Characterizing the spectrum of slip behavior at the Sumatran subduction zone with long-term geodetic records

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**CHARACTERIZING THE SPECTRUM
OF SLIP BEHAVIOR AT THE
SUMATRAN SUBDUCTION ZONE
WITH LONG-TERM GEODETIC
RECORDS**

LOUISA L. H. TSANG

ASIAN SCHOOL OF THE ENVIRONMENT

2016

CHARACTERIZING THE SPECTRUM OF SLIP BEHAVIOR AT THE SUMATRAN SUBDUCTION ZONE WITH LONG-TERM GEODETIC RECORDS

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Summary

I present three studies that depict the spatio-temporal spectrum of slip behavior on the Sunda megathrust. These studies illuminate new characteristics of slip behavior on the Sunda megathrust, which are important for gaining a comprehensive understanding of the earthquake cycle at the Sumatran subduction zone. First, from an ~1100-year-long paleogeodetic record at Simeulue Island, I discover that time-variable vertical deformation rates can be explained by changes in interseismic fault coupling over time. Also, the pattern of fault coupling preceding earthquakes between two seismic cycles may not necessarily be the same, even though the earthquakes themselves have a similar rupture pattern. Our results highlight the need for researchers to treat interseismic fault coupling maps as a snapshot of conditions on the fault that can change with time. Second, I present paleogeodetic time series that reveal a 15-year reversal in vertical displacements from 1966 to 1981. I examine a range of oceanographic and tectonic processes that can potentially explain these observations, and find that the most likely explanation is a 15-year-long slow slip event on the Banyak Islands section of the Sunda megathrust. This event is the longest lasting slow slip event reported at any subduction zone around the world, and is the first to be discovered in Sumatra. Third, I examine long GPS postseismic time series to unravel the spatio-temporal distribution of afterslip following the 2007 M_w 8.4 Bengkulu earthquake, and investigate its spatial relationship with the 2010 M_w 7.8 “tsunami earthquake” that occurred updip of the 2007 event. A key finding is that afterslip borders and overlaps the 2010 rupture area, and that localized increases in strain rates caused by afterslip may have advanced rupture of this patch in 2010.

Introduction

In the early 20th century, the earthquake cycle along active faults was represented by Reid's elastic rebound model [Reid, 1910], which describes a cyclical stick-slip process, whereby during the interseismic period elastic strain gradually accumulates on the fault at a constant rate. The stored elastic energy is then released suddenly in an earthquake. This model adequately described Reid's observations of the surface displacement of the 1906 San Francisco earthquake. In the past few decades, however, the availability of high resolution, accurate geodetic and seismic data has illuminated a diverse range of slip behavior that occurs at various active fault zones during the interseismic period of the earthquake cycle; these include fault creep, (e.g., *Lienkaemper et al.* [2012]), slow slip events, (e.g., *Hirose et al.* [1999]; *Fu and Freymueller* [2013]), afterslip (e.g., *Hsu et al.* [2006]) and viscoelastic relaxation of the mantle after earthquakes, (e.g., *Pollitz et al.* [2006]) , and low frequency earthquakes, (e.g., *Shelly et al.* [2006]). These slip phenomena span a range of timescales, from days to years. The discovery of these additional tectonic slip processes suggests that the earthquake cycle model is more complex than proposed by Reid [1910]. Also, it calls into question how we define interseismic deformation: does it consist of only the strain accumulation rate on the fault, or is it a combination of the strain accumulation rate and one or more longer-term slip processes? Answering this question is important because the results matter for seismic hazard analyses. Present GPS networks span at most one to two decades, and researchers frequently invert interseismic GPS time series to determine coupling maps, which are used to assess the maximum size of earthquakes. However, these coupling maps are often assumed to represent the state of affairs throughout the interseismic period, when in fact they may represent only a snapshot of conditions on the fault that could vary in time, due to longer-term slip processes. Also, because these slip processes potentially change the

stress state on faults, they may play an important role in either advancing or retarding in time the occurrence of earthquakes. In order to gain a comprehensive picture of the earthquake cycle at an active fault, we need long-term geodetic records and techniques to model the longer-term slip processes on faults.

At the Sumatran subduction zone, the combination of paleogeodetic, geodetic and seismic data makes the Sunda megathrust one of the best-studied subduction zone faults around the world. Researchers have obtained a long-term paleogeodetic record of deformation by studying coral microatolls fringing the forearc islands of Sumatra. From analyzing these paleogeodetic records, researchers have pieced together the story of seismic ruptures along the Sunda megathrust over the past millennia [*Zachariassen et al.*, 1999; *Natawidjaja et al.*, 2004; 2006; 2007; *Sieh et al.*, 2008, *Philibosian et al.*, 2012; 2014; *Meltzner et al.*, 2010; 2012; 2015]. In addition, the Sumatran GPS Array (SuGAR) has now been operating for ~13 years, and since the start of the modern cluster of giant earthquakes in 2004, the data have enabled researchers to better understand the modern sequence of ruptures and the segmentation of the Sunda megathrust [*Chlieh et al.*, 2008; *Hill et al.*, 2012; *Konca et al.*, 2007; *Konca et al.*, 2008; *Prawirodirdjo et al.*, 2010]. However, several questions remain to be answered regarding the slip behavior of the Sunda megathrust and its earthquake cycle:

- From a record of interseismic deformation spanning multiple earthquake cycles, what is the character of interseismic deformation? Does it vary in time, and why or why not?
- Do slow slip events occur in the Sumatran subduction zone?
- How does the seismic, postseismic and interseismic deformation history affect the timing and spatial pattern of seismic ruptures?

In this thesis, I examine paleogeodetic and long-term GPS data to address these questions in three chapters. The chapters are organized in research paper format. In chapter 1, I develop models to explain temporal variations in vertical deformation rates over the past ~267 years at Simeulue Island, Sumatra. I discuss the implications of these results for interseismic coupling maps derived from GPS data. In chapter 2, I present evidence for a long-duration slow-slip event (SSE) along the Banyak Islands section of the Sunda megathrust, as recorded by coral microatolls. I examine various scenarios of oceanographic and tectonic processes that can explain the coral data, and find that a 15-year SSE is the most likely explanation for the coral observations. This SSE represents the first observed in the Sumatran subduction zone, and the longest duration SSE observed at any subduction zone around the world. In chapter 3, I examine ~6.3-year-long GPS postseismic time series to estimate the spatio-temporal distribution of afterslip following the 2007 Bengkulu earthquake. I investigate the role of afterslip in relieving stress on this section of the megathrust, and in promoting rupture of the shallow region of the megathrust in the 2010 M_w 7.8 Mentawai “tsunami earthquake”. Lastly, I tie together the main conclusions of this work.

Chapter 1

A paleogeodetic record of variable interseismic rates and megathrust coupling at Simeulue Island, Sumatra

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Abstract

An ~1100-year-long paleogeodetic record of land-height change along the Simeulue section of the Sumatran subduction zone reveals significant variations in vertical motion rates. From an ~267-year-long record, we develop models to explain rate variations in the decades before the 1861, 2004 and 2005 great earthquakes. The record shows that rates accelerated by a factor of four to ten in the decades before the 1861 earthquake; one plausible explanation is a significant increase in the depth of interseismic coupling on the Sunda megathrust under Simeulue. Despite similarity of the 1861 and 2005 coseismic rupture patterns, the pattern of coupling during the decades before the two earthquakes may have been different. Most GPS observations of interseismic deformation at subduction zones span only a decade or two; our results highlight the need to treat GPS-derived coupling maps as only a snapshot of fault conditions that are temporally variable.

1.1 Introduction

Fault coupling represents the pattern of interseismic strain accumulation on a fault, expressed as the ratio of the slip rate deficit during the interseismic period to the plate convergence rate. Highly coupled patches are regions of the fault that can potentially release a large portion of strain seismically in large earthquakes. Therefore, assessing the location and size of future earthquakes along active faults requires characterizing the patchwork quilt of interseismic fault coupling patterns. Maps of interseismic fault coupling are usually derived from GPS data spanning one or two decades, and are useful for identifying rupture segments based on matching interseismic coupling patterns with past earthquake rupture areas (e.g., *Kaneko et al.* [2010]; *Loveless and Meade* [2011]; *Metois et al.* [2012]; *Miura et al.* [2004]; *Moreno et al.* [2010]), identifying potential rupture barriers (e.g., *Chlieh et al.* [2008]; *Prawirodirdjo et al.* [2010]), and understanding various stages of the seismic cycle on adjacent fault segments (e.g., *Freymueller et al.* [2000]; *Moreno et al.* [2011]). However, some researchers have found little correlation between areas of fault coupling and earthquake rupture, and advocate for augmenting GPS-inferred coupling maps with multi-cycle slip histories along megathrusts (e.g., *Lorito et al.* [2011]; *Nalbant et al.* [2013]; *Romano et al.* [2012]; *Thirumalai et al.* [2015]; *Wesson et al.* [2015]). This advocacy is well justified in light of what we now understand about the host of tectonic processes that occur on megathrusts during the interseismic period of the seismic cycle. Transient processes such as slow slip events, afterslip and viscoelastic relaxation of the mantle (that can potentially extend over time periods beyond a seismic cycle) can alter deformation patterns and stress states on megathrusts over long time periods (e.g., *Fu and Freymueller* [2013]; *Meade and Loveless* [2009]; *Paul et al.* [2012]; *Suito and Freymueller* [2009]; *Tsang et al.* [2015]), and thus complicate the interpretation of interseismic deformation.

Furthermore, long duration geodetic records in the Alaska, Chile, Japan and Sumatra subduction zones reveal that strain may not accumulate linearly during the interseismic period [Li *et al.*, 2014; Meltzner *et al.*, 2015; Nishimura *et al.*, 2004; Philiposian *et al.*, 2014; Prawirodirdjo *et al.*, 2010; Wesson *et al.*, 2015]. Such spatiotemporal variability needs to be characterized holistically, along with the tectonic processes that control that variability, if we are to improve assessment of seismic hazard along megathrusts. To achieve this objective, we need to study geodetic records spanning several seismic cycles. In contrast, most GPS coupling maps represent only a snapshot of the interseismic period.

Along the Sumatran subduction zone, Meltzner *et al.* [2010; 2012; 2015] reconstructed interseismic vertical deformation histories by analyzing coral microatolls from multiple sites located across Simeulue Island, which span an ~1100-year-long period from 870 to 2005 AD (Figure 1.1). This long paleogeodetic record reveals abrupt, sometimes coeval changes in long-term subsidence rates across the island. These observations implore the question: What tectonic mechanisms and physical processes control these rate variations?

In this study, we explain with physical models the observed rate changes during the past ~267 years – the time period spanning the past two seismic cycles (labeled periods P1, P2 and P3 in Figure 1.1). During this time, we find near-synchronous long-term rate changes above the northern limit of the 2005 M_w 8.6 Nias earthquake rupture patch. Meltzner *et al.* [2015] also examined the rate changes over this time period and presented simple models indicating that these observations can be explained as temporal changes in the width of the locked zone. In our study, we expand on those results with more sophisticated models, and include data from additional coral sites located in the northwestern part of Simeulue. Our results suggest that the observed interseismic rate

changes can be explained by tectonically feasible spatiotemporal changes in interseismic coupling patterns.

Figure 1.1 (a) Simeulue Island is located offshore of north Sumatra (inset). Sim: Simeulue, BI: Banyak Islands, Ni: Nias, Ba: Batu Islands. *Meltzner et al.* [2010; 2012; 2015] reconstructed interseismic uplift histories from coral microatolls sampled at 13 sites across Simeulue Island (main). Contours indicate the coseismic uplift (in cm) during the 2004 M_w 9.2 Sumatra-Andaman earthquake (dashed grey, from *Meltzner et al.* [2012]) and 2005 M_w 8.6 Nias earthquake (solid grey, from *Meltzner et al.* [2015]), defining the Simeulue saddle. (b) History of interseismic vertical displacement at each coral site. The width of the bars is proportional to the vertical displacement rates. The dates of paleogeodetically-

recorded earthquakes are marked in red. The rate at PPY with a '+' sign (late in P3) indicates interseismic uplift. In this study, we model rate changes during periods P1, P2 and P3.

