6</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>6</td>
<td>2008/5/7, 16:2:2.6</td>
<td>36.18/141.54/19</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>7</td>
<td>2008/5/7, 16:16:36.2</td>
<td>36.16/141.76/23.3</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>8</td>
<td>2008/6/13, 23:43:45.4</td>
<td>39.03/140.88/7.8</td>
<td>6.9</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>9</td>
<td>2008/7/19, 2:39:28.7</td>
<td>37.55/142.21/22</td>
<td>6.9</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>10</td>
<td>2008/12/20, 10:29:23.1</td>
<td>36.54/142.43/19</td>
<td>6.3</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>11</td>
<td>2009/6/5, 3:30:33.1</td>
<td>41.82/143.45/29</td>
<td>6.3</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>12</td>
<td>2010/7/4, 21:55:52</td>
<td>39.7/142.37/27</td>
<td>6.3</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>13</td>
<td>2011/3/9, 18:16:16.4</td>
<td>38.31/142.43/22</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>14</td>
<td>2011/3/9, 21:24:2.7</td>
<td>38.29/142.8/22</td>
<td>6.5</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>15</td>
<td>2011/3/11, 18:59:16.5</td>
<td>37.01/138.38/9.3</td>
<td>6.3</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>16</td>
<td>2011/3/11, 19:2:59.2</td>
<td>39.34/142.87/27.1</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>17</td>
<td>2011/3/11, 20:11:24.1</td>
<td>39.01/142.63/17.4</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>18</td>
<td>2011/3/12, 1:47:15.4</td>
<td>37.59/142.65/20</td>
<td>6.5</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>19</td>
<td>2011/3/13, 1:26:4.2</td>
<td>35.72/141.64/8</td>
<td>6.2</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>20</td>
<td>2011/3/14, 6:12:36.1</td>
<td>37.78/142.46/14</td>
<td>6.0</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>22</td>
<td>2011/3/15, 15:23:54.1</td>
<td>40.33/143.29/19.8</td>
<td>6.0</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>23</td>
<td>2011/3/22, 7:18:45.4</td>
<td>37.24/144/11</td>
<td>6.4</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>24</td>
<td>2011/3/22, 9:19:6.2</td>
<td>37.33/141.79/31</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>25</td>
<td>2011/3/22, 9:44:28.4</td>
<td>39.85/143.44/7.3</td>
<td>6.3</td>
<td>N</td>
<td>Y</td>
</tr>
</tbody>
</table>
### Table 8.S1: M6 earthquakes investigated in the study

