Understanding megathrust behaviour over the earthquake cycle through inversion of geodetic data and dynamic modelling

Qiu, Qiang

2016

Qiu, Q. (2016). Understanding megathrust behaviour over the earthquake cycle through inversion of geodetic data and dynamic modelling. Doctoral thesis, Nanyang Technological University, Singapore.

<http://hdl.handle.net/10356/69080>

<https://doi.org/10.32657/10356/69080>

NANYANG
TECHNOLOGICAL
UNIVERSITY

**UNDERSTANDING MEGATHRUST
BEHAVIOUR
OVER THE EARTHQUAKE CYCLE
THROUGH INVERSION OF
GEODETIC DATA AND DYNAMIC
MODELLING**

QIANG QIU

ASIAN SCHOOL OF THE ENVIRONMENT

2016

UNDERSTANDING MEGATHRUST BEHAVIOUR OVER THE EARTHQUAKE CYCLE THROUGH INVERSION OF GEODETIC DATA AND DYNAMIC MODELLING

QIANG QIU

QIANG QIU

Asian School of the Environment

A thesis submitted to the Nanyang Technological University in
partial fulfilment of the requirement for the degree of Doctor of
Philosophy in Earth Sciences

2016

Acknowledgments

I am very grateful to my PhD adviser and my mentor, Emma Hill. I would like to thank her for her support and encouragement of my research, and walking me through my PhD journey. I thank her for always being patient with me and for giving me lots of freedom in my research. Thank you for sharing your experiences in both science and life, which will change my future career. It has been wonderful and a joy to learn from and work with you and always.

I would also like to thank Sylvain Barbot. I thank him for opening the door of geophysics for me to understand tectonic processes from another angle. I am very grateful to him for sharing his research experiences, and techniques to help me understand the physics behind tectonic processes.

Special thanks to all our geodesy group members; they are Eric Lindsey, Lujia Feng, Dongju Peng, Louisa Tsang, Rino Salman, and Paul Morgan. I thank them for their support for my research, and useful comments on my thesis chapters. I would like to also acknowledge earthquake physics group members Sagar Masuti, Valère Lambert, and James Moore. I thank them for the helpful discussions and for sharing techniques and tricks. I am grateful to Aron Meltzner, Yu Wang, Chan Chung-Han, Xu Hua, Lili Cheng, He Shaoneng River, Iwan Hermawan, and Ashar Muda Lubis for useful suggestions on my research. I thank our English teacher Pavel for improving our English. I thank Hsu Ya-Ju for sharing her “slab carpet” software for creating a nice fault geometry, and also her modeling experiences and suggestions on my research projects.

I would like to thank Chi-Hsien Tang for sharing some scripts for calculating Green's functions. Spatial thanks must go to Christina Widiwijayanti and Annie Elizabeth Grace. I thank Christina for introducing GMT to me, and I thank Annie for sharing her Latex thesis format and answering lots of questions. I thank our sweet secretary Maybel D'Silva for her kindness and patience to arrange our trips. Special thanks to Kerry Sieh. I thank him for his encouragement and support for my research study. I would like also to thank our “EOS family” for supporting my PhD study here.

Finally, I would like to express my deepest gratitude to my dear wife, my cute boy and my big family for their unconditional encouragement, love and support over these years and always.

Contents

Acknowledgments

List of Figures	v
------------------------	----------

List of Tables	viii
-----------------------	-------------

1 Summary	1
------------------	----------

2 Introduction	3
-----------------------	----------

3 The mechanism of partial rupture of a locked megathrust: The role of fault morphology	5
--	----------

3.1 Abstract	5
------------------------	---

3.2 Introduction	6
----------------------------	---

3.3 Data	8
--------------------	---

3.3.1 GPS	8
---------------------	---

3.3.2 InSAR	8
-----------------------	---

3.4 Inversion of slip	8
---------------------------------	---

3.5 Dynamic earthquake cycle modeling	9
---	---

3.6 Discussion and Conclusions	10
--	----

3.7 Acknowledgments	12
-------------------------------	----

4 Transient rheology in the Sumatran mantle wedge revealed by a decade of great earthquakes	13
--	-----------

4.1 Abstract	13
------------------------	----

4.2 Introduction	14
----------------------------	----

4.3 Geodetic joint inversion approach	15
---	----

4.4	Geodetic data and inversion model set up	16
4.5	Results	17
4.6	Discussion	20
5	Ongoing postseismic deformation following the 2005 M_w 8.6 Nias-Simeulue earthquake: Joint inversion of long-term GPS time series for afterslip and viscous flow	21
5.1	Abstract	21
5.2	Introduction	22
5.3	GPS data	23
5.4	Geodetic inversion methods	23
5.4.1	Fault geometry and crustal model	24
5.4.2	Coseismic slip inversion	25
5.4.3	Joint inversion of afterslip and viscous strain	26
5.5	Results	27
5.5.1	Coseismic slip	27
5.5.2	Afterslip, viscous strain and effective viscosity	29
5.6	Discussion	30
5.6.1	Coseismic slip	31
5.6.2	Afterslip and afterslip rate	32
5.6.3	Effective viscosity	33
5.6.4	Slip rates and viscous strain rates	34
5.7	Conclusions	35
6	Concluding remarks	37
References		
Appendices		
A	Supplementary material for Nepal coseismic study	
A.1	Text S1	
A.2	Supplementary figures and table	
B	Supplementary material for Sumatra postseismic study	

B.1	Geodetic data sets	
B.1.1	GPS	
B.2	Land-height displacement derived from tide-gauge and satellite- altimetry data	
B.3	Methods	
B.3.1	Kinematic joint inversion	
B.3.2	Modeling of viscous flow	
B.3.3	Spatial smoothing constraints on fault slip and strain components	
B.3.4	Strain direction constraints	
B.3.5	Estimation of viscosities	
B.3.6	Effective viscosity for a Burgers rheology	
B.4	Synthetic tests to validate our technique and understand model weaknesses	
B.4.1	Model set up	
B.4.2	Synthetic slip inversion	
B.4.3	Viscous strain inversion	
B.4.4	Joint inversion of slip and viscous strain	
B.4.5	Summary of these synthetic tests	
B.5	Afterslip estimates along the Sumatra subduction zone over the last decade	
B.5.1	The Sumatra-Andaman rupture segment	
B.5.2	The Nias rupture segment	
B.5.3	The Bengkulu rupture segment	
B.6	Supplementary figures	
 C Supplementary material for Nias postseismic study		
C.1	Text S1	
C.1.1	Coseismic slip uncertainty estimation	
C.2	Text S2	
C.2.1	Extended Network Inversion Filter (ENIF)	
C.3	Text S3	
C.3.1	Deviatoric constraints for the strain components	
C.3.2	Viscous strain direction constraints from coseismic stress changes	

C.4	Text S4	
C.4.1	Building up the stress kernel for estimation of the stress evolution at each block	
C.4.2	Evaluating the effective viscosity	
C.5	Text S5	
C.5.1	Trade-offs in the models	
C.6	Text S6	
C.6.1	Effective viscosity of a Burgers Rheology	
C.7	Supplementary figures and tables	

List of Figures

3.1	Simplified surface geology for the Kathmandu region of Nepal	7
3.2	Relationship between the spatial extent of coseismic ruptures and slab morphology	10
3.3	Rupture segmentation and behaviour in the context of morphological barriers	11
4.1	Cumulative postseismic displacement following great earthquakes of the Sumatra subduction zone over the last decade.	14
4.2	Accumulated afterslip and deviatoric viscous strain in the mantle wedge, and effective viscosity estimates	17
4.3	Final estimates for the accumulated deviatoric viscous strain distribution over the three layers of blocks	18
5.1	Coseismic and cumulative postseismic displacement of the 2005 M_w 8.6 Nias-Simeulue earthquake	24
5.2	Geometry of the fault and viscous strain blocks	25
5.3	Coseismic slip estimation from joint inversion of GPS and coral data . .	27
5.4	Complementary spatial relationship between the afterslip and coseismic stress drop	28
5.5	Final accumulated viscous strain of blocks in a cross-section view of A-A'	28
5.6	Effective viscosity estimation of blocks 1 and 2	29
5.7	Multiple slip styles at the low-slip patch of the 2005 M_w 8.6 Nias-Simeulue earthquake	30
5.8	A comparison between afterslip models that did not and did include viscous flow	31
5.9	Difference in afterslip between models that did and did not include viscous flow	32
5.10	Slip rates on selected fault patches, and viscous strain rates for blocks 1 and 2	34

A.1	Footprint of InSAR data considered in this study
A.2	RADARSAR-2 InSAR data
A.3	ALOS-2 tracks 048 and 047 InSAR data
A.4	Sentinel-1A, and ALOS2-track 157 InSAR data
A.5	ABIC selecting on slip smoothing
A.6	Coulomb stress
A.7	Weighting effects of joint inversion between InSAR and GPS data sets .
B.1	Geographic location of the GPS and Tide gauge locations
B.2	Fault geometry and rectangular blocks used for joint inversion of afterslip and viscous strain
B.3	Postseismic model prediction of surface displacement at selected GPS sites
B.4	Postseismic Afterslip model prediction of surface displacement at Tide gauges
B.5	Difference in afterslip estimates for models that did and did not include viscous flow
B.6	Effective viscosity estimates at 2007 and 2004 earthquake rupture regions
B.7	A grid search example at site a
B.8	Example time series showing our removal of annual cycles signals from the tide-gauge and Aviso data at the Johor Baru (Malaysia) site
B.9	Surface displacement (Green's function) with unit strain along ϵ_{12} component
B.10	Cartoon showing strain (dark blue) direction constraints based on coseismic stress directions (red)
B.11	Fault and viscous blocks geometry of synthetic tests
B.12	Geometry of the fault plane, viscous blocks and the strain reference system
B.13	Synthetic test results for slip inversion
B.14	Synthetic test results for viscous strain inversion
B.15	Synthetic test results for joint inversion of slip and viscous strain
B.16	An example of comparison between joint inversion model predictions and synthetic time series at sites 7 and 12
B.17	An example of the low-pass filtered stress and strain rate at blocks a and b

C.1	Verification of our fault geometry with relocated seismicity from <i>Tilmann et al.</i> (2010)
C.2	Laplacian smoothing weight determination based on L-curve of stress-drop
C.3	Coseismic slip error estimation from a Bootstrap Monte Carlo sampling approach
C.4	he 3D geometry of the fault, viscous blocks and the calculated model resolution on these blocks
C.5	Coseismic slip comparison with <i>Konca et al.</i> (2007)
C.6	Postseismic model prediction compared to GPS time series
C.7	Coseismic stress changes at viscous blocks in the mantle wedge from the slip model of <i>Konca et al.</i> (2007)
C.8	Grid search example for block 1
C.9	The low-passed strain rate, stress and effective viscosity time series estimated for blocks 1 and 2
C.10	Cartoon showing reference systems for strain base vectors (dark blue), and coseismic stress base vectors (red) for one face of a rectangular block
C.11	A Burgers rheology and its effective viscosity

List of Tables

- A.1 Details of the InSAR data used in this study.
- C.1 Crust 1.0 parameters for the Nias region. We use the parameters of Crust 1.0 at the center of the fault plane to represent the entire fault area since the spatial variation is not significant.

Summary

Here I present three studies that depict megathrust behavior over different time periods of the seismic cycle. The first study is geographically focused on Nepal, and the second two on Sumatra.

First, I use geodetic inversion as constrained by InSAR data and dynamic earthquake cycle modeling for the 2015 M_w 7.8 Gorkha earthquake to show that local fault morphology can introduce complexity in rupture dynamics and bound the propagation of seismic rupture. Our results help to explain why we sometimes see only partial rupture of fully locked fault patches. These findings can be applied to megathrusts worldwide; they highlight that imaging the 3D structure of the fault morphology may be a key factor for seismic forecasting.

Second, I investigate long-term postseismic deformation following large earthquakes along the Sumatra subduction zone over the last decade. We have developed a new technique that can invert for afterslip and viscous flow of the asthenosphere simultaneously without a prior assumption on rheology. Our technique addresses the problem of biases being introduced into afterslip distributions when the more traditional technique of estimation is used, in which viscoelastic flow is either ignored or estimated and removed prior to inversion using an assumed rheology and coseismic source model. Our approach allows us to probe the type of rheology of the mantle wedge. Our key finding from the Sumatra subduction zone is that viscous flow in the mantle wedge has a transient behavior similar to a Burgers rheology with $\eta_K \sim 10^{17}$ Pa.s and $\eta_M \sim 10^{19}$ Pa.s.

Lastly, I present an analysis of ~ 9 -year long GPS time series that reveal the spatial and temporal afterslip and viscous flow in the mantle wedge following the 2005 M_w 8.6 Nias-Simeulue earthquake. We carry out this investigation using our new joint inversion technique. A key finding from this study is that the afterslip occurs both updip and downdip of the coseismic rupture and does not migrate in time. Our estimates of downdip afterslip are lower than for studies that do not include viscous flow in the model, while the up-dip estimates are increased. After a decade of afterslip, slip rates on the fault have reduced to very low but are still declining, suggesting the afterslip is ongoing. The viscous strain rates in the mantle wedge, however, remain significantly higher than the background strain rate, suggesting the viscous relaxation is ongoing and may last for years or decades. We also present an updated coseismic slip distribution for

CHAPTER 1. SUMMARY

this event that uses the same fault geometry as for our afterslip distribution, so that the relative locations of coseismic slip and afterslip can be consistently compared. In addition, we show estimates of coseismic slip for the last great earthquake on this patch, in 1861, based on inversion of coral data. We show that the coseismic slip patches for the 1861 and 2005 events are very similar, suggesting some level of persistently in the slip (and probably afterslip) patterns.

Introduction

Megathrusts generate the largest earthquakes worldwide. They release strain accumulated over hundreds or even thousands of years within seconds or minutes, introducing destructive seismic and tsunami hazards. Following these giant earthquakes, coseismic stress changes around the coseismic rupture patch continuously drive the fault to slip – so-called afterslip – and as a response of these stress changes the mantle flow located at asthenosphere depths will be activated. These resulting slip and viscous flow in the postseismic relaxation process redistribute the stress field, they can trigger future earthquakes by loading existing faults and promoting them to failure. This postseismic process lasts months, years, or even decades. Therefore, in order to better assess future seismic and associated tsunami hazards, a key factor is to understand the multiple behaviors of a megathrust within many cycles to gain more insights on where and how the energy is released.

Geodetic measurements are a key tool towards better understanding the behaviors of megathrusts. At the Sumatra subduction zone, the Sumatran GPS Array (SuGAR) has been deployed along the forearc islands. The network has been in operation since 2002, now giving >13 year-long time series and providing us a great opportunity to learn about the behavior of the Sunda megathrust in detail. In addition, thanks to recent developments in InSAR techniques and observations, we can now investigate continental megathrust behaviors with high spatial resolution. And to supplement our GPS measurements (*Feng et al.*, 2015), we can use tide gauges in conjunction with satellite altimetry data to extract tectonic signals (*Trisirisatayawong et al.*, 2011; *Santamaría-Gómez et al.*, 2014; *Nerem and Mitchum*, 2002).

With these fantastic geodetic data sets, this thesis focuses on address the following questions:

1. How does local fault morphology modify rupture dynamics on the Himalaya megathrust? Can it control rupture extent?
2. What are the spatial and temporal patterns of afterslip following large earthquakes along the Sumatra subduction zone? What is the nature of viscous flow in the mantle wedge? Is it a transient or power-law flow?
3. What do the spatial and temporal patterns of afterslip look like following the 2005 M_w 8.6 Nias-Simeulue earthquake? Has the postseismic relaxation finished yet? What is

CHAPTER 2. INTRODUCTION

the viscosity of the mantle wedge at this location? Where is the afterslip relative to the coseismic slip?

I conduct geodetic inversion models, and rate and state stress-driven earthquake cycle models to answer these questions in three chapters. These chapters are arranged in research paper format. In Chapter 1, I examine the coseismic slip distribution for the 2015 M_w 7.8 Gorkha earthquake in Nepal. I conduct a joint inversion of GPS and InSAR for the coseismic slip, and find that the coseismic slip is bounded by two dipping ramps that can limit the rupture extent and modify the rupture dynamics on the megathrust. I use rate-and-state stress-driven earthquake cycle modeling to examine in more detail the role of fault morphology in impacting the size of ruptures on megathrusts such as this. In Chapter 2, I investigate a combination of decade-long GPS and tide-gauge data to estimate the spatial and temporal evolution of afterslip on the megathrust and viscous strain in the mantle wedge along the Sumatra subduction zone. A key finding is that the viscous mantle flow has a transient Burgers-type rheology with $\eta_K \sim 10^{17}$ and $\eta_M \sim 10^{19}$ Pa.s, and is not a power-law flow. In Chapter 3, I examine the ~ 9 -year long SuGAR time series to investigate in more detail the spatial and temporal distribution of afterslip and viscous strains in the mantle wedge following the 2005 M_w 8.6 Nias-Simeulue earthquake. I find that the afterslip rates on the fault have reduced to very low, but are still declining. The viscous strain rates are still significantly above the typical geological strain rate ~ 9 years after the earthquake, indicating that viscous flow is still ongoing. Finally, I tie these chapters together to draw the main conclusions of this thesis.

The mechanism of partial rupture of a locked megathrust: The role of fault morphology

Qiang Qiu, Emma M. Hill, Sylvain Barbot, Judith Hubbard, Wanpeng Feng, Eric O. Lindsey, Lujia Feng, Keren Dai, Sergey V. Samsonov, and Paul Tapponnier. Published in *Geology*.

3.1 Abstract

Forecasts of seismic hazard rely on estimates of how large an area of a tectonic fault could potentially rupture in a single earthquake. Key ingredients to these forecasts are information on which areas of the fault are locked together and how the fault is segmented. Much research has focused on exploring down-dip limits to fault rupture from chemical and thermal boundaries, and along-strike barriers from subducted structural features, yet we regularly see only partial rupture of fully locked fault patches that could have been expected to produce as a whole in a large earthquake. Here, we draw insight into this conundrum from the 25th April 2015 M_w 7.8 Gorkha earthquake. We invert geodetic data with a structural model of the Main Himalayan Thrust (MHT) in the region of Kathmandu, Nepal, showing that this event was generated by rupture of a décollement bounded on all sides by more steeply dipping ramps. The morphological bounds explain why the event was ruptured only a small piece of a large fully locked seismic gap. We then use dynamic earthquake-cycle modeling on the same fault geometry to reveal that such events are indeed predicted by the physics. Depending on the earthquake history and the details of rupture dynamics, however, great earthquakes that rupture the entire seismogenic zone are also possible. These insights from Nepal should be applicable to understanding bounds on earthquake size on megathrusts worldwide.