1.2 Relative sea-level histories from coral microatolls

Relative sea-level histories were derived from coral microatolls of the genera *Porites* and *Goniastrea* [Meltzner *et al.*, 2010; 2012; 2015]. The upper surfaces of these coral microatolls track relative sea-level (RSL) changes over time.

Changes in RSL are a combined effect of both changes in land level and sea surface height, resulting from a combination of tectonic and non-tectonic factors such as tectonic deformation over the earthquake cycle, isostatic changes of the land due to changing glacial and other surface loads, and regional changes in sea surface height due to oceanographic processes. It is therefore important to understand the relative contribution of these factors to RSL changes. As discussed by Meltzner *et al.* [2012], isostatic and oceanographic processes mostly operate on regional to global scales, and since Simeulue Island is only ~100 km long, these processes would be expected to influence all sites across the island approximately uniformly and synchronously. Although regional changes in sea surface height can result in spatially varying magnitudes of rate change at sites across the island, these variations are likely small in comparison to tectonic changes (further discussed in Text S1.1). On the other hand, if rate changes observed on one part of the island are not observed coevally or coherently on another part of the island, then the spatial scale of these changes is more consistent with tectonic uplift and subsidence over the course of the earthquake cycle.

Such spatial variations in tectonic behavior are observed at Simeulue. For example, corals located over the 2004 M_w 9.2 Sumatra-Andaman earthquake rupture have distinctly different RSL histories from corals over the 2005 M_w 8.6 Nias earthquake rupture, suggesting that the central part of Simeulue acts as a persistent rupture barrier [Meltzner

et al., 2012]. Figure 1.1 illustrates the ~1100-year history of interseismic vertical displacement rate changes at 13 coral sites located on Simeulue. Table 1.1 lists the interseismic vertical displacement rates during each time period in the paleogeodetic record, for respective sites, as well as the 2σ uncertainties associated with each rate. During ~1300 to 1450 AD, rate changes observed across the island are non-coeval. Most rate changes occur after major earthquakes, such as the 1394 AD earthquake, which affected sites over the 2004 patch [Meltzner *et al.*, 2010]. However, the paleogeodetic record also shows rate changes that are not associated with known earthquakes. For example, during ~1738 to 1861 AD, sites over the 1861/2005 patch experienced a sudden acceleration in rates during the decades before the 1861 AD Nias-Simeulue earthquake. We discuss these rate changes in the next section, focusing on the ~267-year period from ~1738 to 2005 AD.

1.3 Interseismic vertical displacement rates during the past ~267 years

While it is relevant to investigate the entirety of this ~1100-year-long record, in this study we focus on a particularly interesting period of interseismic rate changes observed during the ~267-year period from ~1738 to 2005 AD (P1, P2 and P3 in Figure 1.1). Above the 1861/2005 patch, four sites (SLR, SMB, UTG, LBJ) experienced an abrupt four to ten fold acceleration in interseismic subsidence rates roughly 20-40 years prior to the 1861 earthquake (Figure 1.1). We note that corals in the Simeulue-Nias region show that the coseismic displacements of the 1861 Nias-Simeulue earthquake were similar to those of the 2005 Nias-Simeulue earthquake, suggesting that the two earthquakes were similar in extent and magnitude [Meltzner *et al.*, 2015]. These observations beg two questions: what patterns of strain accumulation and tectonic mechanisms explain the pre-1861 acceleration in rates? Were patterns and mechanisms of strain accumulation during the pre-2004 period (P3, Figure 1.1) similar to those preceding the 1861 earthquake?

Modeling the interseismic vertical displacement rates during the last ~267 years will provide a better picture of the interseismic strain accumulation patterns over the past two seismic cycles on this section of the megathrust, and of possible controlling tectonic mechanisms. We therefore develop physical models to explain three time periods in the past ~267 years: P1 (pre-1819), P2 (1839–1861) and P3 (pre-2004). We note that the start of the paleogeodetic record at individual coral sites differs for periods P1 and P3 (Table 1.1).

1.4 Modeling methods

We develop physical models to understand the tectonic mechanisms controlling the rate variations during three time periods: P1, P2 and P3 (Figure 1.1). Our models consist of a fault with a dip profile that approximates the Slab 1.0 model along this section of the megathrust [Hayes *et al.*, 2012], and extends to a depth of 100 km (Figure S1.2). The fault was divided into 5-km-long and 2-km-wide subfault patches.

We forward modeled the observed rate changes as spatiotemporal variations in locking depth and fault coupling. For each combination of fault parameters, the vertical displacement at each coral site was calculated with the Okada model of dislocations in an elastic half-space [Okada, 1985], in a backslip framework [Savage, 1983]. In this framework, slip on the fault can vary from zero to the full plate subduction rate. A slip rate of zero corresponds to a locked fault interface (with a coupling ratio of 1), whereas a slip rate equal to the full plate subduction rate corresponds to a freely slipping fault interface (with a coupling ratio of 0). In this study, we assumed that the plate subduction rate is 40 mm/yr [McNeill and Henstock, 2014; Simons *et al.*, 2007].

Our model configuration is similar to that employed by Meltzner *et al.* [2015]. In their study, they produced along-strike vertical displacement profiles based on a hypothetical

row of surface points located equidistant from the trench. In this study, we incorporated sites in the northwestern part of Simeulue (for period P3), which are located at varying distances from the trench. We therefore calculated vertical displacement rates at each site based on their actual geographical locations.

It is believed that the frictional behavior of trench regions may be different from that at deeper seismogenic depths; this is supported by observations of rapid afterslip that occurred updip of the 2005 rupture area [*Hsu et al.*, 2006]. Further, *Tilmann et al.* [2010] relocated aftershocks of the 2004 and 2005 earthquakes, and found a well-defined updip limit of seismicity that lies at a fairly uniform distance from the trench, with the exception of an abrupt ~25 km landward shift under the central part of Simeulue. They inferred that this updip limit of seismicity marks the transition between the deeper seismogenic zone and stable sliding near the trench. In order to match these observations of the updip limit of seismicity, we followed the rationale outlined by *Meltzner et al.* [2015], assigning the shallowest 18 km of the fault to be partially coupled at a coupling ratio of 0.4 (Figure S1.2). We note that model resolution is poor in regions near the trench, and there will inevitably be tradeoffs between the parameters that characterize the coupling distribution in this region. Also, recent shallow, tsunamigenic megathrust ruptures along the Sumatran subduction zone (e.g., *Hill et al.* [2012]) suggest that the trench region is capable of both aseismic and seismic slip behavior.

For all three periods, we tested for along-strike variations by subdividing the fault into two sections, and determined best-fit combinations of a) locking depths for each section, and b) the location of the boundary separating the two sections of different locking depths (Figure S1.2 shows an example model configuration). We systematically tested for downdip limits of locking ranging from 25 to 55 km depth, at 5 km (or, in some cases

finer) intervals (further details are described in Text S1.2). The preferred two section forward models for each time period correspond to the combinations of parameters that yield the least misfit between the data and model vertical displacement rates, based on the weighted mean of the sum of squared residuals (WMSSR). The WMSSR value is similar to the reduced chi-square value, but uses the number of observations rather than the number of degrees of freedom in the model.

For periods P1 and P2, we have only three sites located in the southeastern part of the island, and hence the data do not enable us to determine a well-resolved inverse model of fault coupling under the length of the island. We therefore chose to present for these periods a suite of possible forward models that can explain the data (Figures S1.3–S1.7 and Tables S1.2–S1.4). In contrast, for period P3, additional coral observations from the northwestern part of Simeulue are available, which provide better model resolution under the length of the island. We therefore present for this period both forward and inverse models of fault coupling. For the forward models, we present the two section models described in the previous paragraph. In addition, we tested more complex forward models by incorporating a priori constraints based on observations of regional seismicity at Simeulue (more details in Text S1.3).

For the inverse models, we first designed a fault with the same dip geometry and size as employed for the forward models, but subdivided the fault into 10 km by 10 km subfault patches. Similar to the forward models, we calculated model displacements at each site using Green's functions from *Okada* [1985]. We inverted for the slip rate deficit on each subpatch with a bounds-constrained linear least squares inversion method, assuming trench-normal backslip on the fault. To regularize the inversion, we applied smoothing constraints and selected a Laplacian weight that yields a visually smooth spatial

distribution of coupling (see Figure S1.8 and Text S1.4). The slip rate deficit is equal to some proportion of the full subduction rate of 40 mm/yr, such that the coupling ratio on each patch equals the slip rate deficit divided by the full subduction rate. We applied bounds of 0 mm/yr and 40 mm/yr to the slip rate deficit.

Due to the limited spatial distribution of data, we again emphasize that our models are limited in resolution and therefore nonunique. In particular, tradeoffs exist between the width of the locked zone and coupling ratio of the fault. Hence, even though we present a preferred model for each time period, many other models can explain the coral data within their uncertainties, and our models do not represent fully optimized solutions. Our goal in this study was simply to determine whether tectonically feasible models of spatially variable locking depths and fault coupling can explain the coral observations showing changing vertical displacement rates for each time period.

1.5 Results

Figure 1.2 shows our preferred forward models of spatially variable fault coupling for periods P1 and P2, and Figure 1.3 shows our preferred forward model and inverted coupling distribution for period P3. For period P1, our preferred model in Figure 1.2a has the fault locked to 35 km depth under northwestern Simeulue, while under southeastern Simeulue it is locked to 30 km depth (additional models in Figure S1.3). For period P2 (pre-1861), the four to ten fold acceleration of interseismic subsidence rates can be explained by a deepening of the downdip limit of locking to 55 km under the entire length of Simeulue, with the fault remaining locked to 30 km depth southeast of Simeulue (Figure 1.2b; additional models in Figure S1.4). The downdip limit of locking changes from 55 to 30 km near the southeasternmost tip of Simeulue.

Our preferred two-section forward model for period P3 is shown in Figure 1.3a, which suggests that the fault was locked to 40 km depth under Simeulue, and 50 km under the southeasternmost part of the fault. The 10 km along-strike change in locking depth across the fault yields marginally better fits than a uniform locking depth of 40 km (Table S1.4a). Testing over a range of depths (results listed in Table S1.4a) suggests that a locking depth of 40 km under Simeulue yields the best data-model fits: neither shallower (35 km) nor deeper (50 km) downdip limits of locking yield good fits to the coral observations.

Our inverse model, shown in Figure 1.3b, shows that allowing for greater along-strike variations in fault coupling improves data-model fits, particularly at sites in the central part of Simeulue – USL, PPY, ULB, SLR, BUN and UTG. This model suggests that the northwesternmost part of the fault was locked to ~40 km depth, while the southeasternmost part of the fault was locked to ~60 km. These locking depths are similar to those suggested by our two-section forward model (Figure 1.3a). In contrast, marked variations in locking depth and coupling occur under Simeulue. Under the portion of northwestern Simeulue (marked ‘A’ in Figure 1.3b) the fault is locked to a shallower depth of ~28–30 km. Also, the region under central-southern Simeulue (marked ‘B’ in Figure 1.3b) is generally less coupled than adjacent parts of the fault. Although the details of regions ‘A’ and ‘B’ in the inverse model (Figure 1.3b) may be beyond the limits of what we can resolve with available data, all the models for P3 show less coupling under central Simeulue, regardless of the patch size and Laplacian weight that we tested (Figure S1.8, Text S1.4). This result suggests that lower coupling under central Simeulue in P3 is a robust conclusion. Moreover, we tested forward models with greater along-strike complexity (Text S1.3) and our results suggest that data-model fits indeed improve when

shallower downdip limits of locking, deeper updip limits of full locking and less fault coupling under central Simeulue are integrated into our models (Figure S1.6).