<table>
<thead>
<tr>
<th>No.</th>
<th>Time</th>
<th>Lat(°)/Lon(°)/Depth(km)</th>
<th>Mw</th>
<th>LF</th>
<th>HF</th>
</tr>
</thead>
<tbody>
<tr>
<td>26</td>
<td>2011/3/27, 22:23:58.8</td>
<td>38.42/142.01/19</td>
<td>6.2</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>27</td>
<td>2011/3/29, 10:54:33.2</td>
<td>37.4/142.29/15</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>28</td>
<td>2011/4/11, 8:16:12.7</td>
<td>37/140.4/11</td>
<td>6.7</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>29</td>
<td>2011/4/11, 23:8:16.9</td>
<td>35.42/140.57/15</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>30</td>
<td>2011/4/13, 19:57:25.4</td>
<td>39.58/143.34/22</td>
<td>6.0</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>31</td>
<td>2011/5/5, 14:58:18.7</td>
<td>38.17/144.03/11</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>32</td>
<td>2011/6/3, 0:5:0.8</td>
<td>37.28/143.91/14</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>33</td>
<td>2011/8/17, 11:44:8.4</td>
<td>36.76/143.77/9</td>
<td>6.2</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>34</td>
<td>2011/9/15, 8:0:9.6</td>
<td>36.26/141.34/28</td>
<td>6.2</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>35</td>
<td>2011/9/16, 19:26:41</td>
<td>40.27/142.78/35</td>
<td>6.7</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>36</td>
<td>2011/9/16, 21:58:5.3</td>
<td>40.24/143.01/18</td>
<td>6.0</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>37</td>
<td>2012/3/14, 9:8:35.1</td>
<td>40.89/144.94/12</td>
<td>7.0</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>38</td>
<td>2012/3/14, 10:49:24.6</td>
<td>40.78/144.76/10</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>39</td>
<td>2012/3/27, 11:0:44.5</td>
<td>39.86/142.02/15</td>
<td>6.0</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>40</td>
<td>2012/5/20, 7:20:36.9</td>
<td>39.65/143.16/11</td>
<td>6.4</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>41</td>
<td>2012/6/5, 19:31:33.8</td>
<td>34.94/141.13/15</td>
<td>6.1</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>42</td>
<td>2012/10/1, 22:21:46</td>
<td>39.81/143.1/15</td>
<td>6.0</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>43</td>
<td>2014/11, 19:22:0.1</td>
<td>37.06/142.37/11.1</td>
<td>6.6</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>44</td>
<td>2014/10/11, 2:35:46.2</td>
<td>40.99/143.22/13.5</td>
<td>6.1</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>45</td>
<td>2015/2/16, 23:6:28</td>
<td>39.83/142.89/23</td>
<td>6.8</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>46</td>
<td>2015/2/21, 10:13:54.1</td>
<td>39.85/143.46/10</td>
<td>6.0</td>
<td>Y</td>
<td>Y</td>
</tr>
</tbody>
</table>

“Y” means the frequency band was investigated, and “N” indicates the opposite, (LF: 0.05–0.3 Hz, HF: 0.3–1 Hz).
Figure 8.S1: Back-projection results with time calibrations from Event 33 (Figure 1, 75% energy contour). All integrated back-projection images of the first 20 s. In each subplot, the black crosses show the epicenter locations with the earthquake origin time and moment magnitude listed in the corner. The blue boxes show the back-projection images that did not pass visual inspection with either multiple peaks or peaks too close to the edges.
Figure 8.S2: Deviation distances from epicenters (De) and back-projection peak energy loci (Db), and coherency ratios with linear and 4th-root stacking without visual inspection (0.05–0.3 Hz), as a function of epicentral distance.
Figure 8.S3: Deviation distance density distribution and coherency ratios versus De with linear and 4th-root stacking without visual inspection (0.05–0.3 Hz). (a)–(d), De density distribution, normalized with total event-pair number, with epicenters located in the center for global stations and USArray with linear and 4th-root stacking. (e)–(h), Db density distribution, normalized with total event-pair number, with epicenters located in the center for global stations and USArray with linear and 4th-root stacking. (i)–(l), Coherency ratio versus De obtained for global stations and USArray with linear and 4th-root stacking. USArray is loosely defined as stations within the azimuthal range from 0° to 90°.
Figure 8.S4: Deviation distances from epicenters (De) and back-projection peak energy loci (Db), and coherency ratios with linear and 4th-root stacking without visual inspection (0.3–1 Hz), as a function of epicentral distance.
Figure 8.S5: Deviation distance density distribution and coherency ratio versus De with linear and 4th-root stacking without visual inspection (0.3–1 Hz). Legends are similar to Figure S3.
References


Huang, Y., L. Meng, and J. P. Ampuero (2012), A dynamic model of the frequency-dependent rupture process of the 2011 Tohoku-Oki earthquake. Earth Planet Sp, 64(12), 1061–1066, doi:10.5047/eps.2012.05.011.


Chapter 9

Conclusions

9.1 Summary

We have shown that earthquake kinematic source imaging provides constraints and clues regarding large earthquake rupture propagations, earthquake triggering and interaction, and relationships between local tectonics and large earthquake rupture patterns. We have developed tools, have applied them to address important questions, and have quantified the uncertainties, providing guidelines to understand kinematic source models. In the following two subsections, we briefly summarize the main findings in this thesis.