3.2 Introduction

Earthquake magnitudes are directly related to the spatial extent of the fault patch that ruptures and the amount of stress on that patch. The level of slip available to generate an earthquake is determined by the loading rate, the level of interseismic coupling, or “locking” of the fault, and the history of past earthquakes.

Seismic hazard estimates, therefore, rely on assessments of the potential spatial extent and magnitude of rupture patches. Geodetic data can be used to estimate coupling maps, which give an estimate of the spatial patterns of fault locking at present (*Ader et al.*, 2012; *Stevens and Avouac*, 2015; *Chlieh et al.*, 2008). Analyses of seismological data and numerical modelling have been used to assess the role of thermal and lithological boundaries in controlling the extent of possible rupture, as well as the mode of failure likely at different depths (*Lay et al.*, 2012; *Oleskevich et al.*, 1999; *Kaneko et al.*, 2010). Geological studies have established the existence of persistent barriers to rupture; as an example, paleogeodetic data for the Simeulue to Nias section of the Sunda megathrust (Sumatra) highlight the likely role of subducting fracture zones in bounding the along-strike terminations of great earthquake ruptures over multiple earthquake cycles (*Meltzner et al.*, 2012).

This simple picture, however, does not address the fact that we regularly see only partial rupture of fault patches that were fully locked and at a late stage in the seismic cycle with no documented evidence for large-scale structural discontinuities. Examples include the 2014 M_w 8.2 Iquique, Chile (*Ruiz et al.*, 2014; *Lay et al.*, 2014) and 2007 M_w 8.4 Bengkulu, Sumatra, earthquakes (*Konca et al.*, 2008). How do we explain the occurrence of moderate to large earthquakes in the middle of patches that have – by this simple model – been forecast to produce great earthquakes?

The 2015 M_w 7.8 Gorkha earthquake was one such conundrum. The magnitude of this event was considerably smaller than expected from forecasts of a great earthquake, which were based on the fact that this area of the Main Himalayan Thrust (MHT) is fully locked and at a seismic gap (Figure 3.2; *Ader et al.*, 2012; *Bilham*, 2015; *Stevens and Avouac*, 2015). And the rupture did not reach the surface, despite considerable evidence that neighbouring ruptures have done just that (*Bollinger et al.*, 2016; *Sapkota et al.*, 2013; *Lavé et al.*, 2005). Why did this event not cascade into complete failure of the seismic gap, all the way to the surface, generating a great earthquake?

The answer lies with the morphology of the fault at the local scale (*Grandin et al.*, 2015; *Hubbard et al.*, 2016). In this study, we use the 3D fault morphology of *Hubbard et al.* (2016) for inversion of geodetic data to estimate the extent of rupture – showing that it was bounded by ramps on all sides – and for dynamic models of slip evolution that explain the role of this morphology in generating a complex earthquake history with single and multi-segment ruptures.

The model for fault morphology was determined by studying the surface geology, topography

CHAPTER 3. FAULT MORPHOLOGY CAN CONTROL RUPTURE SIZE

and seismicity of the area (Figure 3.1; *Hubbard et al.*, 2016). While broadly compatible with other geophysical data (e.g., *Duputel et al.*, 2016), the geometry differs by including two blind ramps up- and down-dip of the shallow-dipping décollement that generated the earthquake. The deeper ramp, at depths of ~ 16 to 30 km, is required to produce the large Gorkha-Pokhara anticlinorium, which extends across central Nepal. The shallow ramp, which extends from depths of ~ 8 to 10 km, explains the surface stratigraphy at this location. In some places, the two ramps join together, resulting in a single, large ramp; this is the case at both ends of the Kathmandu Klippe (Figure 3.1).

Figure 3.1: Simplified surface geology for the Kathmandu region of Nepal. The black focal mechanism marks the epicenter of the M_w 7.8 Gorkha earthquake, and the purple focal mechanism its M_w 7.3 aftershock. The black contour represents the inferred slip of the Gorkha event with a 2-m interval. The red lines marked by MFT and MBT represent the surface traces of the Main Frontal Thrust and Main Boundary Thrust, respectively. White areas in the map are outside Nepal. White dots indicate the location of the first two months of aftershock from the National Seismological Centre, Lainchaur, Kathmandu, Nepal. The colored dots show back projection (BP) results from *Avouac et al.* (2015) with color indicating time after the earthquake. Kathmandu city is in the middle of a quaternary basin (yellow) in the Kathmandu Klippe. The bottom panel shows a geological cross section along the profile A-A' (*Hubbard et al.*, 2016).

3.3 Data

3.3.1 GPS

We use the data from 18 GPS stations from the network run by the California Institute of Technology and the Department of Mines and Geology Nepal (*Galetzka et al.*, 2015; *Avouac et al.*, 2015). Details on our data processing procedure are given in *Feng et al.* (2015). Three GPS stations located close to the rupture region show a maximum southward movement reaching ~ 1.8 m, and uplift of ~ 1.3 m. The remaining stations, located further from the source, show negligible displacements.

3.3.2 InSAR

We use all available InSAR data for this event from the Sentinel-1A (an ascending path), RADARSAT-2 (a descending path), and ALOS-2 satellites (ScanSAR descending paths 048 and 047, and ascending path 157) (Figure A.1 in Appendix A, Table A.1). We process the Sentinel-1A and RADARSAT-2 data using GAMMA (*Scheiber and Moreira*, 2000; *Wegmuller et al.*, 2015). For the ALOS-2 data, we use GMTSAR (*Sandwell et al.*, 2011) to generate the unwrapped interferogram *Lindsey et al.* (2015).

3.4 Inversion of slip

To avoid the redundancy of the InSAR data due to the high degree of spatial correlation, and to achieve an efficient inversion, we use a quad-tree down-sampling scheme (*Simons et al.*, 2002) to extract the deformation pattern with a limited number of data points. Our inversion is therefore based on $\sim 9,000$ data points concentrated mostly over the rupture patch. We use triangular elastic half-space dislocation Green’s functions (*Nikkhoo and Walter*, 2015) to capture the fault geometry. We use a linear constrained least-squares inversion scheme to estimate fault slip with positivity constraints and smoothing regularization. The parameter controlling the level of spatial smoothing is determined using Akaike’s Bayesian Information Criterion (Figure A.5 in Appendix A; *Yabuki and Matsuura*, 1992). The model and corresponding residuals for InSAR are shown in Figure A.2, A.3, A.4 in Appendix A. The relative weight between InSAR and GPS data is chosen by trial and error so that fit to both data sets is satisfactory (Figure A.7 in Appendix A).

We find that the slip extended along-strike on the shallow décollement for ~ 130 km (Figure 3.2). The rupture area was bounded at either end by “pinch points” formed by the two-ramp system merging into a single ramp, at either end of the Kathmandu Klippe (Figure 3.2). Similarly, it was bounded up- and down-dip by two ramps. The hypocentre itself was actually located on

the deeper ramp; the intervening décollement appears to have provided prime real estate for rupture expansion. Because the rupture was confined to the décollement, previously published slip models (e.g., *Avouac et al.*, 2015) that were based on planar faults appear relatively similar to ours. We note that published patterns of coupling (*Ader et al.*, 2012; *Stevens and Avouac*, 2015) also seem to correspond to the fault morphology (Figure 3.2). The rupture increased the Coulomb stress on the up-dip section of the Himalayan megathrust by a few MPa, bringing this section closer to failure (Figure A.6 in Appendix A).

3.5 Dynamic earthquake cycle modeling

An important question is the extent to which the MHT at this location sometimes ruptures in a piecemeal fashion with ruptures bounded by morphological barriers, or sometimes cascades into complete rupture to the surface, generating a great earthquake. In the absence of a comprehensive historical earthquake catalogue, we look for insight from fault mechanics, and investigate the seismic potential of the Kathmandu section of the megathrust using dynamic models of slip evolution in the framework of rate and state friction (*Lapusta et al.*, 2000). We discretize the geometry of the MHT using triangular elements, and the associated stress kernels are obtained using the analytic formulation of *Nikkhoo and Walter* (2015). We consider a simple scenario where the seismogenic zone extends from the surface to ~ 20 km depth with uniform dynamic friction properties. The only complexity of the model resides in the fault morphology (see A.1 in Appendix A). Our modeling shows that some large ($M_w > 7$) ruptures can be confined between two ramps, or on the top-most ramp. But, these relatively smaller events can also be preceded or followed by rupture of the entire seismogenic zone in a single, great ($M_w > 8$) earthquake (Figure 3.2B). Our models show areas of high stress accumulation at ramp boundaries, which often initiate ruptures. These ruptures may or may not cascade into larger events, depending on the slip history and current stress status. The rupture sequences are unpredictable in terms of the location, time, and amount of slip. However, the sum of events will close the slip budget on the fault. The scarcity of through-going ruptures in our simulations may explain why megathrust ruptures do not often reach the surface, even in cases where the shallow fault is velocity weakening and capable of sustaining seismic ruptures. Overall, the non-planar geometry introduces complexity in the dynamics of rupture propagation and heterogeneity in the slip distribution of earthquakes that makes single events poorly representative of the full range of events.

Figure 3.2: Relationship between the spatial extent of the coseismic rupture and slab morphology. A) Surface projection of our estimated slip distribution for the mainshock. Colored circles at the GPS stations indicate vertical displacement, and vectors indicate horizontal displacement. The brown dots indicate two months of aftershocks, and the moment tensor for the main aftershock. The orange profile represents the 0.6 coupling contour line from *Ader et al.* (2012). The dashed and labeled solid black lines indicate depth contours of the MHT, based on the morphology of *Hubbard et al.* (2016). The inset figure shows this slip distribution in relation to coarse estimates of historical rupture patches. B) Spatial distribution of ruptures from rate-and-state dynamic modeling. Colored areas surrounded by contours (with arbitrary background color) represent the half-maximum slip contour of ruptured of $M_w \geq 7.3$ events in a 1,200 year period. The magnitude and time of these events are indicated as the vertical bars with the same color in panel C. The shaded area (with bounding magenta contour and the same color bar as used in panel A) shows one $M_w 7.7$ event that ruptures a qualitatively similar area to the 2015 mainshock. The slip distribution contains afterslip, most visible on the deepest ramp. Some ruptures are only on the décollement, some on ramps, and some are through-going. The inset map shows the spatial distribution of dynamic friction properties that control the dynamic model. C) The slip (filled profiles) and velocity (solid profiles) evolution for the two markers indicated by colored triangles in B.

3.6 Discussion and Conclusions

Based on our results, we propose that both single-segment and multi-segment ruptures are possible near Kathmandu (Figure 3.3). We note that the 1833 M_w 7.6-7.7 earthquake had striking similarities to 2015; it also generated shaking in Kathmandu, and did not rupture to the surface (*Bollinger et al.*, 2016). It seems plausible that both events ruptured the same confined décollement. However, the occurrence of earthquakes like the 1833 M_w 7.6 and the 2015 M_w 7.8 do not preclude the possibility of a great earthquake on this segment.

CHAPTER 3. FAULT MORPHOLOGY CAN CONTROL RUPTURE SIZE

Other parts of the MHT have historically behaved differently from this section, with evidence for great, $M_w > 8$ events rupturing all the way to the surface (Figure 3.2 A; *Sapkota et al.*, 2013). Indeed, the Gorkha section may be relatively unique in having such clear morphological constraints, although there are other published inferences. For example, lateral ramps may have played a role in segmentation of the 1905 M_w 7.8 rupture (*Middlemiss*, 1910; *Molnar*, 1987).

Figure 3.3: Rupture segmentation and behavior in the context of morphological barriers. Seismogenic behavior occurs above the thermal boundary. Pinch points introduce along-strike segmentation. Asperities bounded by morphological barriers may fail persistently, generating repeated earthquakes with similar magnitude. Alternatively, the barriers may only hold sometimes, and at other times great earthquakes are generated by rupture across these barriers. In the absence of morphological barriers (as on the left side of the cartoon) earthquakes may rupture a larger patch more frequently.

The effect of fault morphology on earthquake size is well-known in strike-slip systems (*Wesnowsky*, 1988, 2006; *Klinger*, 2010), where the bends are considerably easier to map than on megathrusts. This idea is also grounded in fault mechanics (*Rice*, 1980; *King*, 1986; *Kame et al.*, 2003). For strike-slip systems, there is evidence that fault morphology can influence not only earthquake size, but also rupture behavior. For example, particularly straight segments may be more prone to supershear rupture velocities (e.g., *Bouchon et al.*, 2010). Local megathrust fault morphology may, therefore, be an additional clue to understanding rupture behavior within the broad frictional regimes outlined by *Lay et al.* (2012). In Figure 3.3 we illustrate the idea that fault morphology may result in asperities (within the broad seismogenic zone) that have a different behavior to their neighbors and a tendency to create along-strike and down-dip segmentation.

CHAPTER 3. FAULT MORPHOLOGY CAN CONTROL RUPTURE SIZE

In our quest for more predictive earthquake models, these results suggest that gaining detailed images of fault morphology may be key to better understanding fault segmentation, the variability of earthquake size, and potentially the rupture style. Collaboration between structural geologists and groups modeling rupture propagation may be key to the next generation of earthquake forecasting.

3.7 Acknowledgments

This research was supported by the National Research Foundation Singapore under its Singapore NRF Fellowship scheme (National Research Fellow Award Nos. NRF-NRFF2010-064, NRF-NRFF2013-04 and NRF-NRFF2013-06) and by the EOS and the National Research Foundation Singapore and the Singapore Ministry of Education under the Research Centres of Excellence initiative. ALOS data were provided by JAXA. Sentinel-1A data were copyrighted by ESA. We thank the Canadian Space Agency for providing RADARSAT-2 data. We are grateful to Jean-Philippe Avouac for useful discussions and for making the GPS data immediately available. This is EOS paper number 108.

Transient rheology in the Sumatran mantle wedge revealed by a decade of great earthquakes

Qiang Qiu, Emma M. Hill, Sylvain Barbot, Lujia Feng. Manuscript to be submitted to *Science*.

4.1 Abstract

Postseismic processes involve the response of the Earth's lithosphere and asthenosphere to stress changes produced as a result of earthquake ruptures. Multiple postseismic mechanisms exist, including afterslip on the fault interface and viscoelastic flow of the asthenosphere. The effects of these processes on observations of post-earthquake surface deformation have traditionally been modeled independently - using inversion for afterslip and forward models for viscous flow - resulting in tradeoffs and uncertainties in estimates for both the levels of afterslip and the viscoelastic rheology. Here we present a newly developed technique that is capable of jointly inverting for fault afterslip and viscous strain within a 3D finite volume located at asthenospheric depths, with no need for prior assumptions on the rheology or mechanism. We implement this approach for the Sumatra subduction zone to model postseismic deformation following the recent slew of large earthquakes over the last decade. In this paper we focus primarily on examining the rheology of the mantle wedge. We show that viscous flow in the mantle wedge has transient behavior similar to that expected from a Burgers rheology, with estimated $\eta_K = \sim 10^{17}$ Pa.s and $\eta_k = \sim 10^{19}$ Pa.s; it is not a typical linear Newtonian flow. This transient rheology contributes to significant deformation in the early time stages after the earthquakes.

4.2 Introduction

Large coseismic fault ruptures result in stress changes in the lithosphere and asthenosphere, generating postseismic deformation over the following months (*Hsu et al.*, 2006), years (*Copley*, 2014), decades (*Suito and Freymueller*, 2009), or maybe even longer. Understanding the mechanisms of postseismic deformation can provide insights into stress evolution on faults (from afterslip) and the rheological structure of Earth's uppermost layers (from viscoelastic relaxation of the asthenosphere).

Figure 4.1: (a) Accumulated postseismic displacement over different time periods (red, blue, and green vectors) over the past ~ 9 years. Beach balls represent the moment tensor of earthquakes with color corresponding to magnitude as given in the vertical bars of panel (b). Contours with the same color as the moment tensor show the outline of the estimated coseismic rupture patch for that event, from *Chlieh et al.* (2007); *Konca et al.* (2007, 2008); *Hill et al.* (2012). Inset map shows our study area and the geodetic network used in this study, with the green arrows representing plate motion. (b) Comparison between model predictions and GPS data at a few selected sites. The vertical bars indicate the year and magnitude of earthquakes.

CHAPTER 4. TRANSIENT MANTLE WEDGE FLOW IN SUMATRA

Large differences in postseismic modeling results highlight the continuing challenge to discriminate whether the dominant postseismic process following an event is distributed viscous flow of the asthenosphere over a broad region, more localized ductile shear, or frictional afterslip on the fault. For example, postseismic deformation following the 1992 M_w 7.2 Landers earthquake in California, USA, was explained by afterslip (*Savage and Svarc, 1997; Perfettini and Avouac, 2007*), viscous mantle flow (*Deng et al., 1998; Pollitz et al., 2000*), poroelasticism combined with afterslip (*Fialko, 2004*), and also viscous flow with poroelastic rebound (*Masterlark and Wang, 2002*). Particularly challenging is the fact that the various different mechanisms can in some places have similar deformation patterns, so they trade off with each other in modeling, especially when the deformation time series are short. Clearly, a new approach is needed to better model and differentiate the different postseismic mechanisms at play.

At the Sumatra subduction zone, the forearc island chain sits directly above the seismogenic zone (Figure 4.1). The Sumatran GPS Array (SuGAR; Figure 4.1, Figure B.1) has been recording surface displacement across these islands and the Sumatran mainland since 2002 (*Feng et al., 2015*), affording a unique opportunity to better constrain the kinematics of the megathrust and subsurface rheology. Here, we present our newly developed algorithm that is capable of jointly inverting for both frictional afterslip and viscous strains within 3D finite volumes (rectangular blocks) distributed at asthenosphere depths, simultaneously and self-consistently, and we implement this algorithm to assess a possible range of effective viscosities in the mantle wedge of the Sumatra subduction zone, to better assess the behavior of the local rheology structure.

