Integrating geodetic images and physics-based modeling towards understanding earthquake ruptures in Indonesia

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Indonesia

Rino

Asian School of the Environment
2020

**Integrating geodetic images and
physics-based modeling towards
understanding earthquake ruptures in
Indonesia**

Rino

Asian School of the Environment

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

2020

Candidate Statement of Originality

I hereby certify that the work embodied in this thesis is the result of original research, is free of plagiarised materials, and has not been submitted for a higher degree to any other University or Institution. I confirm that the investigations were conducted in accord with the ethics policies and integrity standards of Nanyang Technological University and that the research data are presented honestly and without prejudice.

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Authorship Attribution Statement

This thesis contains materials from three published papers (under the name Rino Salman) in the following peer-reviewed journals.

Chapter 2 was published as “Salman, R., Hill, E. M., Feng, L., Lindsey, E. O., Mele Veedu, D., Barbot, S., Banerjee, P., Hermawan., Natawidjaja, D. H. (2017). Piecemeal rupture of the Mentawai patch, Sumatra: The 2008 M_w 7.2 North Pagai earthquake sequence, *Journal of Geophysical Research: Solid Earth*, 122, doi: 10.1002/2017JB014341.”

The contributions of the co-authors are as follows:

- Assoc. Prof. Hill provided the initial project direction and guided the modeling.
- Dr. Feng provided the GPS coseismic offsets and time series.
- Dr. Lindsey guided the SAR data processing.
- Asst. Prof. Barbot and Ms. Mele Veedu performed the physical based simulations, and I was involved in analyzing the results.
- Dr. Hermawan, Dr. Banerjee, and Dr Natawidjaja maintained the GPS network and collected the data.
- I performed the modeling, analyzed the results, and prepared the manuscript drafts.
- All co-authors edited the manuscript.

Chapter 3 was published as “Salman, R., Lindsey, E. O., Feng, L., Bradley, K., Wei, S., Wang, T., Daryono, M. R., Hill, E. M., (2020). Structural controls on rupture extent of recent Sumatran Fault Zone earthquake, Indonesia, *Journal of Geophysical Research: Solid Earth*, 125, doi: 10.1029/2019JB018101”

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- Assoc. Prof. Hill provided the initial project direction.
- Dr. Feng provided the SAR data raw.
- Dr. Lindsey guided the SAR data processing .
- Dr. Wang processed the Sentinel-1 SAR data.
- Asst. Prof. Wei guided the modeling and interpretation of the results.
- Dr. Bradley guided interpretation of the results from a geological point of view.
- Dr. Daryono provided data from field observations.
- I performed the modeling, analyzed the results, and prepared the manuscript drafts.
- All co-authors edited the manuscript.

Chapter 4 was published as “Salman, R., Lindsey, E. O., Lythgoe, K. H., Bradley, K., Muzli, M., Yun, S.H., Chin, S. T., Tay, C. W. J., Costa, F., Wei, S., Hill, E. M., (2020). Cascading partial rupture of the Flores thrust during the 2018 Lombok earthquake sequence, Indonesia, Seismological Research Letters, doi:10.1785/0220190378”

The contributions of the co-authors are as follows:

- Assoc. Prof. Hill provided the initial project direction.
- Dr. Lindsey guided the SAR data processing.
- Dr. Lythgoe modeled the seismic data.
- Dr. Muzli provided the aftershock events.
- Dr. Yun, Ms. Chin, and Ms. Tay produced the damage proxy maps. I was involved in analyzing the results.
- Asst. Prof. Wei guided the modeling and interpretation of the results.
- Dr. Bradley guided interpretation of the results from a geological point of view.
- Assoc. Prof. Costa guided interpretation of the results from a volcanological point of view.
- I performed the modeling, analyzed the results, and prepared the manuscript drafts.
- All co-authors edited the manuscript.

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Summary

I present three studies on integrating geodetic data with physics-based modelling, combined with insights from seismology, geology, and volcanology to better understand how tectonic structures and fault properties affect earthquake ruptures. The first two studies investigate earthquake ruptures associated with megathrust and strike-slip earthquakes at the western Indonesian island of Sumatra, and the third study investigate rupture of thrust earthquakes associated with the back-arc thrusting zone north of Lombok island, eastern Indonesia.

The first study investigate the 2008 M_w 7.2 North Pagai earthquake that partially ruptured a small area in the middle of the Mentawai seismic gap, where interseismic coupling has been estimated to be high. I use GPS and InSAR data to develop a coseismic model and 5.6 years of cumulative GPS displacements to develop an afterslip model. The results suggest that the estimated afterslip partially overlaps the coseismic rupture, and a rate-and-state model implies that the part of the coseismic rupture that does not overlap with the afterslip may represent a velocity-weakening region, while the overlapping part may represent a velocity-strengthening region.

The second study investigate four geodetically observed $M_w \sim 6$ strike-slip earthquakes that occurred in mainland Sumatra between 2007 and 2016. Three events ruptured segments of the Sumatran Fault Zone (SFZ), and one ruptured a previously unmapped fault. I use coseismic offsets obtained from InSAR data to develop coseismic models of all these events. The results suggest several implications for future earthquakes in Sumatra: first, the terminations of ruptures along the SFZ can be controlled by structural step-overs that denote segment boundaries; second, earthquakes along the main SFZ produced surface ruptures, but an event off the main SFZ in northern Sumatra did not; and third, the remaining seismic gaps on the SFZ are primarily located between the equator and northern Sumatra (6° N).

The third study investigate four $M_w > 6$ earthquakes that struck the northern region of Lombok, eastern Indonesia, between late July and mid-August 2018. I based the study on seismic and InSAR data analysis to develop coseismic models of all the events, and damage proxy maps to quantify the effect of each event on buildings. I find that: first, the earthquake sequence originated from different hypocenter depths and ruptured south-dipping thrust faults with different strikes and dips, which may explain the cascading behavior of the events; second, the earlier events

in the sequence played a role in promoting the later events; third, the second event caused volumetric expansion in the shallow magma plumbing system of Rinjani volcano and unclamped its magma ascent cone, which may potentially promote future volcanic activity; and fourth, the second earthquake caused the greatest damage to buildings.

Chapter 1

Introduction

Indonesia is one of the most tectonically active areas of the world due to geologic processes resulting from the convergence of the Pacific, Philippine Sea, Indo-Australian, and Sunda plates. These plates convergence creates plate boundaries on almost all sides of the country (Figure 1.1). Due to their vast coverage, the plate boundaries differ in their characteristics from the west to the east of the country. I order the discussion of these boundaries based on their regional tectonic characteristics and earthquake records. The discussion starts from the western side, where the convergence of the Indo-Australian and Sunda plates beneath Sumatra is oblique. I then move to the centre, where the convergence of the Indo-Australian and Sunda plates beneath Java is almost orthogonal. Finally, the discussion ends with the eastern side, where the convergence of the Pacific, Philippine Sea, Indo-Australian, and Sunda plates creates a complex tectonic setting.

FIGURE 1.1: The convergence of the Pacific, Philippine Sea, Indo-Australian, and Sunda plates forms plate boundaries on almost all sides of the country. Black arrows represent rates and directions of the moving Philippine Sea, Pacific, and Indo-Australian plates relative to the fixed Sunda plate (DeMets et al., 2010).

1.1 Sumatra

The approximately northward-moving Indo-Australian plate subducts beneath the Sunda plate under the northwest-southeast oriented island of Sumatra, resulting in oblique plate convergence (Figure 1.2). The oblique plate convergence is partitioned into trench-perpendicular (dip-slip) and trench-parallel (strike-slip) components (Fitch, 1972), and also diffuse deformation within the Indo-Australian plate of the Wharton Basin (Bradley et al., 2017). While most of the dip-slip component is accommodated along a megathrust offshore west Sumatra, most of the strike-slip component is accommodated along a strike-slip fault within the Sumatra island—the Sumatran Fault Zone (Bradley et al., 2017; Fitch, 1972; McCaffrey, 2009). Both the megathrust and the Sumatran Fault Zone have produced a number of destructive earthquakes that pose significant seismic hazard to communities living nearby.

1.1.1 Megathrust earthquakes

The megathrust ruptured on 26 December 2004, generating the M_w 9.2 Aceh Andaman earthquake and a destructive tsunami that claimed casualties and infrastructure damage on a scale not witnessed in modern history. This earthquake ruptured the megathrust from offshore west Aceh in northern Sumatra to the Andaman islands to the north (Figure 1.2), with a rupture length of about 1500 km and a maximum slip exceeding 20 m (Chlieh et al., 2007; Subarya et al., 2006). The 2004 earthquake was likely not the first event of such magnitude, as prehistoric tsunami deposits from a coastal cave in Aceh, northern Sumatra, suggest that a similar event occurred about 5000 years ago (Rubin et al., 2017).

Around three months after the 2004 event, the megathrust to the south of the 2004 event ruptured on 28 March 2005 in the M_w 8.7 Nias-Simeulue earthquake. Unlike the 2004 event, this earthquake ruptured the megathrust from Nias island to Simeulue island to the south (Figure 1.2), with a rupture length of about 400 km and a maximum slip exceeding 10 m (Briggs et al., 2006; Konca et al., 2007). The event generated a smaller tsunami compared to that of the 2004 event (Kerr, 2005). A study of coral microatoll fossils suggests that the same section of the megathrust ruptured in an event of a similar extent and magnitude in 1861 (Meltzner et al., 2015).

FIGURE 1.2: The convergence of the Indo-Australian and Sunda plates beneath Sumatra is oblique, thus the subducted slip is partitioned into trench-perpendicular (dip-slip) and trench-parallel (strike-slip) components (Fitch, 1972), and also diffuse deformation within the Indo-Australian plate of the Wharton Basin (Bradley et al., 2017). Most of the dip-slip component is accommodated along a megathrust (regions between the barbed-lined and the Sumatra island), and most of the strike-slip component is accommodated along the Sumatran Fault Zone within the island (Fitch, 1972; McCaffrey, 2009). Black arrows represent rates and directions of the moving Australian plate relative to the fixed Sunda plate (DeMets et al., 2010).

The northern end of the 2005 earthquake rupture tapered down under the central Simeulue island, at approximately the same location as the southern extent of the 2004 earthquake rupture patch (Briggs et al., 2006). In addition, the southern end of the 2005 earthquake rupture also tapered down under Batu islands (Figure 1.2). Studies of coral microatoll fossils reveal that seven earthquakes in the past 1100 years terminated under central Simeulue (Meltzner et al., 2012), and the 1861 event also terminated under Batu islands (Meltzner et al., 2012; Philibosian et al., 2014), suggesting that the megathrust under Simeulue island and Batu islands hosts persistent barriers to earthquake ruptures. Nevertheless, despite being persistent barriers to earthquake ruptures, the megathrust under central Simeulue island generated the 2002 M_w 7.3 and the 2008 M_w 7.4 earthquakes, indicating that the persistent barrier regions still comprising a locked asperity (Morgan et al., 2017). To explain the barriers, several mechanisms have been proposed, such as a reduction in the width of the seismogenic zone by ~ 25 km under Simeulue island (Tilman et al., 2010) or numerous fracture zones on the plate subducting under Simeulue island and Batu islands (Franke et al., 2008; Henstock et al., 2016; Meltzner et al., 2012; Philibosian et al., 2014).

Around two years after the Nias-Simeulue earthquake, the megathrust ruptured again on 12 September 2007 in the M_w 8.4 and M_w 7.9 Bengkulu earthquakes, just 12 hours apart. These events ruptured about 250 km (Konca et al., 2008) of a 700-km-long segment of the megathrust from Batu islands to Enggano island (Figure 1.2), known as the Mentawai patch. The 2007 events ruptured the southern half of the Mentawai patch between Enggano island and North Pagai islands, leaving the northern half of the Mentawai patch from North Pagai islands to Batu islands unbroken. The maximum slip of the 2007 events exceeded 5 m (Konca et al., 2008; Tsang et al., 2016), and generated a moderate tsunami (Borrero et al., 2009; Gusman et al., 2010). Studies of coral microatoll fossils shows that the 2007 events were the latest in a sequence of large earthquakes that have occurred in the Mentawai patch over at least the last 700 years (Philibosian et al., 2017; Sieh et al., 2008). The studies also suggest that the Mentawai patch ruptures in supercycles of large earthquakes approximately every 200 years. Each supercycle may involve at least two events: one event rupturing the southern half and the other event rupturing the northern half of the Mentawai patch, with each event always terminating in the centre of the Mentawai patch. The studies suggest that these phenomena could

be explained by spatially variable, and perhaps also temporally variable, frictional properties of the megathrust beneath the central Mentawai patch.

After the 2007 event, the megathrust ruptured in two further earthquakes. The first one on 25 February 2008 was the M_w 7.2 North Pagai earthquake, which ruptured the middle section of the Mentawai patch, and the other was the 25 October 2010 M_w 7.8 earthquake, which ruptured the shallow part of the megathrust immediately trenchward of the Pagai islands (Figure 1.2). The investigation of the 2008 earthquake is the focus of Chapter 2 in this thesis, to reveal why this event did not cascade into a larger event although it ruptured the a highly coupled megathrust. The 2010 event ruptured approximately 100 km of a shallow part of the megathrust, with a maximum slip exceeding 9 m, and generated a tsunami (Hill et al., 2012).

1.1.2 Sumatran Fault Zone earthquakes

The Sumatran Fault Zone (SFZ), which is divided into 19 segments (Sieh and Natawidjaja, 2000), extends the length of the Sumatra mainland (Figure 1.2). The SFZ has generated a number of shallow strike-slip earthquakes. For example, since the 2004 M_w 9.2 megathrust earthquake, seven $6 \leq M_w < 7$ strike-slip earthquakes struck the island between 2007 and 2016, with five earthquakes occurring along the main SFZ segments, and two earthquakes being outside of the main SFZ segments. Although these earthquakes were relatively smaller compared to the previously mentioned megathrust events, these strike-slip earthquakes originated from shallower depths and closer to dense population centres than the megathrust events. Thus these strike-slip events still pose significant seismic hazard to local communities.

Despite the seismic hazard, these strike-slip earthquake ruptures have been investigated far less than the megathrust earthquake ruptures, partly because near-field GPS data are not available due to limited road access to most of the SFZ segments. Therefore, we lack knowledge of seismogenic depths and spatial slip distributions of these strike-slip earthquakes, which are important parameters to improve the seismic hazard assessment of the SFZ. Thanks to InSAR technologies that need no additional instruments on Earth's surface, surface deformations associated with

some of these earthquakes are still detected. Investigation of these strike-slip earthquakes using InSAR observations is the focus of Chapter 3 in this thesis.

1.2 Java

The Australian plate is approximately moving northward, and subducting beneath the Sunda plate under the island of Java, which is oriented in the west-east direction (Figure 1.3). The resulting convergence of the Australian and Sunda plates is thus almost orthogonal, where the majority of the trench-perpendicular slip is accommodated by the megathrust, and a much smaller trench-parallel slip is accommodated along the Baribis and Kendeng Thrusts (Koulali et al., 2017).

FIGURE 1.3: The convergence of the Indo-Australian and Sunda plates beneath Java is almost orthogonal. Black arrows represent rates and directions of the moving Indo-Australian plate relative to the fixed Sunda plate (DeMets et al., 2010).

Although the megathrust accommodates the majority of the trench-perpendicular slip, the megathrust has not historically generated earthquakes larger than M_w 7.8 (Okal, 2012). The largest earthquakes that have occurred during the period of the Global CMT catalogue (1976 - present) were the 2 June 1994 M_w 7.8 and

the 17 July 2006 M_w 7.7 tsunami earthquakes ([Abercrombie et al., 2001](#); [Bilek and Engdahl, 2007](#); [Ekström et al., 2012](#); [Yagi and Fukahata, 2011](#)). The 1994 event ruptured the shallow portion of the megathrust offshore south of eastern Java, with a rupture length of ~ 200 km and a maximum slip >1 m ([Abercrombie et al., 2001](#); [Bilek and Engdahl, 2007](#)). The 2006 event ruptured the shallow portion of the megathrust offshore south of western Java, with a rupture length of about 200 km and a maximum slip exceeding 2 m ([Bilek and Engdahl, 2007](#); [Yagi and Fukahata, 2011](#)). During 1900 - 1976, five earthquakes were reported near the trench, with one event greater than magnitude 7, but none of them was associated with the megathrust ([Okal, 2012](#)).

The absence of large earthquakes ($M_w > 8.5$) can be explained in several ways. First, the trench-perpendicular slip along the megathrust could be dominantly aseismic. Assessing the aseismic level is not possible with the current network of GPS stations that are ~ 200 km away from the subduction trench, and hence lack the resolution in the upper ~ 50 km depths of the seismogenic zone ([Koulali et al., 2017](#)). Second, large megathrust earthquakes can recur on centennial to millennial time scales ([Rubin et al., 2017](#); [Satake and Atwater, 2007](#); [Sieh et al., 2008](#)), with the longest seismic gap ever reported being 2000 years ([Rubin et al., 2017](#)). Thus, before 1900, the Java megathrust could have produced large earthquakes which have not been captured due to the short instrumental and historical records.

Besides the two tsunami earthquakes, most other Java earthquakes were associated with shallow crustal faults within the island. For example, the island was struck by eight damaging earthquakes between 1699 and 1875 ([Harris and Major, 2017](#)) and the 26 May 2006 M_w 6.4 Yogyakarta earthquake ([Abidin et al., 2009](#); [Tsuji et al., 2009](#); [Walter et al., 2008](#)). The 2006 event ruptured a left-lateral strike-slip Opak fault trending south-southwest to north-northeast, with a rupture length of about 18 km and a maximum slip of less than 1 m ([Abidin et al., 2009](#)). Despite the small rupture length and slip, this event caused large economic and societal losses: ~ 127000 houses were destroyed, and ~ 6000 people were killed [Tsuji et al. \(2009\)](#).