9.1.1 Theory

Motivated to understand the non-uniqueness of finite-fault slip inversions, we have developed a frequency-based approach to earthquake slip inversion (Fan et al., 2014). The method is free of constraints on rupture velocity or slip-rate functions. Assuming each frequency bin is independent, the method solves the inversion efficiently at each frequency bin and allows over-parameterization to image the finest resolvable spatial details. In principle, the method is able to resolve complicated ruptures, including variable rupture speeds, and even reversals of rupture direction, although these complexities were not analyzed in the study. We benchmarked the method with Source Inversion Validation (SIV) Exercise 1 (Mai et al.,
and found that various physically plausible regularizations obtain robust inversion results. The linear formulation allowed us to obtain the global minimum because of the implementation of convex optimization (Boyd and Vandenberghe, 2004). To understand uncertainties in finite-fault slip inversions, we applied various types of noise to test the robustness of the method, and found that random noise does not greatly affect the inverted source models. But model-induced errors, such as random time shifts due to unknown velocity structures or erroneous fault geometries, significantly degrade the inverted source models by introducing correlated errors, which can cause artifacts that are difficult to distinguish from true features. To mitigate this assumption-induced error, we were motivated to develop robust kinematic imaging approaches without fault geometry assumptions.

Robust source imaging techniques, like back-projection, have the advantage that they make few assumptions about fault geometries. We have improved back-projection methods by exploiting the superior azimuthal coverage of the global seismic network (Fan and Shearer, 2015). The better spatial resolution of the global array allows us to perform back-projection in multiple frequency bands to resolve the complex spatiotemporal evolution of large earthquakes. Moreover, we analyzed both data and model-induced back-projection image uncertainties, aiming to provide quantitative guidelines to understand the obtained source models.

To model possible data-induced errors, we have investigated: (1) potential bias in the back-projection images due to wave interference and depth phases (Fan and Shearer, 2015), (2) the theoretical resolution, or array response functions, which provides an estimate of the resolution of any given station set at the frequency bands of interest (Fan and Shearer, 2016a), (3) robustness of the observed features by performing back-projection with a spatial subset of the network (Fan et al., 2016), (4) potential biases introduced by the complexities of the wavefield by imaging smaller events within the same region with similar station coverage as the mainshock (Fan et al., 2016). (5) realistic complexities, for example, rapidly varying Green’s functions, different focal mechanisms, and complex noise due to multiples, reflections and noise in the P-wave codas, by performing back projection for a “synthetic” data set obtained by combining the recorded data from local...
earthquakes, which are summations of the first 100 s of seismograms with a 20 s delay for the second event (Fan et al., 2016), (6), statistical significances of observe features by boot-strap and jack-knife resampling data (Fan and Shearer, 2016b).

To evaluate possible model induced errors, in Chapter 8, we imaged 46 M6 earthquakes within the Japan subduction zone by applying timing corrections derived from one earthquake to other events to evaluate the effectiveness of the empirically obtained time calibrations. The analysis addresses to what degree the time corrections computed from initial P-wave alignments at hypocenters to neutralize 3D velocity influences are applicable to later and more distant rupture for large earthquakes, which often have extended rupture areas. Our quantitative analysis of both data- and model-induced errors helps in understanding back-projection uncertainties, providing guidelines to interpret back-projection images. In general, our results show that back-projection can serve as a reliable tool to evaluate rupture processes with good confidence.