4.3 Geodetic joint inversion approach

At depth, the asthenosphere is expected to flow in response to stress changes from the coseismic rupture. To model this flow, we discretized the mantle wedge into rectangular blocks (Figure B.2). The dimensions of the blocks are slightly variable, but approximately 100 km in length, 60 km in width and 40 km in depth. We use relative large dimension of blocks to represent the mantle wedge due to sparse GPS network at the main-island. Each block can deform by six components of strain corresponding to six different patterns of surface deformation. Linear summations of these six strain components determine the final deformation of each rectangular block and its corresponding surface displacement pattern. We calculated the surface displacement (Green's function) for each strain component by specifying unit strain for each block. This 3D Green's function maps the strain component on the block in the mantle wedge to the surface displacement field at GPS sites, and then is used to invert for the time variation of strain for each component of every block. We use the Relax dynamic modeling code (*Barbot et al., 2009; Barbot and Fialko, 2010*) to calculate these Green's functions (see the Methods in Appendix B).

CHAPTER 4. TRANSIENT MANTLE WEDGE FLOW IN SUMATRA

We incorporated our 3D Green’s functions for deformation of the blocks into the Extended Network Inversion Filter (ENIF, *Segall and Matthews*, 1997; *McGuire and Segall*, 2003), which is a Kalman-filter based inversion approach capable of distinguishing fault slip from various source of noise, to model the spatial and temporal evolution of both fault slip and strains on each block. The extra model space from the strain components of all the blocks together with the fault slip on fault patches yield a highly underdetermined inverse problem. We therefore include constraints to force smooth distributions for both the fault slip and strain. In addition, no volume changes are allowed for any of the blocks. Finally, the strain is assumed to approximately follow the direction of the coseismic stress changes. This is physically likely, because it is stress turbulences imposed by coseismic slip that relax during the postseismic process (see Methods B.3 in Appendix B). We have done synthetic tests for joint and separate inversions. These results indicate that our new technique can recover the input slip and/or viscous strain very well (see Synthetic tests section B.4 in Appendix B).

Our megathrust geometry is derived from Slab 1.0 (*Hayes et al.*, 2012), as modified and also extended further north by *Hu and Wang* (2012). The rectangular blocks are located directly below the bottom edge of the fault patches, with depths ranging from ~ 50 km to ~ 230 km (Figure B.2), which is consistent with previous results for modeling the viscous flow following the 2004 M_w 9.2 Sumatra-Andaman earthquake (*Hoechner et al.*, 2011; *Hu and Wang*, 2012; *Pollitz et al.*, 2006, 2008; *Han et al.*, 2008; *Panet et al.*, 2010). We invert the geodetic data for both the time evolution of afterslip on the fault and viscous strains for each block in the mantle wedge using our modified version of the ENIF.

4.4 Geodetic data and inversion model set up

We combine several different geodetic data sets to constrain more than a decade of afterslip and viscous strain following recent large earthquakes along the Sumatra subduction zone (Figure 4.1). We use continuous GPS (cGPS) data for 30 sites of the SuGAR network (those with high signal-to-noise ratios), one IGS site in Singapore, and four sites on the Andaman Islands (*Paul et al.*, 2012). We also use survey mode and cGPS data from 12 sites in Thailand from *Broerse et al.* (2015). In addition, we extract information on land-height change from 24 regional tide gauges, in the far field region of the megathrust. To do this we subtract satellite altimetry AVISO time series for sea-surface height changes from tide-gauge measurements of relative sea level (see Geodetic data section B.1 in Appendix B), with the residuals representing land-height change. Data from the SuGAR network and tide gauges are available since the 2004 Sumatra-Andaman earthquake, while the GPS data from *Broerse et al.* (2015) are shorter, with some sites covering only three or four years beginning in 2005, 2006 or 2008. (See Geodetic data section B.1 in Appendix B).

Figure 4.2: Accumulated afterslip and deviatoric viscous strain (top layer blocks and blocks of the last column of the rest two layers, see Figure 4.3 (b)) in the mantle wedge, and effective viscosity estimates. (A) Accumulated afterslip and deviatoric viscous strain over ~ 9 years since 2005. Vectors show a comparison between predicted model displacements and data. Black dashed lines 1 and 2 separate, for discussion, the Sumatran-Andaman and Nias segments, and the Nias and Bengkulu segments respectively. Letters a to h indicate locations for which we show time series for effective viscosity estimates; locations a to d are shown in (B), with e to h shown in Figure B.6. Small black dots represent the center locations of our viscous blocks. The afterslip at Nicobar segment is shaded out due to poor resolution. (B) Examples of estimated effective viscosities, that have been low-pass filtered (blue) and a model based on a Burgers rheology (red), for sites a to d as shown in panel (a). The best-fitting Kelvin η_K and Maxwell η_M viscosities are given below each time series. The shaded region shows the lower and upper bounds of the effective viscosities.

4.5 Results

Our estimated afterslip following earthquakes at Sumatra subduction zone is primarily located in the areas surrounding coseismic rupture patches, except in places where there is too little data to

CHAPTER 4. TRANSIENT MANTLE WEDGE FLOW IN SUMATRA

properly constrain the afterslip model (e.g., the Nicobar section (Figure 4.1)). Our estimates for the spatial distributions of afterslip following these events are generally consistent with previous studies (Hoechner *et al.*, 2011; Paul *et al.*, 2007, 2012; Gunawan *et al.*, 2014; Hsu *et al.*, 2006; Lubis *et al.*, 2012; Tsang *et al.*, 2016). A more detailed afterslip description is given in Text B.5 in Appendix B. Our model predicts reasonable fits with both the near-field and far-field GPS data sets, and far-field tide-gauge data, for Sumatra (Figure B.3).

Figure 4.3: Final estimates for the accumulated deviatoric viscous strain distribution over the three layers of blocks. (a) Surface projection of the fault patches and viscous blocks. A-A', B-B' and C-C' are three cross sections to show the distributed viscous strain. (b) Viscous strain along the three cross sections shown in (a). The blue curves represent the center location of fault subpatches along these cross sections.

The important role of accounting for viscoelastic flow when estimating afterslip has been reported by Sun *et al.* (2014); Sun and Wang (2015). Sun *et al.* (2014) argue that omitting the viscous flow will underestimate the shallow afterslip and overestimate the deep afterslip. As an example, Sun and Wang (2015) report that the shallow afterslip could be underestimated by $>50\%$. To demonstrate whether our new technique can capture this afterslip difference when including the viscous flow estimation, we calculate the difference in afterslip between models which do and do not include viscous flow. Generally, the shallow afterslip is increased and the deep afterslip is decreased (Figure B.5), which are well captured by our new technique.

Below the deep afterslip of the northern Sumatra, Nias and Bengkulu segments, localized viscous strains across layered blocks are observed over the last decade (Figures 4.2 and 4.3). Our estimated viscous strains decay with distance from the megathrust, both horizontally and with

CHAPTER 4. TRANSIENT MANTLE WEDGE FLOW IN SUMATRA

depth (A-A', B-B' and C-C' in Figure 4.3), which is consistent with the coseismic stress changes.

To investigate the temporal evolution of strain in more detail, we select eight blocks that are closest to the local GPS sites, affording our highest resolution. They are four blocks below the 2004 M_w 9.2 Sumatra-Andaman rupture region, and two below the 2005 M_w 8.6 Nias-Simeulue rupture patch, and the others are below the 2007 M_w 8.4 Bengkulu rupture area (Figure 4.2 and Figure B.6). We then estimate time-dependent effective viscosities for these boxes, assuming the coseismic stress changes are released by viscous strains on the blocks. Using the coseismic stress changes and our estimated strain rates, we can estimate the effective viscosities, with the evolution of stress divided by the viscous strain rate. Background stresses and strain rates for the mantle wedge should be also taken into account for estimating effective viscosity, because the dynamics of the mantle flow are a quasi-static process. However, deriving the actual values of the background strain rates and stresses is not possible without direct measurements of the mantle. Thus, we assume a range of typical geological strain rates ($\sim 10^{-16} \text{ s}^{-1}$ to $\sim 10^{-14} \text{ s}^{-1}$) and a Maxwell rheology with varying viscosity to estimate the background stress level. These strain rates and Maxwell viscosities are determined when fitting the time evolution of the effective viscosity curve. The absolute values of effective viscosity and afterslip are not well determined due to the inherent trade-offs, uncertainty of the background stress and strain rate, physical parameters of the ambient rocks, and the limitation of the data coverage. Our strategy is to test different weighting parameters, and to balance the model fits with a plausible/acceptable viscous flow.

Generally, the estimated effective viscosities for the blocks start with small values ($\sim 10^{17} \text{ Pa.s}$) at the beginning, then progressively increase to larger values ($\sim 10^{18}$ to $\sim 10^{19} \text{ Pa.s}$) within a year or two. After that the viscosities maintain approximately constant values over the following several years. This transient behavior of the viscosity in the early time stages and relatively constant values over the later time stages is similar to the effective viscosity prediction of a Burgers rheology. To quantify the similarity, we fit our estimated effective viscosities with the effective viscosity formula of the Burgers body from *Pollitz* (2003) (See Methods section B.3 in Appendix B) by conducting a grid search for parameters that include the background strain rate, Maxwell viscosity, Kelvin viscosity and the shear modulus of a Kelvin body (an example for site A is shown in Figure B.6). Examples of the final estimated viscosities (low-pass filtered) and the Burgers rheology parameters for sites A and B are shown in Figure 4.2 (additional sites are shown in Figure B.6). Notably, we see excellent fits between our effective viscosity estimates and the predictions from the Burgers rheology model that has been widely used for postseismic modeling studies (*Hoechner et al.*, 2011; *Hu and Wang*, 2012; *Pollitz et al.*, 2006, 2008; *Han et al.*, 2008; *Panet et al.*, 2010). Our estimated Kelvin viscosities and Maxwell viscosities are at levels of $\sim 10^{17} \text{ Pa.s}$ and $\sim 10^{19} \text{ Pa.s}$ respectively. Note that the absolute values of the viscosities are less well determined. This is because the partial and temporal resolution of data sets can vary the final viscosity estimations significantly. The more spatially dense the network and the longer the time span of the data, the better the constraints on the viscosity.

4.6 Discussion

We have demonstrated that our new technique, using joint inversion for time-varying afterslip and viscous flow represented by strain of rectangular blocks in the mantle wedge, is capable of imaging afterslip and viscous strains simultaneously. This technique does not require prior assumption on the rheology of the asthenosphere, and our results provide novel information on the viscous behavior of the mantle wedge of Sumatra.

Our estimated Burgers viscosities for the mantle wedge are comparable with a Kelvin viscosity of 4×10^{17} Pa.s and a Maxwell viscosity of 8×10^{18} Pa.s as estimated by *Panet et al.* (2010), who used 3 years of GRACE and GPS data in Thailand. Our estimate is also consistent with a Kelvin viscosity of 10^{18} Pa.s and a Maxwell viscosity of 7.5×10^{18} Pa.s as constrained by 7 years of GPS data from Thailand (*Broerse et al.*, 2015). *Broerse et al.* (2015) also report that the rheology structure constrained by the GPS data fails to explain the GRACE data, suggesting that different data sets may capture different information on the postseismic deformation and have their own limitations. Thus a combination of multiple data sets could compensate with each other and could better resolve the rheology of the asthenosphere.

Within the uncertainties, the estimated effective viscosities at our selected strain blocks are similar regardless of location. This suggests that the viscosity structure of the mantle wedge in the Sumatra subduction zone does not have a power-law rheology, for which we would expect small viscosities where the coseismic stress changes are high and higher values where the stress changes are low. These findings are different to observations of power-law rheologies for the viscoelastic flow following the 1992 M_w 7.3 Landers and 1999 M_w 7.1 Hector Mine earthquakes of California, (*Freed and Bürgmann*, 2004) and the 2002 M_w 7.9 Denali, Alaska (*Freed et al.*, 2006) strike-slip earthquakes. The reasons may be related to the tectonic setting and/or ambient rock environment at the local scale or trade-offs from existing multiple mechanisms.

We find that viscous flow of the mantle wedge along the Sumatra subduction zone has a transient behavior similar to a Burgers rheology. It does not appear to have a typical Maxwell rheology as was assumed in previous studies, and not a power-law rheology. Our new technique can be implemented for other regions to gain a more comprehensive understanding of fault mechanics and rheology of the asthenosphere in both continental regions and subduction zones worldwide.

Ongoing postseismic deformation following the 2005 M_w 8.6 Nias-Simeulue earthquake: Joint inversion of long-term GPS time series for afterslip and viscous flow

Qiang Qiu, Emma M. Hill, Sylvain Barbot, Lujia Feng, Aron Meltzner. Manuscript in preparation for *Journal of Geophysical Research*.

5.1 Abstract

Coexistence of multiple postseismic mechanisms introduces complexity in modeling and understanding observations of postseismic deformation. In particular, difficulties in simultaneously and self-consistently modeling both afterslip on a fault and viscous flow of the asthenosphere can lead to trade-offs and inaccuracies in determining their relative amplitudes and spatial patterns. Here, we present a joint inversion for afterslip and viscous flow, using ~ 9 years of GPS data following the 2005 M_w 8.6 Nias-Simeulue earthquake of Sumatra. Our technique allows us to invert for these processes simultaneously and without prior assumption of a specific rheology. In addition, we present revised estimates of the coseismic distribution for this event. We find that the afterslip occurs primarily in the area surrounding the coseismic rupture patch, which is consistent with expectations from rate-and-state friction laws. Our data suggest that ~ 6 to 7 years following the earthquake the slip rates on the fault have reduced to very low, but are still declining. Our estimated viscous strain rates in the mantle wedge remain significantly higher than typical geological rates; the viscous flow appears to be ongoing, and may continue for years, decades or maybe even longer. The estimated effective viscosities in the mantle wedge suggest the viscous flow here has a transient behavior similar to a Burgers rheology.

5.2 Introduction

Large earthquakes release, in seconds to minutes, strain that has been accumulated on tectonic faults over many years. The coseismic slip, in turn, generates stress changes around the rupture area. As a response to these stress changes, the fault continues to slip following earthquakes (so-called afterslip). In addition, the local asthenosphere flows viscously as a response to the stress changes, causing significant surface displacement. Estimating the spatial distribution and magnitude of afterslip is important towards understanding stress redistribution on the fault and the triggering of aftershocks and future earthquakes. Understanding the viscous response can provide information on the rheology of the area, which is helpful for earthquake cycle investigations.

The M_w 8.6 Nias-Simeulue earthquake occurred on 28 March 2005 offshore northern Sumatra (Figure 5.1). Coseismic studies for this event reported that >11 m of slip occurred on the Sunda megathrust under Nias Island, forming a saddle-shaped rupture area (*Briggs et al.*, 2006; *Konca et al.*, 2007). Following the earthquake, significant postseismic displacements were detected by the Sumatran GPS Array (SuGAr) (Figure 5.1). *Hsu et al.* (2006) modeled the displacements for the 11 months following the earthquake as pure afterslip, with maximum cumulative slip of ~ 1.4 m occurring up-dip of the main coseismic rupture area. They estimated a cumulative magnitude for the afterslip equivalent to M_w 8.2.

Imaging afterslip can be a challenging task due to the co-existence of multiple postseismic mechanisms. For example, afterslip and early viscous flow are likely coupled together. Early-stage viscous flow was detected by a seafloor geodetic network following the 2011 Tohoku-Oki earthquake (*Watanabe et al.*, 2014) - the fact that the seafloor instruments were located over the shallow megathrust made it easier to distinguish between viscoelastic and afterslip deformation, since the displacements are opposite over this area of the fault. These results indicate that viscous flow is prevalent through the whole relaxation time period and not, as often previously assumed, to be important only in the longer time period. *Sun et al.* (2014) argue that the viscous flow coexists with afterslip, and thus both have to be modeled together; if they are not, shallow afterslip following megathrust earthquakes may be underestimated, and deep afterslip may be overestimated (*Sun et al.*, 2014; *Sun and Wang*, 2015).

Viscous flow is a complex process. The flow behavior is highly dependent on the local tectonic setting and the ambient environment of the rocks. Based on laboratory studies, rocks can flow as transient (*Carter and Kirby*, 1978; *Chopra*, 1997; *Bürgmann and Dresen*, 2008), diffusion, and/or dislocation creep (*Karato*, 2010; *Karato and Wu*, 1993; *Bürgmann and Dresen*, 2008). Different mechanisms produce similar surface displacement patterns, and thus introduce inherent trade-offs in modeling their deformation fields, resulting in different mechanisms being used to explain one single event (*Pollitz et al.*, 2000; *Perfettini and Avouac*, 2007; *Fialko*, 2004; *Freed and Bürgmann*, 2004). Thus, a new approach to modeling postseismic deformation is needed.

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

To the best of our knowledge, no previous postseismic studies have used time series longer than 11 months to model the 2005 M_w 8.6 Nias-Simeulue earthquake. Moreover, previous models neglected viscous flow. In our study, we reanalyze the afterslip process by using GPS time series spanning ~ 9 years. The following aspects of our study are novel:

1. We incorporate viscous flow estimation simultaneously with afterslip by inverting strains within finite volume blocks distributed in the mantle wedge, using a new technique as described by *Qiu et al.* (2016); and
2. We reanalyze the coseismic slip and afterslip in a self-consistent way with a geometry that is more realistic than the planar fault geometry used in other studies; and
3. We use coseismic offsets and postseismic time series that were processed to self-consistently and simultaneously estimate the effects of all earthquake signals in the SuGAR time series, resulting in more accurate and more spatially coherent deformation fields than used in previous studies (*Feng et al.*, 2015).

Our results allow us, firstly, to gain deeper insights into the spatial relationships between coseismic and afterslip behavior on the megathrust; second, to answer whether the afterslip has finished or is still ongoing; and third, to try to understand the behaviors of viscous flow in the mantle wedge.

5.3 GPS data

We use continuous GPS time series for ~ 9 years following the 2005 M_w 8.6 Nias-Simeulue earthquake. We use data from 13 stations, with 12 stations part of the SuGAR network, and one station SAMP operated by the Indonesian Geospatial Information Agency (BIG, formerly known as BAKOSURTANAL) (Figure 5.1). We processed the raw time series using GIPSY-OASIS version 6.2, and estimated coseismic and postseismic amplitudes by fitting a model that contains offsets and exponential/logarithmic functions simultaneously for all events that were recorded, plus annual and semi-annual periodic cycles. The technique for processing and analysis of the GPS data is outlined in more detail by *Feng et al.* (2015).

5.4 Geodetic inversion methods

In this section we present the fault geometry and crustal model that were used in our inversion. We then describe our techniques for coseismic slip inversion and joint inversion of afterslip and viscous strain.

5.4.1 Fault geometry and crustal model

Previous coseismic studies of the Nias-Simeulue earthquake used a planar fault geometry for their modeling (*Briggs et al.*, 2006; *Konca et al.*, 2007). The afterslip study of *Hsu et al.* (2006) used a geometry with only two different dips and a sharp change between the two.