1.3 Bali to Papua

From Bali to Papua (the western half of New Guinea island) (Figure 1.4), the interaction of Pacific, Philippine Sea, Indo-Australian, and Sunda plates creates a complex tectonic setting in several ways. The first complexity arises in the south of the Bali-Papua region, where both the subducting Indo-Australian plate and the overriding Sunda plate change in their average density from the west to the east of Sumba island (120° E), which results in transition from a subduction system to a collision system in the vicinity of Sumba island (Silver et al., 1983). The collision system has been suggested to be the mechanism that drives the formation of the Banda arc, which comprises a volcanic inner arc, outer arc islands, and troughs, all in a westward-concave shape (Carter et al., 1976; Hamilton, 1979), and back-arc thrusting zones, which are dominated by the Flores and the Wetar thrusts (Hamilton, 1979; Silver et al., 1983) (Figure 1.4). The distinct westward-concave shape of the Banda arc has been variously explained as a result of either the westward motion of the Pacific plate and the northward motion of the Indo-Australian plate forming two-slab subduction (Cardwell and Isacks, 1978; Katili, 1975; McCaffrey, 1989), or as being caused by rollback of a single slab associated with northward subduction of the Indo-Australian plate (Spakman and Hall, 2010).

FIGURE 1.4: The interaction of the Pacific, Philippine Sea, Indo-Australian, and Sunda plates creates a complex tectonic setting from Bali to Papua regions. Black arrows represent rates and directions of the moving Philippine Sea, Pacific, and Indo-Australian plates relative to the fixed Sunda plate (DeMets et al., 2010).

The Flores thrust extends from Flores to the Bali Basin. The largest instrumentally recorded earthquakes generated by the Flores thrust were the 1992 M_w 7.9 Flores event, which was followed by a destructive tsunami, and a recent sequence of four shallow M_w 6.2 to 6.9 earthquakes that struck Lombok island in 2018. The location of the 2018 earthquake sequence was not surprising, since the section of the Flores thrust under Lombok accumulates a slip deficit of ~ 10 mm/year (Koulali et al., 2016). However, what remains puzzling is why the accumulated strain was released sequentially. The investigation of this sequence is the focus of Chapter 4 in this thesis.

The second complexity arises in the east of the Bali-Papua region, where the westward-moving oceanic Pacific plate is converging obliquely with the approximately northward-moving continental Australian plate, which results in a plate boundary comprising thrust fault and strike-slip fault zones (Hall and Spakman, 2002; Johnson and Molnar, 1972), with the thrust faults extending from east to west of New Guinea island and the strike-slip faults extending from east of New Guinea island to Sulawesi island (Norvick, 1979).

The third complexity arises in the north of the Bali-Papua region, where the convergence of the Indo-Australian, Philippine Sea, and Sunda plates has formed the Minahasa trench, East Sulawesi trench, and strike-slip faults, all of which define the tectonics of the K-shaped island of Sulawesi (Bellier et al., 2001; Socquet et al., 2006; Vigny et al., 2002; Walpersdorf et al., 1998). The triple convergence has also formed the Sangihe double subduction zone, which defines the tectonics of Halmahera island.

1.4 Thesis roadmap

In this thesis, I integrate geodetic data with physics-based modeling, combined with insights from seismology and geology, to better understand how tectonic structures and fault properties affect earthquake ruptures. The next chapters comprise three body chapters, which are research papers published during my PhD candidature, followed by a concluding chapter. In Chapter 2, I use coseismic offsets estimated from continuous Global Positioning System (GPS) stations and Interferometric Synthetic Aperture Radar (InSAR) to estimate the coseismic slip of the 2008 North

Pagai earthquake. Then, I use 5.6 years of cumulative GPS displacements to estimate the event's afterslip. I analyse how both the coseismic slip and afterslip are connected, and develop a simple rate-and-state model to infer fault properties of the megathrust on which the North Pagai earthquake occurred. In Chapter 3, I use InSAR coseismic offsets to estimate coseismic slip for all strike-slip earthquakes detected geodetically on the SFZ. I analyse the nature of structural features that control the rupture extent of such SFZ earthquakes. In Chapter 4, I use InSAR coseismic offsets and seismic data to understand the 2018 Lombok sequence. I analyse why the earthquakes occurred sequentially, and compute the impact of the sequence on a nearby active volcano. In Chapter 5, I write the overall conclusions of the works presented in Chapter 2 - 4, including recommendations for future works.

Chapter 2

Piecemeal rupture of the Mentawai patch, Sumatra: The 2008 M_w 7.2 North Pagai earthquake sequence

Abstract

The 25 February 2008 M_w 7.2 North Pagai earthquake partially ruptured the middle section of the Mentawai patch of the Sunda megathrust, offshore Sumatra. The patch has been forecast to generate a great earthquake in the next few decades. However, in the current cycle the patch has so far broken in a sequence of partial ruptures, one of which was the 2008 event, illustrating the potential of the patch to generate a spectrum of earthquake sizes. We estimate the coseismic slip distribution of the 2008 event by jointly inverting coseismic offsets from GPS and interferometric synthetic aperture radar. We then estimate afterslip with 5.6 years of cumulative GPS displacements. Our results suggest that the estimated afterslip partially overlaps the coseismic rupture. The overlap of coseismic rupture and afterslip can be explained conceptually by a simple rate-and-state model where the degree of overlapping is controlled by the dynamic weakening and the critical nucleation size in the velocity-weakening area. Comparing our rate-and-state model results with our geodetic inversion results, we suggest that the part of the coseismic rupture that does not overlap with the afterslip may represent a velocity-weakening region, while the overlapping part may represent a velocity-strengthening region.

2.1 Introduction

The Mentawai patch, a 700-km-long segment of the Sunda megathrust, has not yet released all the strain it has accumulated since the last great earthquakes in 1797 and 1833 (Natawidjaja et al., 2006; Philiposian et al., 2014). This is worrying, because the Mentawai patch has ruptured in a sequence of great earthquakes ($M_w > 8.5$) about every two centuries over the last 700 years (Sieh et al., 2008). Despite being seismically dormant for a long period, the Mentawai patch has been suggested to be highly coupled during interseismic periods based on coral (1950 - 2000) and GPS (1991 - 2006) data (Chlieh et al., 2008; Philiposian et al., 2014) and capable of producing another great earthquake that could devastate nearby populous cities such as Padang (Sieh et al., 2008). However, instead of producing one great earthquake, the Mentawai patch has so far ruptured only partially in the most recent earthquake cycle, starting in 2007 (Konca et al., 2008; Philiposian et al., 2014).

The southern half of the Mentawai patch ruptured during the 2007 M_w 8.4 Bengkulu earthquake (Konca et al., 2008; Tsang et al., 2016) and the 2010 M_w 7.8 Mentawai tsunami earthquake (Hill et al., 2012; Lay et al., 2011; Yue et al., 2014) (Figure 2.1). These events and their following afterslip (Feng et al., 2016; Lubis et al., 2013) have released only $\sim 30 - 44\%$ of the moment (Tsang et al., 2016) that was released during the 1833 event. To the contrary, the northern half of the Mentawai patch has experienced only $M_w \sim 7$ earthquakes including aftershocks of the 2007 Bengkulu earthquake (M_w 7.9 and 7.0) (Konca et al., 2008; Tsang et al., 2016) and the 2008 M_w 7.2 North Pagai earthquake.

The M_w 7.2 North Pagai earthquake of 25 February 2008 was preceded by several foreshocks in the updip region of the megathrust. The foreshocks began with an M_w 5.4 event on 19 February, which was then followed by three $M_w \sim 5$ earthquakes until 24 February and one M_w 6.5 earthquake 18 hours before the mainshock. On the same day as the mainshock, two aftershocks of M_w 6.6 and M_w 6.7 occurred. This sequence of earthquakes ruptured a small area in the middle of the Mentawai patch (Figure 2.1) and has not been extensively studied to date. The fact that only a small area of the Mentawai patch has ruptured so far could have important implications for understanding the unbroken megathrust in the northern half of the Mentawai patch, particularly given recent results from coral paleogeodetic data that

FIGURE 2.1: The Mentawai patch has been releasing strain in piecemeal ruptures since 2007. The southern half of the Mentawai patch ruptured during the 2007 M_w 8.4 Bengkulu mainshock and its aftershocks of M_w 7.9 and 7.0 (yellow area) (Tsang et al., 2016) and the 2010 M_w 7.8 Mentawai tsunami earthquake (green area) (Hill et al., 2012). The red star indicates the epicenter of the 2008 mainshock, which was preceded by four $M_w \sim 5$ foreshocks 1 week before and one M_w 6.5 foreshock the day before (gray focal mechanisms). The mainshock was then followed by M_w 6.6 and 6.7 aftershocks (denoted by green and blue focal mechanisms, respectively) that occurred on the same day as the mainshock. The red triangles are the locations of the continuous GPS stations, and the triangles with white color inside are stations that also recorded high-rate data. Black squares show the coverage of ALOS-1 images that documented coseismic offsets. The black arrow is the convergence vector after Prawirodirdjo and Bock (2004). The islands to the southeast of Siberut are Sipora, the Pagai Islands (North Pagai and South Pagai), and Enggano. The gray lines with labels are depth contours of the megathrust based on Slab1.0 (Hayes et al., 2012). Focal mechanisms are from the global CMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). Fracture zones are from Wessel et al. (2015).

suggest that segmented ruptures of the patch have happened in past supercycles (Philibosian et al., 2017). That is, the likelihood of a single great earthquake in the near future remains high as previously suggested (Chlieh et al., 2008; Sieh et al., 2008), but the possibility of releasing the cumulative strain in piecemeal cannot be ruled out.

In this first detailed study of the 2008 M_w 7.2 North Pagai earthquake, we first estimate the coseismic slip distribution by jointly inverting offsets from GPS and interferometric synthetic aperture radar (InSAR) data. We then estimate the afterslip distribution based on 5.6 year cumulative GPS postseismic displacements. Our results show that the estimated afterslip following the 2008 event partially overlaps the updip region of the coseismic rupture zone. We lastly conduct conceptual rate-and-state models to explore the physical conditions that would make such overlap happen.

2.2 Data and analysis

The 2008 North Pagai earthquake was recorded by nine stations from the Sumatran GPS Array (SuGAr). We use coseismic offset estimates and 5.6 year (February 2008 to December 2013) postseismic time series derived by Feng et al. (2015), who carefully separated the 2008 signal from that of the other earthquakes. Of the nine GPS stations, only three stations (SLBU, KTET, and SMGY) located within 30 km from the epicenter recorded >5 cm of coseismic and postseismic deformation (Figure 2.2a). We note that because only a white noise term is considered, the GPS uncertainties are likely to be systematically underestimated. These three stations also recorded high-rate (1 s sampling) data. To generate high-rate kinematic solutions, we used the GPS-Inferred Positioning System and Orbit Analysis Simulation Software (GIPSY-OASIS) version 6.2 from the Jet Propulsion Laboratory (JPL) and JPL final precise satellite orbits and clocks (Zumberge et al., 1997). We then estimated the offsets using 7 min of low-pass filtered time series both before and after the earthquake, with the first 270 s after the earthquake removed to avoid the dynamic waveforms of the stations (Figure A.1).

The GPS coseismic offsets show that horizontal motions recorded by both static (daily) and high-rate solutions are generally consistent in their orientations but

FIGURE 2.2: (a) The nine GPS stations used in our inversion. Of the nine, three stations (SLBU, KTET, and SMGY) are located within 30 km from the epicenter; these are the only stations that recorded >5 cm of surface displacements (daily solutions) for both coseismic and postseismic processes. (b) The subsampled unwrapped phase for our InSAR result shows a clear spatial gradient of coseismic offsets, which are mostly located along the western coastline of North Pagai island.

not in their amplitudes (Figure 2.3). At SLBU, KTET, and SMGY the static solutions are larger compared to the high-rate ones. This discrepancy was likely caused by the two aftershocks and early afterslip (Figure 2.4). The static offsets show that station SLBU recorded the largest horizontal displacement of ~ 22 cm and vertical displacement of ~ 15 cm. In contrast to the coseismic offsets, after 5.6 years station SMGY (which initially recorded the smallest coseismic offsets among the three stations) recorded the largest cumulative horizontal postseismic displacements of ~ 17 cm (Figure 2.2a).

In addition to GPS coseismic offsets, we also use InSAR observations from two different tracks of the ALOS-1 satellite: one track covers Sipora island and the other

FIGURE 2.3: Estimated GPS coseismic offsets for both daily and high-rate solutions are generally consistent in their orientation but not in their amplitudes. At SLBU, KTET, and SMGY, the static solutions are larger compared to the high-rate ones.

track covers Pagai islands including North Pagai and South Pagai whose land-mass are almost continuous (black squares in Figure 2.1). To get an unwrapped phase solution, we used an InSAR processing system based on GMT (GMTSAR) (Sandwell et al., 2011). We used Shuttle Radar Topography Mission global 30 m DEM (SRTMGL1) data (NASA JPL, 2013) to reduce topography contributions to the unwrapped phase and unwrapped each image individually using Statistical-Cost, Network-Flow Algorithm for Phase Unwrapping (SNAPHU) (Chen and Zebker, 2000). After the processing, we referenced unwrapped phases with three GPS line-of-sight displacements covering the same time span as the images (e.g., Cavalié et al., 2013). As the InSAR slave images were taken 27 and 44 days after the main-shock (Table A.1), the result (Figure 2.2b) may capture displacements from both the coseismic and 1.5 months of postseismic deformation. In particular, the GPS time series for KTET shows ~ 2 cm of total postseismic displacement (Feng et al., 2015). Finally, we subsampled the detrended unwrapped phase using the quadtree

FIGURE 2.4: Comparison between GPS solutions from daily (black vectors) and high-rate (colored vectors) solutions. High-rate solutions for the mainshock along with the two aftershocks are smaller than the daily solutions. Cyan-colored vectors are the estimated displacements of the two aftershocks and early afterslip, calculated by subtracting high-rate solutions (red) from daily solutions (black).

method (Jónsson et al., 2002), hence reducing the number of measurements from $\sim 10^6$ points to 642.

2.3 Modelling Approach

We modeled both the coseismic offsets and cumulative displacement of postseismic deformation (Table A.2) using analytical solutions of dislocations in a homogeneous elastic half-space (Okada, 1992). We based our fault geometry on the three-dimensional Slab 1.0 model (Hayes et al., 2012), but we reduced the fault depth by 2 km to fit with relocated aftershock locations from a temporary seismic network (Collings et al., 2012). We then discretized the fault geometry into 10 by 10 km patches. In inversion, we assumed the initial rake to be the gCMT rake of 102° and allowed it to vary within $\pm 45^\circ$. We applied a positivity constraint (Du et al., 1992) as well as second-derivative Laplacian smoothing (Du et al., 1992; Harris and Segall, 1987; Jónsson et al., 2002). We used a smoothing factor, determined using an L-curve criterion (Aster et al., 2018) (Figure A.2), to choose the preferred trade-off between slip roughness and model misfit.

To estimate a coseismic slip distribution, we jointly inverted both GPS and InSAR data sets. First, in order to give reasonable weights between the GPS and subsampled InSAR data sets, we inverted each data set individually without using spatial smoothing to obtain the uncertainties that would result in a normalized, reduced chi-square, χ^2_r (defined as the sum of squared weighted residuals divided by the number of observations) of 1. The uncertainties for all InSAR points were estimated to be 1.25 cm, while the scaling for GPS formal uncertainties (on the order of 0.25 cm) was 4. Second, we jointly inverted the data using the uncertainties obtained in the previous step and estimated the best smoothing factor until we obtained an optimal trade-off between data misfit and slip distribution. With regard to postseismic deformation, we modeled the cumulative GPS displacements assuming that they were predominantly controlled only by the afterslip process, but we do consider (and discuss in the next section) the possible implications of neglecting viscoelastic effects.

2.4 Model Results

Our preferred coseismic slip model and its misfit are shown in Figure 2.5. The model (Figure 2.5a) shows ~ 2 m of peak coseismic slip, which is far smaller than expected from the accumulated slip deficit of ~ 10 m stored in the Mentawai patch

since the last great earthquake in 1833 (given a convergence rate of ~ 5.7 cm/year after Prawirodirdjo and Bock (2004)). The amplitude of peak slip, however, is heavily dependent on the smoothing factor. Our model also shows an estimated slip of ~ 0.5 m on the megathrust under the southern part of Sipora island. Modelling results for each data set (Figure A.3) show that including InSAR data in the joint inversion results in the estimation of some slip beneath Sipora island. The slip is presumably needed to explain the 1.5 months of postseismic deformation included in the observations. The coseismic slip, assuming a constant shear modulus of 30 GPa, is equivalent to M_w 7.16. This magnitude agrees with the gCMT magnitude of M_w 7.2.

FIGURE 2.5: (a) We estimate peak coseismic slip during the 2008 event of ~ 2 m, which is far smaller than expected from the accumulated slip deficit of ~ 10 m stored in the Mentawai patch since the last great earthquake in 1833 (given a convergence rate of ~ 5.7 cm/year after Prawirodirdjo and Bock (2004)). Black-colored vectors are coseismic GPS offsets from daily solutions. (b) Comparison of InSAR coseismic observation and model prediction.

FIGURE 2.6: We estimate afterslip to be located updip and downdip of the coseismic slip patch. In the updip area, the estimated afterslip partially overlaps the coseismic rupture area (blue 50 cm contour lines).

Our postseismic model estimates afterslip to be located updip and downdip of the coseismic rupture area (Figure 2.6). A prominent feature of our postseismic model is that the estimated afterslip partially overlaps the updip region of the coseismic rupture area. To determine the robustness of the partial overlap, we conducted several tests. First, an inversion with less smoothing also shows partial overlap (Figure 2.7), suggesting that the overlap is not an artifact caused by the smoothing. Second, an inversion in which we constrain afterslip not to be allowed within the 1 m coseismic slip contour gives a relatively poor fit (albeit not a large difference; χ^2_2 increases from 3.0 to 4.85) in the horizontal component at stations SLBU and SMGY as well as in the vertical component at station SLBU (Figure 2.8). These tests suggest that the data do seem to have a preference toward the overlap, although with this underdetermined problem we cannot rule out nonoverlapping coseismic slip and afterslip. We also consider whether the overlap could be a result of not including the effects of viscoelastic deformation in our models. We therefore ran some simple tests to see whether including viscoelastic processes in our models could result in a change in the estimated location of afterslip or just the amplitude. To run these tests, we produced a suite of viscoelastic forward models using the

FIGURE 2.7: (a) Inversion with less smoothing for both coseismic (GPS and InSAR) and afterslip. (b) The same as Figure 7a but with coseismic slip based on GPS only. These inversion results with less smoothing also show partial overlap, suggesting that the overlap is not due to an artifact of the inversion because of the smoothing factor.