9.1.2 Applications

Understanding earthquake rupture processes is essential to unravel rupture physics and provide realistic constraints for future hazard mitigation. Kinematic earthquake source imaging provides information about how ruptures propagate spatiotemporally. Back-projection has a natural advantage for providing such critical information because of its robustness. We investigated the 2015 Mw 7.8 Nepal earthquake with global back-projection in two frequency bands (high-frequency, 0.2–3 Hz; low-frequency 0.05–0.2 Hz), providing a more complete description of the seismic radiation during the earthquake (Fan and Shearer, 2015). The results revealed frequency-dependent multiple-stage rupture propagation of the earthquake with a high-frequency deficient rupture stage below Kathmandu, which could be a major reason why there was less ground motion there than expected. Using the same method, we also resolved a high-resolution kinematic model for the 2009 Tonga–Samoa earthquake (Fan et al., 2016).

During the 2009 Tonga–Samoa earthquake, the rupture initiated as a normal-faulting earthquake located seaward of the trench-axis that curved into the trench,
then triggered a reverse-faulting subduction thrust event dominated by long-period seismic radiation (Fan et al., 2016). Our source model of the 2009 Tonga-Samoa earthquake reveals a fault system involving seaward-wall intraplate normal faults, megathrust reverse faults, and reactivated fabrics. The observed three-branch rupture is a natural consequence of the curved fault geometries associated with reactivation of pre-existing fabric and fore-arc segmentation. Our observations suggest identifying the fault geometry and fore-arc material segmentation can help in understanding tectonic modulating factors for large earthquake rupture propagation. The results suggest the possibility of improving forecasts of earthquake behavior by investigating surrounding host faults, which is particularly useful for this region because of its high seismic risk and high tsunamigenesis.

Tsunami earthquakes can cause huge devastation. For example, the 2004 Sumatra earthquake induced a tsunami that led to serious damage as far away as Africa (\(\sim 8,000\) km). These earthquakes often show complex seismic radiation patterns, and exact tsunami exciting mechanisms remain debatable (e.g. Fritz et al., 2007; Fujii and Satake, 2006; Mori et al., 2007). The modern global seismic network provides opportunities to investigate these earthquakes in great detail. We applied our methods to investigate the 2006 Java tsunami earthquake in Chapter 7 and found two distinct stages in high-frequency seismic radiation. The first stage lasted \(\sim 65\) s with a rupture speed of \(\sim 1.2\) km/s, while the second stage lasted from \(\sim 65\) to 150 s with a rupture speed of \(\sim 2.7\) km/s. High-frequency radiators resolved with back-projection during the second stage spatially correlate with splay fault traces mapped from residual free-air gravity anomalies. These splay faults also collocate with a major tsunami source associated with the earthquake inferred from tsunami first-crest back-propagation simulation. Although back-projection cannot resolve the depth of the radiators, the abrupt rupture transition and the strong spatial correlations indicate that the splay faults may have been reactivated during the earthquake and acted as the primary source of high-frequency radiation during the second stage rupture.

The observed fault triggering and interactions during the 2006 Java earthquake and the 2009 Tonga–Samoa earthquake suggest these may be common pro-
cesses at subduction zones and plate boundaries. With our improved imaging methods, we systematically imaged 88 large earthquakes ($7 \leq M_w < 8$) from 2004 to 2015, and detected and located 48 previously unidentified early aftershocks buried in the mainshock coda (Fan and Shearer, 2016b). These “hidden” aftershocks ($M \sim 6$) were triggered by 27 M7 earthquakes within a few fault lengths ($\sim 300$ km), during times when high-amplitude surface waves were arriving from the mainshock ($\leq 200$ s). The observations indicate that near-to-intermediate-field dynamic triggering commonly exists and promotes large aftershock occurrence. This study suggests that stress can be near-instantaneously transferred across complex multiple-fault systems and large earthquakes may develop into earthquake sequences, even where the source faults of these earthquakes are structurally disconnected. These observations also address long-standing controversies regarding the relative contributions of static and dynamic triggering, and have important implications for hazard assessment in regions with complex fault systems. Near-instantaneous dynamic triggering is also seen in our analysis of the 2012 Mw 7.2 Sumatra earthquake (Fan and Shearer, 2016a).