Figure 5.1: Coseismic and cumulative postseismic displacement of the 2005 M_w 8.6 Nias-Simeulue earthquake. The purple contour is our preferred coseismic slip model with a 2-m interval. The black beach ball represents the moment tensor of the main shock from GCMT catalog. The Sunda megathrust is represented by the black curve.

In order to use a more realistic fault geometry, we create a fault mesh with a curved interface that is based on Slab 1.0 (*Hayes et al.*, 2012). We verified the Slab 1.0 geometry against relocated seismicity recorded by a temporary local array (*Tilmann et al.*, 2010) and see good agreement (Figure C.1 in Appendix C). Our final mesh is divided into rectangular fault patches with a dimension of $\sim 30 \times 30$ km.

Using this fault mesh, we produce Green's functions for fault slip using EDGRN (*Wang et al.*, 2003). We assume a layered elastic Earth model based on Crust 1.0 (*Laske G. and Pasyanos* (2013); see Table C.1).

5.4.2 Coseismic slip inversion

Slip models estimated by different studies for the same event often vary significantly due to differences in inversion techniques, assumed fault geometries and the data being used in modeling. This can make it difficult to compare models, including comparing coseismic and afterslip distributions produced by different groups. In order, therefore, to get self-consistent slip models we reanalyzed the coseismic slip of the Nias-Simeulue earthquake with the same fault geometry that we use for our afterslip study.

Figure 5.2: Geometry of the fault and viscous strain blocks. (a) Geometry of the fault (e.g., rectangular small patches) and viscous strain blocks (e.g., the large rectangular patches); (b) Cross-section view along A-A' and arrangement (e.g., numbers) of these blocks.

To estimate coseismic slip we use a linear least-squares method. We constrain the rake to $118 \pm 45^\circ$, based on the value given in the GCMT catalog, and apply a Laplacian constraint to ensure spatially smooth slip and non-negative constraints to get rid of backward slip. We

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

select a smoothing weight for our final model based on a average stress-drop curve as shown in Figure C.2.

To quantify uncertainties in our estimated coseismic slip, we use a Bootstrap method (*Feng et al.*, 2016b; *Parsons et al.*, 2006; *Wright et al.*, 2003). This technique uses a Monte Carlo based resampling scheme (see Text C.1 in Appendix C). In our study, we performed 100 iterations for estimating the slip uncertainty; the results are shown in Figure C.3.

5.4.3 Joint inversion of afterslip and viscous strain

In *Qiu et al.* (2016) we presented a new technique that is capable of simultaneously inverting time-variable fault slip and viscous strain within a finite volume (rectangular block) located at asthenosphere depths, without *a priori* assumptions of the rheology. Here, we apply this approach to invert the spatial and time evolution of afterslip and viscous strain at the Nias segment of the Sumatra subduction zone. Beyond *Qiu et al.* (2016), we use the 2005 Nias-Simeulue event as a case study for testing the limitations and inherent trade-offs of this new technique.

To model viscous flow, we discretized the lower crust and the upper mantle beneath the coseismic rupture patch into three layers of rectangular blocks that extend from the bottom of the fault to a depth of 200 km (Figure 5.2). For each layer, there are three blocks extending 180 km perpendicular to the fault. Each block has dimensions of 570 km along-strike (extending the whole length of the Nias-Simeulue rupture area), 60 km in width, and 40 km in depth (Figure 5.2). These blocks are relatively large, but we do not have the resolution to use smaller blocks since each block has six strain components to be resolved (in addition to the afterslip estimates). We have limited data coverage directly above the downdip part of the fault and the model blocks (only two stations PBLI and SAMP, Figure 5.2); therefore, the resolution in this area is poor (see the resolution matrix in Figure C.4).

Our technique implements a modified version of the Extended Network Inversion Filter (ENIF, *Segall and Matthews*, 1997; *McGuire and Segall*, 2003; *Qiu et al.*, 2016, and see Text C.2 in Appendix C). We use this to model the time evolution of both afterslip and viscous strain constrained by the ~ 9 -year time series. We run the filter with a 10-day sampling rate, to reduce the computational burden compared with a faster rate. Our selected sampling rate may have some effects on the estimation of transient slip rates or strain rates in the early time stages, but it should not significantly change the estimates of cumulative slip and strain. We apply spatial smoothing for both slip rates and strain rates. To model a viscous-like strain, we implement a deviatoric (no volume change) constraint in the inversion (see Text C.3 in Appendix C). We also apply a constraint to guide the postseismic viscous strain to point in approximately the same direction as the coseismic stress changes (see Text C.3 in Appendix C), as would be the case for any postseismic dynamic model driven by coseismic stress changes.

5.5 Results

We summarize our results for coseismic slip, afterslip, viscous strain and effective viscosity in the following two sections. We find that coseismic slip occurred on the megathrust underneath of Nias and southern Simeulue. Afterslip was located primarily updip and downdip of the coseismic rupture area, with cumulative viscous strain occurring downdip of the coseismic rupture patch.

5.5.1 Coseismic slip

The coseismic slip distribution has two concentrations separated by the epicenter, forming a saddle-shaped rupture area (Figure 5.3). We estimate maximum slip to have occurred directly under the northern tip of Nias Island with a magnitude of >12 m. The northern part of the rupture saddle has estimated maximum slip located a little downdip of the southern tip of the Simeulue Island (Figure 5.3). We estimate reduced slip (<2 m) between the two higher slip patches.

Figure 5.3: Coseismic slip estimation from joint inversion of GPS and coral data. The color scale is saturated at 14 m.

Overall, the general pattern of our coseismic slip is consistent with published slip models (*Briggs et al., 2006; Konca et al., 2007*). Some differences in slip details do exist; these may be the result

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

of differences in the fault geometry, data sets, displacement estimates, and inversion techniques. To illustrate, we plot our slip model alongside the slip model estimated by *Konca et al. (2007)* which used joint inversion of teleseismic, GPS and coral data sets (Figure C.5).

The general features of the coseismic slip distribution appear to be robust, but some areas suffer from limited resolution. The slip uncertainties for the maximum slip regions are <50 cm, while the slip uncertainties are >1 m under the Banyak Islands and the southern part of Nias, suggesting that the slip is not well constrained in these locations (Figure C.3). The large uncertainties in our slip estimates for these areas are due to paucity of stations there.

Figure 5.4: Complementary spatial relationship between the afterslip and coseismic stress change. Green contour is the 0.5-m contour of the final cumulative afterslip. The color shows the Coulomb stress changes from our preferred coseismic slip model. Black dots are aftershocks for the period 2005 to 2014 from the Advanced National Seismic System (ANSS) catalog.

Figure 5.5: Final cumulative viscous strain of blocks in cross-section view of A-A' (see Figure 5.2).

5.5.2 Afterslip, viscous strain and effective viscosity

In general, our predicted postseismic model displacements have good fits to the time series (Figure C.6). Our estimates suggest that afterslip following the 2005 M_w 8.6 Nias-Simeulue earthquake surrounds the coseismic rupture patch (Figure 5.4). This spatial complementarity of the afterslip and coseismic rupture patches is consistent with predictions from rate-and-state friction laws and the expectation that afterslip is driven by the coseismic stress changes (Figure 5.4). We note that most of the aftershocks are located on the boundary between the updip edge of the coseismic patch and the afterslip (Figure 5.4).

Downdip of the deep afterslip under the Banyak Islands, our model estimates viscous strain primarily across the first two boxes of the first layer (Figure 5.5), decaying both horizontally and vertically away from the fault. This decay is consistent with the spatial decay pattern of the coseismic stress changes (Figure C.7).

Figure 5.6: Effective viscosity estimation of blocks 1 and 2. The blue curves are the low-pass filtered estimates of effective viscosity, with the shaded region showing the lower and upper bounds. The red curves are the predicted effective viscosities from a Burgers rheology with Kelvin viscosity η_K and Maxwell viscosity η_M , respectively.

We evaluate the time evolution of the stress relaxation for the first two blocks in the top layer (Figure 5.2). These blocks experience the most significant viscous strain, and are also closest to the GPS network and therefore better resolved (Figure C.4). This stress evolution is estimated based on the stress kernel at these blocks in conjunction with the inverted viscous strain components for each block from our modified ENIF (see Text C.4 in Appendix C). For this calculation, we assume the coseismic stresses are relaxed by these viscous strains on these blocks (see Text C.4 in Appendix C). Using the stress curves and strain rates estimated using our ENIF, we can evaluate the effective viscosity for these blocks by dividing the stress by the strain rate (see Text C.4 in Appendix C). In addition, the background stress and the strain rate of the mantle flow should also be taken into account for evaluating the effective viscosity, since the dynamics of the mantle are a quasi-static process. However, deriving the actual values of the

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

background stress and strain rates is a challenging task; obviously, we do not have instruments deployed directly in the mantle to measure these values. Thus, we assume a Maxwell rheology with a certain range of strain rates to estimate the background stress. We fit the effective viscosity curve at these blocks with a Burgers rheology. We estimate the background strain rate, Maxwell viscosity, Kelvin rigidity and viscosity using a grid search. We show the final determined effective viscosity and the Burgers rheology fit in Figure 5.6, with the grid search results shown in Figure C.8.

5.6 Discussion

In the following sections, we first discuss the idea that the low-slip patch of the 2005 coseismic slip is likely related to the subducting features on the megathrust. We then discuss temporal and spatial afterslip, and the slip rates which have reduced to very low, but still declining after ~ 7 years. Lastly we conclude that our estimated effective viscosities that have a transient behavior similar to a Burgers rheology.

Figure 5.7: Multiple slip styles at the low-slip patch of the 2005 M_w 8.6 Nias-Simeulue earthquake. The coral sites for the 1861 earthquake are shown with the light blue dots. Color indicates the coseismic slip estimation of the 1861 earthquake inferred from the coral data (light blue dots). The blue contour represents our preferred coseismic slip model of the 2005 event. The Banyaks slip patch was estimated by *Morgan et al.* (2016) and the SSE slip patch was estimated by *Tsang et al.* (2015).

5.6.1 Coseismic slip

An $M_w \sim 8.5$ event in 1861 (*Natawidjaja et al.*, 2004) closely resembles the 2005 Nias-Simeulue event in terms of along-strike rupture extent and the vertical displacement of corals at some sites (*Meltzner et al.*, 2015). To get a first order picture of the coseismic slip patch for the 1861 event, we applied the same approach as we used to invert for the 2005 coseismic slip, using the coral measurements from *Meltzner et al.* (2015). We show the results in Figure 5.7. Only two sites recorded the full coseismic offsets; the remaining sites recorded the minimum uplifts in the 1861 events. Thus, we here present in Figure 5.7 only the minimum coseismic slip of the 1861 event. Notably, the spatial patterns of the 1861 and 2005 ruptures look remarkably similar, with two isolated large slip patches and a low slip patch in between them. Such similarity would be accentuated if we knew the maximum vertical uplifts at the remaining sites (Figure 5.7).

The patch of lower slip between southern Simeulue and northern Nias appears persistent for both the 1861 and the 2005 events. Our error analysis indicates > 1 m uncertainty for this region (Figure C.3), suggesting that this area is poorly resolved. However, the coral measurements do suggest that the feature is robust due to the strong spatial displacement gradient.

Figure 5.8: A comparison between afterslip models that did not (a) and did (b) include viscous flow. The green dots show the locations of the time evolution of the slip rates in Figure 5.10. Note that the vertical and horizontal components are plotted with different scales.

The low-slip patch has been proposed to be related to structural features of the subducted bathymetry of the outer-arc ridge (*Briggs et al.*, 2006; *Jacob et al.*, 2014). In *Qiu et al.* (2016) we demonstrated an example of one way in which structural features can modify rupture patterns as a result of variations in fault morphology, using the M_w 7.8 Gorkha earthquake in Nepal as a case study. In this study, we showed that local fault morphology can bound the rupture extent, resulting in different rupture behaviors. In the Nias-Simeulue region, the 2010 M_w 7.8 Banyak Islands earthquake (Figure 5.7) occurred in the low slip rupture area of the 2005 and 1861 events,

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

as did a 15-year long slow slip event (SSE) that occurred between 1966 and 1981 (*Tsang et al.*, 2015) located down-dip of the 2010 event (Figure 5.7).

5.6.2 Afterslip and afterslip rate

We estimate that afterslip following the 2005 earthquake occurred primarily in the area surrounding the coseismic rupture patch (Figure 5.4). The afterslip does not appear to have migrated spatially over time. This is also suggested by the azimuth of the horizontal displacement at GPS stations, which do not change significantly over time (Figure 5.1). Our afterslip model differs to that of *Hsu et al.* (2006) in details of slip pattern and slip amount, because of the longer period of data, different fault geometry and the fact that we included viscous flow of the asthenosphere.

Figure 5.9: Difference in afterslip between models that did and did not include viscous flow.

Our model that includes viscous flow yields a similar statistical fit to the data as compared with our model that does not include the viscous flow. However, the inclusion of viscous flow improves the fits to the vertical component of displacement at station SAMP, which is located in the far field of the fault (Figure 5.8). The inclusion of viscous flow in our model increases the shallow afterslip and reduces the down-dip afterslip. However, the absolute values are not well determined due to various trade-offs in the modeling (see Text C.5 in Appendix C). Figure 5.9 shows the differences between the two models. Some areas of the shallow afterslip did not increase when viscous flow is included in the model, but these areas have low resolution and poor

data coverage. The reduction in the amount of estimated afterslip for the shallow megathrust is consistent with the investigation of *Sun et al.* (2014).

5.6.3 Effective viscosity

The time evolutions of effective viscosities for our strain blocks 1 and 2 (Figure 5.2) show a similar curvature. They have relatively small values of $\sim 10^{17}$ Pa.s directly after the earthquake, and increase progressively to higher viscosities of $\sim 10^{18}$ Pa.s in the following year (Figure 5.6; and stress and strain rates are shown in Figure C.9). After reaching a viscosity of $\sim 10^{18}$ Pa.s, they stay almost constant for the following years. This time-varying viscosity in the early time stages and almost constant viscosity in the later time period suggests that viscous flow in the mantle wedge has a transient behavior. Such transient behavior and constant long-term viscosity is similar to the predicted effective viscosity of a Burgers rheology. A Burgers body has Kelvin and Maxwell elements connected in series, with the Kelvin element explaining the early fast transient displacements and the Maxwell element explaining the long-term displacement field. A Burgers rheology has been successfully used to model postseismic deformation following the neighboring 2004 M_w 9.2 Sumatra-Andaman earthquake (*Hoechner et al.*, 2011; *Hu and Wang*, 2012; *Pollitz et al.*, 2006; *Han et al.*, 2008; *Panet et al.*, 2010). The relaxation times of the Kelvin and Maxwell elements are decided by their corresponding viscosities and shear moduli. To quantitatively compare our effective viscosity estimates with those predicted from a Burgers rheology, we fit these viscosity curves with the effective viscosity formula of the Burgers rheology (see Text C.6 in Appendix C). These estimated viscosities fit quite well with a Burgers rheology that has Kelvin viscosity $\sim 10^{17}$ Pa.s and Maxwell viscosity $\sim 10^{18}$ Pa.s (Figure 5.6).

Our estimates for Kelvin and Maxwell viscosities for blocks 1 and 2 fall well within the range of $10^{17} - 10^{19}$ Pa.s, which is a similar range as estimated for postseismic deformation following the 2004 M_w 9.2 event (*Hoechner et al.*, 2011; *Hu and Wang*, 2012; *Pollitz et al.*, 2006; *Han et al.*, 2008; *Panet et al.*, 2010). We note, however, that our estimates of viscosity are poorly constrained in that different data coverage and time spans can vary the final viscosity estimation significantly. Our estimated viscosity is also comparable with a Kelvin viscosity of 4×10^{17} Pa.s and Maxwell viscosity of 8×10^{17} Pa.s as estimated by *Panet et al.* (2010), who used 3 years of GRACE and GPS data over Thailand for the 2004 postseismic study, and also consistent with estimates of a Kelvin viscosity of 10^{18} Pa.s and Maxwell viscosity of 7.5×10^{18} Pa.s from a recent study by *Broerse et al.* (2015), who used 7 years of GPS data from Thailand for their 2004 postseismic investigation. *Broerse et al.* (2015) report that the same viscous structure constrained by the 7 years of GPS data fails to explain 7 years of regional gravity change from GRACE data. A combination of multiple data sets that have high resolution both in space and time, together with our joint inversion approach, should in future give us better estimates of the viscosity range.

Overall, the viscosity estimation in our study falls within the range of estimates from previous studies. The Burgers-like rheology of our estimated effective viscosity curves suggests that the

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

viscous flow in the mantle wedge has a transient behavior, and it is not a typical linear Newtonian flow.

5.6.4 Slip rates and viscous strain rates

We select four patches to explore the time evolution of slip rate on the megathrust; the locations of these patches are shown in Figure 5.8b. The shallow patches (1 and 2), located in the maximum afterslip region, slipped much faster than the deep patches (3 and 4). These four patches are separated tens to hundreds of kilometers apart, however the slip rates at these locations have similar logarithmic curvature (Figure 5.10a). This could suggest that the friction properties for both the shallow and downdip regions of the fault are similar. Such frictional similarities of the shallow and deep parts of the fault are consistent with the investigation from *Hsu et al.* (2006). After ~ 9 years following the earthquake, the slip rates for these patches have reduced to very low, but still declining, suggesting that afterslip on the fault is ongoing.

In order to understand whether the postseismic mantle flow has dissipated the coseismic stress changes over ~ 9 years, we plot the upper bound of typical geological strain rate ($\sim 10^{-14} \text{ s}^{-1}$, typical geological strain rate is from 10^{-16} to 10^{-14} s^{-1}) as the red dashed line in Figure 5.10b. We note that until 2014 the strain rate in the mantle wedge remains higher than the background flow rate, suggesting that the acceleration of the mantle flow after the earthquake has not yet returned to steady state. Complete relaxation may take decades.

Figure 5.10: Slip rates on selected fault patches, and viscous strain rates for blocks 1 and 2. Fault patch locations are shown in Figure 5.8b. Figure 5.2 shows the location of the blocks. Note that plots are in log scale thus some detailed variations are exaggerated.