FIGURE 2.8: Inversion result when afterslip is constrained not to slip within the coseismic rupture area. The inversion result gives a relatively poor fit at stations SLBU and SMGY, reflecting that the data do seem to prefer afterslip within the rupture area.

open-source program Relax (version 1.0.7) (Barbot and Fialko, 2010a,b; Barbot et al., 2009) with a linear rheology. We used lithospheric thicknesses of 60 km (Lubis et al., 2013) and 100 km (Kumar et al., 2007), with three different mantle viscosities; 10^{17} , 10^{18} , and 10^{19} Pa s. We removed the displacements predicted by these models from the GPS displacements then examined whether the residual vectors indicate a change in orientation compared to the originals. We assume that if the direction of the corrected displacements does not change, the estimated location of the afterslip would not change in the case that we were able to accurately remove the viscoelastic effects. The results, Figure 2.9, show that only the residual vector at station SMGY changes its direction compared to the original but still within the measurement error. This test implies that even after considering a viscoelastic component, our estimated location for the afterslip should be relatively robust and would still partially overlap the updip region of the coseismic rupture area. It also highlights that in addition to the uncertainty imposed by the use of spatial smoothing, the estimated amplitude of the afterslip signal is not robust for our models.

FIGURE 2.9: Comparison between original postseismic displacements from GPS (black vectors) and residuals with viscoelastic deformation removed (colors). The removed contribution of the viscoelastic deformation is modeled using a lithospheric thickness of 60 km (left) and 100 km (right) with three different viscosities in each: 10^{17} (red), 10^{18} (green), and 10^{19} (blue) Pa s.

2.5 Resolution Tests

We performed checkerboard tests to assess the resolution limits of our data sets. We used checkerboard-like slip distributions (Figure 2.10a) to generate synthetic surface displacements and then inverted these synthetic displacements using the same smoothing factor as that employed in the real data inversion. The test with only the GPS data set (Figure 2.10b) shows good resolution on the shallower portion of the megathrust, particularly under the islands where at least two GPS stations are available in the along-dip direction. In addition, the combination of GPS and InSAR data sets provides much better resolution (Figure 2.10c).

We also conducted other tests to investigate how well slip in key coseismic and afterslip regions is recovered. We defined a synthetic rupture which has uniform slip of 1 m, located in coseismic (Figure 2.11a) as well as updip and downdip afterslip (Figure 2.11c) areas, then we used this uniform slip to calculate synthetic surface displacements. Similar to the procedure we applied in the checkerboard tests, we made sure that these synthetic displacements were inverted using the same smoothing factor as that used in the real data inversion. Our results show that although the inverted slip is smoothed out, most of the slip magnitude in coseismic

FIGURE 2.10: Checkerboard resolution tests for GPS only and combination of GPS and InSAR data sets. (a) Synthetic input slip with alternating 0 or 1 m of dip-slip motion. (b) Afterslip resolution for the GPS data set only (triangles). (c) Coseismic resolution for combination of GPS and InSAR data sets. The tests show that the addition of InSAR data provides much better recovery of the input slip pattern.

(Figure 2.11b) and afterslip (Figure 2.11d) regions is well recovered. These tests show that the coseismic and afterslip models fall within areas of reasonably good resolution, suggesting that our estimated slip distributions for coseismic rupture and afterslip are reasonably robust.

2.6 Numerical Simulations to Investigate Overlap of Coseismic Slip and Afterslip

We explore why we might expect overlap between coseismic slip and afterslip using a conceptual model based on rate-and-state friction laws (Dieterich, 1979; Marone, 1998; Rice and Ruina, 1983; Ruina, 1983). The dependence on slip rate and the state variable is controlled by the dimensionless parameters a and b , respectively. The state evolution can be described by the aging law (Dieterich, 1979):

$$\dot{\theta} = 1 - \frac{V\theta}{L} \quad (2.1)$$

where V is the slip rate, θ is the state variable for rate-and-state friction, and L is the characteristic weakening distance which controls the rate of weakening that occurs during frictional instabilities (Lapusta and Barbot, 2012).

FIGURE 2.11: Assessment of slip recovery in coseismic and afterslip rupture areas. Left column shows synthetic input slip for coseismic (a) and afterslip (c) rupture areas. The tests show that the slip distribution in coseismic (b) and afterslip (d) rupture areas is well estimated.

We consider the dynamics of frictional resistance of a fault plane (24 m both along strike and dip) with a velocity-strengthening (VS , $a - b > 0$) area surrounding a circular ($R = 3$ m) velocity-weakening (VW , $a - b < 0$) asperity (Figure 2.12a). To simulate the slip evolution on the fault, we use the fully dynamic Boundary Integral Cycle of Earthquakes (BICycle) program (Barbot et al., 2012; Jiang and Lapusta, 2016; Lapusta and Liu, 2009; Noda et al., 2013; Veedu and Barbot, 2016). The friction parameter ($a = 0.01$), the plate loading rate ($V_{pl} = 3$ cm/year), the effective confining pressure ($\bar{\sigma} = 50$ MPa), and the rigidity ($G = 30$ GPa) are considered uniform. The stress drop in the velocity-weakening area is strongly controlled by the combination of parameters $(b - a)\bar{\sigma}$, and the dynamics of fault

slip are unstable when the asperity size is greater than the critical nucleation size (Dieterich, 1979; Marone, 1998; Rice and Ruina, 1983; Rubin and Ampuero, 2005; Ruina, 1983):

$$h^* = \frac{\pi G b L}{2(b - a)^2 \bar{\sigma}} \quad (2.2)$$

We present three models with different parameters $(b - a)\bar{\sigma}$ and critical nucleation size h^* (Table 2.1) to illustrate the effect of these parameters on the extent of the overlap. We consider the coseismic interval as the period during which significant energy is radiated, that is, for slip velocities that exceed the critical slip rate (Noda and Hori, 2014):

$$V_{dyn} = 2V_s \frac{a\bar{\sigma}}{G} \quad (2.3)$$

where V_s is a shear wave speed of 3,000 m/s. In all cases, we ignore the first two events of the simulation to mitigate the impact of the choice of initial conditions.

TABLE 2.1: Rate-and-state friction parameters used in dynamic simulations and resulting average stress drop

Model	L (μm)	b	h^* (m)	$(b-a)\bar{\sigma}$ (MPa)	stress drop (MPa)
1	2	0.014	1.6	0.2	3
2	5.5	0.02	1.0	0.5	6
3	20	0.05	0.6	2	15

In all the models presented here (Figure 2.12), the overlap is caused by the coseismic slip extending into the velocity-strengthening regions, not by afterslip propagating into the velocity-weakening region. For reference, the model with the largest h^* in the VW asperity and the smallest parameter $(b - a)\bar{\sigma}$ produces a rupture that propagates little into the VS regions (model 1), resulting in no significant overlap between coseismic and postseismic slip. Models with decreasing values of h^* and larger parameter $(b - a)\bar{\sigma}$ result in more energetic ruptures that propagate farther into the velocity-strengthening regions (models 2 and 3). In model 2, the coseismic slip extends to a radius that is 50% larger than the velocity-weakening region in the along-strike direction. In model 3, which has the lowest value of h^* and the largest parameter $(b - a)\bar{\sigma}$, the rupture extends around twice the width of the velocity-weakening region. We estimate the average stress drop of the simulated

FIGURE 2.12: (a) A velocity-strengthening (VS , $a - b > 0$) fault plane contains a single velocity-weakening (VW , $a - b < 0$) asperity at the center of the fault. We simulate slip evolution on this fault in three different cases with different $(b - a)\bar{\sigma}$ and critical nucleation size h^* . We then plot cross sections of coseismic and afterslip profiles along the dashed lines. (b) Yellow area shows normalized coseismic slip, and blue lines are plotted afterslip profiles for selected times listed in hours after the coseismic event. Hatched area denotes overlap between the coseismic and postseismic slips. The VW asperity becomes more unstable from model 1 to model 3 as $(b - a)\bar{\sigma}$ increases and h^* decreases, leading to increasing average stress drops ($\Delta\tau$, calculated using the spatial average of stress drop distribution approach of Noda et al., 2013). These conditions result in coseismic slip that propagates farther to the VS region, causing larger overlap between coseismic slip and afterslip.

ruptures (Table 2.1) using the method of Noda et al. (2013) whereby the stress change is weighted by a quantity proportional to the local coseismic slip. When the rupture extends over both velocity-weakening and velocity-strengthening regions, the rupture produces a static stress drop that is an average of negative stress change in the velocity-weakening regions and positive stress change in the velocity-strengthening regions. Hence, the static stress drop associated with the rupture is reduced compared to that of ruptures confined to velocity-weakening regions. As a result, the static stress drop estimated from the spatial distribution of coseismic slip may not always be representative of the frictional properties of the velocity-weakening regions. The excursion of the elastodynamic rupture into the velocity-strengthening region is facilitated in some cases by the rupture directivity. The direction of propagation is updip in models 2 and 3, and we also find more overlap in the updip direction.

2.7 Discussion

The 2008 M_w 7.2 North Pagai earthquake ruptured the middle section of the Mentawai patch only partially. Possible explanations for the partial rupture include the following. First, Philibosian et al. (2014, 2017) suggest that heterogeneities in fault frictional properties in the Mentawai patch act as barriers to rupture, and thus the Mentawai patch persistently breaks in a sequence of partial ruptures. Second, Hubbard et al. (2016) and Qiu et al. (2016) show that local fault morphology can result in partial rupture of fully locked patches; they show that the rupture that generated the 2015 Gorkha (Nepal) earthquake may have been bounded on all sides by ramps on the fault and that the halting effect of geometry on rupture propagation can be predicted by earthquake cycle models. It is possible that the 2008 event was also bounded by morphological features on the slab, but we do not currently have the information on slab geometry needed to assess this. Third, Michel et al. (2017) suggest that partial rupture is possible even without structural control; they show that numerical simulations using a simple fault can produce partial ruptures if the ratio of the nucleation size and width of the velocity-weakening patch is smaller than ~ 0.08 . Distinguishing the true mechanism from among these possible explanations will require additional data such as a high-quality seismic

network, high-resolution bathymetry, and seismic reflection data to reveal more details about the subducting slab.

We show that the estimated afterslip following the 2008 event partially overlaps the updip region of its coseismic rupture area. Similar overlapping cases were also inferred for afterslip following the 2011 Tohoku-Oki earthquake (Johnson et al., 2012, 2016), the 2010 Maule earthquake (Bedford et al., 2013), the 2015 Illapel earthquake (Barnhart et al., 2016; Shrivastava et al., 2016), the 2007 Bengkulu earthquake (Tsang et al., 2016), and the 2010 Mentawai earthquake (Feng et al., 2016). In the Mentawai patch, afterslip following the recent earthquakes does seem to have a preference to overlap. In Figure 2.13, we plot slip distributions for all the recent earthquakes in the Mentawai patch and their afterslip. A moderate amount of afterslip ($\sim 1 - 2$ m) following the 2007 event is estimated to overlap the 2007 mainshock rupture area (Tsang et al., 2016), and the estimated afterslip following the 2010 event partially overlaps the 2010 coseismic area, the mainshock and afterslip of the 2008 event, and the area that slipped during the mainshock and afterslip of the 2007 Bengkulu earthquake (Feng et al., 2016).

The models presented in Figure 2.12 corresponding to a single asperity embedded in a velocity-strengthening region represent a conceptual framework to understand the dynamics of the Pagai earthquake sequence, and in particular how the coseismic and afterslip can overlap to such a degree. However, the geometry of the megathrust (Qiu et al., 2016), the spatial distribution of frictional properties (Noda et al., 2013), and the complex history of faulting in the region (Feng et al., 2016; Tsang et al., 2016; Wiseman and Bürgmann, 2011) complicate the mechanical behavior. In addition, some of the overlap in the coseismic and postseismic slip may be due to the smoothing applied in our geodetic inversion. Despite these caveats, we interpret the earthquake sequence of the Pagai earthquake in light of the simplified rate-and-state models presented above (Figure 2.12). The stress drop of the Pagai earthquake, estimated in the range 1 - 2 MPa, is relatively low and is probably the result of averaging stress change in velocity-strengthening and velocity-weakening coseismic slip areas. The extent of the coseismic rupture that does not overlap with the afterslip distribution may represent the velocity-weakening region near Pagai, and the critical nucleation size must be significantly smaller. Similar to our simplified models, we observe more overlap in the direction of rupture propagation, which may be updip in this case as inferred by high-rate GPS solutions in Figure 2.4.

FIGURE 2.13: Slip distributions of the recent earthquakes in the Mentawai patch are plotted as 1 m slip interval black contour lines. Afterslip of the 2007 event (background color) (Tsang et al., 2016) overlaps with the 2007 mainshock area. To the west of Pagai Islands, afterslip of the 2010 event (outlined with black solid line) (Feng et al., 2016) partially overlaps the 2010 coseismic area, the mainshock and afterslip of the 2008 event (outlined with black-dashed line), and also the area that slipped during the mainshock and afterslip of the 2007 event. Fracture zones are from Wessel et al. (2015).

The in situ conditions in the Mentawai patch that promote this behavior are unclear. However, it is possible that subducted topographic highs, such as a seamount (Singh et al., 2011) or a fracture zone (Graindorge et al., 2008; Lange et al., 2010), may cause heterogeneous stresses that result in locally large values of $(b - a)\bar{\sigma}$ (Wang and Bilek, 2014). Then resulting smaller but more energetic ruptures would likely propagate further into the surrounding velocity-strengthening regions.

2.8 Conclusion

The 2008 M_w 7.2 North Pagai earthquake partially ruptured a small area in the middle of the Mentawai patch, where interseismic coupling has been estimated to be high. We use GPS and InSAR data to develop a coseismic model and 5.6 years of cumulative GPS displacements to develop an afterslip model. The results suggest that the estimated afterslip partially overlaps the coseismic rupture, and our rate-and-state models imply that part of the coseismic rupture that does not overlap with the afterslip may represent a velocity-weakening region, while the overlapping part may represent a velocity-strengthening region.

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Chapter 3

Structural controls on rupture extent of recent Sumatran Fault Zone earthquakes, Indonesia

Abstract

We present geodetically derived coseismic slip models for four $M_w \sim 6$ strike-slip earthquakes that struck the mainland island of Sumatra, Indonesia, between 2007 and 2016. Three of these earthquakes ruptured segments of the Sumatran Fault Zone: the 2007 M_w 6.3 and 6.4 Lake Singkarak doublet, and the 2009 M_w 6.6 Dikit event. The fourth earthquake, the 2016 M_w 6.6 Pidie Jaya event, ruptured an unmapped active fault in a region of historically low seismicity near the northern coast of Aceh. Our results suggest that: (1) the estimated rupture extents of the 2007 Lake Singkarak doublet and 2009 Dikit event were limited by geological structures such as step-overs that define segment boundaries; (2) the 2016 Pidie Jaya event did not produce surface rupture despite the fact that it produced large shallow slip of more than 2.5 m; (3) the remaining seismic gaps on the Sumatran Fault Zone are primarily located between the equator and northern Sumatra. We strongly recommend that seismic hazard assessment for densely populated areas in Sumatra include threats not only from the Sumatran Fault Zone but also from additional potentially active faults outside of the main Sumatran Fault Zone.

3.1 Introduction

Active faulting on the island of Sumatra, Indonesia, primarily occurs along the 1700-km-long Sumatran Fault Zone (SFZ) (Genrich et al., 2000; Sieh and Natawidjaja, 2000). This zone of dominantly right-lateral strike-slip faults accommodates a significant portion of the trench-parallel component of the oblique plate convergence between the Indo-Australian and Sunda plates (Fitch, 1972; McCaffrey, 2009), at an average slip rate of $\sim 15 - 16$ mm/year along much of its length (Bradley et al., 2017; Natawidjaja et al., 2017). The SFZ can be divided into 19 segments with at least 12 separated by ~ 3 to 12-km-wide step-overs (Sieh and Natawidjaja, 2000), which may limit the rupture area that can break in a single event (Segall and Pollard, 1980; Wesnousky, 2006). However, due to their shallow depths and proximity to dense population centers located within fault-bounded valleys, earthquakes generated by even the shortest SFZ segment still pose a significant threat to local residents and infrastructure.

Concerns about the potential seismic hazard of the SFZ have escalated after a series of $M_w > 8$ earthquakes (the 26 December 2004 M_w 9.2 Sumatra-Andaman event, the 28 March 2005 M_w 8.6 Nias-Simeulue event, and the 12 September 2007 M_w 8.4 Bengkulu event) were generated by the Sunda megathrust offshore Sumatra, starting in 2004. These large megathrust events increased stress loading on the SFZ (Cattin et al., 2009; Nalbant et al., 2005; Sørensen and Atakan, 2008), and likely triggered more frequent moderate seismicity along the SFZ, including five $6 \leq M < 7$ earthquakes: the 2007 M_w 6.4 and 6.3 earthquakes (Daryono et al., 2012; Nakano et al., 2010) that caused 69 fatalities and more than 100 injuries (hereafter referred to as the 2007 Lake Singkarak doublet); the 2008 M_w 6.0 Toru earthquake that broke the Toru segment (Hurukawa et al., 2014); the 2009 M_w 6.6 earthquake that broke the Dikit seismic gap (Hurukawa et al., 2014) (hereafter referred to as the 2009 Dikit event); and the January 2013 M_w 6.1 earthquake that ruptured the Aceh creeping segment (Ito et al., 2016) (Figure 3.1). The 2008 Toru, the 2009 Dikit, and the January 2013 Aceh events did not cause casualties. Of these events, the 2008 Toru event was not detected geodetically, and the January 2013 Aceh event was recorded by eight GPS stations, but only two stations on one side of the fault recorded displacements that are large enough to be above the data uncertainties (Ito et al., 2016); hence they are excluded from our study (See Text B1 for more information).

FIGURE 3.1: (a) We study three $M_w \sim 6$ strike-slip earthquakes (blue dots with plus symbols inside) that ruptured the SFZ (drawn as a multicolor line based on Sieh and Natawidjaja (2000)), and one (blue dot with a cross symbol inside) that ruptured a blind fault near the northern coastal region of Aceh. Red stars and colored contours represent epicenter location and areas of 1 m coseismic slip for three great earthquakes ($M_w > 8$) on the Sunda megathrust (Chlieh et al., 2007; Konca et al., 2007; Tsang et al., 2016). The other colored dots are relocated shallow (non-megathrust) earthquakes during the period 1960 - 2009 (Pesicek et al., 2010) and 2009 - 2015 (Nugraha et al., 2018). Plate convergence rates in inset map are from DeMets et al. (2010). (b) Block modelling result based on an updated decadal GPS velocity field, suggesting a nearly constant slip rate along the SFZ of ~ 15 mm/year (Bradley et al., 2017). (c) The occurrence of large earthquakes on the Sunda megathrust (red stars with dashed lines showing the estimated length of the earthquake ruptures) has increased stress loading on the SFZ (Cattin et al., 2009; Nalbant et al., 2005; Sørensen and Atakan, 2008) and hence triggered more seismic activity along the SFZ. Vertical axis reflects distance along the SFZ from the equator.