9.2 Future research directions

9.2.1 Kinematic source models of moderate to small earthquakes

Great to large earthquakes ($M \geq 7$) occur infrequently and rupture differently on a case-by-case basis because of their unique hosting faults (e.g., Ji et al., 2002b; Ishii et al., 2005; Walker and Shearer, 2009; Minson et al., 2014). On the other hand, moderate to small earthquakes occur frequently and show similarities among events in the same region. Understanding moderate to small earthquake kinematic behavior is important for resolving the complex rupture propagation of large earthquakes. More importantly, the large number of the moderate events makes it possible to statistically test and evaluate earthquake physics. Traditionally, moderate earthquakes ($5 \leq M \leq 7$) are taken as simple events and are often
modeled as point sources. But as shown by the 2004 M6 Parkfield earthquake (Allmann and Shearer, 2007), moderate earthquakes can be as complex as large earthquakes. However, such detailed source models are rarely available because finite-fault source modeling of moderate earthquakes requires very near-field observations.

Rapid developments in instrumentation and data collection have made possible the imaging of moderate earthquakes as finite-fault sources. Publicly available data of large continental and nodal arrays (e.g., USArray and Long Beach array) has led to unprecedented discoveries. Furthermore, seafloor geodetic observations have become available at Japan, Cascadia, and Hawai’i in recent years. It is a great time to utilize these large datasets to resolve kinematic behaviors of moderate earthquakes. Systematically extending finite-fault kinematic imaging to moderate earthquakes will provide an observational base to develop a unified theory of earthquake physics and help unravel the complex rupture behaviors of large earthquakes. In the future, we plan to apply the methods developed here to moderate subduction zone earthquakes in conjunction with available seafloor geodetic data. It will be particularly beneficial to build up a catalog of earthquake rupture length, average rupture velocity, and stress drop with the developed methods (e.g., back-projection and finite-slip inversion). Investigating these key kinematic source parameters will significantly advance the fundamental understanding of earthquake physics and regional tectonics where these earthquakes occurred.

9.2.2 Role of fluids in earthquake triggering

Large aftershocks of mainshocks can cause significant damage. Aftershocks are caused by static and/or dynamic stress perturbations from mainshocks. Whereas static-stress triggering is most effective in the near field, dynamic-stress triggering has been widely reported to cause earthquakes remotely. However, in the near-to-intermediate field, the relative importance of static and dynamic triggering is not well understood. Recent observations indicate that near-to-intermediate field dynamic triggering commonly exists and promotes large aftershocks (Fan and Shearer, 2016a,b). These observations lead to questions such as: When and how
do the receiver faults get triggered? What are the physical properties that control the triggering process? What are the magnitude limits of the triggering/triggered earthquakes?

Most of the observed near-instantaneously dynamically triggered earthquakes occurred at subduction zones, where the complex fault systems are fluid-saturated. Recent studies imply a strong feedback between fault strength and permeability and suggest spatiotemporal variability in susceptibility to dynamic triggering (e.g. Xue et al., 2013; Aiken and Peng, 2014; Barbour, 2015). Therefore, fluids must play an important role in the observed earthquake triggering. Quantifying the hydraulic response of faults to stress perturbations is critical to deciphering mechanisms of dynamic earthquake triggering and understanding the nonlinear frictional behavior of faults. More and more marine geophysical data, especially from dense ocean bottom seismograph (OBS) and seafloor geodetic arrays, provide critical observations to constraining hydromechanical properties of triggering/triggered faults at subduction zones (e.g. McGuire et al., 2012). New hypotheses for how hydromechanical fault properties affect earthquake triggering can be tested using detailed surveys of active fault systems. The fast advances in observations provide excellent opportunities to address these questions. In the near future, we intend to investigate and quantify the hydromechanical properties of faults hosting triggered quakes. Characterizing the temporal evolution of such feedback systems during passing seismic waves will also enhance our understanding of rupture physics and provide clues to evaluate subduction zone faults as a dynamically interactive complex system.
References