5.7 Conclusions

We have jointly inverted for afterslip and viscous flow for the 2005 M_w Nias-Simeulue earthquake, using ~ 9 -year-long SuGAR time series following the event. We model viscous flow by inverting for viscous strain across finite rectangular blocks distributed in the mantle wedge. We find that for models that do and do not include viscous flow, the estimated afterslip differs significantly. With inclusion of viscous flow in the inversion, the level of deep afterslip we estimate is reduced, while shallow afterslip is increased. Following the earthquake, slip rates (for both the shallow and deep parts of the megathrust) have reduced to very low, but still declining, suggesting afterslip on the fault is ongoing. The viscous strain rate is still significantly above the typical geological strain rate ~ 9 years after the earthquake, indicating that viscous flow is still ongoing. Effective viscosities estimated for blocks down-dip of coseismic rupture patch indicate that the viscous flow in the mantle wedge has a transient behavior like a Burgers rheology, and it is not a typical linear Newtonian flow. Our estimated Burgers viscosity falls within the range of $\sim 10^{17}$ Pa.s to $\sim 10^{19}$ Pa.s. We also estimated a revised coseismic slip distribution, using the same fault geometry as for the afterslip distribution. The persistent low-slip patch of the Nias-Simeulue event is likely related to subducting structures on the megathrust, which introducing different rupture behaviors at this location.

CHAPTER 5. COSEISMIC AND POSTSEISMIC DEFORMATION FROM NIAS-SIMEULUE EARTHQUAKE

Concluding remarks

Megathrusts generate large earthquakes, posing significant seismic and tsunami hazard threats to many areas around the world. In order to better assess seismic hazards, we must understand megathrust behavior over many cycles. Behaviors of megathrusts at different time periods are complex, requiring multiple geodetic data sets and innovative techniques to separate the different processes involved. In this thesis, I have developed inversion techniques and dynamic models to understand behaviors of two megathrust segments: the Main Himalaya Thrust and the Sunda megathrust along the Indo-Australia and Sunda plate boundary, using GPS, InSAR, tide gauge and satellite altimetry data sets.

Chapter 3 tackles the following questions: How does fault morphology modify rupture dynamics on the Himalaya megathrust? Can it control the rupture extent? We conducted inversion models constrained by various tracks of InSAR data to examine the coseismic slip of the 2015 M_w 7.8 Gorkha earthquake. We also implemented dynamic earthquake cycle simulations to investigate the rupture dynamics on the thrust. We found that morphological bounds can modify the rupture dynamics; sometimes the fault ruptures as smaller events and sometimes the ruptures span the whole seismogenic zone of the megathrust generating giant earthquakes. The presence of these morphological bounds can control rupture propagation and confine the rupture into a small area. Our results highlight the importance of imaging the 3D structure of fault morphology for seismic forecasting.

Chapter 4 addresses these questions: Is viscous flow activated in the mantle wedge? What is the nature of the viscous flow and local rheology? To tackle these questions, we developed a new approach that can jointly invert for afterslip and viscous flow without prior assumptions of the rheology structure. Our new technique can even estimate which rheology the viscous flow may be. By examining ~ 9 -year long GPS, satellite altimetry and tide gauge data sets for Sumatra, Malaysia and Thailand, we show that our technique is capable of modeling afterslip and viscous flow simultaneously. We also found that the viscous flow in the mantle wedge has a transient behavior similar to a Burgers rheology with $\eta_K \sim 10^{17}$ Pa.s and $\eta_M \sim 10^{19}$ Pa.s.

Lastly, Chapter 5 addresses the following questions: What is the spatial and temporal pattern of afterslip following the 2005 M_w 8.6 Nias-Simeulue earthquake? Has the postseismic relaxation finished yet? What is the viscosity of the mantle wedge? By using decade-long SuGAR GPS

CHAPTER 6. CONCLUDING REMARKS

time series, we found that the afterslip rates have reduced to very low, but are still declining. The viscous strain rates remain significantly above the typical geological strain rate in the mantle wedge, suggesting that the viscous relaxation is ongoing and may take years, decades or even longer, and long-term geodetic monitoring could help us to answer this. Our results highlight the capability of the joint inversion approach, and in future can help us to learn more about fault mechanics and the rheology of the asthenosphere along other subduction zones as well.

These studies together demonstrate the complex behaviors of megathrusts over different time stages within their seismic cycles. We suggest that a key to future understanding of megathrust behavior will be collaboration between geodesists, structural geologists, and geophysicists, to better understand linkages between rupture segmentation, rupture behavior, and the structural and morphological characteristics of the faults.

References

- Ader, T., et al., Convergence rate across the Nepal Himalaya and interseismic coupling on the Main Himalayan Thrust: Implications for seismic hazard, *Journal of Geophysical Research: Solid Earth*, 117(4), 1–16, doi:10.1029/2011JB009071, 2012.
- Avouac, J.-P., L. Meng, S. Wei, T. Wang, and J.-P. Ampuero, Lower edge of locked Main Himalayan Thrust unzipped by the 2015 Gorkha earthquake, *Nature Geosci*, 8(9), 708–711, 2015.
- Barbot, S., and Y. Fialko, A unified continuum representation of post-seismic relaxation mechanisms: semi-analytic models of afterslip, poroelastic rebound and viscoelastic flow, *Geophysical Journal International*, 182(3), 1124–1140, doi:10.1111/j.1365-246X.2010.04678.x, 2010.
- Barbot, S., Y. Fialko, and D. Sandwell, Three-dimensional models of elastostatic deformation in heterogeneous media, with applications to the Eastern California Shear Zone, *Geophysical Journal International*, 179(1), 500–520, doi:10.1111/j.1365-246X.2009.04194.x, 2009.
- Bilham, R., Seismology: Raising Kathmandu, *Nature Geoscience*, 8(8), 582–584, doi:10.1038/ngeo2498, 2015.
- Bollinger, L., P. Tapponnier, S. N. Sapkota, and Y. Klinger, Slip deficit in central Nepal: omen for a repeat of the 1344 AD earthquake?, *Earth, Planets and Space*, 68(1), 1–12, doi:10.1186/s40623-016-0389-1, 2016.
- Bouchon, M., H. Karabulut, M.-P. Bouin, J. Schmittbuhl, M. Vallée, R. Archuleta, S. Das, F. Renard, and D. Marsan, Faulting characteristics of supershear earthquakes, *Tectonophysics*, 493(3-4), 244–253, doi:10.1016/j.tecto.2010.06.011, 2010.
- Briggs, R. W., et al., Deformation and slip along the Sunda megathrust in the Great 2005 Nias-Simeulue earthquake., *Science (New York, N.Y.)*, 311(5769), 1897–901, doi:10.1126/science.1122602, 2006.
- Broerse, T., R. Riva, W. Simons, R. Govers, and B. Vermeersen, Postseismic GRACE and GPS observations indicate a rheology contrast above and below the Sumatra slab, *Journal of Geophysical Research: Solid Earth*, 120(7), 5343–5361, doi:10.1002/2015JB011951, 2015.
- Burgette, R. J., R. J. Weldon, and D. A. Schmidt, Interseismic uplift rates for western Oregon and along-strike variation in locking on the Cascadia subduction zone, *Journal of Geophysical Research: Solid Earth*, 114(B1), n/a—n/a, doi:10.1029/2008JB005679, 2009.
- Bürgmann, R., and G. Dresen, Rheology of the Lower Crust and Upper Mantle: Evidence from Rock Mechanics, Geodesy, and Field Observations, *Annual Review of Earth and Planetary Sciences*, 36(1), 531–567, doi:10.1146/annurev.earth.36.031207.124326, 2008.
-

REFERENCES

- Cabanes, C., A. Cazenave, and C. Le Provost, Sea Level Rise During Past 40 Years Determined from Satellite and in Situ Observations, *Science*, 294(5543), 840–842, doi:10.1126/science.1063556, 2001.
- Carter, N. L., and S. H. Kirby, *Transient Creep and Semibrittle Behavior of Crystalline Rocks*, 807–839 pp., Birkhäuser Basel, Basel, doi:10.1007/978-3-0348-7182-2_16, 1978.
- Cazenave, A., K. Dominh, S. Guinehut, E. Berthier, W. Llovel, G. Ramillien, M. Ablain, and G. Larnicol, Sea level budget over 2003–2008: A reevaluation from {GRACE} space gravimetry, satellite altimetry and Argo, *Global and Planetary Change*, 65(12), 83–88, doi:http://dx.doi.org/10.1016/j.gloplacha.2008.10.004, 2009.
- Chlieh, M., J. P. Avouac, K. Sieh, D. H. Natawidjaja, and J. Galetzka, Heterogeneous coupling of the Sumatran megathrust constrained by geodetic and paleogeodetic measurements, *Journal of Geophysical Research*, 113(B5), B05,305, doi:10.1029/2007JB004981, 2008.
- Chlieh, M., et al., Coseismic Slip and Afterslip of the Great Mw 9.15 Sumatra-Andaman Earthquake of 2004, *Bulletin of the Seismological Society of America*, 97(1A), S152–S173, doi:10.1785/0120050631, 2007.
- Chopra, P. N., The Adolphe Nicolas Volume High-temperature transient creep in olivine rocks, *Tectonophysics*, 279(1), 93–111, doi:http://dx.doi.org/10.1016/S0040-1951(97)00134-0, 1997.
- Church, J. A., and N. J. White, Sea-Level Rise from the Late 19th to the Early 21st Century, *Surveys in Geophysics*, 32(4), 585–602, doi:10.1007/s10712-011-9119-1, 2011.
- Cnes, SSALTO/DUACS User Handbook, doi:10.1126/science.1163589, 2014.
- Copley, A., Postseismic afterslip 30?years after the 1978 Tabas-e-Golshan (Iran) earthquake: observations and implications for the geological evolution of thrust belts, *Geophysical Journal International*, doi:10.1093/gji/ggu023, 2014.
- Deng, J., M. Gurnis, H. Kanamori, and E. Hauksson, Viscoelastic Flow in the Lower Crust after the 1992 Landers, California, Earthquake, *Science*, 282(5394), 1689–1692, doi:10.1126/science.282.5394.1689, 1998.
- Duputel, Z., J. Vergne, L. Rivera, G. Wittlinger, V. Farra, and G. Hetényi, The 2015 Gorkha earthquake: A large event illuminating the Main Himalayan Thrust fault, *Geophysical Research Letters*, 43(6), 2517–2525, doi:10.1002/2016GL068083, 2016.
- Feng, L., E. M. Hill, P. Banerjee, I. Hermawan, L. L. H. Tsang, D. H. Natawidjaja, B. W. Suwargadi, and K. Sieh, A unified GPS-based earthquake catalog for the Sumatran plate boundary between 2002 and 2013, *Journal of Geophysical Research: Solid Earth*, 120(5), 3566–3598, doi:10.1002/2014JB011661, 2015.
- Feng, L., S. Barbot, E. M. Hill, I. Hermawan, P. Banerjee, and D. Natawidjaja, Footprint of past earthquakes revealed in the afterslip of the 2010 {Mw} 7.8 Mentawai tsunami earthquake, *To be submitted to Journal of Geophysical Research Letters*, 2016a.
- Feng, W., et al., Source characteristics of the 2015 {Mw} 7.8 Gorkha (Nepal) earthquake and its {Mw} 7.2 aftershock from space geodesy, *Tectonophysics*, pp. –, doi:http://dx.doi.org/10.1016/j.tecto.2016.02.029, 2016b.

REFERENCES

- Fialko, Y., Evidence of fluid-filled upper crust from observations of postseismic deformation due to the 1992 Mw7.3 Landers earthquake, *Journal of Geophysical Research: Solid Earth*, 109(B8), n/a—n/a, doi:10.1029/2004JB002985, 2004.
- Freed, A. M., and R. Bürgmann, Evidence of power-law flow in the Mojave desert mantle, *Nature*, 430(6999), 548–551, 2004.
- Freed, A. M., R. Bürgmann, E. Calais, and J. Freymueller, Stress-dependent power-law flow in the upper mantle following the 2002 Denali, Alaska, earthquake, *Earth and Planetary Science Letters*, 252(34), 481 – 489, doi:http://dx.doi.org/10.1016/j.epsl.2006.10.011, 2006.
- Fukuda, J., S. Miyazaki, T. Higuchi, and T. Kato, Geodetic inversion for space-time distribution of fault slip with time-varying smoothing regularization, *Geophysical Journal International*, 173(1), 25–48, doi:10.1111/j.1365-246X.2007.03722.x, 2008.
- Galetzka, J., et al., Slip pulse and resonance of the Kathmandu basin during the 2015 Gorkha earthquake, Nepal, *Science*, 349(6252), 1091–1095, doi:10.1126/science.aac6383, 2015.
- García, D., I. Vigo, F. B. Chao, and C. M. Martínez, Vertical crustal motion along the Mediterranean and Black Sea coast derived from ocean altimetry and tide gauge data, *Pure and Applied Geophysics*, 164(4), 851–863, doi:10.1007/s00024-007-0193-8, 2007.
- Grandin, R., M. Vallée, C. Satriano, R. Lacassin, Y. Klinger, M. Simoes, and L. Bollinger, Rupture process of the Mw 7.9 2015 Gorkha earthquake (Nepal): Insights into Himalayan megathrust segmentation, *Geophysical Research Letters*, 42(20), n/a—n/a, doi:10.1002/2015GL066044, 2015.
- Gunawan, E., et al., A comprehensive model of postseismic deformation of the 2004 Sumatra-Andaman earthquake deduced from {GPS} observations in northern Sumatra, *Journal of Asian Earth Sciences*, 88, 218 – 229, doi:http://dx.doi.org/10.1016/j.jseaes.2014.03.016, 2014.
- Han, S.-C., J. Sauber, S. B. Luthcke, C. Ji, and F. F. Pollitz, Implications of postseismic gravity change following the great 2004 Sumatra-Andaman earthquake from the regional harmonic analysis of GRACE intersatellite tracking data, *Journal of Geophysical Research: Solid Earth*, 113(B11), n/a—n/a, doi:10.1029/2008JB005705, 2008.
- Hayes, G. P., D. J. Wald, and R. L. Johnson, Slab1.0: A three-dimensional model of global subduction zone geometries, *Journal of Geophysical Research: Solid Earth*, 117(B1), n/a—n/a, doi:10.1029/2011JB008524, 2012.
- Hill, E. M., et al., The 2010 Mw 7.8 Mentawai earthquake: Very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS data, *Journal of Geophysical Research: Solid Earth*, 117(B6), n/a—n/a, doi:10.1029/2012JB009159, 2012.
- Hoechner, A., S. V. Sobolev, I. Einarsson, and R. Wang, Investigation on afterslip and steady state and transient rheology based on postseismic deformation and geoid change caused by the Sumatra 2004 earthquake, *Geochemistry, Geophysics, Geosystems*, 12(7), n/a—n/a, doi:10.1029/2010GC003450, 2011.
- Holgate, S. J., et al., New Data Systems and Products at the Permanent Service for Mean Sea Level, *Journal of Coastal Research*, pp. 493–504, doi:10.2112/JCOASTRES-D-12-00175.1, 2013.

REFERENCES

- Hsu, Y.-J., M. Simons, J.-P. Avouac, J. Galetzka, K. Sieh, M. Chlieh, D. Natawidjaja, L. Prawirodirdjo, and Y. Bock, Frictional Afterslip Following the 2005 Nias-Simeulue Earthquake, Sumatra, *Science*, 312(5782), 1921–1926, doi:10.1126/science.1126960, 2006.
- Hu, Y., and K. Wang, Spherical-Earth finite element model of short-term postseismic deformation following the 2004 Sumatra earthquake, *Journal of Geophysical Research: Solid Earth*, 117(B5), n/a—n/a, doi:10.1029/2012JB009153, 2012.
- Hubbard, J., R. Almeida, A. Foster, S. Sapkota, P. Burgi, and P. Tapponnier, Structural segmentation controlled the 2015 Mw 7.8 Gorkha earthquake rupture in Nepal, *Geology*, doi:10.1130/G38077.1, 2016.
- Jacob, J., J. Dymant, and V. Yatheesh, Revisiting the structure, age, and evolution of the Wharton Basin to better understand subduction under Indonesia, *Journal of Geophysical Research: Solid Earth*, 119(1), 169–190, doi:10.1002/2013JB010285, 2014.
- Kame, N., J. R. Rice, and R. Dmowska, Effects of prestress state and rupture velocity on dynamic fault branching, *Journal of Geophysical Research: Solid Earth*, 108(B5), 2003.
- Kaneko, Y., J.-P. Avouac, and N. Lapusta, Towards inferring earthquake patterns from geodetic observations of interseismic coupling, *Nature Geoscience*, 3(5), 363–369, doi:10.1038/ngeo843, 2010.
- Karato, S.-i., Rheology of the Earth's mantle: A historical review, *Gondwana Research*, 18(1), 17–45, doi:http://dx.doi.org/10.1016/j.gr.2010.03.004, 2010.
- Karato, S.-i., and P. Wu, Rheology of the upper mantle: A synthesis, *Science*, 260(5109), 771–778, 1993.
- Kato, N., Repeating slip events at a circular asperity: Numerical simulation with a rate- and state- dependent friction law, *Bull. Earthq. Res. Inst.*, 78, 151–166, 2003.
- King, G. C. P., Speculations on the geometry of the initiation and termination processes of earthquake rupture and its relation to morphology and geological structure, *Pure and Applied Geophysics PAGEOPH*, 124(3), 567–585, doi:10.1007/BF00877216, 1986.
- Klinger, Y., Relation between continental strike-slip earthquake segmentation and thickness of the crust, *Journal of Geophysical Research: Solid Earth*, 115(B7), n/a—n/a, doi:10.1029/2009JB006550, 2010.
- Konca, A. O., V. Hjorleifsdottir, T.-R. A. Song, J.-P. Avouac, D. V. Helmberger, C. Ji, K. Sieh, R. Briggs, and A. Meltzner, Rupture Kinematics of the 2005 Mw 8.6 Nias-Simeulue Earthquake from the Joint Inversion of Seismic and Geodetic Data, *Bulletin of the Seismological Society of America*, 97(1A), S307–S322, doi:10.1785/0120050632, 2007.
- Konca, A. O., et al., Partial rupture of a locked patch of the Sumatra megathrust during the 2007 earthquake sequence., *Nature*, 456(7222), 631–5, doi:10.1038/nature07572, 2008.
- Lapusta, N., J. R. Rice, Y. Ben-Zion, and G. Zheng, Elastodynamic analysis for slow tectonic loading with spontaneous rupture episodes on faults with rate-and state-dependent friction, 2000.
- Laske G., M. G. M. Z., and M. Pasyanos, Update on CRUST1.0 - A 1-degree Global Model of Earth's Crust, *Journal of Geophysical Research: Solid Earth*, pp. Abstract EGU2013–2658, 2013.