Earthquakes produced by rupture of faults outside the SFZ pose an additional hazard, with recent examples being the July 2013 M_w 6.1 Tanah Gayo earthquake (Daryono and Tohari, 2016) and the 2016 M_w 6.6 Pidie Jaya earthquake (Muzli et al., 2018). The Tanah Gayo event ruptured a previously unidentified active fault in Aceh, ~ 30 km away from the nearest SFZ segment. This event triggered landslides and liquefaction, and killed six people (Daryono and Tohari, 2016). Three years later, the Pidie Jaya event ruptured another previously unmapped active fault near the coast of northern Aceh, ~ 50 km away from the nearest SFZ segment (Muzli et al., 2018). This event killed more than 100 people, and thousands of buildings collapsed partially or completely. The 2013 Tanah Gayo event was recorded by five GPS stations, but only one station recorded displacements that are large enough to be above the data uncertainties (Ito et al., 2016); it is therefore excluded from our study.

To better assess seismic hazard potential of earthquakes in mainland Sumatra, published studies have focused on paleo-earthquakes (Bellier et al., 1997), epicenter relocations (Hurukawa et al., 2014; Newcomb and McCann, 1987; Nugraha et al., 2018; Pesicek et al., 2010), aftershock distributions (Muzli et al., 2018; Widiwijayanti et al., 1996), field observations of surface ruptures (Daryono et al., 2012; Daryono and Tohari, 2016; Untung et al., 1985), deterministic and probabilistic hazard assessments (Natawidjaja and Triyoso, 2007), active fault mapping (Fernández-Blanco et al., 2016; Muksin et al., 2019; Natawidjaja et al., 2017; Sieh and Natawidjaja, 2000; Weller et al., 2012), fault slip rates (Bellier and Sebrier, 1995; Bradley et al., 2017; Genrich et al., 2000; Ito et al., 2012; Natawidjaja et al., 2017; Prawirodirdjo et al., 2000; Tong et al., 2018), in-situ stress analysis (Sahara et al., 2018), source characteristics of ruptures (Duquesnoy et al., 1996; Gunawan et al., 2018; Ito et al., 2016; Prawirodirdjo et al., 2000; Reid, 1913) and post-seismic processes (Gunawan et al., 2019). However, none of these studies used well-constrained near-field geodetic data to address the seismogenic depth and spatial distribution of coseismic slip, or to study the nature of the tectonic features that control earthquake rupture sizes along the SFZ.

Here, we use observations of coseismic deformation obtained from InSAR data to investigate the seismogenic depth and spatial distribution of coseismic slip associated with the 2007 Lake Singkarak doublet, the 2009 Dikit, and the 2016 Pidie Jaya events. We find that: (1) the terminations of rupture along SFZ segments

commonly coincide with structural step-overs, (2) the 2016 Pidie Jaya event did not produce surface rupture despite the fact that it produced large shallow slip of more than 2.5 m, and (3) the SFZ has several important remaining seismic gaps that have no historical rupture recorded since 1835.

3.2 Geodetic observational data and fault modelling methodology

We used InSAR observations from the Japanese Aerospace Agency’s (JAXA) Advanced Land Observing Satellites (ALOS-1 and -2) to study the 2007 Lake Singkarak doublet, the 2009 Dikit, and the 2016 Pidie Jaya events ([Table B.1](#)). For the 2016 Pidie Jaya event, we used additional InSAR observations from the Copernicus Sentinel-1 satellites operated by the European Space Agency (ESA). For ALOS-1 and -2 images, we produced coseismic interferograms using the Generic Mapping Tools Synthetic Aperture Radar (GMTSAR) software ([Sandwell et al., 2011](#)). We formed the interferograms from alignment of Single-Look Complex (SLC) images to get phase difference solutions. To remove topographic contributions to the phase, we used Shuttle Radar Topography Mission global 30 m DEM (SRTMGL1) data ([NASA JPL, 2013](#)). We filtered the phase using an adaptive Goldstein filter ([Goldstein and Werner, 1998](#)), and unwrapped the phase using the Statistical-Cost, Network-Flow Algorithm of Phase Unwrapping (SNAPHU) algorithm ([Chen and Zebker, 2000](#)), with an improved algorithm for interpolating incoherent areas ([Lindsey et al., 2015](#); [Shanker and Zebker, 2009](#)) and recovering near-fault phase ([Xu et al., 2016](#)). Finally, we applied a model-assisted phase unwrapping approach to correct unwrapping errors near the fault trace of the 2009 interferograms ([Figure B.1](#)). In this approach, we generated synthetic coseismic interferograms based on a uniform slip model, then subtracted the synthetic interferograms from the original interferograms, to reduce fringe rate near the fault trace so that the fringes can be unwrapped correctly. The residuals of wrapped phase were then unwrapped using SNAPHU, and the results were then added back to the synthetic deformation to obtain the coseismic deformation correctly.

For the Sentinel-1A images, we processed each burst independently, and then merged all bursts into one interferogram after correcting for phase jumps between

bursts using burst-overlap interferometry (Jiang et al., 2017). We multilooked the interferogram by a factor of 23 in range and 6 in azimuth, resulting in a ground pixel size of ~ 90 m. Then, we unwrapped selected pixels with coherence larger than 0.2 using SNAPHU (Chen and Zebker, 2000). For all the ALOS and Sentinel unwrapped images, we removed the best-fitting plane to reduce the effects of orbital errors and possible ionospheric distortions. Lastly, we down-sampled all the unwrapped results using the quadtree method (Jónsson et al., 2002), reducing each of the unwrapped results to less than 1000 points with typical data point spacing varying between 0.7 - 3 km.

We modeled each earthquake using planar dislocations embedded in a homogeneous elastic half-space (Okada, 1992), and used a shear modulus of 30 GPa for moment magnitude calculation. First, we performed a nonlinear inversion assuming uniform slip on a single rectangular fault, to search for the best-fit solution of unknown fault parameters (longitude, latitude, and depth associated with the upper left corner of the rectangular fault, strike, dip and rake) based on our down-sampled InSAR observations. For each event, we fixed the length and width of the rectangular fault based on empirical relationships of earthquake source parameters (Wells and Coppersmith, 1994) (Table B.2); our test shows that fixing the length and width results in similar results to the results without fixing them (See Text B2 for more information). We then determined the full probability distributions of the unknown fault parameters using a Monte Carlo-type slice sampling algorithm (Neal, 2003), and selected the preferred solutions based on the mean of marginal probability distributions with 95% confidence interval bounds.

Second, we conducted a linear inversion to estimate the coseismic slip distribution for each event. In this step, we placed the fault models touching the surface and fixed the strikes and dips to the values obtained in the first step. For the 2007 Lake Singkarak doublet, we ran additional slip models where the dips of the fault models were fixed to seismological solutions of Nakano et al. (2010) (Table B.2). We assumed the initial rake to the values obtained in the first step but allowed the rake to vary within 45° , and expanded the fault models size along strike and dip to give space for distributed slip model. For the 2007 Lake Singkarak doublet, we allow the fault models to have some overlap in the northern end of Lake Singkarak, following the overlap from the mapped fault trace (Figure 3.2). We discretized the

FIGURE 3.2: The 2007 M_w 6.3 and 6.4 Lake Singkarak doublet ruptured the Sumani and Sianok segments of the SFZ, which are separated by a dilatational step-over around lake Singkarak. The M_w 6.4 event occurred first and two hours later followed by the second M_w 6.3 event. Dashed rectangle shows the coverage of the InSAR frame that documented coseismic offsets. Epicenter location and its associated focal mechanisms are from the global CMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). Epicenter locations from other catalogues are located even further away from the ruptured faults (Figure B.4). Historical events and aftershocks are from Pesicek et al. (2010). Faults segments are from Sieh and Natawidjaja (2000). Plate convergence rates in inset map are from DeMets et al. (2010). The solid black lines are the surface projection of the two inverted planar fault models with their subsurface outlines drawn as dashed lines. Transparent arrows represent sense of right-lateral strike-slip fault.

fault into small rectangular sub-patches approximately 2 by 2 km, applied a second-derivative Laplacian smoothing (Du et al., 1992; Harris and Segall, 1987; Jónsson et al., 2002) to avoid abrupt variations in slip estimation, determined a smoothing factor, which controls a tradeoff between slip roughness and model misfit, based on an L-curve criterion (Aster et al., 2018) (Figure B.2), and used data variance estimated by fitting a functional form of:

$$C_{i,j} = \sigma^2 e^{-\left(\frac{d(i,j)}{\lambda}\right)} \quad (3.1)$$

on covariograms inferred from non-deforming regions in the InSAR data (Figure B.3), where σ^2 is the maximum variance, $d(i,j)$ is the distance between pixels i and j , and λ is the characteristic length scale of maximum variance (Jolivet et al., 2012; Sudhaus and Jónsson, 2009). Then, we compared the data and model misfit by using percentage of variance reduction p , defined as:

$$p = \left[1 - \frac{(Gm - d)^2}{d^2} \right] \times 100 \% \quad (3.2)$$

where G is the Green's matrix, m is our model and d is our InSAR observations. For the 2007 Lake Singkarak doublet, we do not impose boundary constraints other than smoothing at the end of the fault segments, and do not attempt to model the inelastic deformation between the faults.

3.3 Coseismic slip modelling results

3.3.1 The 6 March 2007 M_w 6.4 and M_w 6.3 Lake Singkarak doublet

The M_w 6.4 earthquake ruptured the northern part of the Sumani segment on 6 March 2007; two hours later and ~ 15 km to the northwest, a M_w 6.3 event ruptured the neighboring southern part of the Sianok segment (Figure 3.2). The northern end of the Sumani segment and the southern end of the Sianok segment run parallel to each other for ~ 20 km, separated by Lake Singkarak and at least a 3-km-wide step-over (Figure 3.2). Field investigations by Daryono et al. (2012)

FIGURE 3.3: (a) Both field investigations (white dots) and LOS data indicate right-lateral deformation due to the 2007 Lake Singkarak doublet. North of the lake, the LOS data show clear near-fault displacements over a distance up to 10 km. Radar decorrelation along the flank of Marapi volcano east of the fault trace obscures the northern limit of surface rupture. Further northwestward, the LOS data do not show clear evidence of right-lateral deformation. South of the lake, the LOS data near the fault are masked out due to very large values associated with unwrapping errors. We draw two cross sections perpendicular to the fault segments to compare the observed LOS displacement with our model. (b) The cross sections show that the first and second events produced surface rupture of similar amplitude, and our models reproduce the observed LOS displacement well.

revealed that surface rupture extended for a total length of ~ 44 km (Figure 3.3a). South of the lake, surface rupture over a distance up to 22 km produced right-lateral offsets of 8 - 41 cm. The surface rupture included a location with ~ 23 cm of uplift on the western side of the fault. North of the lake, surface rupture observed over a distance up to 22.5 km and was indicated by right-lateral offsets of ~ 46 cm, with a dip slip component only along the northern part of the surface rupture. In addition to field observations, waveform analyses of the events suggest that the first rupture initiated at the northern end of the Sumani segment and propagated southeastward, while the second rupture started close to the initiation of the first rupture, but propagated northwestward (Nakano et al., 2010). The waveform analyses also suggest that the first and second events ruptured a near-vertical strike-slip fault dipping 80° to the southwest.

Our InSAR line-of-sight (LOS) observations, which span the first three months after the event (Table B.1), indicate right-lateral deformation along the Sianok and Sumani segments (Figure 3.3a). North of the lake, the LOS data show clear near-fault displacements over a distance up to 10 km. Decorrelation of pixels along the Marapi volcano flank, east of the fault trace, obscures the northern limit of surface rupture. Further northwestward, the LOS data do not show right-lateral deformation. South of the lake, we masked out the LOS data near the fault due to very large values associated with unwrapping errors.

Our results show that although we constrained the Sumani segment to be dipping southwest (following the solutions from global CMT, Nakano et al. (2010) and Wong et al. (2019)), the probability distribution for the dip peaked at a northeast-dipping fault (Figure B.5b), indicating a change in dip direction from southwest along the Sianok segment to northeast along the Sumani segment (Figure B.5). For the coseismic models, while the Sumani segment dipping northeast (Figure B.6) results in more variance reduction (by approximately 5% than if the segment dipping southwest, Figure 3.4), we prefer the coseismic models using the Sumani segment with a southwest dip, because there is no independent evidence (such as relocated aftershock distribution) for the change in dip direction. In addition, the LOS displacements that are recorded only along the descending track, may not be providing enough resolution to constrain the dip direction. Overall, whether the Sumani segment dips southwest (Figure 3.4) or northeast (Figure B.6), this ambiguity does not affect our estimated rupture extent and pattern.

Our preferred coseismic models estimate that most slip occurred at depths of 0 - 15 km, with a maximum slip of ~ 40 cm at the surface along the Sianok segment, and ~ 60 cm along the Sumani segment (Figure 3.4c-d). The models suggest that both segments had nonzero slip in their overlapping area; this could reflect the presence of some un-modelled inelastic deformation or slip on linking faults between the two segments. While the model for the Sumani segment reproduces the global moment-tensor solution, the model for the Sianok segment does not, due to the slip being mostly oblique (Figure 3.4c-d). The unmodeled residuals are potentially due to small unwrapping errors near the fault and the impact of atmospheric delays. Our preferred coseismic models result in more than 80% variance reduction (Figure 3.4e).

We compare the shallow slip component of the coseismic models with observed fault offsets along the Sianok and Sumani segments (Figure 3.4a-b). Along the Sianok segment, the shallow slip model and the observed fault offsets are in good agreement. Along the Sumani segment, our estimated shallow slip model is fairly consistent with the observed fault offsets except at two locations: along the lake where the estimated shallow slip is larger, and just south of the lake where the estimated shallow slip is smaller. Along the lake, the larger shallow slip is possibly due to the inclusion of any off-fault deformation into our simplified straight fault (Wang et al., 2014) (Figure 3.2). South of the lake, two locations within a distance of ~ 1.25 km have very different observed displacements, suggesting that the region experienced highly variable slip on the fault. These two locations are within the masked region in our LOS data, likely resulting in smaller shallow slip in the model compared to the field observations. The total moment release of our coseismic models is equivalent to M_w 6.5 for the first event, and M_w 6.4 for the second event.

3.3.2 The 1 October 2009 M_w 6.6 Dikit event

The M_w 6.6 earthquake was produced by rupture of the northern part of the Dikit segment on 1 October 2009 (Figure 3.5). With only 10 $M_w \leq 5$ earthquakes recorded over the past 130 years (Pesicek et al., 2010), this segment had been classified as a seismic gap along the SFZ (Hurukawa et al., 2014; Natawidjaja and Triyoso, 2007). The Dikit segment is limited by a ~ 3 -km-wide step-over between its southeastern end and the Ketaun segment (Figure 3.5), and an ~ 11 -km-wide step-over between

FIGURE 3.4: (a - d) Our preferred coseismic models of the 2007 doublet estimate that most slip occurred at depths of 0 - 15 km, with maximum slip at the surface. The coseismic models show no slip at the hypocenter depths (colored stars downloaded from global CMT catalog). We interpret the reported hypocenter depths may not be resolved well due to a lack of local seismic stations. White arrows indicate slip direction on the western side of the fault. The shallow slip of the coseismic models is in good agreement with the observed fault offsets, except at the region by the lake which could be explained by inclusion of any off-fault deformation into our simplified straight fault. (e) Observed, modeled, and residual LOS velocities with 0.7 km typical data point spacing. Our preferred coseismic models result in more than 80% variance reduction.

FIGURE 3.5: The 2009 M_w 6.6 Dikit event was the largest recorded earthquake to occur on the Dikit segment, one of the seismic gaps along the SFZ. While the other fault segments to the south and north produced $M_w \sim 6$ earthquakes (blue focal mechanisms), only ten $M_w \leq 5$ earthquakes (blue circles) occurred along the Dikit segment in the past 130 years (from Pesicek et al., 2010). Along this segment, the most evident geological features are step-overs at both ends of the fault, and a volcanic chain to the east of the fault. At the northwestern end, the Dikit and Siulak segments form the widest step-over (~ 11 km) along the SFZ. Dark blue lines are our newly mapped linking faults based on mapping of Shuttle Radar Topography Mission global 30 m DEM. Dashed rectangles show the InSAR frames that documented coseismic offsets. Epicenter location and its associated focal mechanism (in red) are from the global CMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). Epicenter locations from other catalogues are located even further away from the ruptured faults (Figure B.7). Historical events and aftershocks are from Pesicek et al. (2010). Faults segments are from Sieh and Natawidjaja (2000). Plate convergence rates in inset map are from DeMets et al. (2010). The solid black line is the surface projection of the inverted planar fault model with its subsurface outlines drawn as dashed line. Gray arrows represent sense of right-lateral strike-slip fault.

its northwestern end and the Siulak segment (Sieh and Natawidjaja, 2000) (Figure 3.5). To the east of the Dikit segment, a chain of volcanic edifices extends along approximately two thirds of the segment, trending almost parallel to the fault trace (Figure 3.5).

FIGURE 3.6: (a) The 2009 Dikit event produced surface rupture along a length of ~ 15 km. The surface rupture on the ascending track that had a satellite look angle of $\sim 70^\circ$ relative to the fault strike (top panel), is less obvious than the surface rupture on the descending track that had a satellite look angle of $\sim 40^\circ$ (bottom panel). This is because the look angle (black vector) along the ascending track is almost perpendicular to the strike of the slip motion. In addition, near-fault unwrapping errors in areas of low correlation could also explain the less obvious surface rupture on the ascending track. We draw three cross sections perpendicular to the fault segments to compare the observed LOS displacement with our model. (b) The cross sections show that maximum surface rupture is estimated to be ~ 60 cm along the BB' profile on the descending track (green dots), and our model reproduces the observed LOS displacement well.