As noted in the previous section, our models are limited in resolution, and hence nonunique. We therefore present various other possible models for each of the three time periods in Figures S1.3–S1.7, and Tables S1.2–S1.4 (Text S1.2 details a discussion of these figures and tradeoffs between various model parameters, among our two-section forward models). Our main conclusion from these models is that there exist tectonically feasible ranges of locking depths that can explain the coral observations for each time period. Based on this, we argue that the time-varying vertical displacement rates on Simeulue are likely tectonically controlled.

Figure 1.2 Preferred models of fault coupling for periods P1 (a) and P2 (b). The four to ten fold acceleration in interseismic subsidence rates from period P1 to P2 can be explained by a deepening of the downdip limit of locking from 35 to 55 km under Simeulue. Grey circles: Seismicity from *Tilmann et al.* [2010].

Figure 1.3 Preferred models of fault coupling for period P3. (a) Our preferred two-section forward model suggests that the fault was locked to a 40 km depth under Simeulue, and 50 km under the southeasternmost part of the fault. (b) Our inverse model suggests that allowing for greater along-strike variations in locking depth and coupling under Simeulue yields better data-model fits. Four features observed in central Simeulue are overlain on the model: 1) The segment boundary location proposed by *Franke et al.* [2008] (blue dashed lines), 2)

contours of coseismic uplift during the 2004 and 2005 earthquakes [*Meltzner et al.*, 2012; 2015] (dashed and solid magenta lines, respectively), 3) updip limit of seismicity [*Tilman et al.*, 2010] (white circles), and 4) locations of four M~7 earthquakes (stars from the Advanced National Seismic System (ANSS) catalog). Regions marked 'A' and 'B' are discussed in the text.

1.6 Discussion

Our suites of models suggest that for periods P2 and P3, the patterns of fault coupling were different (compare Figures 1.2b and 1.3a–b). The differences in the models of fault coupling between the two periods may in part reflect the more limited spatial distribution of data across the island for period P2, but different vertical displacement rates (although not significant at 2σ) at sites with rates estimated before each earthquake suggest that patterns of strain accumulation in the decades before the 1861 and 2005 earthquakes may have been different.

The along-strike variations in locking depths and fault coupling suggested by our inverse model for period P3 may be a manifestation of the persistent rupture barrier in central Simeulue. This barrier is characterized by low cumulative slip during the 2004 and 2005 ruptures, and localized M7+ earthquakes in 1976, 2002, 2008 and 2010 (Figure 1.3b). In addition, *Meltzner et al.* [2012] showed that at least seven large ruptures were arrested at this barrier over the past ~1100 years. Researchers have previously associated this barrier with structural heterogeneities on the subducting plate and different frictional properties on the fault interface. Notably, *Franke et al.* [2008] conducted wide angle refraction and multichannel reflection seismic surveys along this part of the megathrust, and their results reveal an elevated NNE–SSW oriented ridge of oceanic basement under central Simeulue, which projects to the location of one of the main fracture zones in the Wharton basin, as mapped by *Singh et al.* [2011b] and *Jacob et al.* [2014]. They interpret this ridge to reflect the structural relief caused by a fracture zone on the subducting plate at central Simeulue. The orientation of this ridge is consistent with the orientation of the Simeulue saddle in the middle of the island, as defined by contours of uplift from the coseismic

rupture ends of the 2004 and 2005 earthquakes [Briggs *et al.*, 2006; Meltzner *et al.*, 2015].

Figure 1.3b shows an overlay of the proposed location of the aforementioned features onto our coupling model. A visual comparison shows that the shallower downdip limit of locking under northwestern Simeulue (marked ‘A’ in Figure 1.3b) lies in the region of the proposed location of the subducted fracture zone. The region of lower coupling that is marked ‘B’ in Figure 1.3b coincides spatially with the eastern flank of the subducted fracture zone, where Franke *et al.* [2008] proposed that the megathrust is faulted and torn. The eastern flank is also marked by a ~25 km landward shift in the updip limit of seismicity, a feature that may reflect the structural heterogeneities in this region. We speculate that the lower apparent coupling in region ‘B’ may be attributed to a fault or tear in the megathrust at this location.

The time-varying interseismic rates suggest that interseismic fault coupling and slip behavior may be influenced by frictional conditions along the fault that can change with time. Localized fluid release from the subducted fracture zone, as proposed by Tilmann *et al.* [2010], might lead to transient changes in pore pressure on the fault interface. This increased fluid pressure in turn reduces the effective normal stress and changes the frictional properties on the megathrust, since fault friction is controlled by factors such as normal stress, temperature, type of rock material and distribution of fluids in the crust [Kaneko *et al.*, 2010]. Numerical simulations of the seismic cycle will enable us to gain further insights into how these factors control friction and interseismic behavior of faults. For example, Noda and Hori [2014] conducted simulations of a sequence of earthquakes assuming a rate and state friction law, and reported that the A/B parameter (direct effect/evolution effect) controls the complexity of interseismic behavior on seismogenic patches. Until now, it remains poorly understood how tectonic and physical processes

vary over multiple seismic cycles at subduction zones, which justifies the need for combining observations of multi-cycle geodetic records with modeling efforts in order to obtain a holistic picture of the physical factors that control time-varying interseismic behavior of megathrusts.

1.7 Conclusions

Coral microatolls on Simeulue Island reveal vertical displacement rates that vary with time over the past ~1100 years. Our results suggest that these rate variations are likely tectonically controlled, and that the strain accumulation process is not uniform during the interseismic period of the seismic cycle. In the decades prior to the 1861 Simeulue-Nias earthquake, coral observations reveal a four to ten fold acceleration in interseismic subsidence rates. One way to explain these observations is with a significant deepening of the downdip limits of locking under the island. In addition, despite similar coseismic rupture areas of the 1861 and 2005 earthquakes, the pattern of interseismic fault coupling in the decades prior to the two earthquakes may have been different. Up to now, multi-cycle geodetic records at subduction zones are sparse, but obtaining such records are essential to improve our understanding of how interseismic behavior on megathrusts varies with time.

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1.9 Supporting Information

Introduction

This supporting information contains additional details of our analysis of the coral data, our modeling methods and our suite of coupling models for each time period. In Text S1.1, we detail our method of calculating weighted mean rates for coral sites with multiple rates in period P3. We also present in Text S1.1, Table S1.1 and Figure S1.1 complete details of our analysis of the Aviso satellite altimetry data and the calculation of additional rate uncertainties due to spatial variations in sea level. In Text S1.2, we provide further details and justification for our method of modeling time-varying coral rates as spatiotemporal variations in fault coupling. Figure S1.2 shows that our fault geometry is a good approximation of both the Slab 1.0 model [Hayes *et al.*, 2012] and relocated seismicity in SE Simeulue [Tilman *et al.*, 2010]. We also present and discuss the suite of possible forward models that can explain the coral observations for each of the three time periods (P1, P2 and P3), and tradeoffs between various model parameters (see Figures S1.3, S1.4 and S1.5, and Tables S1.2, S1.3 and S1.4, for periods P1, P2 and P3 respectively). In Text S1.3, Figures S1.6 and S1.7, we present more complex forward models that we tested for period P3. Text S1.4 and Figure S1.8 provide further details of our inversion methods.

Text S1.1

Weighted average rates for period P3 (pre -2004)

For period P3 (pre-2004), *Meltzner et al.* [2010; 2012; 2015] reported multiple rates for overlapping time periods at seven coral sites: BUN, LWK, LKP, LDL, USG, LNG and USL. For each of these sites, we calculated a weighted mean rate (WMR):

$$WMR = \frac{\sum_{i=1}^N \frac{v_i}{\sigma_i^2}}{\sum_{i=1}^N \left(\frac{1}{\sigma_i}\right)^2}$$

where N is the total number of uplift rates, v_i is the individual uplift rate and σ_i is the 1σ uncertainty in the individual uplift rates.

The 2σ uncertainty in the WMR accounts for the dispersion of the observed rates. This uncertainty was obtained by first calculating the standard deviation of the mean (SDOM, also known as the “formal error”):

$$SDOM = \sqrt{\frac{1}{\sum_{i=1}^N \left(\frac{1}{\sigma_i^2}\right)}}$$

Second, the reduced chi-square value, χ_r^2 , was calculated:

$$\chi_r^2 = \left(\frac{1}{N-1}\right) \sum_{i=1}^N \frac{(v_i - WMR)^2}{\sigma_i^2}$$

If $\chi_r^2 \leq 1$, the SDOM was adopted as the 1σ uncertainty in the WMR.

On the other hand, if $\chi_r^2 > 1$, the 1σ uncertainty in the WMR (or the standard deviation of the weighted mean, SDOWM) was calculated as follows:

$$SDOWM = \sqrt{\chi_r^2 * (SDOM)^2}$$

Figure 1.1 shows the interseismic vertical displacement rates, and Table 1.1 lists the interseismic vertical displacement rates together with their 2σ uncertainties.

Spatial variations in sea surface height from modern satellite altimetry data

In order to quantify the likely contribution due to spatial variations in sea surface height across Simeulue, we analyzed modern satellite altimetry data in our small study area, following the method of *Tsang et al.* [2015]. Their method entails calculating an additional uncertainty due to spatial variations of sea surface height, adding this to the uncertainties determined by *Meltzner et al.* [2012], and determining whether the modified rate uncertainties overlap at the 1σ confidence level. We note that modern satellite altimetry data spanning the past two decades may not adequately represent the patterns of sea surface height over the whole period from 870 AD, but in the absence of longer records, analyzing this data is the best that we can do.

We used the globally gridded sea level anomalies (MSLA) product from Ssalto/DUACS, distributed by Aviso, with support from Cnes (www.aviso.altimetry.fr/duacs). We first extracted time series of sea level anomalies (SLA) from 30 grid points in a 1.5° by 1.25° degree region centered on Simeulue (Figure S1.1a). Second, we calculated the time series

of mean SLA for each grid point, using a running time window of ± 30 days (Figure S1.1b). This length of time window theoretically reduces the contribution of aliased signals with less than two months [Ponte and Lyard, 2002]. Third, we differenced the 2-month mean SLA between each pair of grid points, to obtain time series of differences in 2-month mean SLA (Figure S1.1c). Finally, we determined the 95th percentile of these differences — this value represents the maximum magnitude of spatial variations in sea level in our study area, at the 2σ confidence level. This value is 17.7 mm (Figure S1.1d).

For each period (P1, P2 and P3), we calculated an additional uncertainty due to spatial variations of sea surface height, equal to 17.7 mm divided by the time period over which each rate was calculated. This additional uncertainty was added in quadrature to the uncertainties determined by *Meltzner et al.* [2012], to obtain a total uncertainty.

Table S1.1 lists values of the additional and total uncertainties for each coral rate. The change in the uncertainties as a result of accounting for spatial variations in sea level ranges from ± 0 to ± 0.2 mm/yr. The mean change in the total uncertainty is ± 0.06 mm/yr. These changes are small and suggest that spatial variations in sea level do not contribute significantly to the observed rate changes. Rather, the observed rate changes are more consistent with tectonic, land-level changes.

Text S1.2

Modeling method: Two-section forward models

For each of the three periods (P1, P2, P3), we tested for along-strike variations by subdividing the fault into two sections and testing a range of values to determine the best fit combination of three model parameters: (1) the locking depth of the NW section, (2) the locking depth of the SE section, and (3) the location of the boundary separating the two sections of different locking depths. An example configuration of the two-section model shown in Figure S1.2 illustrates locking depths of 50 km and 30 km on the NW and SE sections, respectively, with the boundary between the sections (the location of the locking depth change) located at the center of the modeled fault. We note that the lateral boundaries of the fault model do not significantly affect the calculated uplift rates.

We forward modeled many combinations of locking depths for each section. For a given locking depth on one section, we incrementally varied the locking depth on the other section from 30 to 55 km depth, at 5 km depth increments. For example, for a locking depth of 30 km on the NW section, we tested a total of seven locking depths on the SE section: 25, 30, 35, 40, 45, 50 and 55 km.