REFERENCES

- Lavé, J., D. Yule, S. Sapkota, K. Basant, C. Madden, M. Attal, and R. Pandey, Evidence for a great medieval earthquake (1100 a.d.) in the central himalayas, nepal, *Science*, 307(5713), 1302–1305, doi:10.1126/science.1104804, 2005.
- Lay, T., H. Kanamori, C. J. Ammon, K. D. Koper, A. R. Hutko, L. Ye, H. Yue, and T. M. Rushing, Depth-varying rupture properties of subduction zone megathrust faults, *Journal of Geophysical Research*, 117(B4), B04,311, doi:10.1029/2011JB009133, 2012.
- Lay, T., H. Yue, E. E. Brodsky, and C. An, The 1 April 2014 Iquique, Chile, Mw 8.1 earthquake rupture sequence, *Geophysical Research Letters*, 41(11), 3818–3825, doi:10.1002/2014GL060238, 2014.
- Lindsey, E. O., R. Natsuaki, X. Xu, M. Shimada, M. Hashimoto, D. Melgar, and D. T. Sandwell, Line-of-sight displacement from ALOS-2 interferometry: Mw 7.8 Gorkha Earthquake and Mw 7.3 aftershock, *Geophysical Research Letters*, 42(16), 6655–6661, doi:10.1002/2015GL065385, 2015.
- Lubis, A. M., A. Hashima, and T. Sato, Analysis of afterslip distribution following the 2007 September 12 southern Sumatra earthquake using poroelastic and viscoelastic media, *Geophysical Journal International*, doi:10.1093/gji/ggs020, 2012.
- Luu, Q. H., P. Tkalich, and T. W. Tay, Sea level trend and variability around Peninsular Malaysia, *Ocean Science*, 11(4), 617–628, doi:10.5194/os-11-617-2015, 2015.
- Masterlark, T., and H. F. Wang, Transient Stress-Coupling Between the 1992 Landers and 1999 Hector Mine, California, Earthquakes, *Bulletin of the Seismological Society of America*, 92(4), 1470–1486, doi:10.1785/0120000905, 2002.
- McGuire, J. J., and P. Segall, Imaging of aseismic fault slip transients recorded by dense geodetic networks, *Geophysical Journal International*, 155(3), 778–788, doi:10.1111/j.1365-246X.2003.02022.x, 2003.
- Meltzner, A. J., K. Sieh, H.-W. Chiang, C.-C. Shen, B. W. Suwargadi, D. H. Natawidjaja, B. Philibosian, and R. W. Briggs, Persistent termini of 2004- and 2005-like ruptures of the Sunda megathrust, *Journal of Geophysical Research*, 117(B4), B04,405, doi:10.1029/2011JB008888, 2012.
- Meltzner, A. J., et al., Time-varying interseismic strain rates and similar seismic ruptures on the NiasSimeulue patch of the Sunda megathrust, *Quaternary Science Reviews*, 122, 258–281, doi:http://dx.doi.org/10.1016/j.quascirev.2015.06.003, 2015.
- Merrifield, M. A., S. T. Merrifield, and G. T. Mitchum, An Anomalous Recent Acceleration of Global Sea Level Rise, *Journal of Climate*, 22(21), 5772–5781, doi:10.1175/2009JCLI2985.1, 2009.
- Middlemiss, C. S., *The Kangra earthquake of 4th April 1905*, Memoirs of the Geological Survey of India, Geological survey of India, 1910.
- Molnar, P., The Distribution of Intensity Associated with the 1905 Kangra Earthquake and Bounds on the Extent of the Rupture Zone, *Geological Society of India*, 29(2), 1987.
- Morgan, P., L. Feng, A. J. Meltzner, E. M. Hill, L. L. H. Tsang, and K. Sieh, Piecemeal and time variant slip behavior at a structural discontinuity of the sunda megathrust, *Manuscript in preparation for Journal of Geophysical Research*, 2016.

REFERENCES

- Natawidjaja, D. H., K. Sieh, S. N. Ward, H. Cheng, R. L. Edwards, J. Galetzka, and B. W. Suwargadi, Paleogeodetic records of seismic and aseismic subduction from central Sumatran microatolls, Indonesia, *Journal of Geophysical Research: Solid Earth*, 109(B4), n/a—n/a, doi:10.1029/2003JB002398, 2004.
- Nerem, R. S., and G. T. Mitchum, Estimates of vertical crustal motion derived from differences of TOPEX/POSEIDON and tide gauge sea level measurements, *Geophysical Research Letters*, 29(19), 40–44, doi:10.1029/2002GL015037, 2002.
- Nicholls, R. J., and A. Cazenave, Sea-Level Rise and Its Impact on Coastal Zones, *Science*, 328(5985), 1517–1520, doi:10.1126/science.1185782, 2010.
- Nikkhoo, M., and T. R. Walter, Triangular dislocation: an analytical, artefact-free solution, *Geophys. J. Int.*, 201(2), 1119–1141, doi:10.1093/gji/ggv035, 2015.
- Okada, Y., Internal deformation due to shear and tensile faults in a half-space, *Bulletin of the Seismological Society of America*, 82(2), 1018–1040, 1992.
- Oleskevich, D. A., R. D. Hyndman, and K. Wang, The updip and downdip limits to great subduction earthquakes: Thermal and structural models of Cascadia, south Alaska, SW Japan, and Chile, *Journal of Geophysical Research*, 104(B7), 14,965, doi:10.1029/1999JB900060, 1999.
- Ostanciaux, É., L. Husson, G. Choblet, C. Robin, and K. Pedoja, Present-day trends of vertical ground motion along the coast lines, *Earth-Science Reviews*, 110(14), 74–92, doi:http://dx.doi.org/10.1016/j.earscirev.2011.10.004, 2012.
- Panet, I., F. Pollitz, V. Mikhailov, M. Diament, P. Banerjee, and K. Grijalva, Upper mantle rheology from GRACE and GPS postseismic deformation after the 2004 Sumatra-Andaman earthquake, *Geochemistry, Geophysics, Geosystems*, 11(6), n/a—n/a, doi:10.1029/2009GC002905, 2010.
- Parsons, B., et al., The 1994 Sefidabeh (eastern Iran) earthquakes revisited: new evidence from satellite radar interferometry and carbonate dating about the growth of an active fold above a blind thrust fault, *Geophysical Journal International*, 164(1), 202–217, doi:10.1111/j.1365-246X.2005.02655.x, 2006.
- Paul, J., A. R. Lowry, R. Bilham, S. Sen, and R. Smalley, Postseismic deformation of the Andaman Islands following the 26 December, 2004 Great Sumatra-Andaman earthquake, *Geophysical Research Letters*, 34(19), n/a—n/a, doi:10.1029/2007GL031024, 2007.
- Paul, J., C. P. Rajendran, A. R. Lowry, V. Andrade, and K. Rajendran, Andaman Postseismic Deformation Observations: Still Slipping after All These Years?, *Bulletin of the Seismological Society of America*, 102(1), 343–351, doi:10.1785/0120110074, 2012.
- Perfettini, H., and J.-P. Avouac, Modeling afterslip and aftershocks following the 1992 Landers earthquake, *Journal of Geophysical Research: Solid Earth*, 112(B7), n/a—n/a, doi:10.1029/2006JB004399, 2007.
- Pollitz, F., P. Banerjee, K. Grijalva, B. Nagarajan, and R. Brügmann, Effect of 3-D viscoelastic structure on post-seismic relaxation from the 2004 M= 9.2 Sumatra earthquake, *Geophysical Journal International*, 173(1), 189–204, doi:10.1111/j.1365-246X.2007.03666.x, 2008.
- Pollitz, F. F., Transient rheology of the uppermost mantle beneath the Mojave Desert, California, *Earth and Planetary Science Letters*, 215(12), 89 – 104, doi:http://dx.doi.org/10.1016/S0012-821X(03)00432-1, 2003.

REFERENCES

- Pollitz, F. F., G. Peltzer, and R. Bürgmann, Mobility of continental mantle: Evidence from postseismic geodetic observations following the 1992 Landers earthquake, *Journal of Geophysical Research: Solid Earth*, 105(B4), 8035–8054, doi:10.1029/1999JB900380, 2000.
- Pollitz, F. F., R. Brgmann, and P. Banerjee, Post-seismic relaxation following the great 2004 Sumatra-Andaman earthquake on a compressible self-gravitating Earth, *Geophysical Journal International*, 167(1), 397–420, doi:10.1111/j.1365-246X.2006.03018.x, 2006.
- Qiu, Q., E. M. Hill, S. Barbot, and L. Feng, Transient rheology in the sumatra mantle wedge revealed by a decade of great earthquakes, *Manuscript in preparation for Science*, 2016.
- Rice, J. R., The mechanics of earthquake rupture, in *Physics of the Earth's Interior*, edited by A.M. Dziewonski and E. Boschi, pp. 555–649, 1980.
- Rice, J. R., N. Lapusta, and K. Ranjith, Rate and state dependent friction and the stability of sliding between elastically deformable solids, *Journal of the Mechanics and Physics of Solids*, 49(9), 1865–1898, doi:10.1016/S0022-5096(01)00042-4, 2001.
- Ruiz, S., M. Metois, A. Fuenzalida, J. Ruiz, F. Leyton, R. Grandin, C. Vigny, R. Madariaga, and J. Campos, Intense foreshocks and a slow slip event preceded the 2014 Iquique Mw 8.1 earthquake, *Science*, 345(6201), 1165–1169, doi:10.1126/science.1256074, 2014.
- Sandwell, D. T., R. Mellors, T. Xiaopeng, W. Matt, and P. Wessel, Open Radar Interferometry Software for Mapping Surface Deformation, *Eos, Transactions American Geophysical Union*, 92(2), 2011, 2011.
- Santamaría-Gómez, A., M. Gravelle, and G. Wöppelmann, Long-term vertical land motion from double-differenced tide gauge and satellite altimetry data, *Journal of Geodesy*, 88(3), 207–222, doi:10.1007/s00190-013-0677-5, 2014.
- Sapkota, S. N., L. Bollinger, Y. Klinger, P. Tapponnier, Y. Gaudemer, and D. Tiwari, Primary surface ruptures of the great Himalayan earthquakes in 1934 and 1255, *Nature Publishing Group*, 6(2), 152, doi:10.1038/ngeo1669, 2013.
- Savage, J. C., and J. L. Svarc, Postseismic deformation associated with the 1992 Mw=7.3 Landers earthquake, southern California, *Journal of Geophysical Research: Solid Earth*, 102(B4), 7565–7577, doi:10.1029/97JB00210, 1997.
- Segall, P., and M. Matthews, Time dependent inversion of geodetic data, *Journal of Geophysical Research: Solid Earth*, 102(B10), 22,391–22,409, doi:10.1029/97JB01795, 1997.
- Simons, M., Y. Fialko, and L. Rivera, Coseismic deformation from the 1999 Mw 7.1 Hector Mine, California, earthquake as inferred from InSAR and GPS observations, *Bull. Seism. Soc. Am.*, 92(4), 1390–1402, doi:10.1785/0120000933, 2002.
- Stevens, V. L., and J. Avouac, Interseismic Coupling on the Main Himalayan Thrust, *Geophysical Research Letters*, pp. n/a–n/a, doi:10.1002/2015GL064845, 2015.
- Suito, H., and J. T. Freymueller, A viscoelastic and afterslip postseismic deformation model for the 1964 Alaska earthquake, *Journal of Geophysical Research: Solid Earth*, 114(B11), n/a—n/a, doi:10.1029/2008JB005954, 2009.
- Sun, T., and K. Wang, Viscoelastic relaxation following subduction earthquakes and its effects on afterslip determination, *Journal of Geophysical Research: Solid Earth*, 120(2), 1329–1344, doi:10.1002/2014JB011707, 2014JB011707, 2015.

REFERENCES

- Sun, T., et al., Prevalence of viscoelastic relaxation after the 2011 Tohoku-oki earthquake, *Nature*, 514(7520), 84–87, 2014.
- Tilmann, F. J., T. J. Craig, I. Grevenmeyer, B. Suwargadi, H. Kopp, and E. Flueh, The updip seismic/aseismic transition of the Sumatra megathrust illuminated by aftershocks of the 2004 Aceh-Andaman and 2005 Nias events, *Geophysical Journal International*, 181(3), 1261–1274, doi:10.1111/j.1365-246X.2010.04597.x, 2010.
- Trisirisatayawong, I., M. Naeije, W. Simons, and L. Fenoglio-Marc, Sea level change in the gulf of thailand from gps-corrected tide gauge data and multi-satellite altimetry, *Global and Planetary Change*, 76(34), 137 – 151, doi:http://dx.doi.org/10.1016/j.gloplacha.2010.12.010, 2011.
- Tsang, L. L. H., A. J. Meltzner, E. M. Hill, J. T. Freymueller, and K. Sieh, A paleogeodetic record of variable interseismic rates and megathrust coupling at simeulue island, sumatra, *Geophysical Research Letters*, 42(24), 10,585–10,594, doi:10.1002/2015GL066366, 2015GL066366, 2015.
- Tsang, L. L. H., E. M. Hill, S. Barbot, Q. Qiu, L. Feng, I. Hermanwan, P. Banergee, and H. D. Natawidjaja, Afterslip following the m_w 8.4 bengkulu earthquake in sumatra borders and overlaps the 2010 m_w 7.8 mentawai earthquake rupture zone, *Manuscript in preparation for Journal of Geophysical Research: Solid Earth*, 2016.
- Wang, R., F. L. Martn, and F. Roth, Computation of deformation induced by earthquakes in a multi-layered elastic crust: FORTRAN programs EDGRN/EDCMP, *Computers & Geosciences*, 29(2), 195–207, doi:http://dx.doi.org/10.1016/S0098-3004(02)00111-5, 2003.
- Watanabe, S.-i., M. Sato, M. Fujita, T. Ishikawa, Y. Yokota, N. Ujihara, and A. Asada, Evidence of viscoelastic deformation following the 2011 tohoku-oki earthquake revealed from seafloor geodetic observation, *Geophysical Research Letters*, 41(16), 5789–5796, doi:10.1002/2014GL061134, 2014GL061134, 2014.
- Wesnowsky, S. G., Seismological and structural evolution of strike-slip faults, *Nature*, 335(6188), 340–343, 1988.
- Wesnowsky, S. G., Predicting the endpoints of earthquake ruptures., *Nature*, 444(7117), 358–360, doi:10.1038/nature05275, 2006.
- Wright, T. J., Z. Lu, and C. Wicks, Source model for the Mw 6.7, 23 October 2002, Nenana Mountain Earthquake (Alaska) from InSAR, *Geophysical Research Letters*, 30(18), n/a—n/a, doi:10.1029/2003GL018014, 2003.
- Yabuki, T., and M. Matsuura, Geodetic Data Inversion Using a Bayesian Information Criterion for Spatial-Distribution of Fault Slip, *Geophysical Journal International*, 109(2), 363–375, doi:10.1111/j.1365-246X.1992.tb00102.x, 1992.

Appendix A: Supplementary material and figures for Nepal coseismic study

A.1 Text S1

Dynamic earthquake cycle model

We simulate the slip evolution on an isolated segment of the Main Himalayan Thrust (MHT) using a rate-and-state friction framework. We employ the boundary integral method with radiation damping (*Rice et al.*, 2001; *Kato*, 2003). As with the inversion, we discretize the geometry of the MHT using triangular elements. We obtain the stress kernels using the analytic formulation of *Nikkhoo and Walter*, 2015. We use uniform friction properties with the static friction coefficient $\mu_0 = 0.6$, the reference velocity $V_0 = 10^{-6}$ m/s, $a = d\mu/d(\ln V) = 10^{-2}$, the effective confining pressure $\sigma = 300$ MPa, and the characteristic weakening distance $L = 7.5$ cm (μ is the frictional resistance and V is the slip velocity).

We define the region that could potentially rupture as an area larger than the rupture patch of the 2015 M_w 7.8 Gorkha earthquake, extending from the surface to 20 km depth and about the length of the Kathmandu Klippe (Figure 3.2). This region is modeled with velocity-weakening properties, with $(a - b) = d\mu_{ss}/d(\ln V) = -4 \times 10^{-3}$, where μ_{ss} is the frictional resistance at steady state. Outside this region, the fault is modeled with the velocity-strengthening properties of $(a - b) = 4 \times 10^{-3}$. We load the fault at a uniform and constant rate of $V_{pl} = 0.02$ m/yr. In the simulation, we obtain a series of earthquakes of varying magnitudes and with ruptures covering sometimes a fraction of the velocity-weakening patch and some other times the entirety of the patch. The complexity of the slip history emerges spontaneously from the dynamics of the system. In the context of our simulation, the complexity of the slip history comes from the morphology of the megathrust, not from heterogeneous initial conditions or spatial distribution of friction properties. (If the fault were perfectly planar, such a distribution of frictional properties would lead to a single type of earthquakes with a characteristic size and a single recurrence time and both recurrence times and rupture slip would be predictable.) The simulated fault history contains events with characteristics that resemble those of the 2015 M_w 7.8 Gorkha

APPENDIX A. SUPPLEMENTARY MATERIAL FOR NEPAL COSEISMIC STUDY

earthquake. The Gorkha-like ruptures extend through the deep décollement but are confined by the two ramps. Other single-segment ruptures occupy the upper décollement or the topmost ramp, resulting in earthquakes of $M_w < 8$. Occasionally, ruptures spread from the surface to the bottom of the seismogenic zone in a through-going, multi-segment event, resulting in a great earthquake with $M_w \geq 8$. The simulation is only intended to explain the role of geometry on the complexity of earthquake dynamics in the context of a long sequence of earthquakes. In reality, strong heterogeneities in composition, rheology and other physical properties make the earthquake history even more complex.

A.2 Supplementary figures and table

Table A.1: Details of the InSAR data used in this study.

Inf	Mission	Mode	Flight Direction	Track	Master	Slave
1	ALOS-2	ScanSAR	DSC	T047	20150331	20150428
2	ALOS-2	ScanSAR	DSC	T048	20150222	20150503
3	ALOS-2	StripMap	ASC	T157	20150221	20150502
4	S1A	TOPS	DSC	-	20150417	20150429
5	RS2	StripMap	DSC	-	20150405	20150429
6	RS2	StripMap	DSC	-	20150412	20150506

Note: a) S1A, RS2 represent Sentinel-1A, and RADARSAT-2, respectively.