Our LOS data provide surface deformation observations from both descending and ascending tracks (Figure 3.6a). Along the descending track, the LOS data show surface rupture over a distance of ~ 10 km, with a maximum LOS offset of ~ 60 cm (Figure 3.6b). Along the ascending track, the LOS data show displacements across the fault but do not show clear surface offsets, probably due to the remaining near-fault unwrapping error in an area of low correlation.

FIGURE 3.7: (a) Our preferred coseismic model for the 2009 Dikit event estimates that most slip occurred at depths of 0 - 10 km, with a maximum slip at a depth of ~ 3 km. The seismic hypocenter (red star) is from global CMT catalog. The coseismic model shows no slip at the hypocenter depth; we suggest the depth may not be resolved well due to a lack of local seismic stations. White arrows indicate slip direction on the eastern side of the fault. Inverted blue triangles are crossing locations of three profiles plotted in Figure 3.6a (b) Observed, modeled and residual LOS displacements from the two tracks, with 1.6 km typical data point spacing. Our preferred coseismic model result in more than 90% variance reduction.

We estimate that the ruptured fault segment dips $\sim 77^\circ$ toward the northeast (Figure B.8 and Table B.2), and that most slip occurred at depths of 0 - 10 km, with a maximum slip of ~ 2 m at a depth of ~ 3 km, and reproduces the global CMT moment-tensor solution (Figure 3.7a). The total moment release of our coseismic model is equivalent to M_w 6.6 and our model results in more than 90% variance reduction (Figure 3.7b).

3.3.3 The 7 December 2016 M_w 6.6 Pidie Jaya event

The M_w 6.6 earthquake on 7 December 2016 was caused by rupture of a previously unmapped active fault with a NNE-SSW strike (Muzli et al., 2018), in a region of historically low seismicity near the northern coast of Aceh (Figure 3.8). The closest historical rupture was the 1967 M_w 6.1 earthquake, ~ 10 km to the southeast (Pesicek et al., 2010) (Figure 3.8). The 1967 event and the Pidie Jaya earthquake could have been produced by the same fault, as the 1967 epicenter is not well constrained due to poor resolution of global seismic networks at the time.

Our LOS data provide surface deformation observations from both ascending and descending tracks (Figure 3.9). Along the ascending track, the surface deformation extends along a northeast-southwest trending axis, consistent with the strike of the aftershock distribution relocated by Muzli et al. (2018) (Figure 3.8). In addition, the LOS deformation is asymmetric across the fault, suggesting that the earthquake ruptured a southeast-dipping plane. Along the descending track, the Sentinel-1 satellite observed the area just one day after the mainshock. However, due to the satellite's use of C-band radar, the interferogram has low coherence in regions of steep topography or dense vegetation. Therefore, our Sentinel-1 LOS observations are limited only to low-lying regions near the epicenter and further east. In comparison, ALOS-2 descending data spanned the first two months after the mainshock. The deformation on the eastern side of the epicenter is larger by almost two times than that recorded by Sentinel-1, likely due to an additional component of postseismic displacement. Both sets of LOS data indicate (from peak displacements occurring at the coast) that a significant amount of deformation occurred offshore and this part of the rupture is therefore not resolved.

Our results show that the event ruptured an estimated 78° southeast dipping fault plane (Figure B.9 and Table B.2). Along this fault, our preferred coseismic model

FIGURE 3.8: The 2016 M_w 6.6 Pidie Jaya event (red star) struck the northern coastal region of Aceh, ~ 50 km north from the nearest segments of the Sumatran fault. Before this event, the epicentral region experienced an M_w 6.1 earthquake in 1967, ~ 12 km to the southeast (blue star) (Pesicek et al., 2010). Between these two events, most seismicity (blue circles and focal mechanisms) was associated with the Sumatran fault segments (drawn as multicolor lines), including the January 2013 M_w 6.1 earthquake in Aceh segment (Ito et al., 2016) and the July 2013 M_w 6.1 Tanah Gayo earthquake ~ 30 km from the nearest segment of the Sumatran fault (Daryono et al., 2016; Gunawan et al., 2018). Historical events are from Pesicek et al. (2010) and Nugraha et al. (2018). Aftershock distributions (M_L 1 - 4) are for the period of 14 December 2016 - 15 January 2017 (Muzli et al., 2018). Dashed rectangles show the InSAR frames that documented coseismic offsets. Epicenter location is from Muzli et al. (2010) and its associated focal mechanism is from the global CMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). Fault segments are from Sieh and Natawidjaja (2000). Plate convergence rates in inset map are from DeMets et al. (2010). The solid black line is the surface projection of the inverted planar fault model with its subsurface outlines drawn as a dashed line. Gray arrows represent sense of right-lateral strike-slip fault.

FIGURE 3.9: (a) Along the ascending track, LOS displacements show an asymmetric feature, suggesting that the earthquake ruptured on a southeast dipping plane. (b - c) Along the descending track, while the ALOS-2 data were recorded two months after the mainshock, the Sentinel-1 satellite recorded data one day after the mainshock. Thus, the larger ALOS-2 LOS deformation on the eastern side of the epicenter might be attributed to additional displacement from post-seismic process. In addition, the Sentinel-1 LOS observations experienced severe decorrelation issue (only pixels with coherence larger than 0.2 were unwrapped) and hence the LOS data cover only the regions near the epicenter and further east.

estimates that most slip occurred at depths of 2 - 12 km, with a maximum slip of ~ 3 m at a depth of ~ 6 km (Figure 3.10a). The estimated amplitude and depth of our slip is different from that estimated by Muzli et al. (2018) using teleseismic data; they estimated a maximum slip of 80 cm at a depth of ~ 10 km. The difference can partly be explained by a lack of teleseismic resolution at shallow depths. The total moment release of our coseismic model is equivalent to M_w 6.6, and our model results in $\sim 83\%$ variance reduction (Figure 3.10b). The model mostly indicates strike slip motion with some normal component in the main slip patch, but then becomes almost pure reverse slip in the northeast (Figure 3.10a), indicating the need for a SE-NW striking right-lateral or reverse faulting slip in the north. This reverse slip, together with the observation that the coseismic slip based on a single fault model cannot reproduce the global CMT moment-tensor solution (Figure 3.10a), justifies the addition of a secondary fault.

We remodel the data by adding a secondary fault based on two scenarios: 1) it is a buried SW-NE thrust fault dipping northwest, as suggested by the relocated aftershock distribution in Muzli et al. (2018), or 2) it is a SE-NW right-lateral fault dipping southwest. We optimized the dip angle of the secondary faults by trial and error in our linear inversion. The results show that the main slip patch in the first fault still has the same slip characteristics as the single fault model but without the reverse slip (Figure 3.11a), and instead the reverse slip is accommodated in the secondary fault (Figure 3.11b). However, this first scenario does not improve the moment-tensor solution (Figure 3.11c) and variance reduction (Figure 3.11d). We find that the second scenario results in the same slip characteristics and slightly improves the variance reduction, but still does not improve the moment-tensor solution (Figure B.10). Based on our modelling, although the second scenario results in more variance reduction by approximately 4%, we prefer the first scenario for the secondary fault, because it is consistent with the relocated aftershocks distribution estimated by Muzli et al. (2018). The unmodeled residuals, despite the added fault complexity, are potentially due to unwrapping errors near the fault and the impact of atmospheric delays.

All coseismic models suggest that aftershocks were mostly concentrated outside the main slip patch, and may have been triggered by the stress transfer of the coseismic slip. In addition, the coseismic slip model (Figures 3.10 - 3.11) estimates no slip at shallow depths of 0 - 2 km. The absence of slip may be due to a higher

friction coefficient (Byerlee, 1978) and a higher density of cracks (Manning and Ingebritsen, 1999) at shallow depths, all of which impede strain accumulation.

3.3.4 Summary of coseismic models

All the coseismic models we have presented above are our preferred models among many possible models that can fit the underdetermined problem. The preferred models well resolve rupture dimensions of the slip distributions in the upper 10 km, based on our checkerboard tests (Figures B.11 - B.12). However, the amplitude of the maximum slip is heavily dependent on the chosen smoothing factor. In addition, our preferred coseismic models show little to no slip at the hypocentral depths (>10 km for all the events); we suspect the hypocenter depth as estimated from seismic data may not be resolved well due to a lack of local seismic stations.

3.4 Discussion

3.4.1 Structural controls on rupture extents

Step-overs have been suggested to be the primary structural discontinuities that separate the SFZ into the most highly segmented strike-slip fault zone in the world (Sieh and Natawidjaja, 2000). Fault segmentation can act as a major structural barrier to rupture (Segall and Pollard, 1980; Wesnousky, 2006). For example, in 1926 and 1943 earthquake doublets started at the Sumani and Suliti segments, and the first event of each doublet terminated at the northwestern tip of the segments that share an ~ 4 km step-over with the adjacent segment (Daryono et al., 2012; Natawidjaja and Triyoso, 2007; Sieh and Natawidjaja, 2000) (Figure 3.12). A similar case was also observed in the 1994 M_w 6.8 Liwa earthquake, whose rupture did not even reach the Ranau lake caldera which is an old pull-apart basin produced by a step-over (Widiwijayanti et al., 1996).

Our results show that the rupture extent of the first event of the 2007 Lake Singkarak doublet terminated at the northwestern part of the Sumani segment (Figure 3.4) that shares an ~ 4 km step-over with the Sianok segment, similar to the pattern of events in 1926 and 1943 (Figure 3.12). In addition, the rupture

FIGURE 3.10: (a) Our coseismic model based on one planar fault for the Pidie Jaya event estimates that slip occurred at depths of 2 - 15 km, with a maximum slip at a depth of ~ 6 km. The coseismic model shows that aftershocks were mostly concentrated outside of the rupture area, which may be triggered by the stress transfer of the coseismic slip. White arrows indicate slip direction on the eastern side of the fault. (b) Observed, modeled and residual LOS displacements from the three tracks, with 1 km typical data point spacing along descending and 3 km along ascending. This coseismic model result in more than 80% variance reduction.

FIGURE 3.11: (a) Our preferred coseismic model based on two planar faults (the first scenario) for the Pidie Jaya event estimates that the first fault has a similar main slip patch to the previous model based on one planar fault. (b) The reverse slip on the first fault disappears, and is accommodated on this secondary fault as reverse slip. (c) A cross section along profile AA' showing that the secondary fault is buried at a depth of 8.5 km. In addition, coseismic modelling in this first scenario still does not reproduce the global CMT solution. (d) Observed, modeled, and residual LOS displacements from the three tracks. Our preferred coseismic models result in more than 80% variance reduction, except for some residuals which may be due to poor constraints on both depth and lateral variation in elastic basin structure.

FIGURE 3.12: Summary of the surface rupture lengths associated with earthquake doublets along the SFZ segments, modified from Daryono et al. (2012). The plot shows that the first event of the doublets ruptured the southern segments, and the second event ruptured the northern segments.

extent of the 2009 Dikit event also terminated ~ 7 km from the northwestern tip of the Dikit segment, that shares an ~ 11 km step-over with the Siulak segment (Figure 3.7). The termination of these ruptures at structural step-overs is likely due to the presence of complex tectonic structures and the absence of a mature localized fault, due to a combination of strike-slip and extensional deformation at these boundaries (Gürbüz, 2010). Nevertheless, although the Sumatran fault zone earthquakes have not ruptured multiple fault segments to date, multifault rupture in one earthquake is still possible given the recent examples of the 2010 Mw 7.2 El Mayor-Cucapah earthquake in Mexico (Fletcher et al., 2014; Wei et al., 2011), the 2012 Mw 8.6 Wharton Basin sequence in Indonesia (Hill et al., 2015; Wei et al., 2013), the 2016 Mw 7.9 Kaikōura earthquake in New Zealand (Hamling et al., 2017; Wang et al., 2018), and the 2018 Mw 7.5 Palu earthquake in Indonesia (Bao et al., 2019; Socquet et al., 2019).

The 2009 Dikit rupture terminated ~ 7 km from the segment's northwestern tip (Figure 3.7), near the southern edge of a structurally complex magmatic step-over that occupies an area of rugged topography south of Lake Kerinci (Figure 3.5). This part of the SFZ is characterized by intense geothermal activity (Figure 3.5) (Hochstein and Sudarman, 1993; JICA, 1989) and high temperatures (above $\sim 200^\circ$ C) at a depth of 1 - 2 km (JICA, 1989; Prastika et al., 2016; Silitonga et al., 2015).

The relative absence of deep slip during the Dikit rupture (Figure 3.7) may reflect a shallower maximum seismogenic depth associated with this high geothermal gradient (Yang et al., 2018), and an even higher heat flux in the area of the magmatic step-over could have prevented northward continuation of the rupture. Such local relationships between structural step-overs, volcanic centers, and earthquakes may be particularly important for the arc-related Sumatran Fault Zone.

3.4.2 Seismic gaps along the SFZ

The slip models presented in this paper fill in several previously identified ‘seismic gaps’ (areas without significant historical earthquakes) along the Sumatran Fault Zone (Hurukawa et al., 2014; Natawidjaja and Triyoso, 2007). The remaining seismic gaps on the SFZ are primarily located between the Equator and northern Sumatra (Figure 3.13), onshore of the large portion of the Sunda megathrust that has ruptured in the cascade of great earthquakes that initiated with the 26 December 2004 M_w 9.2 Sumatra-Andaman earthquake. The SFZ fault segments north of the Equator are significantly longer and straighter than in the highly segmented fault zone of southern Sumatra. If fault segmentation does exert a fundamental control on earthquake size along the Sumatran Fault Zone, this suggests a greater maximum magnitude and possibly a longer return period in this region. Thus, one possible explanation for the larger number of seismic gaps along the northern part of the SFZ is that the ~ 180 year historical record of earthquakes (Bellier et al., 1997) is still too short to observe these more infrequent and potentially larger events. At the same time, this implies that a more complete understanding of the relative hazard along a particular segment requires some information about its average recurrence interval; for example, recently ruptured segments could actually be more hazardous if their average recurrence interval is short.

The factors that influence damaging seismicity off the main Sumatran Fault Zone in northern Sumatra remain unclear. The 2016 M_w 6.6 Pidie Jaya event struck the northern coastline of the Aceh region, affecting a low-lying coastal area with soft soil and thick sediments (Muzli et al., 2018). Surface geomorphic expressions (Figure 3.9) lie nearly parallel with the estimated strike in our fault model, suggesting that the earthquake reactivated old tectonic features, or that the local geologic surface structure reflects previously un-recognized active tectonic processes. The

NRF2018NRF-NRFI001-21 to EH), the Earth Observatory of Singapore (EOS), the National Research Foundation of Singapore, and the Singapore Ministry of Education under the Research Centers of Excellence initiative. We are grateful to the Alaska Satellite Facility (ASF) for providing access to ALOS-1 and Sentinel-1 SAR data (<https://search.asf.alaska.edu/#/?flightDirs=>), and to JAXA for providing ALOS-2 data through two data proposals granted to Lujia Feng under the 4th and 6th Research Announcement for ALOS-2 (PI No. 1436 for RA4 and 3205 for RA6; ownership and copyright of raw data maintained by JAXA). Our coseismic slip models, line-of-sight SAR displacements from ALOS-1, ALOS-2 and Sentinel-1, and newly mapped fault coordinates are all freely available at <https://doi.org/10.21979/N9/BDOEYS>. We express our thanks to Editor Yehuda Ben-Zion, Associate Editor, and two anonymous reviewers for their comments that improved the paper. RS would like to thank Pavel Adamek for all his invaluable linguistic suggestions that greatly improved the English writing of this paper, and Rishav Mallick, Nina Lin, Muzli for insightful discussions. Figures were made using the Generic Mapping Tools (GMT) (Wessel et al., 2013) with some color maps from Crameri (2018). This work comprises EOS contribution no. 264.

Chapter 4

Cascading partial rupture of the Flores thrust during the 2018 Lombok earthquake sequence, Indonesia

Abstract

A series of four $M_w > 6$ earthquakes struck the northern region of Lombok, eastern Indonesia, in a span of three weeks from late July to mid-August 2018. The series was thought to be associated with the Flores thrust, but the exact mechanism causing the unusual earthquake series has remained elusive. Our InSAR analysis, combined with insights from seismology, indicate that the events originated at different hypocenter depths with differing fault geometries, which may explain the cascading behaviour of the events, and indicates that better imaging of active fault geometry might provide some insight into future rupture behaviour on other similar thrust systems. Our static-stress-change calculations suggest that the earlier events in the sequence played a role in promoting the later events. In addition, the second event brought the most significant impact on a nearby volcano, by causing volumetric expansion at its shallow magma plumbing system and unclamped its magma ascent zone. However, no volcanic activity has so far occurred after the earthquakes. Lastly, our damage proxy maps suggest that the second event caused the greatest damage to buildings.

4.1 Introduction

The eastern Indonesian island of Lombok is tectonically bounded by the Sunda megathrust to the south and the Flores thrust to the north (Figure 4.1). The Flores thrust is an active, back-arc thrusting zone dipping to the south, extending from Flores to the Bali Basin (Hamilton, 1979; Koulali et al., 2016; McCaffrey, 1988; McCaffrey and Nabelek, 1984, 1987; Nugroho et al., 2009; Silver et al., 1986, 1983). The slip rate of the Flores thrust, based on the block modelling results of both campaign (1993 - 2014) and continuous (2009 - 2014) GPS velocity measurements, has been estimated at ~ 26 mm/year near Flores, decreasing to ~ 10 mm/year towards the west near Lombok (Koulali et al., 2016), accumulating slip deficit that has been the source of 19 $M_w > 6$ earthquakes over the period 1976 - 2018 (U.S Geological Survey, USGS), including the 1992 M_w 7.9 Flores earthquake that generated a destructive tsunami (Beckers and Lay, 1995; Hidayat et al., 1995; Pranantyo and Cummins, 2019; Yeh et al., 1993), and the 2018 Lombok earthquake sequence.

The Lombok earthquake sequence comprised a series of shallow M_w 6.2 to 6.9 earthquakes that struck the northern region of Lombok (Figure 4.1) over three weeks in July and August 2018. The earthquake series started with an M_w 6.4 event on 28 July (hereafter referred to as “28J”), which was followed by several smaller events in the following seven days, as recorded by the regional seismic network of Indonesia’s Meteorological, Climatological, and Geophysical Agency (BMKG) (Figure 4.1). On 5 August, an M_w 6.9 event occurred 10 km to the west (hereafter referred to as “5A”). This earthquake was followed by several aftershocks in the subsequent two weeks, with the largest one being an M_w 5.9 event on 9 August. Then, on 19 August, two earthquakes of M_w 6.2 and 6.9 (hereafter referred to as “19A”) occurred within 10 hours of each other, 15 km to the east of the 28 July event. Overall, the BMKG network recorded more than 200 events of $M_w > 4$ between 28 July and 21 December 2018.