For a particular combination of along-strike locking depth differences, e.g., 30 km on the NW section, 40 km on the SE section (which we abbreviate as the “30–40 km” combination), we tested a range of locations for the boundary of locking depth change, by shifting the location of the boundary in 5 km along-strike increments. For example, a particular model configuration might consist of a 30–40 km along-strike locking depth difference, with the boundary of locking depth change located 45 km from the SE edge of the fault. For each boundary location, we determined the weighted mean of the sum of

squared residuals (WMSSR). The boundary location with the lowest WMSSR corresponds to the best model for the 30–40 km combination.

The suite of plausible models for each period (P1, P2, P3) corresponds to the models that have the lowest WMSSR out of all combinations tested. These models are presented in Figures S1.3 and S1.4 for periods P1 and P2, respectively. Of this suite, the preferred model presented in the main paper is the model that yields the lowest WMSSR.

Suite of results: Period P1 (pre-1819)

Table S1.2 lists the best models of all along-strike combinations of locking depths tested for period P1, and the suite of possible models that yield the lowest WMSSR. This suite of models is also shown in Figures S1.3a–c, has WMSSR ranging from 0.22 to 0.38, and suggests that the fault was locked to a depth of 30 km under SE Simeulue, while it was locked to a depth of 33 to 35 km under NW Simeulue. The boundary of locking depth change is located near central Simeulue.

Our preferred model, which has a deeper locking depth to the NW (shown in Figures 1.2a and S1.3c), differs from that presented by *Meltzner et al.* [2015]. Both studies find little along-strike difference in locking depths during this period, but *Meltzner et al.* [2015] suggested a shallower locking depth of 26 km under the entire length of Simeulue, with deeper locking farther SE. Some of these differences likely arise from slight differences in the model fault geometry used in each study and reflect limitations in the model resolution. That said, the difference between our result of deeper locking to the NW and the earlier result of *Meltzner et al.* [2015] (which suggested deeper locking to the SE) reflects a robust improvement in our modeling technique: we adopted a 2.5-dimensional

fault geometry in this study, whereas the 2-dimensional geometry used by *Meltzner et al.* [2015] required the simplified assumption that the sites were equidistant from the trench.

Our latest results suggest that the slight differences in each site's distance from the trench result in significant differences in rates at the various sites, such that a deeper locking depth under NW Simeulue is required to compensate for these differences. In Figures S1.3d–f, we show that models with uniform locking depths (30 to 32 km) across the length of the fault generate more subsidence at sites farther SE, simply because those sites are 1–5 km closer to the trench than the more northwesterly sites; these models yield poor data-model fits. In contrast, the similarity of the data vectors at the three sites during this period requires deeper locking to the NW.

Suite of results: Period P2 (1839–1861)

Table S1.3 lists the best models of all along-strike combinations of locking depths tested for period P2. The suite of possible models consists of the combinations that yield the lowest WMSSR values of all combinations tested (WMSSR ranging from 0.34 to 1.52). A subset of the suite of possible models is shown in Figures S1.4a–h. Our models suggest that the fault can be locked to depths ranging from 40 to 55 km under the length of Simeulue. However, our preferred model (the model in the suite that yields the lowest WMSSR value of 0.34) suggests that the fault is locked to a depth of 55 km under the length of Simeulue, while it remained locked to a depth of 30 km under the SE part of the fault (Figure 1.3b). The spatial pattern of interseismic coupling suggested by our preferred model is similar to that presented by *Meltzner et al.* [2015].

As a means of exploring the trade-offs between various parameters in our models, we took our preferred model, perturbed one parameter one at a time from the best-fit value,

and evaluated how the WMSSR value changed. Figures S1.4i–p show how the WMSSR changes as a result of perturbing one parameter at a time from the values in the preferred model (results also listed in Table S1.3).

Suite of results: Period P3 (pre-2004)

Table S1.4a lists the best two-section forward models of all combinations of locking depths tested for period P3, and the suite of possible models that yield the lowest WMSSR. Our results suggest that a locking depth of 40 km under Simeulue best fits the coral observations across the island. The WMSSR of these models range from 1.73 to 1.79. Our preferred two-section model (with a WMSSR value of 1.73) suggests that the fault is locked to 40 km under the length of Simeulue and 50 km under the SE part of the fault.

Figure S1.5 shows how the WMSSR changes as each model parameter is perturbed from the values shown in the preferred model (results also listed in Table S1.4a). Figure S1.5b suggests that a slightly shallower updip limit of full locking of 16 km yields a lower WMSSR, but this does not change our main conclusions.

Text S1.3

Period P3 (pre-2004): More complex forward models

For period P3, we have data that span the length of Simeulue, which enables us to explore models that are more complex than the two-section forward model shown in Figure 1.3a. We therefore determined models that assume added levels of complexity, which are based on observations of along-strike variations in seismicity under Simeulue. Notably, *Tilmann et al.* [2010] observed a ~25 km landward shift in the updip limit of relocated seismicity at central Simeulue. This may reflect along-strike changes in coupling of the shallow portions of the fault. Using this a priori constraint, we therefore determined two different sets of forward models for period P3: 1) simple models with a uniform updip limit of full locking fixed at 18 km depth and two sections along strike that were each allowed to have their own downdip limit of locking (these models were already discussed in Text S1.2, and the preferred model of this set is shown in Figure 1.3a), and 2) complex models with a variable updip limit of full locking (mostly fixed at 18 km depth but that follows the ~25 km landward shift of the updip limit of seismicity at central Simeulue) and three sections along strike that are each allowed to have their own downdip limit of locking.

To determine a three-section forward model, we first adjusted the updip limit of full locking at central Simeulue to match in map view the landward shift in the updip limit of relocated seismicity. The deeper updip limit of full locking at central Simeulue lies at a depth of 26 km on our fault geometry. Starting from the preferred two-section model shown in Figure 1.3a, we added a section in the central part of the fault and tested a range of locking depths, along-strike locations and along-strike widths of the central section. We chose the model with the lowest WMSSR as our preferred model.

Table S1.4b lists the different configurations of the three-section models that we tested. Figure S1.6 shows a subset of our suite of three-section models that can explain the coral observations. Our results suggest that shallower locking depths of 30 and 35 km under the central part of Simeulue is needed to yield good data-model fits. The WMSSR of these models are lower than those of our two-section forward models, suggesting that greater along-strike variations in fault coupling are needed to improve data-model fits.

Figure S1.7 shows how the WMSSR changes as each model parameter is perturbed from the values shown in the preferred models. Because tradeoffs between parameters will increase as more complexity is added to these forward models, we chose to perform an inversion of the data in order to determine the spatial distribution of fault coupling for this period.

Text S1.4

Inverse models: Testing for the Laplacian weight

We applied Laplacian smoothing constraints to regularize the inversion. The strength of the smoothing constraints is tuned via the Laplacian weight. We tested a range of Laplacian weight values and selected the weight that yields a smooth spatial distribution of fault coupling. Figure S1.8 shows how the spatial distribution changes with various values of the Laplacian weight – all models suggest the same three key spatial patterns of coupling. First, the fault is locked downward to a greater depth in the SE part of the fault compared to the NW. Second, the fault is locked downward to a shallower depth under NW Simeulue, particularly near sites PPY and SLR. Third, the fault is less coupled under central Simeulue, near sites BUN, UTG and ULB. For the latter region of lower coupling, the degree of coupling increases with the Laplacian weight, ranging from uncoupled (freely slipping at the full plate subduction rate) for lower weights, to ~50% coupled for a weight of 50. We selected a Laplacian weight of 40, because it visually yields a smooth distribution of coupling.

Figure S1.1 (a) SLA are extracted from 30 grid points around Simeulue. (b) The time series of mean SLA for each of the 30 grid points are calculated with a +/- 30 day running time window (TW). (c) Each of the time series of mean SLA from (b) are differenced between each pair of grid points, in order to obtain a time series of mean SLA differences. (d) The 95th percentile of mean SLA differences is 17.7 mm.

Figure S1.2 (a) The dip profile of the fault approximates the Slab 1.0 model of *Hayes et al.* [2012] along the Simeulue section of the Sunda megathrust. Regional seismicity is based on *Tilmann et al.* [2010]’s earthquake catalog for SE Simeulue. (b) An oblique view showing a sample geometry of along-strike changes in locking depth: from 50 km depth on the NW section of the fault, to 30 km depth on the SE section. The shallowest 18 km of the fault is assigned a

coupling ratio of 0.4. The transition to this partially coupled zone is consistent with the updip limit of seismicity reported by *Tilman et al.* [2010] (white circles).

Figure S1.3 (a)–(c): Suite of possible models that can explain coral rates in period P1 (prior to 1819). Our results suggest that that the fault was locked to a depth of 33 to 35 km in the NW section, and 30 km in the SE section. (d)–(f): Models of uniform locking depths of (a) 30 km, (b) 31 km, and (c) 32 km across the fault yield poor data-model fits. Additional models are listed in Table S1.2.

Figure S1.4 (a)–(h): Subset of suite of possible models that can explain the coral rates for period P2 (1839–1861). Our results show that the data can be explained by a fault locked to 40 to 55 km under the length of Simeulue. (i)–(p): Models showing how data-model fits (WMSSR) change as model parameters are perturbed one at a time from the best fit model values shown in (i). The best fit model shows an updip limit of full locking at 18 km depth, locking depths of 55 and 30 km in the NW and SE sections of the fault respectively, and the location of the boundary of locking depth change located 55 km from the SE end of the fault. (j) The updip limit of full locking is perturbed from a depth of 18 km to 16 km. (k)–(l): The locking depth under the NW section is perturbed from a depth of 55 km to: (k) 50 km, and (l) 60 km. (m)–(n): The locking depth under the SE section is perturbed from a depth of 30 km to: (m) 25 km, and (n) 35 km. (o)–(p): The location of the boundary of locking depth change is perturbed from a position 55 km from the SE end of the fault to: (o) 65 km, and (p) 45 km from the SE end of the fault. Additional models are listed in Table S1.3.

Figure S1.5 Two-section models for period P3 (pre-2004) showing how data-model fits change as model parameters are perturbed one at a time from the best-fit two-section model values shown in (a). The best-fit model shows an updip limit of full locking at 18 km depth, locking depths of 40 and 50 km under the NW and SE sections, respectively. (b) The updip limit of full locking is perturbed from a depth of 18 km to 16 km. (c)–(d): The locking depth under the NW section is

perturbed from a depth of 40 km to: (c) 35 km, and (d) 45 km. (e)–(f): The locking depth under the SE section is perturbed from a depth of 50 km to: (e) 45 km, and (f) 55 km. (g)–(h): The boundary of locking depth change is perturbed from a position 55 km from the SE end of the fault, to (g) 65 km, and (h) 45 km from the SE end of the fault. Additional models are listed in Table S1.4a.

Figure S1.6 (a)–(f) Subset of suite of possible models that can explain the coral rates for period P3 (pre-2004), assuming a model with three sections, and a variable updip limit of full locking. Our results show that the data can be fit better with a section under the central part of Simeulue with a locking depth of 30 to 35 km. Models are referred to in Table S1.4b.

Figure S1.7 Models for P3 (pre-2004) showing how data-model fits change as model parameters are perturbed one at a time from the best-fit three-section model values shown in (a). The best-fit model shows an updip limit of full locking at 18 km depth except under central Simeulue, and locking depths of 40, 30 and 50 km in the NW, central, and SE sections of the fault, respectively. The central

section is 20 km wide. (b) The updip limit of full locking is perturbed from a depth of 18 km to 16 km. (c) The locking depth under the NW section is perturbed from 40 to 35 km. (d) The locking depth under the central section is perturbed from 30 to 35 km. (e) The locking depth under the SE section is perturbed from 50 to 45 km. (f)–(g): The boundary of the central section is moved an along-strike distance of 10 km (f) farther SE, and (g) farther NW. (h) The central section is widened by 10 km to the northwest. Additional models are listed in Table S1.4b.