APPENDIX A. SUPPLEMENTARY MATERIAL FOR NEPAL COSEISMIC STUDY

Figure A.1: Footprint of InSAR data considered in this study, including from the Sentinel-1A, RADARSAT-2, and ALOS-2 satellites. The beach balls represent the moment tensors of the mainshock and largest aftershock. The coloured circles correspond to aftershocks (same as in Figure 3.1).

Figure A.2: InSAR data (left), model (middle) and residuals (right).

APPENDIX A. SUPPLEMENTARY MATERIAL FOR NEPAL COSEISMIC STUDY

Figure A.3: InSAR data (left), model (middle) and residuals (right)

Figure A.4: InSAR data (left), model (middle) and residuals (right).

APPENDIX A. SUPPLEMENTARY MATERIAL FOR NEPAL COSEISMIC STUDY

Figure A.5: The Laplacian weight for spatial smoothing in the inversion for slip was selected using Akaike's Bayesian Information Criterion. The red star indicates the Laplacian weight used for our preferred model.

Figure A.6: Coulomb stress changes on the megathrust from our preferred slip model, assuming a coefficient of friction of $\mu_0 = 0.6$. The dashed lines show the location of the shallow (8-10 km) and deep ramp (16 km), respectively. The coloured dots correspond to aftershocks (same as Figure 3.1). The beach balls represent the moment tensor of the mainshock and largest aftershock.

APPENDIX A. SUPPLEMENTARY MATERIAL FOR NEPAL COSEISMIC STUDY

Figure A.7: Examples of estimated slip distributions obtained using different relative weights for data input to the slip inversion. The GPS data prefer a little more shallow slip, but in general the distributions remain similar regardless of data weight.

Appendix B: Supplementary material for Sumatra postseismic study

B.1 Geodetic data sets

B.1.1 GPS

We used GIPSY-OASIS version 6.2 to process the raw GPS time series spanning from 2005 to 2014. Tidal effects and tropospheric delays were estimated and removed during processing. To separate the various tectonic signals we first estimated the coseismic offsets for large earthquakes, to obtain the residual time series. We then further separated the interseismic and postseismic signals by modeling a secular rate and exponential or logarithmic function simultaneously and systematically in a self-consistent manner. Our model also included annual and semiannual seasonal cycles. Further details about the technique and GPS processing procedure are outlined in *Feng et al. (2015)*.

B.2 Land-height displacement derived from tide-gauge and satellite-altimetry data

Tide gauges measure the relative sea level between the sea surface and the land. Thus, changes of the land elevation will be captured by tide gauges. To extract the land-height changes, information on sea-surface changes are required; satellite altimetry measurements can provide this information. A combination of tide-gauge and satellite-altimetry data has been widely used to separate the land-height changes (*García et al., 2007; Nerem and Mitchum, 2002; Oostanciaux et al., 2012; Santamaría-Gómez et al., 2014*). Land-height changes following the 2004 Aceh-Andaman earthquake were extracted from tide gauges by *Trisirisatayawong et al. (2011)*.

We carefully selected tide gauges data that have as few gaps as possible. Multiple data gaps or shorter tide recordings might bias the long-term trend estimation. Our selected tide gauges

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

are located in Singapore, Malaysia and Thailand, covering a wide region scale. We downloaded the tide-gauge data from the Permanent Service for Mean Sea Level database (*Holgate et al.*, 2013;PSMSL, 2016) from website <http://www.psmsl.org/data/>.

For the satellite altimetry data, we used the AVISO data product from <ftp.aviso.altimetry.fr>. The altimeter products were produced by Ssalto/DUACS and distributed by Aviso, with support from CNES (<http://www.avso.altimetry.fr/duacs/>). These data were processed by a combination of all altimeter missions, and corrected for instrument, orbit errors and errors coming from the signal propagating through atmosphere, and perturbation errors from surface reflection. Tides, such as pole tides, solid, ocean, and loading tides were corrected as well. Cross-calibration was also applied to yield a consistent and homogenous high-quality data set (see *Cnes*, 2014). This data set has been successfully and widely implemented on sea level studies (*Luu et al.*, 2015; *García et al.*, 2007; *Cabanes et al.*, 2001; *Cazenave et al.*, 2009; *Church and White*, 2011; *Merrifield et al.*, 2009; *Nicholls and Cazenave*, 2010). We used the delay-time mode of the global mean sea level anomalies product of the Aviso data set because of its accuracy as compared with real-time mode. These global, gridded, multi-mission data span from 1993 until May of 2014 with a daily solution. In order to extract the land-elevation information by subtracting altimetry data with the monthly tide gauge data, we used the daily solution to estimate monthly average sea surface height anomalies.

A previous study for this region from *Trisirisatayawong et al.* (2011) reports that if the distance between the tide gauge and the satellite grid is less than 100 km, a good estimate of land-height changes could be achieved. Under this rule, we selected 24 tide gauges with nearby Aviso grid points for extracting the land-height changes following the 2004 M_w 9.2 Sumatra-Andaman earthquake until 2014.

Before subtracting the Aviso data from the tide-gauge data to obtain the land-height changes, we removed seasonal cycles from both data sets. To do this we used a simple technique as described by (*Burgette et al.*, 2009). That is, we average the whole time period data for each month, and then subtract the monthly averages from each month of the time series. This technique is better than a least-squares approach, because the least-squares method estimates one single seasonal parameter for the whole time period at each tide gauge. Using our current method, we find that the monthly seasonal signals vary a lot. We show one example of seasonal signals from both Aviso and tide gauges at Johor Baru tide gauges station in Singapore in Figure B.8.

B.3 Methods

B.3.1 Kinematic joint inversion

We used a modified version of the Extended Network Inversion Filter (ENIF, *Segall and Matthews, 1997; McGuire and Segall, 2003*) to jointly invert for both fault slip and viscous strain in a finite volume of rectangular block distributed at asthenosphere depths. The fault plane was discretized into rectangular subpatches. We calculated the Green's functions that map the fault slip on the patch buried in a homogeneous, isotropic, elastic half-space to surface displacement at GPS stations using the analytic solution of Okada (*Okada, 1992*). We numerically solved for the 3D strain Green's functions that map the strain components to surface displacement at GPS stations using the Relax code (*Barbot et al., 2009; Barbot and Fialko, 2010*). For each block, there are six components of strain. We show one example of surface displacement associated with the ϵ_{12} component in Figure B.9.

B.3.2 Modeling of viscous flow

In order to simulate a viscous flow, a deviatoric constraints (no volume change within the rectangular block) need to be applied to strain components in each block located at asthenosphere depth. This requires the summation of the trace of the strain tensor to be equal to zero e.g.,

$$\epsilon_{11} + \epsilon_{22} + \epsilon_{33} = 0 \quad (\text{B.1})$$

B.3.3 Spatial smoothing constraints on fault slip and strain components

As we are facing an underdetermined inverse problem, we apply spatial smoothing to both afterslip and strain components. We use a Laplacian matrix to regularize the fault slip that does not allow sudden jumps in slip between neighboring patches. For the strain blocks, we also applied this matrix to smooth each strain component, forcing the strain to change smoothly from one block to another in 3D space.

B.3.4 Strain direction constraints

Viscous flow following earthquakes is driven by coseismic stress perturbations, as is the case for any postseismic dynamic models. In our inversion, we also implemented direction constraints for the strain tensor of each block to encourage the viscous strain to flow in approximately the same direction to the coseismic stress change for the block. For each block, we calculate

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

the coseismic stress change at the center using the Relax code (*Barbot et al., 2009; Barbot and Fialko, 2010*). As an example, we show the strain bases vectors and coseismic stress at one face of a rectangular block along the ε_{11} component (See Figure B.10). Following are the constraints we use to encourage the strain to approximately follow the direction of stress σ along this component. We use the same rule for other strain components.

$$\begin{aligned}\widehat{\varepsilon} = (\varepsilon_{11}, \varepsilon_{12}, \varepsilon_{13}), \widehat{\sigma}' = (\sigma'_1, \sigma'_2, \sigma'_3), \widehat{\sigma}'' = (\sigma''_1, \sigma''_2, \sigma''_3) \widehat{\varepsilon} \cdot \widehat{\sigma}' = 0 \\ \widehat{\varepsilon} \cdot \widehat{\sigma}'' = 0\end{aligned}\quad (\text{B.2})$$

Where σ' , σ'' and σ are orthogonal to each other.

B.3.5 Estimation of viscosities

We estimate effective viscosity at each block by using the equation below,

$$\eta_{effective} = \frac{\Delta\sigma(t) + \sigma_0}{\Delta\dot{\varepsilon}(t) + \dot{\varepsilon}_0} \quad (\text{B.3})$$

Here, $\Delta\sigma$ is the deviatoric stress evolution for each block; $\Delta\dot{\varepsilon}$ is the deviatoric strain rate estimated by the Kalman filter; σ_0 and $\dot{\varepsilon}_0$ are the stress and strain rate of the steady-state mantle wedge flow. It is impossible to evaluate the magnitude of stress and strain rate in the mantle, thus we assume a certain range of strain rates and a Maxwell rheology with varying viscosity to estimate the stress based on typical geological strain rate and geological estimated viscosity. We conducted a grid search to estimate these values when we fitted our estimated effective viscosity. We show one example at block a (Figure 4.2) in Figure B.7.

B.3.6 Effective viscosity for a Burgers rheology

The strain of a Burgers rheology has been proposed by *Pollitz (2003)*, and is shown as below.

$$\varepsilon = \frac{\sigma_i}{2\mu_M} + \frac{\sigma_i}{2\mu_K} (1 - e^{-\frac{t}{\tau_K}}) + \frac{\sigma_i}{2\mu_M} \frac{t}{\tau_M} \quad (\text{B.4})$$

Here σ_i is the initial stress; μ_K , η_K is the shear modulus and viscosity of the Kelvin body with $\tau_K = \eta_K/\mu_K$ representing the characteristic relaxation time; μ_M and η_M are the viscosity and shear modulus of a Maxwell fluid with $\tau_M = \mu_M/\eta_M$ showing the Maxwell relaxation time. Thus the derivative of the strain yields the strain rate as below;

$$\dot{\varepsilon} = \frac{1}{\tau_K} \frac{\sigma_i}{2\mu_K} e^{-\frac{t}{\tau_K}} + \frac{\sigma_i}{2\mu_M} \frac{1}{\tau_M} \quad (\text{B.5})$$

and the effective viscosity is as

$$\eta_{effective} = \frac{\eta_K \eta_M}{\eta_M e^{\frac{-t}{\tau_K}} + \eta_K} \quad (B.6)$$

This indicates that the effective viscosity of Burgers rheology is a function of η_K , η_M and τ_K changing with time.

B.4 Synthetic tests to validate our technique and understand model weaknesses

B.4.1 Model set up

We conducted a synthetic test to validate our joint inversion algorithm. To do this we assumed a synthetic rupture area that has a planar fault with 150 km in length and 100 km in width. The fault strikes at 0° and dips at 17.5° to the east (Figure B.11b), and it has a Gaussian shape transient of slip behavior over 20 days, starting at day 20 and ending at day 40 (Figure B.11b). We discretized the fault plane into 30 x 20 km subpatches with 5 patches along both the strike and dip directions. The transient slip rate also has a Gaussian spatial distribution centered on the fault plane at 50 km along the dip direction and 75 km along-strike (Figure B.11b). We assume the same transient slip process along both the thrust and right-lateral slip components. We also assume synthetic GPS sites; these are shown in Figure B.11a.

To model the strain, we discretized the region below 36 km as rectangular blocks with $100 \times 40 \times 40$ km dimensions, as shown in Figure B.11b. There are three layers with 3 blocks for each layer, resulting in 9 blocks in total. We assign a spatial Gaussian shape of the strain distributed on these blocks, with the center at the second block on the top layer. Temporally, the transient strain also has the same Gaussian shape centered on the same day as for the transient slip (Figure B.11b). In our synthetic tests, we implemented this transient strain along the ϵ_{11} , ϵ_{13} and ϵ_{33} components to simulate a thrust-like motion. The reference system of the fault and the viscous strain blocks is shown in Figure B.12.

For this synthetic test, we assume the fault plane is buried in a homogenous, isotropic, elastic half-space. We calculate the Green's functions using Okada (1992). We calculate the Greens's function for the strain components of the blocks using the Relax code (*Barbot et al.*, 2009; *Barbot and Fialko*, 2010). We calculate Green's function for the 6 strain components individually. Once we have the Green's function for each strain component for each block, we reorganize them to build up the design matrix for inversion in the Kalman filter.

To validate our code and the joint inversion technique, we conducted three individual synthetic tests. First, we invert for slip on the assumed fault geometry without including the viscous strain

in the inversion. Second, we invert for the viscous strain alone with the fault slip turned off. Last, we jointly invert the slip and viscous strain simultaneously. The results of each individual test are shown in the following sections.

B.4.2 Synthetic slip inversion

Figure B.13 shows the input cumulative slip and the inverted slip from our modified ENIF. The filter recovers the slip very well in terms of the spatial pattern, although the maximum slip amount is slightly underestimated; this is a typical limitation of the filter (*McGuire and Segall, 2003*). We also show examples of time series comparison between the model predictions and the synthetic data for sites 7 and 12 in Figure B.11. The displacement prediction at these two sites fit nicely with the synthetic data.

B.4.3 Viscous strain inversion

Figure B.14 shows the input final deviatoric strains and the inverted deviatoric strains at blocks in the subsurface. The filter can recover the deviatoric strains very well from the synthetic data in terms of both the spatial distribution and the magnitude. Again, we calculate the model predictions for sites 7 and 12 (refer to Figure B.11a for locations). The model predictions fit very well except at time epochs where there are strong gradient of displacement change. These misfits are because the hyper-parameter that controls the temporal smoothing is constant for our final estimation, thus it may not be appropriate for these sharpen signal changes. A better algorithm that can tackle this issue by allowing for temporal variations in the hyper-parameter is outlined in *Fukuda et al. (2008)*; this is a modification we can test in future work.

B.4.4 Joint inversion of slip and viscous strain

Figure B.15 shows the results of our joint inversion of slip and viscous strain on the blocks. Our new technique recover well both the slip on the fault plane and the viscous strain in the subsurface. Especially, the spatial patterns of the slip and strains are recovered very well, while the magnitude of the afterslip is slightly underestimated as compared with the input. When we compare the slip distribution from joint inversion with that from slip inversion alone (Figure B.13), using the same spatial smoothing weight for the slip, the former case is slightly smoother as compared with the later case. We again show the model predictions at site 7 and 12 in comparison with the synthetic data as shown in Figure B.16. Overall, the model recovers most components of displacement very well except some components where the displacement magnitude is comparable to the noise level.

B.4.5 Summary of these synthetic tests

Based on these tests, we believe that our Kalman filter based inversion technique can recover slip on the fault and viscous strain on the blocks in the subsurface, both individually and jointly. This filter can recover the spatial pattern of them very well, although the absolute maximum magnitudes of both slip and viscous strains are slightly underestimated.

B.5 Afterslip estimates along the Sumatra subduction zone over the last decade

We have not focused on the afterslip distributions in the main text, because the focus of the paper is on the mantle-wedge rheology. In addition, we cannot examine the details of the afterslip distributions for this study because we use a relatively large spatial smoothing weight. However, we here provide some additional descriptions of the distributions we estimate, and how they compare to previous studies.

B.5.1 The Sumatra-Andaman rupture segment

Afterslip on the megathrust over the last decade is generally distributed in areas surrounding coseismic rupture patches, except for the middle segment of the 2004 Sumatra-Andaman earthquake (Figure 4.1). In the Andaman section of the 2004 rupture region (Figure 4.2), the afterslip occurs mainly downdip of the coseismic slip patch, which is consistent with previous studies (*Hoechner et al.*, 2011; *Paul et al.*, 2007, 2012; *Gunawan et al.*, 2014). Meanwhile at the Nicobar section between the Andaman Islands and north Sumatra, the afterslip occurs updip, downdip and also overlapping of the coseismic rupture patch (Figure 4.2). The afterslip estimates for the northern Sumatra segment by *Gunawan et al.* (2014) extend to the shallow part of the megathrust, overlapping the coseismic rupture area. This overlapping of the coseismic rupture patch is probably due to lack of data coverage in this area. Besides, afterslip on the shallow megathrust is poorly resolved due to lack of data coverage. Overall, the relative location of the downdip afterslip feature is consistent with previous postseismic studies. However, the absence of viscous flow in previous studies may have overestimated the downdip afterslip as what we found in our study.

B.5.2 The Nias rupture segment

At the Nias segment, we estimate that afterslip occurs both updip of maximum coseismic slip at Nias Island and downdip of the earthquake epicenter (Figure 4.2), generally in the area

surrounding the coseismic rupture patch. These locations are consistent with an earlier afterslip study from *Hsu et al.* (2006). The spatial afterslip distribution did not migrate in time and stayed stationary on the megathrust, which is also suggested by almost no azimuth changes for GPS offset over early and late time period (Figure 4.1).

B.5.3 The Bengkulu rupture segment

The general pattern of our estimated afterslip distribution is similar to that by *Lubis et al.* (2012) who examined the afterslip process using the first 15 months GPS time series following the earthquake. Our afterslip estimates concentrates both downdip and also updip of the main coseismic rupture (Figure 4.2), which is consistent with expectations from rate-and-state friction laws. Shallow afterslip from the 2007 event overlaps with the 2010 Pagai tsunami earthquake; this may have generated time-varying stress loading on the 2010 rupture patch that contributed to its eventual failure (*Tsang et al.*, 2016). The downdip part of the 2010 rupture patch that have been slipped following the 2007 event have been continually slipped after the 2010 Pagai, which is consistent with the rate- and state stress driven afterslip investigation from *Feng et al.* (2016a).

B.6 Supplementary figures

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.1: Geographic location of the GPS and Tide gauge locations.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.2: Fault geometry and rectangular blocks used for joint inversion of afterslip and viscous strain. The fault geometry is based on a slightly modified version of Slab 1.0 (Hu and Wang, 2012; Hayes et al., 2012).

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.3: Postseismic model prediction of surface displacement at selected GPS sites.

Figure B.4: Postseismic Afterslip model prediction of surface displacement at tide gauges.

Figure B.5: Difference in afterslip estimates for models that did and did not include viscous flow.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.6: Examples of low-pass filtered effective viscosity and predictions from a Burgers rheology for sites e, f, g, and h (refer to Figure 4.2 for the location). Blue and black broken lines show the time of the 2004 Sumatra-Andaman earthquake and 2007 Bengkulu earthquake.