Based on the USGS catalog, all $M_w > 6$ event epicenter locations were located in regions mostly made of lahars, tuffs, and other volcanic deposits from the active Rinjani volcano (Ganas et al., 2018; Mangga et al., 1994) (Figure C.1). The fact that volcanic deposits are generally soft and could amplify seismic waves (Walter et al., 2008), and that the residential areas were developed on these volcanic

FIGURE 4.1: Four $M_w > 6$ earthquakes (colored stars; epicenter locations are from the USGS earthquake catalog) struck the northern region of Lombok in a span of three weeks from late July to mid-August in 2018. These events were followed by tens of smaller earthquakes with magnitude larger than four (purple dots), recorded by the regional seismic network of BMKG (inverted triangles). Black and blue dashed rectangles show the InSAR frames that documented co-seismic offsets. Focal mechanisms are from our moment tensor solutions.

deposits, worsened the earthquakes' impact: 515 people were killed, 7,733 were injured, 431,416 were displaced, and 76,765 buildings were damaged ([ASEAN Humanitarian Assistance \(AHA\) Centre, 2018](#)).

All the $M_w > 6$ events ruptured either a steep north-dipping fault or a shallow south-dipping fault, based on focal mechanisms generated by inversion of regional seismic waveforms ([Figure C.2](#)). A steep northward dipping fault can reasonably be excluded as it is inconsistent with the geology of the region and previous seismic reflection images ([Silver et al., 1986, 1983](#)). Thus, the Lombok sequence of shallow events was most likely generated by the south-dipping fault associated with the Flores thrust. What then remains unanswered is why the earthquakes occurred as a sequence rather than a single rupture.

In this study, we investigate the mechanism that controlled the sequence, using InSAR and seismic observations to study the underlying fault geometry and rupture

pattern of each event. For all the $M_w > 6$ events, we estimate the focal depth and best-fitting planar fault geometry, and model the coseismic slip distribution. Then, we use the slip results to compute the Coulomb stress changes that each $M_w > 6$ event likely caused at the rupture area of the next event. Finally, we use the estimated slip distributions to compute the normal stress changes and dilatational stress to evaluate the effect of each event on the nearby Rinjani volcano (Figure 4.1). In addition to studying the source mechanism, we also investigate the effect of each $M_w > 6$ event on buildings by performing quantitative analysis using damage proxy maps (DPMs). We find that: (1) the earthquake sequence originated from different hypocenter depths and ruptured separate south-dipping thrust faults with different geometries, which may explain the cascading behaviour of the events, (2) the previous events in the sequence increased Coulomb stress changes at the hypocentre location of the next events, which may promote the failure of the next events, (3) the 5A event caused volumetric expansion in the shallow magma plumbing system of Rinjani volcano and unclamped its magma ascent zone, which may potentially have an impact on its future eruptive activity, (4) the second event caused the greatest damage to buildings.

4.2 Data analysis and earthquake slip modelling methodology

4.2.1 InSAR data processing

We used InSAR observations from the European Space Agency (ESA) that provides Copernicus Sentinel-1 satellites data [2018] and the Japanese Aerospace Agency’s (JAXA) Advanced Land Observing Satellite (ALOS-2) (Table C.1) to study these events. We used the Generic Mapping Tools Synthetic Aperture Radar (GMT-SAR) software (Sandwell et al., 2011) to form the interferograms and removed topographic contributions to the phase with the Shuttle Radar Topography Mission global 30 m DEM (SRTMGL1) data (NASA JPL, 2013). We filtered the phase using an adaptive Goldstein filter (Goldstein and Werner, 1998), and unwrapped the phase using the Statistical-Cost, Network-Flow Algorithm of Phase Unwrapping (SNAPHU) algorithm (Chen and Zebker, 2000), with an improved algorithm for interpolating incoherent areas (Lindsey et al., 2015; Shanker and Zebker, 2009).

We succeeded in unwrapping all the interferograms except for the ALOS-2 phases along the descending track for the 5A and 19A events (these events' interferograms could not be unwrapped due to severe atmospheric noise). Then, we removed the atmospheric noise component that correlated with high topography (Cavalié et al., 2008) (Figure C.3), and removed the best-fitting plane to reduce the effects of orbital errors. Lastly, we down-sampled the unwrapped results using the quadtree method (Jónsson et al., 2002) to provide a smaller set of data points for inversion (Figure C.4).

4.2.2 Seismic analysis for moment tensors and hypocenter depths

We jointly modeled regional and teleseismic waves to obtain moment tensors and hypocenter depths for the $M_w > 6$ events. Although global Centroid Moment Tensor (CMT) solutions are available, we prefer to perform our own analysis using a dense network of regional stations to ensure their robustness and to place uncertainty bounds on the results. After obtaining moment tensors, we searched for the hypocenter depths using carefully selected teleseismic stations (Figure C.2). We performed the waveform modelling at higher frequencies than used in global CMT, allowing better resolution of fault geometry and hypocenter depth.

To obtain moment tensors, we modeled waveforms from regional broadband seismic stations of the BMKG network (Figure 4.1). In addition, we used waveforms from several stations of the Global Seismic Network (GSN) to improve the azimuthal coverage (Figure C.2c). We used the Cut-And-Paste (CAP) method (Zhu and Helmberger, 1996) to perform the inversion, with a 1D velocity structure extracted from the P-wave model of Simmons et al. (2012). Then, we estimated uncertainties of the moment tensors by using bootstrapping resampling, whereby a random 5% of stations were removed and a new result obtained. This resampling processes was repeated with different data subsets, such that each station is absent from one inversion. The resulting suite of models provides an estimate of the uncertainty in the focal mechanism solution (Figure 4.3a). The bootstrapping method generally underestimates uncertainties in the solutions of moment tensor inversion. However, this method can effectively emphasize the relative errors between the

moment tensor solutions (e.g., the dip angle is relatively well constrained). More comprehensive and in-depth error analysis is left for future efforts.

To further refine hypocenter depths, we used the best focal mechanisms obtained from the previous step and modeled teleseismic P-waves for the best fitting hypocenter depth. To determine accurate depths, we used only high quality broadband vertical seismograms of the direct P wave plus depth phases. There are significant azimuthal variations in waveform complexity due to the earthquake radiation patterns and source-side velocity structure; we used only stations along an azimuth between 100° to 200° (Figure C.2d) as these stations are located towards the strongest radiation of P-waves and where the depth phases sampled mainly the islands rather than the offshore basins. The best-fitting depth has the lowest RMS misfit, and we used a 10% range to estimate the uncertainty.

4.2.3 Modelling earthquake slip

We modeled the InSAR data using fault dislocations embedded in a homogeneous elastic half-space (Okada, 1992), and used a shear modulus of 30 GPa for moment magnitude calculation. For each event, we fixed the strike and dip angle of the fault to our moment tensor solution, optimized the fault depth by trial and error, discretized the fault into smaller rectangular of approximately 3 by 3 km sub-patches, then compared the data and model misfit by using the percentage of variance reduction p , defined as:

$$p = \left[1 - \frac{(Gm - d)^2}{d^2} \right] \times 100 \% \quad (4.1)$$

where G is the Green's matrix, m is our model and d is our InSAR observations. To avoid abrupt variations in slip estimation, we applied second-derivative Laplacian smoothing (Du et al., 1992; Harris and Segall, 1987; Jónsson et al., 2002), and determined a smoothing factor, which controls a tradeoff between slip roughness and model misfit, based on an L-curve criterion (Aster et al., 2018) (Figure C.5).

4.3 Quantitative analysis on building damage using damage proxy maps

We evaluate the effect of each $M_w > 6$ event on buildings. To do this, we created damage proxy maps (DPMs) and extracted damage related only to buildings. We derived damage proxy maps (DPMs) due to each $M_w > 6$ event from Copernicus Sentinel-1 data (Table C.2) by following the method of (Yun et al., 2015). We initially estimated coherence difference between a pair of interferometric synthetic aperture radar (InSAR) images before (image 1 and 2) and spanning each event (image 2 and 3) (Table C.2), using the Advanced Rapid Imaging and Analysis Singapore system (ARIA-SG; a clone of the ARIA system developed by NASA-JPL and Caltech). The estimated coherence difference with large positive values correspond to greater loss in coherence, indicating more damage is detected. Then, we determined a threshold value by comparing the estimated coherence difference with ground-truth reports of sample areas with and without damaged buildings. Thus, pixels associated with buildings that have coherence difference values greater than the threshold are considered as damaged buildings. Finally, we combined the DPMs with building footprint information from OpenStreetMap (<https://www.openstreetmap.org>) to calculate the number of buildings damaged. In the calculation, we selected buildings that intersect only with pixels indicating the highest confidence of significant damage (saturated red pixels in Figure C.6a), and excluded pixels associated with other source of incoherence such as vegetation changes (Figure C.6b) and layover/shadow effect or landslides (Figure C.6c).

4.4 Results

Our line-of-sight (LOS) data from both ascending and descending tracks show surface deformation caused by each event (Figure 4.2). The 28J event produced ~ 5 - 10 cm LOS displacements confined to the coastline close to the epicenter (Figure Figure 4.2a-d). The 5A event deformed a larger area (Figure 4.2e-g), with ~ 5 - 10 cm displacements near the epicenter to ~ 30 cm displacements along the coastline ~ 20 km west of the epicenter. The 19A events showed a different deformation pattern along the ascending and descending tracks. Along the ascending track, surface deformation is observed in a large area, with LOS displacements peaked

at ~ 30 cm along the NE coastline (Figure 4.2h and j). In comparison, along the descending track, the deformation is seen only in a small area, with LOS displacements peaking at ~ 15 cm along the NE coastline (Figure 4.2i). In addition, the 19A event also deformed the western region of Sumbawa, with LOS displacements of < 5 cm confined only along the coastline (Figure 4.2h-i). Overall, the LOS data that spanned all the events show that the 2018 sequence deformed the northern region of Lombok by at least ~ 15 cm of LOS displacements (Figure 4.2k-l). In addition, the cumulative LOS data also show that the surface deformation extends southward on both the western and eastern sides of the volcano (Figure 4.2k-l). However, the surface deformation is observed north of the volcano, indicating that the volcano may act as a barrier for the slip to propagate deeper.

FIGURE 4.2: Our InSAR results show LOS displacements due to each event (a - j) and the cumulative events (k - l). Each event deformed the northern region of Lombok by at least ~ 10 cm of LOS displacements. Positive LOS values represent motion toward the satellite. Contour intervals of LOS displacements are plotted every 5 cm.

Our seismic analysis suggests that the events originated at different hypocenter depths, and ruptured south-dipping thrust faults with different geometries (Figure 4.3a and Table C.3). The 28J event, with an estimated hypocenter depth of 10 ± 2 km, ruptured a fault trending NNE and dipping 15° . The 5A event, with an estimated hypocenter depth of 12 ± 3 km, ruptured a fault trending between N and NNW with a dip angle of 21° . For the 19A events, while the first M_w 6.2 event

had a hypocenter depth of 13 ± 2 km and ruptured a fault dipping 22° , the second M_w 6.9 event had a hypocenter depth of 17 ± 2 km and ruptured a fault dipping 25° . The two 19A events ruptured faults that trend similarly between N and NNE. The 5A event has a complex source time function, which we did not account for in our waveform modelling, thus resulting in a larger depth uncertainty.

Our LOS modelling results show that the southern limits of the 28J and 5A coseismic slips (Figure C.7) differ from the southern limit of the cumulative LOS data (Figure 4.2k-l): The southern limit of the coseismic slip for event 5A is estimated further south than the limit of event 28J, while the cumulative data shows the LOS displacements of both events are at a similar southern limit. Thus, we produce another coseismic slip model by varying the fault dip of the 5A event. We find that changing the fault dip from 21° to 25° results in a more consistent southern limit of the coseismic slips (Figure 4.3b). Furthermore, changing the fault dip places the fault 2.5 km deeper than the hypocenter depth, but the fault depth is still within the hypocenter depth uncertainty (Figure 4.3c).

The coseismic slips show that the 28J event has a maximum coseismic slip of ~ 125 cm, the 5A event has a maximum slip of ~ 200 cm, and the 19A events have a maximum slip of ~ 200 cm (Figure 4.3b). The coseismic slip of the later events may be a little incorrect as the InSAR measurements potentially include afterslip from the previous events. Nevertheless, our estimated coseismic slip distributions result in earthquake sizes that are similar to GCMT solutions (Table C.3) and more than 90% variance reduction (Figure C.4) for each event, and slip magnitudes are well resolved based on our checkerboard tests (Figure C.8). The surface projection of the estimated faults places the surface expression of these faults between the large seafloor escarpment and surface expression of the Flores thrust (Figure 4.3b), suggesting that these events occurred on a deeper ramp rather than a décollement. In addition, the fact that faults do not project onto the surface expression of the Flores thrust suggests that the Flores thrust transitions from a décollement beneath the sediments in the deep Bali Basin into a ramp beneath Lombok Island (Figure 4.3c).

FIGURE 4.3: (a) Moment tensor solutions for all $M_w > 6$ events show that the events ruptured faults at different strikes and dip angles. (b) The estimated $M_w > 6$ fault planes trend similarly to the general trend of the seafloor escarpment, suggesting that the results of our seismic- and InSAR-based analysis are consistent with the geological structure in this region. Black solid lines are surface projections of the estimated fault planes. Dashed lines are north-south cross sections through the point of estimated maximum slip for all the coseismic models. (c) A vertical profile from south (S) to north (N) shows that our estimated fault planes do not project onto the surface expression of the Flores thrust, suggesting that the Flores thrust transitions from a décollement beneath the sediments in the deep Bali Basin into a ramp beneath Lombok Island. The profile also shows that the down-dip end of the 5A event intersects with the magma plumbing system for Rinjani that roots at the base of the crust (Zubaidah et al., 2014), meaning that heat from the magma ascent zone could have prevented the 5A rupture from propagating farther down-dip (e.g., Miyakawa et al., 2016; Yang et al., 2018). Black vertical lines on top of focal mechanisms are depth uncertainties based on bootstrapping resampling. Magma plumbing system after Métrich et al. (2017).

4.5 Discussion

4.5.1 The effect of stress changes induced by individual events on the later events

As all four events occurred in a sequence within three weeks, we hypothesize that each event may have changed the stress state in the hypocentre location of the subsequent event, thus precipitating the coseismic failure. To evaluate our hypothesis, we calculate static Coulomb stress changes due to a previous $M_w > 6$ event (‘source’) at the location of the next event fault (‘receiver’) (Text C1 in supporting information), noting that positive values encourage failure and negative values discourage failure (Stein, 1999). Our calculation shows that the 28J event caused positive Coulomb stress changes of >100 KPa at the hypocentre location of the 5A event (Figure 4.4a). Subsequently, the 5A event caused positive Coulomb stress changes of >20 KPa at the hypocentre location of the 19A events (Figure 4.4b).

To test the robustness of these results, we repeat the calculations using rougher slip models and different friction coefficient values. The calculations show consistent results both in amplitude and spatial overlap of the Coulomb stress changes and hypocentre locations (Figures C.9 - C.10). Therefore, bearing in mind that stress changes as low as 10 KPa can promote earthquakes (King et al., 1994), all available evidence suggests that the increased Coulomb stress changes at the hypocentre location of the 5A and 19A events supports our hypothesis that stress changes induced by a previous event played a role in promoting the later event. Nevertheless, we have seen that there was a delay instead of an immediate triggering effect. The delayed triggering can possibly be explained by the role of afterslip. Although our computations have shown that the coseismic Coulomb stress changes are potentially large enough to trigger the next events, the precise amount needed to trigger a given event is unknown. However, the events are likely to have been followed by afterslip, and the cumulative stress changes due to early afterslip that occurred within days to a week may have further promoted failure at the hypocenter location of the subsequent events. Unfortunately, the temporal sampling of the InSAR observations is limited (only one image acquired between each event), thus we cannot compute Coulomb stress changes due to the afterslip. A robust separation of the coseismic slip and afterslip would require time series analysis using

multiple SAR images between each event, or other geodetic observations sampled densely in time.

FIGURE 4.4: (a) The 28J event caused positive Coulomb stress changes of >100 KPa at the hypocenter location of the 5A event (see also Figure C.11). (b) The 5A event caused positive Coulomb stress changes of >20 KPa at the hypocenter location of the 19A events (see also Figure C.11). (c) Normal stress changes along A-A' profile, computed using coseismic slip of the 28J event, indicate that the 28J event unclamped the magma ascent zone to depths of ~ 14 km, but unclamped the magma ascent zone in the deeper depths of $\sim 16 - 25$ km. (d) Normal stress changes along A-A' profile, computed using coseismic slip of the 5A event, indicates that the 5A event clamped the magma ascent zone in the upper ~ 3 km, but unclamped the magma ascent zone in the deeper depths of $\sim 5 - 25$ km. (e) Normal stress changes along A-A' profile, computed using coseismic slip of the 19A events, indicates that the 19A events clamped the magma ascent zone to depths of 25 km. Epicenter locations are from the USGS earthquake catalog, crustal thickness after Zubaidah et al. (2014) and magma plumbing system after Métrich et al. (2017). (f) Dilatational stress along a vertical profile under Rinjani, computed using coseismic slip of the 28J, 5A, and 19A events. The results show that the second event (5A) caused the most significant volumetric expansion at the shallow magma plumbing system.