Figure S1.8 (a)–(e): Inverse models with values of the Laplacian weight (LW) ranging from 10 to 50. For all models, the same key patterns of coupling can be observed. See Text S1.4 for a discussion.

Table 1.1 Coral site locations and interseismic vertical displacement rates

Site	Longitude	Latitude	Time period	Uplift rate (mm/yr) [#]	Site name (from Meltzner et al. [2010,2012,2015])	Notes
Langi LNG	95.72130	2.82592	1961-1997 AD	-0.1 ±2.2	LNG-A	Weighted average ^{##}
			1962-1998 AD	-3.3 ±2.2	LNG-A	
			1961-1998 AD	-1.7 ±3.2	-	
Lewak LWK	95.79091	2.92827	1301-1355 AD	-6.1 ±1.5	LWK-B	Weighted average ^{##}
			1397-1450 AD	-1.5 ±1.5	LWK-B	
			1951-1997 AD	-3.3 ±1.7	LWK-A	
			1962-1998 AD	-1.4 ±2.2	LWK-A	
			1951-1998 AD	-2.6 ±1.8	-	
Lhok Pauh LKP	95.76324	2.86160	1346-1394 AD	-3.6 ±1.7	LKP-B	Weighted average ^{##}
			1409-1430 AD	-4.4 ±3.8	LKP-B	
			1430-1450 AD	-3.5 ±4.0	LKP-B	
			1945-1997 AD	-5.3 ±1.5	LKP-A	
			1962-1998 AD	-5.2 ±2.2	LKP-A	
			1945-1998 AD	-5.3 ±1.2	-	
Lhok Dalam LDL	95.70072	2.74984	1327-1394 AD	-5.9 ±1.2	LDL-B	Weighted average ^{##}
			1982-1997 AD	-6.2 ±5.3	LDL-A	
			1983-1998 AD	-8.7 ±5.3	LDL-A	
			1982-1998 AD	-7.5 ±3.8	-	
Ujung Sanggiran USG	95.86741	2.91213	1413-1450 AD	-7.2 ±2.2	USG-A	Weighted average ^{##}
			1958-1997 AD	-2.2 ±2.1	USG-A	
			1962-1998 AD	-2.2 ±2.2	USG-A	
			1958-1998 AD	-2.2 ±1.5	-	
Ujung Salang USL	95.75935	2.70612	886-956 AD	-4.7 ±1.1	USL-A	Weighted average ^{##}
			1961-1997 AD	-7.6 ±2.2	USL-A	
			1962-1998 AD	-7.1 ±2.2	USL-A	
			1961-1998 AD	-7.4 ±1.6	-	
Pulau Penyu PPY	95.94407	2.85309	1449-1488 AD	-1.6 ±2.1	PPY-A	
			1488-1537 AD	-1.2 ±1.6	PPY-A	
			1537-1565 AD	-5.6 ±2.9	PPY-A	
			1928-1980 AD	-3.9 ±1.5	PPY-A	
			1980-2002 AD	+0.3 ±3.6	PPY-A	
Ujung Lambajo ULB	95.99508	2.56620	1961-1997 AD	-7.4 ±2.2	ULB-A	
Silinggar SLR	96.14749	2.70812	1800-1849 AD	-0.9 ±1.6	SLR-A	
			1849-1861 AD	-4.8 ±6.7	SLR-A	
			1961-1997 AD	-4.2 ±2.2	SLR-A	
Sambay SMB	96.19470	2.65109	1760-1819 AD	-0.8 ±1.4	SMB-A	
			1819-1861 AD	-8.7 ±1.9	SMB-A	
Bunon	96.14427	2.51291	870-1024 AD	-0.5 ±0.5	BUN-A	

BUN			1311-1340 AD	-2.2 ±2.8	BUN-A	Average rate
			1353-1422 AD	-6.6 ±1.2	BUN-A	
			1433-1466 AD	-0.3 ±2.4	BUN-A	
			1481-1516 AD	-5.8 ±2.3	BUN-A	
			1516-1538 AD	-10.1 ±3.6	BUN-A	
			1545-1576 AD	-5.6 ±2.6	BUN-A	
			1481-1576 AD	-6.0 ±0.8	BUN-A	
			1982-1997 AD	-5.3 ±5.3	BUN-A	
			1982-1997 AD	-8.9 ±5.3	BUN-B	
			1982-1997 AD	-7.1 ±3.8	-	
Ujung Tinggi UTG	96.27640	2.57304	1757-1839 AD	-1.1 ±1.0	UTG-A	Weighted average ^{##}
			1839-1861 AD	-7.0 ±3.6	UTG-A	
			1982-1997 AD	-4.2 ±5.3	UTG-A	
Labuhan Bajau LBJ	96.48623	2.40321	1738-1838 AD	-1.6 ±0.8	LBJ-A	
			1838-1861 AD	-6.1 ±3.5	LBJ-A	
			1945-1997 AD	-8.2 ±1.5	LBJ-A	

= The 2σ total uncertainty equals the 2σ level of uncertainties due to differential coral erosion and growth rates (calculated with the methodology of *Meltzner et al.* [2012]).

= For period P3, two independent estimates of the pre-2004 rate were reported at seven coral sites: LNG, LWK, LKP, LDL, USG, USL and BUN.

For each of these sites, we calculated a weighted mean rate (WMR) (discussed in Text S1.1).

Table S1.1 Additional uncertainties in coral vertical displacement rates, as a result of taking into account spatial variations in sea level within this region, as quantified with modern satellite altimetry data (Text S1.1, Figure S1.1). Total uncertainties equal the uncertainties reported by *Meltzner et al.* [2015] and the additional uncertainties, added in quadrature. The maximum change in the uncertainties after accounting for spatial variations in sea level is ± 0.2 mm/yr.

Site	Longitude	Latitude	Time period	No. years	Uplift rate (mm/yr) [#]	Additional uncertainty	Total uncertainty
Langi LNG	95.72130	2.82592	1961-1997 AD		-0.1 $\pm$ 2.2		
			1962-1998 AD		-3.3 $\pm$ 2.2		
			1961-1998 AD	37	-1.7 $\pm$ 3.2	0.5	3.2
Lewak LWK	95.79091	2.92827	1301-1355 AD	54	-6.1 $\pm$ 1.5	0.3	1.5
			1397-1450 AD	53	-1.5 $\pm$ 1.5	0.3	1.5
			1951-1997 AD		-3.3 $\pm$ 1.7		
			1962-1998 AD		-1.4 $\pm$ 2.2		
			1951-1998 AD	47	-2.6 $\pm$ 1.8	0.4	1.8
Lhok Pauh LKP	95.76324	2.86160	1346-1394 AD	48	-3.6 $\pm$ 1.7	0.4	1.7
			1409-1430 AD	21	-4.4 $\pm$ 3.8	0.8	3.9
			1430-1450 AD	20	-3.5 $\pm$ 4.0	0.9	4.1
			1945-1997 AD		-5.3 $\pm$ 1.5		
			1962-1998 AD		-5.2 $\pm$ 2.2		
			1945-1998 AD	53	-5.3 $\pm$ 1.2	0.3	1.2
Lhok Dalam LDL	95.70072	2.74984	1327-1394 AD	67	-5.9 $\pm$ 1.2	0.3	1.2
			1982-1997 AD		-6.2 $\pm$ 5.3		
			1983-1998 AD		-8.7 $\pm$ 5.3		
			1982-1998 AD	16	-7.5 $\pm$ 3.8	1.1	4.0
Ujung Sanggiran USG	95.86741	2.91213	1413-1450 AD	37	-7.2 $\pm$ 2.2	0.5	2.3
			1958-1997 AD		-2.2 $\pm$ 2.1		

			1962-1998 AD		-2.2 ±2.2		
			1958-1998 AD	40	-2.2 ±1.5	0.4	1.6
Ujung Salang USL	95.75935	2.70612	886-956 AD	70	-4.7 ±1.1	0.3	1.1
			1961-1997 AD		-7.6 ±2.2		
			1962-1998 AD		-7.1 ±2.2		
			1961-1998 AD	37	-7.4 ±1.6	0.5	1.7
Pulau Penyu PPY	95.94407	2.85309	1449-1488 AD	39	-1.6 ±2.1	0.5	2.1
			1488-1537 AD	49	-1.2 ±1.6	0.4	1.6
			1537-1565 AD	28	-5.6 ±2.9	0.6	3.0
			1928-1980 AD	52	-3.9 ±1.5	0.3	1.5
			1980-2002 AD	22	+0.3 ±3.6	0.8	3.7
Ujung Lambajo ULB	95.99508	2.56620	1961-1997 AD	36	-7.4 ±2.2	0.5	2.3
Silinggar SLR	96.14749	2.70812	1800-1849 AD	49	-0.9 ±1.6	0.4	1.6
			1849-1861 AD	12	-4.8 ±6.7	1.5	6.9
			1961-1997 AD	36	-4.2 ±2.2	0.5	2.3
Sambay SMB	96.19470	2.65109	1760-1819 AD	59	-0.8 ±1.4	0.3	1.4
			1819-1861 AD	42	-8.7 ±1.9	0.4	1.9
Bunon BUN	96.14427	2.51291	871-1024 AD	153	-0.5 ±0.5	0.1	0.5
			1311-1340 AD	29	-2.2 ±2.8	0.6	2.9
			1353-1422 AD	69	-6.6 ±1.2	0.3	1.2
			1433-1466 AD	33	-0.3 ±2.4	0.5	2.5
			1481-1516 AD	35	-5.8 ±2.3	0.5	2.4
			1516-1538 AD	22	-10.1 ±3.6	0.8	3.7
			1545-1576 AD	31	-5.6 ±2.6	0.6	2.7
			1481-1576 AD	95	-6.0 ±0.8	0.2	0.8
			1982-1997 AD		-5.3 ±5.3		

			1982-1997 AD		-8.9 ±5.3		
			1982-1997 AD	15	-7.1 ±3.8	1.2	4.0
Ujung Tinggi UTG	96.27640	2.57304	1757-1839 AD	82	-1.1 ±1.0	0.2	1.0
			1839-1861 AD	22	-7.0 ±3.6	0.8	3.7
			1982-1997 AD	15	-4.2 ±5.3	1.2	5.4
Labuhan Bajau LBJ	96.48623	2.40321	1738-1838 AD	100	-1.6 ±0.8	0.2	0.8
			1838-1861 AD	23	-6.1 ±3.5	0.8	3.6
			1945-1997 AD	52	-8.2 ±1.5	0.3	1.5

= The 2σ total uncertainty equals the 2σ level of uncertainties due to differential coral erosion and growth rates (calculated with the methodology of *Meltzner et al.* [2012]).

Table S1.2 Best models and suite of possible two-section models for period P1, from testing all combinations of along-strike locking depth changes.

Change in locking depth (km)	Location of along-strike boundary (in km, relative to SE edge of fault)	Updip limit of locking (km)	WMSSR value (n = 3)	Figure number
Suite of models				
33 -- 30	85	18	0.38	S3a
34 -- 30	95	18	0.26	S3b
35 -- 30	100	18	0.22	S3c/2a
Models with a uniform locking depth (poor data-model fits)				
30 -- 30	-	18	10.41	S3d
31 -- 31	-	18	2.93	S3e
32 -- 32	-	18	3.94	S3f
Other tested models (poor data-model fits)				
25 -- 30	155	18	10.08	-
25 -- 31	175	18	2.92	-
25 -- 33	115	18	11.57	-
25 -- 34	105	18	20.35	-
25 -- 35	105	18	29.71	-
26 -- 30	165	18	10.20	-
26 -- 31	185	18	2.92	-
27 -- 31	185	18	2.93	-
28 -- 31	195	18	2.93	-
29 -- 31	195	18	2.93	-
30 -- 25	5	18	10.48	-
30 -- 31	205	18	2.93	-
30 -- 33	45	18	9.25	-
30 -- 34	45	18	9.25	-
30 -- 35	35	18	9.27	-
33 -- 31	105	18	1.61	-

Table S1.3 Best models and suite of possible two-section models for period P2, from testing all combinations of along-strike locking depth changes. In bold: preferred model for period P2.