Figure B.7: An example of our grid results for site a (refer to Figure 4.2 for location details) to determine the background strain rate ($Epsilon_0$), Maxwell viscosity (Eta_0), Kelvin viscosity (Eta_k) and shear modulus of a Kelvin body (G_k).

Figure B.8: Example time series showing our removal of annual cycles signals from the tide-gauge and Aviso data at the Johor Baru (Malaysia) site.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.9: Surface displacement (Green's function) with unit strain along ϵ_{12} component.

Figure B.10: Cartoon showing strain (dark blue) direction constraints based on coseismic stress directions (red). Dark blue vectors are the bases of the strain on this plane. The red and pink stress vectors form a possible stress bases vectors.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.11: Geometry of the fault plane and strain blocks in the subsurface for our synthetic tests. (a) Surface projection of the fault plane (green rectangle) with numbers showing the synthetic GPS sites for inversion. (b) Cross-section view of the geometry of the fault and blocks. The blue Gaussian shape plots indicate the time evolution of slip and viscous strain. Numbers indicate the arrangement of these blocks.

Figure B.12: Geometry of the fault plane, viscous blocks and the strain reference system. (a) Map view of the fault and blocks. The green rectangle shows the surface projection of the fault plane. (b) Upper panel shows the strain system at block 1 in a map view angle. Lower panel shows a cross-section view (A-A') of the strain reference system.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.13: Synthetic test results for slip inversion. Upper left panel shows the cumulative input slip on the fault plane. Upper right panel represents the recovered slip from our Kalman filter, with white dots showing the synthetic GPS sites. Lower panels show an example comparison between model predictions at sites 7 and 12 and the synthetic time series.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.14: Synthetic test results for viscous strain inversion. Upper left panel shows the cumulative input deviatoric strain on the blocks; upper right panel represents the recovered deviatoric strain. Lower panels show an example comparison between the model predictions at sites 7 and 12 and the synthetic time series.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.15: Synthetic test results for joint inversion of slip and viscous strain. Upper left panel is the cumulative input slip and the upper right panel shows the recovered slip from the filter, with the white dots showing synthetic GPS sites. Lower left panel shows the final cumulated input deviatoric strain, and lower right panel represents the deviatoric recovery from the filter.

APPENDIX B. SUPPLEMENTARY MATERIAL FOR SUMATRA POSTSEISMIC STUDY

Figure B.16: An example of comparison between joint inversion model predictions and synthetic time series at sites 7 and 12.

Figure B.17: An example of the low-pass filtered stress and strain rate at blocks a and b (refer to Figure 4.2 for location details).

Appendix C: Supplementary material for Nias-Simeulue postseismic study

C.1 Text S1

C.1.1 Coseismic slip uncertainty estimation

To quantify uncertainties in our estimated slip, we use a Bootstrap method. This technique has been widely used for slip uncertainty studies (*Feng et al.*, 2016b; *Parsons et al.*, 2006; *Wright et al.*, 2003). The technique uses a Monte Carlo based resampling scheme. It firstly perturbs the geodetic data that were used for inversion by randomly resampling the residuals from the best slip estimation and the data. It then inverts the slip based on this perturbed data set. If the perturbation from the residuals brings large difference in the slip magnitude at some places, it indicates that the slip at that place is not well resolved. In contrast, where the slip difference is small, it may indicate that the slip is relatively well resolved by the geodetic network. We performed 100 sets of perturbations for estimating the slip uncertainties that are shown in Figure C.3.

C.2 Text S2

C.2.1 Extended Network Inversion Filter (ENIF)

We used a modified version of the Extended Network Inversion Filter (ENIF; *Segall and Matthews*, 1997; *McGuire and Segall*, 2003; *Qiu et al.*, 2016) to estimate the time evolution of the afterslip and viscous strain. The ENIF models the cumulative surface displacement ($u_i(r_j, t)$)

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

of a GPS site at location r_j , time t , as follows;

$$u(r_j, t) = \sum_{a=1}^2 \int s_k(\xi, t) G_{ik}(r_j, \xi) dA(\xi) + \sum_{ml=11,12,13,22,23,33}^{33} \varepsilon_{ml}(\vartheta, t) G_{iml}(R_j, \vartheta) + L(r_j, t) + f(t) + error \quad (C.1)$$

where $u_i(r_j, t)$ represents the i th component of displacement at time epoch t ; the first term on the right side is the surface displacement at time epoch t due to slip $s_k(\xi, t)$ on a fault surface $A(\xi)$; $G_{ik}(r_j, \xi)$ represent the Green's functions that map unit slip on a fault patch to surface displacement at a given location; the second term is the surface displacement at time t due to strain component $\varepsilon_{ml}(\vartheta, t)$ of a rectangular block located at R_j in the subsurface; each block has six components of strain (i.e., $ml = 11, 12, 13, 22, 23, 33$) pointing in different directions; $G_{iml}(R_j, \vartheta)$ are the 3D Greens functions that map the unit strain components to the surface displacement; $L(r_j, t)$ represents the local non-tectonic random benchmark motions such as the local random motion of GPS monument; $f(t)$ is the common mode error of the whole GPS network such as seasonal loading, ocean loading, hydraulic loading etc; error contains the observation errors that are not taken into account during the GPS processing (e.g., multiple path errors and azimuthally varying path delays). This error is assumed to follow a Gaussian distribution with zero mean and covariance $\sigma^2 \Sigma_n$; here Σ_n is the GPS data covariance matrix derived from GPS data processing. More details about the original version of the ENIF are presented in *Segall and Matthews (1997)* and *McGuire and Segall (2003)*.

C.3 Text S3

C.3.1 Deviatoric constraints for the strain components

To model a viscous strain, we apply a deviatoric (no volume changes) constraint for each block in the subsurface. This requires that the summation of the trace of the strain tensor to be zero;

$$\varepsilon_{11} + \varepsilon_{22} + \varepsilon_{33} = 0 \quad (C.2)$$

C.3.2 Viscous strain direction constraints from coseismic stress changes

We apply a constraint that guides the strain components to point in approximately the same direction as the coseismic stress changes at each rectangular box. We calculate the stress tensor using the Relax software (*Barbot et al., 2009; Barbot and Fialko, 2010*). We use a reference frame with the first principle strain perpendicular to the general strike of the fault, the second

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

principle strain aligned to the strike of the fault, and the third one pointing upwards. We define extension to correspond to positive principle strains and negative to represent compression. Each pair of principle strains occupies one pair of faces of a rectangular block (Figure C.10). Therefore if we understand the constraints on one pair of principle strains, then we can apply the same constraints for other pairs of principle strains. Take the face with e_{11} principle strains as an example (Figure C.10). The base vectors are represented as ϵ_{11} , ϵ_{12} , ϵ_{13} with the black vectors; the stress vector σ on this face is indicated as the red vector. We find two vectors σ' , σ'' that are perpendicular to the stress vector as shown by the pink vectors. We constrain, firstly, the base strain vectors to be perpendicular to σ' , the solution space is the plane (e.g., plane 1) that goes through base σ and σ'' . Then we constrain the base strain vectors to be perpendicular to σ'' , the solution space is the plane (e.g., plane 2) that goes through σ , and σ' . After this two constraints, the final solution is the intersection of plane1 and plane 2 forming a vector that is parallel with the stress vector σ' . The mathematical formula is as follows.

$$\begin{aligned}\epsilon_{11}\sigma'_1 + \epsilon_{12}\sigma'_2 + \epsilon_{13}\sigma'_3 &= 0 \\ \epsilon_{11}\sigma''_1 + \epsilon_{12}\sigma''_2 + \epsilon_{13}\sigma''_3 &= 0\end{aligned}\tag{C.3}$$

C.4 Text S4

C.4.1 Building up the stress kernel for estimation of the stress evolution at each block

We calculate the stress tensor at the center of each block by using the Relax software (*Barbot et al.*, 2009; *Barbot and Fialko*, 2010). Once we have these components, it is straightforward to build up the stress kernel for stress components of each block. Here we show one example of how to build stress kernel along one stress component at a block. Then we apply the same rule for other components, and finally for all blocks. As an example of stress kernel component k_{11}^1 at block 1, it is the summation $\sigma_{(kl,mn)}^{ij}$, here i, j represent the j th block acting on the i th receiver block along stress $kl = 11, 12, 13, 22, 23, 33$ direction when j th block acting along $mn = 11, 12, 13, 22, 23, 33$ direction. Thus, for one component at stress acting on the i th block, it stresses all the remaining blocks along all the components. As an example of block 1 along 11 component, it is shown as below

$$k_{11}^1 = [\sigma_{11,11}^{11} + \sigma_{11,11}^{12} \dots \sigma_{11,11}^{1j} + \sigma_{11,12}^{11} + \sigma_{11,12}^{12} \dots \sigma_{11,12}^{1j} + \sigma_{11,33}^{11} + \sigma_{11,33}^{12} \dots \sigma_{11,33}^{1j}] \tag{C.4}$$

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

For the rest of other components at block 1, they are shown as follows

$$\begin{aligned}
 k_{11}^1 &= [\sigma_{11,11}^{11} + \sigma_{11,11}^{12} \dots \sigma_{11,11}^{1j} + \sigma_{11,12}^{11} + \sigma_{11,12}^{12} \dots \sigma_{11,12}^{1j} + \sigma_{11,33}^{11} + \sigma_{11,33}^{12} \dots \sigma_{11,33}^{1j}] \\
 k_{12}^1 &= [\sigma_{12,11}^{11} + \sigma_{12,11}^{12} \dots \sigma_{12,11}^{1j} + \sigma_{12,12}^{11} + \sigma_{12,12}^{12} \dots \sigma_{12,12}^{1j} + \sigma_{12,33}^{11} + \sigma_{12,33}^{12} \dots \sigma_{12,33}^{1j}] \\
 &\dots \\
 k_{33}^1 &= [\sigma_{33,11}^{11} + \sigma_{33,11}^{12} \dots \sigma_{33,11}^{1j} + \sigma_{33,12}^{11} + \sigma_{33,12}^{12} \dots \sigma_{33,12}^{1j} + \sigma_{33,33}^{11} + \sigma_{33,33}^{12} \dots \sigma_{33,33}^{1j}]
 \end{aligned} \tag{C.5}$$

Thus the stress kernel for j th block is shown as below,

$$k_{kl}^j = \begin{bmatrix} \sigma_{11,11}^{11} + \sigma_{11,11}^{12} \dots \sigma_{11,11}^{1j} \dots \sigma_{11,12}^{11} + \sigma_{11,12}^{12} \dots \sigma_{11,12}^{1j} \dots \sigma_{11,33}^{11} + \sigma_{11,33}^{12} \dots \sigma_{11,33}^{1j} \\ \sigma_{12,11}^{11} + \sigma_{12,11}^{12} \dots \sigma_{12,11}^{1j} \dots \sigma_{12,12}^{11} + \sigma_{12,12}^{12} \dots \sigma_{12,12}^{1j} \dots \sigma_{12,33}^{11} + \sigma_{12,33}^{12} \dots \sigma_{12,33}^{1j} \\ \dots \\ \sigma_{33,11}^{11} + \sigma_{33,11}^{12} \dots \sigma_{33,11}^{1j} \dots \sigma_{33,12}^{11} + \sigma_{33,12}^{12} \dots \sigma_{33,12}^{1j} \dots \sigma_{33,33}^{11} + \sigma_{33,33}^{12} \dots \sigma_{33,33}^{1j} \end{bmatrix} \tag{C.6}$$

and therefore the stress kernel K for N_s blocks are shown as below

$$K = [k_{kl}^j]_{6N_s \times 6N_s} \tag{C.7}$$

and this stress kernel K is used to estimate the time-varying stress tensor for N_s blocks and is shown as below

$$K\sigma = K \times [\varepsilon_{kl}^j]_{6N_s} \tag{C.8}$$

Here ε_{kl}^j represents the inverted cumulative strain components for all N_s blocks.

C.4.2 Evaluating the effective viscosity

We evaluate the effective viscosity with the following formula;

$$\eta_{effective} = \frac{\Delta\sigma(t) + \sigma_0}{\Delta\dot{\varepsilon}(t) + \dot{\varepsilon}_0} \tag{C.9}$$

Here, $\Delta\sigma$ is the deviatoric stress evolution at each block, $\Delta\dot{\varepsilon}$ are the deviatoric strain rate estimates from our ENIF. σ_0 and $\dot{\varepsilon}_0$ are the stress level and the strain rate in the mantle wedge, respectively. It is not possible to get the actual values of the stress and strain rate in the mantle, thus we assume a certain range of strain rates and a Maxwell rheology to estimate the steady-state stress level based on typical geological strain rate and geologically estimated viscosity. We estimate this strain rate and Maxwell viscosity through a grid search to fit this effective viscosity curve.

C.5 Text S5

C.5.1 Trade-offs in the models

Due to the coexistence of both afterslip and viscous flow in the postseismic relaxation time period, and also their significant contributions to surface displacement in many areas, it is not possible to distinguish the absolute contribution to the displacement field for each mechanism with available geodetic measurements; there are inherent trade-offs between them. In order to better understand the limitations of our new technique, we ran many models to examine how the trade-offs affect our final estimation of afterslip and viscous strain. To consistently compare models is a little tricky, because the addition of the Green's functions for strain over the blocks changes the overall weight of the design matrix, thus making it difficult to use consistently the same level of spatial smoothing for models with and without the viscous strains included. With the same smoothing parameters for the fault slip, inclusion of the Green's function from the blocks will modify the effective smoothing weight; for the same weight, the estimates of afterslip from joint inversion are smoother than the estimates when we invert for afterslip alone.

Based on many tests we have to use a higher level of spatial smoothing than we might like for the viscous strains, in order to get a reasonable stress evolution curve (Figure C.9). This is likely because the resolution for strain estimates for the blocks here is poor due to limited data coverage. Our strategy here is to test different smoothing weights that can balance the model fits with data, and meanwhile yield a reasonable shape of stress decay for the blocks. We find pairs of parameters that can satisfy the stress decay requirements, and we present the one that yields the best model fits in this study. Hopefully, high spatial and temporal data sets can help to better understand the model trade-offs, and to better determine the relative smooth weighting and the slip, strain magnitude as well.

C.6 Text S6

C.6.1 Effective viscosity of a Burgers Rheology

A Burgers rheology has Kelvin and Maxwell elements connected in series as shown in Figure C.11a. The total strain of the Burgers rheology has been proposed by Pollitz (2003) as

$$\varepsilon = \frac{\sigma_i}{2\mu_M} + \frac{\sigma_i}{2\mu_K} (1 - e^{-\frac{t}{\tau_K}}) + \frac{\sigma_i}{2\mu_M} \frac{t}{\tau_M} \quad (\text{C.10})$$

where σ_i is the initial stress, $\tau_M = \eta_M/\mu_M$ represents the characteristic relaxation of the Maxwell fluid, with η_M and μ_M the corresponding viscosity and shear modulus; $\tau_K = \eta_K/\mu_K$ refers to the characteristic relaxation time of the Kelvin body, with η_K and μ_K the corresponding viscosity

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

and shear modulus. Thus, the time derivative yields strain rate

$$\dot{\epsilon} = \frac{1}{\tau_K} \frac{\sigma_i}{2\mu_K} e^{\frac{-t}{\tau_K}} + \frac{\sigma_i}{2\mu_M} \frac{1}{\tau_M} \quad (\text{C.11})$$

and thus the effective viscosity is the stress divided by the strain rate as follows,

$$\eta_{effective} = \frac{\eta_K \eta_M}{\eta_M e^{\frac{-t}{\tau_K}} + \eta_K} \quad (\text{C.12})$$

which has a time-dependent behavior with the magnitude decided by the combination of Kelvin and Maxwell viscosities (η_K and η_M respectively). As an example, in the mantle wedge of the Nias segment, our best-fitting model gives an effective viscosity at block 1 of $\eta_K = 1.2 \times 10^{17}$ Pa.s, $\eta_M = 1.1 \times 10^{18}$ Pa.s. $\eta_K = 1.2 \times 10^{17}$ Pa.s / 20 GPa $\approx$ 0.2 year, thus the effective initial viscosity is $\sim 1 \times 10^{17}$ Pa.s when $t=0$, $\sim 1.05 \times 10^{18}$ Pa.s when $t=1$ year, and eventually 1.1×10^{18} Pa.s when $t \geq 2$ years (see Figure C.11).

C.7 Supplementary figures and tables

Table C.1: Crust 1.0 parameters for the Nias region. We use the parameters of Crust 1.0 at the center of the fault plane to represent the entire fault area since the spatial variation is not significant.

Depth (km)	Vp (km.s ⁻¹)	Vs (km.s ⁻¹)	Density (kg.m ⁻³)
0-2	2.44	1.01	2.09
2-10.4	5.8	3.5	2.63
10.4-18.6	6.3	3.62	2.74
18.6-26.7	6.9	3.94	2.92

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

Figure C.1: Verification of our fault geometry with relocated seismicity from *Tilmann et al.* (2010).

Figure C.2: Laplacian smoothing weight determination based on an L-curve of average stress-drop. The red star shows our preferred weight.

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

Figure C.3: Coseismic slip error estimation by using Bootstrap Monte Carlo sampling approach.

Figure C.4: The 3D geometry of the fault, viscous blocks and the calculated model resolution on these blocks.

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC
STUDY

Figure C.5: Comparison between our preferred coseismic slip estimates with *Konca et al. (2007)* (the green contour) model. The color scale is saturated at 14 m.

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

Figure C.6: Postseismic model prediction compared to GPS time series.

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

Figure C.7: Coseismic stress changes at viscous blocks in the mantle wedge from the slip model of *Konca et al. (2007)*.

Figure C.8: One grid search example for parameters η_K , η_0 , G_K , and ϵ_0 at block 1.

APPENDIX C. SUPPLEMENTARY MATERIAL FOR NIAS POSTSEISMIC STUDY

Figure C.9: The low-passed strain rate, stress and effective viscosity time series estimated for blocks 1 and 2. The shaded region indicates the lower and upper bounds of the estimated uncertainties.

Figure C.10: Cartoon showing reference systems for strain base vectors (dark blue), and coseismic stress base vectors (red) for one face of a rectangular block.

Figure C.11: A Burgers rheology and its effective viscosity. (a) The Burgers rheology is composed of Kelvin and Maxwell elements; (b) One example of our best-fitting effective viscosity for the Burgers rheology.