4.5.2 The effect of stress changes induced by individual events on Rinjani volcano

Given that all the four events occurred within ~ 50 km of Rinjani volcano (its magma plumbing system is made of melts and crystal mush at a variety of levels,

but primarily at 4 km and 25 km below sea level (Métrich et al., 2017)) (Figure 4.3b) and that earthquakes have been reported to trigger volcanic activity in similar circumstances (e.g., Bebbington and Marzocchi, 2011; Bonali, 2013; Bonali et al., 2013; Delle Donne et al., 2010; Hill et al., 2002; Manga and Brodsky, 2006; Walter and Amelung, 2007), we considered the effect of the individual events on the volcano. For this purpose, we calculate normal stress changes and dilatational stress (Text C1 in supporting information), which have been widely used to evaluate earthquake-volcano interactions (e.g., Bonali, 2013; Bonali et al., 2013; Walter and Amelung, 2006). We use our coseismic slip distributions to calculate the normal stress changes in a depth profile along dykes that trend from east to west (Figure 4.4c-e). This trend is inferred from east-west orientation of the two volcanic vents and one caldera of Rinjani volcano, and is also the most likely orientation to experience extension due to ruptures on the south-dipping Flores thrust system. We then use the normal stress changes to evaluate the effect of earthquakes on the magma ascent zone, with positive values unclamping the magma ascent zone and encouraging volcanic activity, while negative values discourage volcanic activity (Bonali et al., 2013; Walter and Amelung, 2006). Furthermore, we also use our coseismic slips to calculate dilatational stress along a vertical profile below the volcano’s base. We use the dilatational stress to evaluate the effect of earthquakes on the wider magma plumbing system, with positive values correspond to volumetric expansion and encouraging volcanic activity, while negative values corresponding to volumetric contraction and discourage volcanic activity (Walter and Amelung, 2006, 2007).

Our calculation suggests that each of the $M_w > 6$ events caused different effects on the volcano. The 28J event unclamped the magma ascent zone by ~ 250 KPa to a depth of 15 km below the volcano’s base (Figure 4.4c). In comparison, the 5A event clamped the upper magma ascent zone by ~ 250 KPa in the upper ~ 3 km, but unclamped the magma ascent zone by $\sim 250 - 750$ KPa at deeper depths of $\sim 6 - 15$ km (Figure 4.4d). In contrast, the 19A events clamped the magma ascent zone by ~ 100 KPa at depths of 0 - 15 km (Figure 4.4e). In addition, the 5A event caused the most significant volumetric expansion at a shallow magma plumbing system than the 28A and 19A events, which may facilitate magmatic overpressure and promote magma replenishment into a shallow magma plumbing system (Walter and Amelung, 2006).

Although the 5A event brought the most significant impact on the volcano by causing volumetric expansion at the shallow magma plumbing system, and unclamped the magma ascent zone, there were no volcanic manifestations immediately following these earthquakes ([Global Volcanism Program, 2018](#)). However, the events' effect on volcanic activity may not be instantaneous. The earthquake effect also depends on the state of the magmatic system prior to the earthquake: the volcano needs to be already at a critical state in order to be triggered (e.g., [Bonali et al., 2013](#); [Hill et al., 2002](#); [Manga and Brodsky, 2006](#); [Walter and Amelung, 2007](#)). In other areas, a triggered eruption has been observed to occur anytime within zero to five years after earthquakes ([Marzocchi, 2002](#)).

4.5.3 The role of Rinjani volcano in controlling the rupture extent

The slip magnitudes of the two M_w 6.9 events are different at the down-dip ends of the faults, when they intersect with a vertical profile of magma plumbing system below Rinjani volcano. While we estimate the 19A event has ~ 100 cm of slip at the down-dip end of the fault, the 5A event has almost zero slip at similar depth ([Figure 4.3c](#)). One possibility for this difference is the location of the two events relative to the location of Rinjani volcano ([Figure 4.3b](#)). The estimated rupture for the 19A events is located ~ 30 km east of the volcano, hence there is no potential barrier to the southward propagation of the slip ([Figure 4.3b](#)). In comparison, the 5A event ruptured an area just north of the volcano, where the down-dip end of the fault intersects with the magma plumbing system that roots at the base of the crust ([Zubaidah et al., 2014](#)) ([Figure 4.3c](#)). Therefore, heat at this depth could have prevented the down-dip propagation of the 5A slip. Similarly, the 28J event north of the volcano may have also encountered the same geothermal anomaly when the rupture propagated southwards, hence the 28J event stopped at a shallow depth.

4.5.4 Possible reasons for the cascading behavior of the sequence

The earthquake energy for this sequence was released in four moderate events within three weeks instead of a single large event at one time. One possible explanation is the irregular geometries of the Flores thrust both along-strike and-dip, which could act as rupture barriers (e.g., [Aki, 1979](#); [Yang et al., 2013](#)). In addition, dynamic simulations show that variation of the hypocenter location of an earthquake can affect the size of the earthquake (e.g., [Yang et al., 2019](#)). However, whether the entire slip area could rupture as a single event if initiated from a different hypocenter is still an open question. More detailed inspections and analysis to reveal the in-situ conditions of the Flores thrust are needed to further address the kinematic and dynamic rupture properties of the earthquake sequence.

4.5.5 Building damage caused by individual events

Our calculations show that the first M_w 6.9 (5A) event damaged more than 5000 buildings, almost 28 times larger than the number of buildings damaged by the second M_w 6.9 (19A) event ([Figure 4.5](#)). This difference could be explained either by the hypocenter depth of the first M_w 6.9 event that is ~ 5 km shallower than the second M_w 6.9 event or possibly the sediment layers in the western Lombok is thicker than the eastern Lombok, so that the thick sediments heavily amplified the seismic shaking of the first M_w 6.9 event.

The number of damaged buildings from our calculation are far smaller than the numbers reported by [ASEAN Humanitarian Assistance \(AHA\) Centre \(2018\)](#). This difference could be explained by several reasons. First, our DPMs may underestimate damaged buildings that are smaller than DPMs' pixels resolution of 30 m. Second, while our DPMs observed only heavy damages (total collapse of buildings), the AHA centre reported light to moderate damages as well (the roof of buildings remain but with structural damages in wall section) ([Figure C.12](#)). Third, the calculated coherence difference for the 5A event is likely biased toward lower values because the reference coherence is already low due to the 28J event ([Table C.2](#)), thus underestimating the DPMs. Fourth, uncertainties in damage extents may arise from satellite look direction and homogeneous threshold value we used -the

FIGURE 4.5: Spatial distribution of damaged buildings, based on Sentinel-1 SAR DPM and building footprints from OpenStreetMap, caused by individual events suggest that the first M_w 6.9 (5A) event caused the greatest damage to buildings than the second M_w 6.9 (19A) event.

use of spatially variable threshold across the whole spatial extent is beyond the scope of this study.

4.6 Conclusion

We study four $M_w > 6$ earthquakes that struck the northern region of Lombok, eastern Indonesia, between late July and mid-August 2018. We based the study on seismic and InSAR data analysis to develop coseismic models of all the events, and damage proxy maps to quantify the effect of each event on buildings. We find that: first, the earthquake sequence originated from different hypocenter depths and ruptured south-dipping thrust faults with different strikes and dips, which may explain the cascading behavior of the events; second, the earlier events in the sequence played a role in promoting the later events; third, the second event caused volumetric expansion in the shallow magma plumbing system of Rinjani volcano and unclamped its magma ascent cone, which may potentially promote future volcanic activity; and fourth, the second event caused the greatest damage to buildings.

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Chapter 5

Concluding thoughts

Earthquake hazards in Indonesia are posed partly by the Sunda megathrust, which runs along the entire length of the country, the Sumatran Fault Zone (SFZ), which spans the length of Sumatra island, and the back-arc Flores thrust in eastern Indonesia. To help with efforts to mitigate the earthquake hazards, we need to understand how these faults have behaved in the past, for example, how far earthquake ruptures on these faults propagated along strike and along depth, and what mechanisms controlled the extent of earthquake ruptures, so that we can ultimately provide insight into their future rupture behavior. In my thesis, I contribute to better understanding of how tectonic structures and fault properties affect earthquake ruptures by integrating geodetic data with physics-based modeling and with insights from seismology, geology, and volcanology.

The first chapter of my thesis addresses the question: Why did the 2008 North Pagai earthquake not cascade into a larger event, although the Mentawai patch was a seismic gap and highly coupled? I use GPS and InSAR data to model the coseismic slip of the 2008 event, and 5.6 years of cumulative GPS displacements to model the event's afterslip. Additionally, I run a conceptual model based on rate-and-state friction laws to explain my coseismic and afterslip results. The geodetic inversion results show that the estimated afterslip partially overlaps the coseismic rupture, and the rate-and-state model suggests that the non-overlapping part represents a velocity-weakening region while the overlapping part represents a velocity-strengthening region. The existence of a velocity-strengthening region has been reported to act as a systematic barrier to earthquake propagation. Thus, I suggest this velocity-strengthening region is one explanation as to why the 2008 North Pagai earthquake did not cascade into a larger event. My study in this chapter highlights that incorporating a physics-based model allows us to more robustly infer the frictional properties of faults (e.g., velocity strengthening and velocity weakening regions).

The second chapter of my thesis addresses the questions: What tectonic features controlled the rupture extent of the recent earthquakes along the Sumatran Fault Zone (SFZ)? and where are the remaining seismic gaps along the SFZ? I use InSAR data to model the coseismic slips of the SFZ earthquakes that occurred between 2007 and 2016. The geodetic inversion results show that the rupture extents were controlled by structural step-overs along the SFZ, which served as barriers to slip. In addition, the earthquake slip models presented in this chapter fill in several identified seismic gaps along the SFZ. My results suggest that the remaining seismic gaps on the SFZ are located primarily between the equator and northern Sumatra. My study in this chapter highlights the importance of incorporating a detailed map of fault structures (e.g., fault step-overs) in explaining the extent of earthquake ruptures.

The third chapter of my thesis addresses the question: Why did the 2018 Lombok earthquakes rupture the Flores thrust sequentially? I use InSAR and seismic data to model the coseismic slip of four earthquakes that occurred between 28 July and 19 August, 2018. The geodetic inversion results show that the first two earthquakes just north of active Rinjani volcano ruptured the Flores thrust from about a 5-km depth to 15 km, while the last two earthquake ruptures (~ 30 km northeast of the volcano) propagated to a depth of about 23 km. I suggest that the thermal fluid associated with the magma plumbing system of the volcano prevented the down-dip propagation of the first two earthquake ruptures. In addition, the results show that the four earthquakes ruptured faults with different strikes and dips, indicating changes of the along-strike dipping angles of the Flores thrust under Lombok island, which could act as rupture barriers. Additional work is needed to clarify these geothermal anomaly and irregular fault geometry findings, such as tomography studies and relocating micro-seismicity recorded during the sequence. My study in this chapter highlights the importance of combining geodetic and seismic data, and insights from volcanology when explaining geodetic inversion results of an earthquake sequence that occurred in close proximity to an active volcano.

The three studies I have presented show that earthquake rupture extent can be controlled by either internal factors (velocity-strengthening regions) or external factors (step-overs and geothermal anomalies). Identification of the internal and

external factors was possible only by drawing on the expertise of several earth science disciplines, which highlights the importance of multi-disciplinary approaches to studying seismic hazard.

Appendix A

Supporting information for North Pagai earthquake study

FIGURE A.1: Time series of the high-rate solution for the mainshock and the two aftershocks. Vertical dashed lines show the time of the earthquakes. Time series plotted in red are portions used to estimate coseismic offsets using zero-slope horizontal black lines.

FIGURE A.2: L-curve plot represents the trade-off between model misfit and slip roughness for coseismic slip model (a) and afterslip (b). In the top panel, red cross represents trade-off for coseismic model in Figure ??a, and the green cross represents trade-off for coseismic model with less smoothing in Figure ??a. For smoothing in the coseismic slip model, instead of choosing a smoothing factor located in the corner of the L-curve, we picked a smoothing factor that results in a bit smoother slip distribution.

FIGURE A.3: Modelling results showing the contribution of each data set: (a) based on GPS only, (b) InSAR only, and (c) both GPS and InSAR. The results show that including InSAR data in the joint inversion results in some slip estimated beneath Sipora island. In (b), we used the estimated InSAR slip to produce synthetic GPS surface displacements and compare to the observed GPS.

TABLE A.1: InSAR data used to produce coseismic displacements

Area	Frame	Path	Days before mainshock	Days after mainshock	Acquisition time	
					Master	Slave
Pagai	7130	448	64	27	22 Dec. 2007	23 Mar. 2008
Sipora	7140	449	47	44	08 Jan. 2008	09 Apr. 2008

TABLE A.2: GPS displacements used in this study

Site Name	Lon.	Lat.	Elev.	E	N	U	σ_e	σ_n	σ_u
Coseismic offsets									
BSAT	100.28	-3.07	6.25	0.68	-0.91	-0.07	0.15	0.15	0.42
LNNG	101.16	-2.29	40.09	-0.76	-0.25	0.22	0.15	0.12	0.39
MKMK	101.09	-2.54	0.54	-0.79	-0.26	0.69	0.18	0.18	0.6
PKRT	99.54	-2.15	31.82	-0.07	-0.35	-0.13	0.36	0.36	1.08
PPNJ	99.60	-1.99	34.87	0.63	-1.44	-0.27	0.18	0.21	0.39
PRKB	100.39	-2.97	21.84	0.36	-0.38	1.50	0.18	0.15	0.6
SLBU	100.01	-2.77	2.79	4.08	-21.45	14.51	0.19	0.34	1.08
SMGY	100.10	-2.61	6.94	-9.73	-5.81	-2.87	0.08	0.09	0.57
KTET	99.85	-2.36	35.20	-4.04	-14.47	-7.31	0.15	0.13	1.23
Cumulative postseismic (5.6 years)									
BSAT	100.28	-3.07	6.25	0.00	0.00	0.10	1.00	1.00	0.50
LNNG	101.16	-2.29	40.09	0.00	0.00	0.10	1.00	1.00	0.50
MKMK	101.09	-2.54	0.54	0.00	0.00	0.10	1.00	1.00	0.50
PKRT	99.54	-2.15	31.82	0.00	0.00	0.10	1.00	1.00	0.50
PPNJ	99.60	-1.99	34.87	0.80	-1.46	0.00	0.84	0.84	0.96
PRKB	100.39	-2.97	21.84	0.00	0.00	0.10	1.00	1.00	0.50
SLBU	100.01	-2.77	2.79	-3.23	-6.57	4.13	4.96	5.60	5.35
SMGY	100.10	-2.61	6.94	-9.42	-14.94	4.32	1.93	2.40	2.16
KTET	99.85	-2.36	35.20	-6.10	-11.22	4.49	0.69	1.06	1.30

Abbreviations:

Lon., Lat., Elev. = Longitude, latitude, elevation (m)

E, N, U = Displacement (cm) in east, north, vertical directions

σ_e , σ_n , σ_u = Uncertainties (cm) in east, north, vertical directions

Appendix B

Supporting information for SFZ earthquakes study

Text B1. The 2 July 2013 M_w 6.1 Aceh event

The M_w 6.1 earthquake occurred on 21 January 2013, produced by rupture of approximately a 50-km-long segment of the Sumatran Fault in Aceh province ([Figure B.14](#)). This fault segment is located between two prominent bifurcations of the Sumatran Fault, and has been estimated to be creeping ([Ito et al., 2016](#)), although non-uniformly, with maximum creep near the GEUM GPS station and very little creep towards the northwest and southeast bifurcations ([Tong et al., 2018](#)) ([Figure B.14](#)). Despite the shallow creep, the section is still accumulating strain at depth ([Ito et al., 2016](#); [Tong et al., 2018](#)).

The 2013 event occurred during a gap between SAR satellite missions: ALOS-1 (2006 - 2011), ALOS-2 (2014 - present), and Sentinel-1 (2014 - present). However, the coseismic offsets were recorded by the Aceh GNSS network for the Sumatran Fault Zone (AGNeSS) ([Ito et al., 2016](#)). The network recorded coseismic offsets at two GPS stations on the eastern side of the fault and one GPS station on the western side of the fault. While the western station is 20 km from the fault trace, the two eastern stations are ~ 100 m and ~ 1.5 km away ([Figure B.14](#)). The two eastern GPS stations recorded 5 - 7 cm of coseismic offsets.

We modeled the coseismic offsets by performing a nonlinear inversion assuming uniform slip on a single rectangular fault, to search for the best-fit solution of unknown fault parameters associated with the 2013 event (longitude, latitude, and depth associated with the upper left corner of the rectangular fault, length and width of the rectangular fault, rake, and slip). We fixed the geometry of the rectangular fault (strike = 314° and dip = 85°) based on seismicity observed using a local seismic network ([Muksin et al., 2019](#)). We made sure to provide wide ranging

for the searched parameters: 0 - 5 km for the depth, 25 - 65 km for the length, 1 - 8 km for the width, 125° - 250° for the rake, and 0 - 1 m for the slip. We then determined the full probability distributions of the unknown fault parameters using a Monte Carlo-type slice sampling algorithm (Neal, 2003), and selected the preferred solutions based on the mean of marginal probability distributions with 95% confidence interval bounds.

Our tradeoff diagrams for the unknown fault parameters (Figure B.15) show that the retrieved values (Table B.3) are poorly constrained. We thus exclude this event from the main text.

Text B2. Nonlinear inversion with and without fixing fault length and width

We ran a non-linear inversion using LOS data of the 2009 M_w 6.6 Dikit event, with and without the fault dimension (length and width) of the rectangular fault fixed. The comparison below shows fixing the fault dimension resulting in similar results to the results without fixing the fault dimension.

Fault dimension free:

Length: 33 ± 0.16 km

Width: 10.1 ± 0.39 km

Longitude: $101.732 \pm 0.001^\circ$

Latitude: $-2.559 \pm 0.001^\circ$

Depth: 0.3 ± 0.02 km

Strike: $321.2 \pm 0.1^\circ$

Dip: $77.2 \pm 1.5^\circ$

Rake: $-179 \pm 0.3^\circ$

Fault dimension fixed:

Length: 33.2 km ([Wells and Coppersmith, 1994](#))

Width: 10.5 km ([Wells and Coppersmith, 1994](#))

Longitude: $101.732 \pm 0.002^\circ$

Latitude: $-2.559 \pm 0.002^\circ$

Depth: 0.4 ± 0.08 km

Strike: $321.2 \pm 0.6^\circ$

Dip: $77.6 \pm 1.6^\circ$

Rake: $180 \pm 0.7^\circ$

FIGURE B.1: (a) Original wrapped phase of the 2009 M_w 6.6 Dikit event. (b) Initial unwrapped phase of the event showing unwrapping errors near the fault that obscure the true surface rupture trace. (c) Synthetic unwrapped phase of the event based on a uniform slip model fit to (b). (d) The residuals of wrapped phase were then unwrapped using SNAPHU. (e) The uniform slip model was added back to the result of (d) to obtain correctly unwrapped original phase with no remaining influence of the model. The final unwrapped phase of the 2009 event used in this study is that shown in (e). (f) The difference between (b) and (e).