Change in locking depth (km)	Location of along-strike boundary (in km, relative to SE edge of fault)	Updip limit of locking (km)	WMSSR value (n = 3)	Figure number
Suite of models				
55 -- 30	55	18	0.34	2b/S4i
55 -- 35	75	18	0.46	-
55 -- 40	95	18	0.75	-
55 -- 45	115	18	1.02	-
55 -- 50	175	18	1.25	-
50 -- 30	55	18	0.49	S4a
50 -- 35	65	18	0.63	-
50 -- 40	85	18	0.86	-
50 -- 45	105	18	1.07	S4b
45 -- 30	45	18	0.97	S4c
45 -- 35	55	18	1.10	-
45 -- 40	65	18	1.23	S4d
30 -- 55	165	18	1.52	-
30 -- 45	195	18	1.34	-
35 -- 55	165	18	1.51	S4e
50 -- 50	-	18	1.26	S4f
45 -- 45	-	18	1.31	S4g
40 -- 55	155	18	1.49	S4h
45 -- 55	5	18	1.32	-
50 -- 55	5	18	1.26	-
Other tested models (poor data-model fits)				
40 -- 30	15	18	3.02	-
40 -- 35	15	18	3.03	-
35 -- 30	5	18	12.32	-
30 -- 35	215	18	12.31	-
40 -- 40	-	18	3.03	-
55 -- 30	55	15	0.51	-

55 -- 30	55	16	0.45	S4j
55 -- 30	55	17	0.36	-
55 -- 30	55	20	0.77	-
50 -- 30	55	18	0.49	S4k
60 -- 30	55	18	0.36	S4l
55 -- 25	55	18	2.19	S4m
55 -- 35	55	18	0.76	S4n
55 -- 30	65	18	0.48	S4o
55 -- 30	45	18	0.53	S4p

Table S1.4 Best models and suite of possible models for period P3, assuming models with (a) two sections, and a uniform updip limit of full locking, (b) three sections, and a variable updip limit of full locking that matches the updip limit of regional seismicity. In bold: preferred models out of the two sets of forward models that we tested for period P3 (details in Text S1.3).

Change in locking depth (km)	Location of along-strike boundary (in km, relative to SE edge of fault)	Updip limit of locking (km)	WMSSR value (n = 12)	Figure number
Suite of models				
35 -- 40	195	18	1.73	-
40 -- 40	-	18	1.78	-
40 -- 30	5	18	1.79	-
40 -- 35	5	18	1.79	-
40 -- 45	65	18	1.73	-
40 -- 50	55	18	1.73	3a, S5a
40 -- 55	55	18	1.73	-
Other tested models (poor data-model fits)				
30 -- 55	155	18	7.43	-
30 -- 50	175	18	3.42	-
30 -- 45	175	18	2.59	-
35 -- 35	-	18	6.85	-
35 -- 55	155	18	3.20	-
35 -- 50	155	18	2.85	-
35 -- 30	5	18	6.87	-
35 -- 45	165	18	2.38	-
45 -- 55	5	18	4.48	-
45 -- 50	5	18	4.48	-
45 -- 45	-	18	4.48	-
45 -- 30	5	18	4.49	-
45 -- 35	145	18	2.59	-
50 -- 55	5	18	7.37	-
50 -- 30	35	18	7.30	-
50 -- 35	155	18	2.51	-

50 -- 50	-	18	7.37	-
55 -- 30	125	18	11.29	-
55 -- 35	155	18	2.67	-
40 -- 50	55	15	1.34	-
40 -- 50	55	16	1.30	S5b
40 -- 50	55	17	1.41	-
40 -- 50	55	20	4.31	-
35 -- 50	55	18	6.00	S5c
45 -- 50	55	18	4.50	S5d
40 -- 45	55	18	1.74	S5e
40 -- 55	55	18	1.73	S5f
40 -- 50	65	18	1.74	S5g
40 -- 50	45	18	1.73	S5h

Chapter 2

A 15 year slow-slip event on the Sunda megathrust offshore Sumatra

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Abstract

In the Banyak Islands of Sumatra, coral microatoll records reveal a 15-year-long reversal of interseismic vertical displacement from subsidence to uplift between 1966 and 1981. To explain these coral observations, we test four hypotheses, including regional sea-level changes and various tectonic mechanisms. Our results show that the coral observations likely reflect a 15-year-long slow slip event (SSE) on the Sunda megathrust. This long duration SSE exceeds the duration of previously reported SSEs and demonstrates the importance of multi-decade geodetic records in illuminating the full spectrum of megathrust slip behavior at subduction zones.

2.1 Introduction

‘Silent earthquakes’, or slow slip events (SSEs), release strain on tectonic faults at slip rates faster than plate rates, but slowly enough that seismic shaking is not generated. They are manifested along subduction zone megathrusts either as pre-seismic slip prior to large earthquakes, afterslip following large earthquakes, or slip during the interseismic period of the earthquake cycle [Schwartz and Rokosky, 2007]. In this study, we use the term SSE to refer only to the last of these three manifestations.

SSEs are visible in geodetic time series as reversals of displacements associated with a locked fault interface. SSEs have been observed to typically last from a few days (e.g., Douglas *et al.* [2005]) to a few months (e.g., Hirose *et al.* [1999]), but longer duration SSEs lasting several years have also been detected: ~2–3 years [Ohta *et al.*, 2006], ~4 years [Fu and Freymueller, 2013], ~5 years [S Miyazaki *et al.*, 2006; Ochi and Kato, 2013; Ohta *et al.*, 2004; Ozawa *et al.*, 2002] and ~10 years [Li *et al.*, 2014]. In order to detect SSEs of longer duration and thus illuminate the full temporal spectrum of megathrust slip behavior, multi-decade geodetic records are required. However, existing geodetic networks have not been operating long enough to detect such events.

One way to extend our geodetic observations is to take advantage of paleogeodetic data preserved in the geological record. For example, coral microatolls along the Sumatran subduction zone provide high-resolution records of vertical deformation over the past several centuries (e.g., Meltzner *et al.* [2010]; Philibosian *et al.* [2014]; Zachariasen *et al.* [2000]). Using these data, Meltzner *et al.* [2015] reconstructed interseismic uplift and subsidence at multiple sites over the past three centuries. At one site on Bangkaru Island, in the Banyak Islands (site PBK, Figure 2.1), coral records show an abruptly reversed

displacement trend from interseismic subsidence to interseismic uplift between 1966 and 1981 (Figures 2.1b–d); an SSE on the underlying megathrust is a potential explanation.

In this paper we extend the work of *Meltzner et al.* [2015] to examine the potential physical processes that can explain the coral observations. We show that indeed they likely reflect a very long duration SSE on the Sunda megathrust, where, to date, no SSEs other than afterslip following major earthquakes have been recorded. We include previously unpublished coral observations for the Simeulue-Nias section of the megathrust, including additional sites in the Banyak Islands and neighboring islands of Simeulue and Nias. These added observations allow us to model the possible locations and sizes of this SSE, both with improved constraints.

Figure 2.1 (a) The Banyak Islands section was estimated to have experienced less coseismic slip than elsewhere along strike during the 2005 M_w 8.6 Nias rupture [Konca *et al.*, 2007]. Part of this section ruptured during the 2010 M_w 7.8 Banyak Islands earthquake (source model from the USGS: http://earthquake.usgs.gov/earthquakes/eventpage/usp000hat0\#scientific_finite_fault). M_w 6+ earthquakes from the NEIC catalog are plotted. (b) to (d) Spatiotemporal distribution of vertical displacement rates recorded by coral microatolls in the Simeulue-Nias section. From 1966 to 1981, coral microatolls on the Banyak Islands reversed their sense of displacement from interseismic subsidence to interseismic uplift. Black lines: Slab contours at 20 km depth intervals from Slab 1.0 [Hayes *et al.*, 2012].

2.2 Displacement rate changes recorded by corals

We reconstructed interseismic uplift histories from coral microatolls at six sites along the Simeulue-Nias section of the Sunda megathrust (Figure 2.1). Microatolls are coral colonies that grow upward to a limit near mean low water springs (MLWS), and their upper surfaces can track changes in relative sea level (RSL) over time [Meltzner *et al.*, 2010; Scoffin *et al.*, 1978; Taylor *et al.*, 1987; Zachariasen *et al.*, 2000]. Microatoll morphologies form because prolonged subaerial exposure at times of extreme low water limits the highest level to which the coral colonies can grow [Briggs *et al.*, 2006; Meltzner and Woodroffe, 2014; Meltzner *et al.*, 2010]. Flat-topped microatolls record RSL stability; colonies that rise radially outward toward their perimeter reflect rising RSL during their decades of growth, and colonies with progressively lower diedowns as they grow outward reflect gradually falling RSL.

Changes in RSL are themselves the combined effect of changes in land level and changes in local absolute sea level. As islands and their fringing reefs subside or rise in response to tectonic elastic strain accumulation and release, microatolls record changes in RSL. Because these corals' skeletons have annual growth bands, we can precisely calculate rates of change in RSL, when those changes are gradual.

The coral microatolls were slabbed, x-rayed, and analyzed following the methods described by Meltzner *et al.* [2010; 2012] and Meltzner and Woodroffe [2014]. The interpreted coral cross sections and corresponding time series for the relevant corals are presented in Figures S2.1–S2.6, and Table S2.1. The displacement rates at each site (and their 2σ level of uncertainties) over various time periods are shown in Figures 2.1b–d. Three sites (PBK, PLM, and PUB) define a trench-normal transect through the Banyak

Islands, and three additional sites are located near the southeastern tip of Simeulue (LBJ) and the northern tip of Nias (MZL and SBA).

The RSL history of site PBK is particularly well resolved as a result of various idiosyncrasies (as discussed by *Meltzner et al.* [2015]). Modern microatoll PBK-4 (Figure S2.1) started growing in the 1950s and first recorded a diedown in 1956. The first few diedowns were successively higher on the coral, but around 1966 there was an abrupt change, and for ~15 years successive diedowns were lower and lower. Around 1981, the trend reversed again, and successive diedowns were higher. After correcting for an assumed eustatic sea-level rise of 2 mm/yr, *Meltzner et al.* [2015] estimated tectonic deformation rates at PBK-4 of -2.7 ± 3.4 mm/yr (subsidence) from the beginning of the record in 1956 until 1966; $+7.8 \pm 2.1$ mm/yr (uplift) from 1966 to 1981; and -2.4 ± 0.8 mm/yr (subsidence) from 1981 until coseismic uplift in 2005 (Table S2.1). Although rates of sea-level change are not well known in the eastern Indian Ocean in the decades prior to 1992, there is no compelling evidence that they differed markedly from the global average of $+2 \pm 1$ mm/yr. In this paper, we adopt the rates (and correction) of *Meltzner et al.* [2015], but we explore alternatives in the supplementary materials (Text S2.2).

The abruptness of each rate change at site PBK is well constrained. *Meltzner et al.* [2015] estimate that they can resolve the timing of rate changes at PBK-4 to within ± 2 years of 1966 and 1981, respectively. Hence, these changes must have been rapid; each transition from the initial rate to the subsequent rate occurred over no more than 2 to 4 years, but it may have occurred over a period of a few months or even less.

Because of a combination of greater erosion and less frequent diedowns on all the other corals, the rates at the other sites are not as well constrained. However, they still provide

useful information. At sites LBJ, MZL, and SBA on Simeulue and Nias, no rate change occurs between the beginning of the modern record (1945 at LBJ, 1970 at MZL, and 1962 at SBA) and 2005 (Figures S2.4–S2.6). At PLM and PUB in the Banyak Islands, however, the time series are more ambiguous; rate changes may occur at each site around 1966 and 1981, but the uncertainties in the various rates are large enough that the rates before and after each change are statistically indistinguishable from one another at 2σ . At both PLM and PUB, we separately estimated rates for 1966–1981 and for 1981–2005 (Figure S2.2–2.S3), and we used those rates to constrain our models; we also estimated average rates over the duration of the modern coral record at each site.