FIGURE B.2: We picked a smoothing factor (red circle) based on a tradeoff curve (L-curve) between slip roughness and model misfit. This L-curve plot is for the 2016 Pidie Jaya event that uses one single fault model.

FIGURE B.3: Two-dimensional empirical covariance (black dots) and the estimated covariance function (red lined) based on non-deforming regions of the LOS data in Figures 2.3, 2.6 and 2.9 of the main text.

FIGURE B.4: Global CMT solutions estimated the 2007 M_w 6.4 and 6.3 Lake Singkarak doublet epicenter locations very close to the ruptured faults, different from other solutions that estimated the locations further away from the ruptured faults. GFZ solutions estimated the first and the second events exactly at the same location.

a) Sianok segment

FIGURE B.5: Tradeoff diagrams and probability distributions for the unknown fault parameters of the 2007 M_w 6.3 and 6.4 Lake Singkarak doublet estimated from InSAR data. The solutions are based on the mean of probability distributions which are drawn as red vertical lines, with 95% confidence interval bounds drawn as yellow vertical lines. (a) The Sianok segment is estimated dipping southwest. (b) The Sumani segment is estimated dipping northeast, although we constrained the fault to dipping southwest, following the seismological solutions from both global CMT and Nakano et al. (2010).

FIGURE B.6: Coseismic slip model of the 2007 Lake Singkarak doublet where fault models were fixed to our nonlinear inversion results: the Sianok segment dipping southwest and the Sumani segment dipping northeast. This slip model results in more misfit reduction by approximately 5% than the slip model based on the southwest-dipping Sumani segment (Figure 2.4) in the main text.

FIGURE B.7: Global CMT solutions estimated the 2009 M_w 6.6 Dikit epicenter locations very close to the ruptured faults, different from other solutions that estimated the locations further away from the ruptured faults.

FIGURE B.8: Tradeoff diagrams and probability distributions for the unknown fault parameters of the 2009 M_w 6.6 Dikit event estimated from InSAR data. The solutions are based on the mean of probability distributions which are drawn as red vertical lines, with 95% confidence interval bounds drawn as yellow vertical lines.

FIGURE B.9: Tradeoff diagrams and probability distributions for the unknown fault parameters of the 2016 M_w 6.6 Pidie Jaya event estimated from InSAR data. The solutions are based on the mean of probability distributions which are drawn as red vertical lines, with 95% confidence interval bounds drawn as yellow vertical lines.

FIGURE B.10: (a - b) A coseismic model based on two planar faults (the second scenario) for the Pidie Jaya event results in similar slip characteristics as the first scenario. (c) This second scenario shows that the moment-tensor solution is even worse. (d) Nevertheless, the slip model slightly improves the misfit reduction.

FIGURE B.11: Checkerboard test for the 2007 M_w 6.4 and 6.3 Lake Singkarak doublet: along the segments of Sianok (left) and Sumani (right). (a - b) Synthetic input slip with alternating 0 or 1 m of strike-slip motion on a ~ 25 km wide fault, with individual sub-fault size of $\sim 2 \times 2$ km. (c - d) The retrieved slip distribution.

FIGURE B.12: Checkerboard test for the 2009 M_w 6.6 Dikit event. (a) Synthetic input slip with alternating 0 or 1 m of strike-slip motion on a ~ 23 km wide fault, with individual sub-fault size of $\sim 2 \times 2$ km. (b) The retrieved slip distribution.

FIGURE B.13: Checkerboard test for the 2016 M_w 6.5 Pidie Jaya event. (a) Synthetic input slip with alternating 0 or 1 m of strike-slip motion on a ~ 23 km wide fault, with individual sub-fault size of $\sim 2 \times 2$ km. (b) The retrieved slip distribution.

FIGURE B.14: The 2013 M_w 6.1 Aceh event ruptured the southern half of the Aceh creeping segment, located between two bifurcations. The creeping rate is estimated to be ~ 2 cm/year along the fault section near the GEUM GPS station, and tapers down to zero towards both the northwest and southeast bifurcations (Tong et al., 2018). The 2013 event produced ~ 6 cm coseismic displacements at two GPS stations located on the eastern side of the fault. The fault dips approximately at 85° northeast, based on seismicity observed using a local seismic network (Muksin et al., 2019). Epicenter location and its associated focal mechanism are from the global CMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). Historical events are from Pesicek et al. (2010) and Nugraha et al. (2018). Aftershocks are from Nugraha et al. (2018). Fault segments are from Sieh and Natawidjaja (2000). Plate convergence rates in inset map are from DeMets et al. (2010). Gray arrows with ellipses are GPS displacements from Ito et al. (2016) and red arrows are predicted displacements based on our nonlinear inversion. The solid black line is the surface projection of the inverted planar fault model with its subsurface outlines drawn as dashed line. Grey arrows represent sense of right-lateral strike-slip fault.

FIGURE B.15: (a) Tradeoff diagrams and probability distributions for the unknown fault parameters of the 2013 M_w 6.1 Aceh event estimated from GPS data. The solutions are based on the mean of probability distributions which are drawn as red vertical lines, with 95% confidence interval bounds drawn as yellow vertical lines. (b) Focal mechanisms comparison between global GMT solution and this study.

TABLE B.1: InSAR data used to produce coseismic displacements

Data	Track	Path	Frame	Acquisition time	
				Master	Slave
The 2007 M_w 6.3 & 6.4 Lake Singkarak					
ALOS-1	Dsc.	117	3630	19 Oct. 2006	06 Jun. 2007
The 2009 M_w 6.6 Dikit					
ALOS-1	Asc.	445	7130	21 Sept. 2009	06 Nov. 2009
ALOS-1	Dsc.	114	3670	19 Jan. 2009	22 Oct. 2009
The 2016 M_w 6.6 Pidie Jaya					
ALOS-2	Asc.	153	0090	04 Dec. 2016	12 Feb. 2017
ALOS-2	Dsc.	040	3500	29 Nov. 2016	10 Jan. 2017
Sentinel-1A	Dsc.	135	0574	20 Nov. 2016	08 Dec. 2016

Abbreviations:

Asc., ascending; Dsc., descending; ALOS, Advance Land Observation Satellite

TABLE B.2: Estimated fault parameters resolved in this study

Inversion type	Longitude ($^{\circ}$) ^a	Latitude ($^{\circ}$) ^a	Depth (km) ^a	Strike ($^{\circ}$)	Dip ($^{\circ}$)	Rake ($^{\circ}$)	Length (km)	Width (km)	M _w
The 2007 M _w 6.3 & 6.4 Lake Singkarak									
Nonlinear (Sianok)	100.415 ± 0.017	-0.385 ± 0.032	1.5 ± 0.9	147.2 ± 0.13 ^b	81.4 ± 8.4 ^d	176.5 ± 5.8	21.7 ^f	8.7 ^f	6.3
Nonlinear (Sumani)	100.481 ± 0.009	-0.554 ± 0.018	1.2 ± 0.5	146.8 ± 2.66 ^b	97.2 ± 7.13 ^e	184.13 ± 11.53	25 ^f	9.3 ^f	6.3
Linear (Sianok)	100.342	-0.293	0	147.52	81	158	33.85	25.39	6.3
Linear (Sumani)	100.445	-0.503	0	326.37	83	-173	44.72	25.39	6.4
Linear (Sianok)	100.342	-0.293	0	147.52	80	147	33.85	25.39	6.4
Linear (Sumani)	100.445	-0.503	0	146.37	80	-178	44.72	25.39	6.5
The 2009 M _w 6.6 Dikit									
Nonlinear	101.732 ± 0.002	-2.559 ± 0.002	0.4 ± 0.08	321.2 ± 0.6 ^c	77.6 ± 1.6 ^d	180 ± 0.7	33.2 ^f	10.5 ^f	6.6
Linear	101.778	-2.628	0	321	77	179	38.8	22.7	6.6
The 2016 M _w 6.6 Pidie Jaya									
Nonlinear	96.1707 ± 0.003	5.236 ± 0.003	3.9 ± 0.2	53.6 ± 1.6 ^d	78.65 ± 1.9 ^d	14.4 ± 1.1	28.8 ^f	9.8 ^f	6.6
Linear (one fault)	96.098	5.183	0	53.5	78	9	26.9	22.6	6.6
Linear: fault 1	96.098	5.183	0	53.5	78	0	26.9	22.6	6.5
fault 2	96.418	5.244	8.5	233.5	12	70	20	13	6.0

Abbreviations:

^aUpper left corner of the rectangular faultSearch range: ^b120 - 180, ^c300 - 360, ^d0 - 90, ^e50 - 130^fFixed values based on Wells and Coppersmith (1994)

TABLE B.3: Estimated fault parameters resolved for the 2013 Aceh event

Longitude ($^{\circ}$) ^a	Latitude ($^{\circ}$) ^a	Depth (km) ^a	Strike ($^{\circ}$)	Dip ($^{\circ}$) ^b	Rake ($^{\circ}$)	Length (km)	Width (km)	M _w
96.263 $\pm$ 0.018	4.687 $\pm$ 0.017	2.4 $\pm$ 1.3	314	85	177.5 $\pm$ 9.2	46.3 $\pm$ 7.7	4.5 $\pm$ 1.7	6.3

Abbreviations:

^aUpper left corner of the rectangular fault

^bFixed values based on Muksin et al. (2019)

Appendix C

Supporting information for Lombok earthquakes study

Text C1. Static stress change and dilatational stress calculation

We evaluate the effect of an event on promoting failure of the later event based on static Coulomb stress change calculations. The Coulomb stress changes can be calculated through the following equation ([King et al., 1994](#); [Stein, 1999](#)):

$$\Delta CFF = \Delta\tau + \mu\Delta\sigma_n \quad (C.1)$$

where $\Delta\tau$ is the shear stress on a fault, $\Delta\sigma_n$ is the normal stress change, and μ is the friction coefficient with range 0 - 1. We obtain the shear stress and the fault-normal stress changes based on stress tensors calculated using the analytical solution of [Okada \(1992\)](#), assuming a shear modulus of 30 GPa and a Poisson's ratio of 0.25. We compute the Coulomb stress changes by using a standard value of $\mu = 0.4$ ([King et al., 1994](#)). Noting that positive values of the Coulomb stress changes encourage failure, and negative values discourage failure ([Stein, 1999](#)). We compute the Coulomb stress changes using our coseismic slip of the 28J event as a source, on the 5A fault as a receiver. Similarly, we repeat our calculation using our coseismic slip of the 5A event as a source, on the 19A fault as the receiver.

We evaluate the effect of each event on magma plumbing system of Rinjani volcano, by calculating the elastic dilatational stress under the volcano's base as a function of depth, using our coseismic slips, through the following equation ([Walter and Amelung, 2006](#)):

$$\varepsilon = e_{xx} + e_{yy} + e_{zz} \quad (C.2)$$

Where e_{xx} , e_{yy} and e_{zz} are diagonal elements of the stress tensors calculated using the analytical solution of [Okada \(1992\)](#). $\varepsilon > 0$ corresponds to volumetric expansion and encouraging volcanic activity, while $\varepsilon < 0$ corresponds to volumetric contraction and discourage volcanic activity ([Walter and Amelung, 2006, 2007](#)).

FIGURE C.1: All $M_w > 6$ event epicenter locations were located in regions mostly made of lahars, tuffs, and other volcanic deposits from the active Rinjani volcano. Modified after Mangga et al. (1994) and Ganas et al. (2018).

(a) 19A event: M_w 6.9 (19/08/2018 14:56)

(b) Regional waveforms

(c) Regional stations

(d) Teleseismic stations

(e) Event depth uncertainty

(f) Teleseismic waveforms

FIGURE C.2: CAP inversion result of the 19 August M_w 6.9 event. (a) Lower hemisphere projection of the best-fitting focal mechanism (strike/dip/rake = $114^\circ/15^\circ/116^\circ$), triangles are station projections on the lower hemisphere according to their first arrival of P-wave takeoff angle. (b - c) Waveform fits and distribution of seismic stations used to estimate the mechanism in figure a. (d - f) Distribution of teleseismic stations, point-source solutions using different source durations/centroid depths, and the waveforms fit to estimate the event depth.

FIGURE C.3: (a) The unwrapping result of an interferogram along the descending track shows that the LOS displacements for the 5 August M_w 6.9 event contain the atmospheric noise component that correlated with high topography, ranging from 0.5 to 3.5 km. (b - e) The atmospheric noise component was removed in two steps by first estimating the correlation at elevations above 0.5 km (b) and then above 2 km (d). (f) The corrected LOS displacements.

FIGURE C.4: Comparison of LOS displacements based on InSAR observations and coseismic model predictions. The coseismic models result in more than 90% variance reduction (p) for each event. The number of InSAR observations (Obs) has been downsampled to less than 1000 points to make the inversion computations more feasible.

FIGURE C.5: We picked a smoothing factor (red circle) based on a tradeoff curve (L-curve) between slip model roughness and model misfit. This L-curve plot is for the 28 July M_w 6.4 event.

FIGURE C.6: Original damage proxy map (DPM) showing the extent of damage in northern Lombok due to the 5 August M_w 6.9 event. (a) Yellow to red pixels show different level of coherence differences. Rectangles overlaid on top of pixels are building footprint information where cyan rectangles intersect with saturated red pixels are considered damage buildings and black rectangles are not considered as damage buildings. Coherence differences from other sources such as vegetation changes (b) and layover/shadow effect or landslides (c) are excluded.

FIGURE C.7: Inversion results where fault geometries were fixed to our moment tensor solutions show that the southern limits of the 28J and 5A coseismic slips in (b), differ from the southern limit of the cumulative LOS data in Figure 3.2k-l): The southern limit of the coseismic slip for event 5A is estimated further south than the limit of event 28J, while the cumulative data shows the LOS displacements of both events are at a similar southern limit.

FIGURE C.8: Checkerboard test for all $M_w > 6$ events. (Top) Synthetic input slip with alternating 0 or 100 cm of dip-slip motion. (Bottom) The retrieved slip distribution. White contour lines are the estimated coseismic slips plotted every 25 cm.

FIGURE C.9: Calculation of the Coulomb stress changes using rougher slip models shows consistent results where the earlier events in the sequence caused positive Coulomb stress changes at the hypocenter location of the next events.

FIGURE C.10: Calculation of the Coulomb stress changes using different friction values shows consistent results where the earlier events in the sequence caused positive Coulomb stress changes at the hypocenter location of the next events.

FIGURE C.11: The same as Figure 4a and 4b in the main text, but without contour lines of coseismic slips.

FIGURE C.12: Comparison of damaged buildings due to the 28J event reported by AHA (a) and based on our DPMs study (b), shows that our DPMs underestimated number of damaged buildings. AHA reported there were at least 129 damaged buildings (red dots in Figure S13a). However, in the same area, our DPMs (Figure S13b) result in only eight pixels that intersect with ~ 30 houses. In addition, Figure S13a shows that buildings considered damaged in AHA report are not only buildings that totally collapsed, but buildings with walls or pillars damage/collapse.

TABLE C.1: InSAR data used to produce coseismic displacements

Data	Track	Path	Frame	Acquisition time		Figure 2
				Master	Slave	
28 July M_w 6.4 event (28J)						
Sentinel-1	Asc.	156	1149	27 Jul. 2018	02 Aug. 2018	a
	Dsc.	032	0619	18 Jul. 2018	30 Jul. 2018	b
ALOS-2	Asc.	129	7020	12 May 2018	04 Aug. 2018	c
	Dsc.	027	3800	15 Jul. 2018	29 Jul. 2018	d
05 August M_w 6.9 event (5A)						
Sentinel-1	Asc.	156	1149	02 Aug. 2018	08 Aug. 2018	e
	Dsc.	032	0619	30 Jul. 2018	05 Aug. 2018	f
ALOS-2	Asc.	129	7020	04 Aug. 2018	18 Aug. 2018	g
19 August M_w 6.2 & 6.9 event (19A)						
Sentinel-1	Asc.	156	1149	14 Aug. 2018	20 Aug. 2018	h
	Dsc.	032	0619	17 Aug. 2018	23 Aug. 2018	i
ALOS-2	Asc.	129	7020	18 Aug. 2018	05 Jan. 2019	j
All events						
Sentinel-1	Asc.	156	1149	27 Jul. 2018	20 Aug. 2018	k
	Dsc.	032	0619	24 Jul. 2018	23 Aug. 2018	l

Abbreviations:

Asc., ascending; Dsc., descending; ALOS, Advance Land Observation Satellite

TABLE C.2: InSAR data used to produced Damage Proxy Maps (DPMs)

Path	Pre-event image 1 (Satellite / Frame)	Pre-event image 2 (Satellite / Frame)	Post-event image 3 (Satellite / Frame)
28 July M_w 6.4 event (28J)			
156	21 Jul. 2018 (S1B / 1152)	27 Jul. 2018 (S1A / 1149)	02 Aug. 2018 (S1B / 1148)
05 August M_w 6.9 event (5A)			
156	27 Jul. 2018 (S1A / 1149)	02 Aug. 2018 (S1B / 1148)	08 Aug. 2018 (S1A / 1148)
19 August M_w 6.2 & 6.9 event (19A)			
156	08 Aug. 2018 (S1A / 1148)	14 Aug. 2018 (S1B / 1148)	20 Aug. 2018 (S1A / 1149)

Abbreviations:

S1A, Sentinel-1A; S1B, Sentinel-1B

TABLE C.3: Estimated fault parameters for all $M_w > 6$ events

Strike ^o	Dip ^o	Depth (km)	M_w	Reference
28 July M_w 6.4 event (28J)				
83	28	11.5	6.4	USGS
91	14	13	6.5	GFZ
86	21	12	6.5	GCMT
114	15	10 ± 2	6.4	This study
05 August M_w 6.9 event (5A)				
93	28	30.5	6.9	USGS
79	22	15	6.8	GFZ
92	20	15.7	6.9	GCMT
80	21	12 ± 3	6.9	This study
19 August M_w 6.2 1 st event (19A)				
101	30	23.5	6.3	USGS
107	31	20	6.3	GFZ
		Not available		GCMT
100	22	13 ± 2	6.2	This study
19 August M_w 6.9 2 nd event (19A)				
85	30	30.5	6.9	USGS
93	23	29	6.9	GFZ
96	21	24.6	7.0	GCMT
95	25	17 ± 2	6.9	This study

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