2.3 Explaining the coral displacement rate changes

We identified and tested four scenarios that could explain the observed pattern of vertical displacement reversals recorded by the corals.

2.3.1 Scenario 1: Changes in sea level due to oceanographic processes

Although evidence exists for global or ocean-basin-wide fluctuations in sea level on multi-decadal timescales (e.g., *Chambers et al.* [2012]; *Woodworth et al.* [2009]), these cannot, by themselves, explain the changes in RSL observed in the Banyak Islands, because similar, contemporaneous changes would be expected at nearby sites on Simeulue and Nias but did not occur (details in Text S2.2).

If RSL varies significantly from site to site as a result of oceanographic processes, this could explain some and perhaps all of the observed site-to-site differences in the displacement rates. We therefore sought to quantify the magnitudes of spatial variations of sea level in this region, and treated them as an independent error for the coral rate uncertainties. We added this error in quadrature to the 2σ coral rate uncertainties determined by the method of *Meltzner et al.* [2012], and then calculated the overall

uncertainty associated with the displacement rate change at each site (details in Text S2.1). Our hypothesis is that if the summed uncertainties in the displacement rate changes at each site do not overlap at the 1σ confidence level, then the spatiotemporal changes in the coral displacement rates are unlikely due to oceanographic sea-level changes, and more likely to be consistent with tectonic changes on the underlying megathrust.

2.3.1.1 Scenario 1 methods: Investigating ocean variability with satellite altimetry data

We use satellite altimetry data from the past two decades as a proxy for the spatial variability of sea level over the 20th century. The data, distributed by Aviso (Archiving, Validation, and Interpretation of Satellite Oceanographic data), span January 1993 to April 2014, and consist of daily sea level anomalies (SLA) estimated on a global grid, at a spatial resolution of 0.25° by 0.25° (more details in Text S2.1). Due to the spatial resolution of the dataset, the data may not capture local coastal effects and related seasonal signals. Nevertheless, since regional tide-gauge records are not yet available to provide better constraints on such local effects, analysis of satellite altimetry data is the best alternative.

We first extracted the time series of daily SLA at a set of grid points in the Simeulue-Nias region (Figure S2.8a), and estimated time series of 2-month mean SLA (Figure S2.8b). Second, we calculated the difference in the 2-month mean SLA between each pair of grid points (Figure S2.8c). Third, we determined the 95th percentile of mean SLA differences between grid points (Figure S2.8d) – this represents the estimated maximum magnitude (at 2σ) of spatial variations in sea level in this region. Fourth, we calculated for each coral displacement rate an added amount of uncertainty, which is equal to the maximum magnitude of spatial variations in sea level divided by the length of time over which each displacement rate was determined. Finally, we ascertained whether any of the observed temporal rate changes overlapped between the sites, at the 1σ confidence level. Further

details of this analysis method, as well as analysis of the time series of SLA rates, are provided in Text S2.1.

2.3.1.2 Scenario 1 results: Uncertainty in coral displacement rates due to spatial variations of sea level

We estimated a mean SLA difference of 17.3 mm, at the 2σ confidence level (Figure S2.8d). The added uncertainty to each coral displacement rate, and the summed uncertainties, are detailed in Table S2.2, and discussed in Text S2.1. Our results suggest that the site-to-site and temporal variations in the coral displacement rates do not overlap at the 1σ uncertainties – that is, differences in sea-level change across our small study area seem unlikely to contribute significantly to the observed rate changes.

2.3.2 Scenario 2: Triggered postseismic deformation from regional earthquakes

Since viscoelastic deformation has a larger spatial signature than elastic coseismic deformation, distant large magnitude earthquakes have the potential to cause long-term changes in vertical displacement rates, without any discernable coseismic signals. An analogous example is the postseismic deformation recorded by the Sumatran GPS Array (SuGAR) following the 2012 M_w 8.6 Wharton Basin earthquake, in which most GPS stations experienced smaller coseismic than postseismic vertical displacements [Feng *et al.*, 2015b]. We tested the hypothesis that the 1966–1981 coral displacement rate changes are caused by viscoelastic deformation from a regional earthquake.

2.3.2.1 Scenario 2 methods and results

To support this hypothesis, we would first expect evidence of a large ($M > 8$) earthquake in the region, before the onset of the displacement reversal in 1966. However, we find no evidence of large, $M > 8$ earthquakes in the region from 1956 to 2004 (prior to the M_w 9.2 Sumatra-Andaman earthquake), from either instrumental, historical or paleoseismic records [Newcomb and McCann, 1987] (Figure 2.1a and Text S2.3). Second, viscoelastic relaxation processes would need to be able to explain the 15-year-long linear uplift rate at

site PBK. It is possible to observe an apparent linear uplift rate, if the mantle is sufficiently viscous (e.g., $>5 \times 10^{18}$ Pa s) and our observation time window represents a short time snapshot of the total length of the viscoelastic signal. However, the abrupt reversal of displacements back to subsidence in 1981 is inconsistent with what is expected from viscoelastic processes. We therefore refute the hypothesis that the rate changes observed at site PBK are significantly influenced by viscoelastic processes.

2.3.3 Scenario 3: Temporal changes in fault coupling patterns and downdip limits of locking

In the Sumatran subduction zone, temporal changes in long-term deformation rates have been reported to occur before or after major megathrust earthquakes (e.g., *Meltzner et al.* [2012]; *Meltzner et al.* [2010]; *Meltzner et al.* [2015]; *Philibosian et al.* [2014]; *Prawirodirdjo et al.* [2010]). Most of these rate changes maintained a sense of displacement consistent with locking on the underlying megathrust, and have been modeled as spatiotemporal variations in interseismic coupling and/or downdip limits of locking.

The downdip limit of locking at subduction zones controls the location of the pivot line that separates regions of interseismic subsidence from interseismic uplift. If the pivot line shifts under a particular site, the sense of displacement at that site would reverse. To test whether spatial variations in coupling and downdip limits of locking could explain the 1966–1981 uplift rates in the Banyak Islands, we developed a suite of forward models of fault coupling. We hypothesize that if the displacement rate changes are due to temporal changes in coupling patterns, then tectonically feasible coupling parameters and downdip limits of locking should be able to explain the observations for each of the three periods (1956–1966, 1966–1981 and 1981–2005).

2.3.3.1 Scenario 3 methods: Forward models of fault coupling–coupling models

To model deformation from spatial variations in fault coupling, we assumed a plate subduction rate of 40 mm/yr [McNeill and Henstock, 2014; Simons *et al.*, 2007], and calculated surface displacements with the Okada model of elastic dislocations [Okada, 1985], in a backslip framework [Savage, 1983] (other fault and model parameters are described in Text S2.4). In this framework, fully locked regions have a coupling ratio of 1, while parts of the fault that slip freely at the full plate subduction rate (40 mm/yr) have a coupling ratio of 0.

For all time periods, we assigned the coupling ratio along the shallowest 22 km of the fault to 0.4. Our rationale for including this partially coupled shallow region is discussed in Text S2.4. Its location is consistent with that of the updip aseismic/seismic transition zone reported by Tilmann *et al.* [2010], though we note that the shallow region of the megathrust here may well be capable of rupturing seismically, as suggested by the M 7.6 tsunami earthquake of 1907 [Kanamori *et al.*, 2010].

In order to fit the subsidence rates in the Banyak Islands, we included in the models a region of partial coupling in the downdip portion of the fault (a downdip transition zone). We tested for its updip limit, downdip limit and coupling ratio, with a range of values consistent with previous regional and global studies [McCaffrey *et al.*, 2000; Tichelaar and Ruff, 1993] (Text S2.4).

In addition, we estimated separate downdip limits of full coupling under Simeulue and the Banyak Islands, respectively, because such along strike variation was needed to substantially improve data-model fits at site LBJ (Figure S2.10, Text S2.4).

In summary, we determined best-fit models by varying the a) downdip limit of the fully coupled zone under Simeulue and the Banyak Islands, b) downdip limit of the downdip transition zone, and c) coupling ratio of the downdip transition zone. We selected the best-fit range of models based on the weighted mean of the sum of squared residuals (WMSSR), which is similar to a reduced chi-square value, but uses the number of observations rather than the number of degrees of freedom in the model.

Although we incorporated constraints based on regional seismicity [Tilman *et al.*, 2010] and global studies of transition zone depths at subduction zones [McCaffrey *et al.*, 2000; Tichelaar and Ruff, 1993], we are aware that the models are limited in resolution and include tradeoffs between the fault geometry, downdip limits of locking, downdip transition zone location, and coupling ratios. The models therefore do not represent optimized or full-scale inversions of the data, but rather a range that fit the coral observations within their 2σ uncertainties.

2.3.3.2 Scenario 3 results: Coupling models

Coupling models can explain the observed displacement rates during 1956–1966 and 1981–2005; Figures 2.2a and 2.2b show one such model. In this model, the portion of the fault under the Banyak Islands and northern Nias is locked to 30 km depth, while under southern Simeulue it is locked to 45 km depth. The downdip transition zone extends from depths of 30–60 km and 45–60 km under the Banyak Islands and southern Simeulue, respectively, and is coupled at 50% of the plate subduction rate. This model yields a WMSSR of 0.9. This represents our preferred coupling model, given the good data-model fits and comparisons with locking depth parameters determined from previous studies, although we note again that other models are possible given the limited model resolution (alternative models that fit the coral observations are listed in Table S2.3).

In contrast to models of deformation during periods of subsidence in the Banyak Islands, none of the tested coupling models do a good job of fitting the 1966–1981 displacement rates. Since the subsidence rates at sites on Simeulue and Nias remain unchanged between 1966 and 1981, we kept the same along-strike parameters under these sites (as shown in Figures 2.2a and 2.2b), and only varied parameters under the Banyak Islands in order to fit the uplift rates. To maximize uplift in the Banyak Islands, the underlying megathrust needs to be freely slipping at the full plate subduction rate (uncoupled). Thus, we included in the models a freely slipping patch under the Banyak Islands, and consecutively tested for the updip, downdip and along-strike limits of this patch.

Figure 2.2c shows our best attempt at such a model (additional models are described in Figure S2.12 and Table S2.4), which has a locking depth of 30 km along a 90 km section under the Banyak Islands, and creeps at the full plate subduction rate at depths greater than 30 km. This model yields a WMSSR of 2.5, and fails to explain the observed uplift rate at PBK. It is therefore difficult to explain the 1966–1981 uplift rates in the Banyak Islands with a model that only includes spatially variable coupling of the megathrust.

In addition, we performed an inversion of the data, to ensure that we had not overlooked any distribution of coupling that could explain the 1966–1981 displacement rates. The results of the inversion are available in Text S2.5 and Figure S2.14. Due to the limited spatial coverage of the data, and therefore limited model resolution, we refrain from interpreting the details of the slip distribution suggested by the inverse models, and prefer the relative simplicity of the forward models. However, on fault patches with reasonable model resolution, the first order spatial patterns of the inverse models compare well with those of the forward models. Also, they show that while the 1981–2005 displacement rates are fit well (Figure S2.14a), there are large misfits to the 1966–1981 uplift rate at

site PBK (Figure S2.14b). This supports our conclusion based on the forward models, that changes in coupling on the megathrust cannot explain the 1966–1981 uplift rates in the Banyak Islands.

Figure 2.2 Preferred model of spatially variable coupling on the megathrust, during (a) 1956–1966, and (b) 1981–2005. In both periods, the fault is locked down to 30 km depth under the Banyak Islands and northern Nias, and 45 km depth under southern Simeulue. The downdip transition zone extends to a maximum depth of 60 km, and is coupled at 50% of the subduction rate. Below 60 km, the fault is freely slipping at the full subduction rate. Everywhere along strike above 22 km depth, the fault is coupled at 40% of the subduction rate. (c) This model shows our best attempt to fit the 1966–1981 uplift rates with spatially variable coupling on the megathrust, but it nonetheless shows poor data-model

fits at site PBK. (d) At least one model with a slow slip event (slip rate exceeding the subduction rate) fits significantly better than any of our coupling models. Our preferred explanation for the 1966–1981 displacement rates includes a SSE localized along a 60 km section of the megathrust under the Banyak Islands, between 30 and 55 km depth.

2.3.4 Scenario 4: A slow slip event from 1966 to 1981

Since we could not explain the 1966–1981 displacement rates as spatially variable changes in coupling along the megathrust, we tested the hypothesis of an SSE on the underlying megathrust.

2.3.4.1 Scenario 4 methods

We forward modeled the 1966–1981 displacement rates as a superposition of a) backslip on the megathrust and b) an SSE with thrust motion on the megathrust, with a slip rate greater than the plate subduction rate. For the backslip component, we assumed the spatial distribution of locking depths and transition zone parameters previously presented in our preferred model for the periods of subsidence (Figures 2.2a and 2.2b). This assumption leads to the desirable consequence that the only difference between our preferred model for 1956–1966 and 1981–2005, and our preferred model for 1966–1981 (as discussed in the following paragraphs), is the SSE itself.

We determined a range of best-fit SSE models by testing for the best combination of the a) upper and lower depth ranges of the SSE, b) along-strike length and location of the SSE, and c) SSE slip rates. For the depth ranges of the SSE, we tested upper and lower SSE depths ranging from 15 to 50 km and 25 to 65 km, respectively, at 5 km depth increments. Sites LBJ, MZL and SBA allowed us to constrain the along-strike length and location of the SSE, to the nearest 10 km. We selected the range of SSE models that yield good fits between the data and model displacement rates. We again note that the models do not represent optimized inversions of the data, and the aim of our modeling approach was to explore whether an SSE could explain the 1966–1981 displacement rates within their 2σ uncertainties.

2.3.4.2 Scenario 4 results: Our slow slip event model

We found a range of SSE models that fit the 1966–1981 displacement rates (Table S2.5). Figure 2.2d shows our preferred SSE model, which yields the lowest WMSSR value. The results suggest that the SSE was localized along a 60 km section under the Banyak Islands, at depths of 30 to 55 km. The suite of best-fit models (Table S2.5) suggests alternative downdip limits of the SSE ranging from 45 to 70 km, and SSE slip rates ranging from 55 to 61 mm/yr (WMSSR values range from 1.2 to 1.3). Thus, the SSE likely occurred within the partially coupled downdip transition zone. Our preferred model suggests an SSE slip rate of 56 mm/yr, and an equivalent cumulative seismic moment of 1.15×10^{20} N m, corresponding to a moment magnitude of $\sim M_w$ 7.3 (assuming a rigidity of 30 GPa). The suite of best-fit models (Table S2.5) suggests SSE moment magnitudes ranging from M_w 7.2 to 7.4.

The inversion results also support our hypothesis for an SSE on the megathrust (Text S2.5 and Figure S2.14). Although the inversion models are sensitive to slip constraints, the results show that data-model fits improve significantly when the fault is allowed to slip at rates greater than the plate subduction rate.

2.4 Discussion

Our results show that an SSE is the most likely explanation for the 15-year-long interseismic displacement reversal from 1966 to 1981, and that the event occurred on the Banyak Islands portion of the megathrust, at depths of 30 to 55 km. This depth range is consistent with the majority of SSE depth ranges found at other subduction zones, within the downdip zone of transitional friction properties from locked to creeping behavior [Schwartz and Rokosky, 2007].

The Banyak Islands portion of the megathrust appears to release its strain via a combination of several distinct slip phenomena. In addition to the SSE, this portion has ruptured in localized patches (e.g., the 2010 M_w 7.8 Banyak Islands earthquake (Figure 2.1)), participated in large coseismic ruptures (e.g., the 2005 M_w 8.6 Nias earthquake), and experienced afterslip following large earthquakes (e.g., *Hsu et al.* [2006]). This range of slip behavior may reflect distinctly different frictional properties on the Banyak Islands portion of the megathrust [*Briggs et al.*, 2006]. Probing the physical processes and/or structures controlling these frictional properties remains a relevant topic for future work.

The SSE model suggests an equivalent cumulative seismic moment of 1.15×10^{20} N m, and corresponding moment magnitude of $\sim M_w$ 7.3, assuming a rigidity of 30 GPa. However, we note that equivalent seismic moment estimates are sensitive to modeling methods and assumed rigidities. For example, if we assume a higher rigidity of 75 GPa (representing the minimum rigidity estimate at >30 km depths from the CRUST2.0 model in the Sumatran subduction zone), our SSE model yields an equivalent seismic moment of 2.87×10^{20} N m, corresponding to $\sim M_w$ 7.6. Figure 2.3 shows that our moment estimates for this SSE do not fall far from *Ide et al.* [2007] 's SSE moment-duration scaling relations. This result is reasonable, given that with our limited dataset, moments cannot be estimated robustly.

There are a number of other studies reporting SSEs with durations longer than expected from current moment-duration relations, especially in Japan and Alaska. For example, *Ochi and Kato* [2013] obtained a revised equivalent moment magnitude of M_w 6.6 for the ~ 5 -year long Tokai SSE, corresponding to an equivalent seismic moment at least 16 times smaller than predicted by the lower-limit envelope of *Ide et al.* [2007]'s moment-duration

relations. As more long duration SSEs are detected and studied, we speculate that we may gain new insights into the scaling relations of long duration SSEs.

The Sumatran GPS Array (SuGAR) has been operating for 12 years, from 2002 to 2014. *Feng et al.* [2015a] have carefully analyzed these 12-year-long time series, and find no evidence for short-term SSEs. From the currently available data, it appears that the megathrust hosts long duration SSEs during the interseismic period, but no SSEs that coincide with the modern cluster of seismic ruptures. Why might this be? Perhaps the megathrust switches its mode of strain release in the downdip region between afterslip and long duration SSEs: afterslip during the modern stage of the seismic cycle, and SSEs during more quiescent time periods. Alternatively, perhaps similar long duration SSEs do occur during large earthquake clusters along portions of the fault with varying frictional properties, but they have not yet been detected. Resolving these possibilities will require continued and expanded geodetic monitoring.

Figure 2.3 SSE moment-duration scaling relations of *Ide et al.* [2007] (upper and lower limit envelopes shown by black lines) and SSEs at various subduction zones, compiled by *Peng and Gomberg* [2010] and *Schwartz and Rokosky* [2007] (events listed in Table S2.6). The lower and upper moment estimates of our SSE model assume rigidities of 30 and 75 GPa, respectively. The SSE does not fall far from the scaling relations of *Ide et al.* [2007]. SSEs with a range of reported moment estimates are shown by crosses on the circles, with connecting dashed lines.

2.5 Conclusions

Our results show that an SSE is the most likely explanation for a 15-year-long interseismic displacement reversal from 1966 to 1981, as recorded by coral data. The SSE

occurred on the Banyak Islands portion of the megathrust, at 30 to 55 km depth, within the downdip transition zone. A range of slip behavior occurs on the portion of megathrust under the Banyak Islands, likely reflecting its different structural and frictional properties compared to adjacent parts of the Simeulue-Nias section of the megathrust.

2.6 Acknowledgements

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2.7 Supporting Information

Introduction

This supplementary information contains additional details of the coral data and the modeling methods for the various scenarios that we present in the paper. We include sketches of the coral slabs and time series at each of the sites in our study area (Figures S2.1 to S2.6; displacement rates are also listed in Table S2.1). We provide complete details of our analysis of the Aviso data and calculation of additional rate uncertainties due to spatial variations in sea level (Text S2.1, Table S2.2). We also present our analysis of the time series of mean SLA and SLA rates from the Aviso data in Text S2.1, Figures S2.8 and S2.9. A discussion of corrections for modern sea-level rise can be found in Text S2.2. We provide further details of our fault geometry and justification of our modeling assumptions (Text S2.4, Figures S2.7 and S2.10), alternative backslip and SSE models (Figures S2.12, Tables S2.3 to S2.5), and sensitivity tests of our models (Figures S2.11 and S2.13). Methods and results of our inversion models can be found in Text S2.3 and Figure S2.14. Table S2.6 lists the subduction zone slow slip events that are plotted in Figure 2.3.

Text S2.1

Investigating local differences in sea level due to oceanographic processes

An intriguing observation of our coral dataset is the spatial variation of the rate changes: rates at sites in the Banyak Islands changed around 1966 and 1981, but significant changes are not observed at these times on southern Simeulue or northern Nias. Because rates of RSL change reflect changes both in land level and in absolute sea level, it is important to consider how much of this could result from changes in local sea level.

The 2σ (95%) level uncertainties listed in Table S2.1 were determined following the methodology of *Meltzner et al.* [2012; 2015] and consider uncertainties in the calculated rates that may arise from coral irregularities, but not from variability in sea level itself. At a single site, individual coral microatolls may have experienced differential erosion rates, or may have contrasting morphologies due to differential growth rates. These two physical factors lead to potential irregularities in a single coral colony that could bias its RSL history. *Meltzner et al.* [2012] addressed this issue by comparing RSL histories of coeval coral microatolls at a particular site, and they determined the maximum difference in the HLG (highest level of growth) between corals, within any year. They found a worst-case scenario error of 80 mm (which they conservatively treated as an error at the 95% confidence level) in the apparent elevation gain of a single coral colony due to differential erosion and growth rates. Based on these observations, *Meltzner et al.* [2012] suggested a formal method to calculate the 2σ error in the rates of RSL change determined from coral microatolls off the west coast of Sumatra: this worst-case scenario error (80 mm) is divided by the length of time over which each coral displacement rate was determined.

We also wish to consider the extent to which spatial variations in the coral vertical displacement rates can be explained by spatial variations in sea level. In a manner parallel to that of *Meltzner et al.* [2012], we considered the difference between the elevation gains recorded by corals at two sites over a particular time interval, and we desired to estimate, at the 95% confidence level, how much of that difference could be attributable to differences in absolute sea level at the two sites.

For this purpose, we analyzed modern satellite altimetry data to quantify spatial variations of sea level in the Simeulue-Nias region. We selected the delayed-time mode, all-satellite merged product of globally gridded sea level anomalies (SLAs), produced by Ssalto/DUACS, and distributed by Aviso, with support from Cnes (www.aviso.altimetry.fr/duacs). The data span January 1993 to April 2014, and consist of daily SLAs estimated on a global grid, at a spatial resolution of 0.25° by 0.25° . We first extracted the time series of daily SLAs at 25 grid points in a 1.25° by 1.25° region centered between Simeulue and Nias (Figure S2.8a). In order to smooth out unmodeled high frequency ocean signals and aliasing effects in the data, we estimated the mean SLA for each day with a ± 30 day running time window, to obtain time series of 2-month mean SLAs at daily resolution (Figure S2.8b). In principle, this reduces the contribution of aliased signals with periods less than 2 months [*Ponte and Lyard*, 2002]. Next, we calculated the difference in the SLA between each pair of grid points—for example, grid point 1 was differenced with each of the remaining 24 grid points (Figure S2.8c). We then determined the 95th percentile of mean SLA differences between grid points. Figure S2.8d shows the cumulative frequency histogram of mean SLA differences between grid points, and the values containing 65 and 95 percent (65th and 95th percentile) of the data. The 95th percentile of mean SLA differences is 17.3 mm. We caution that near-coastal shallow-water areas can have deviations from deeper-water open-ocean sea levels that are not

captured by Aviso's gridded product, but these deviations are not well understood or modeled. Nonetheless, most of our sites (PBK, PUB, SBA, MZL, and LBJ) are within 3 km of water that is 20 m deep or deeper, and they are within 10 km of 50-m-deep water. PLM is in a somewhat shallower setting, being 7 km from the 20-m bathymetric contour, and 11 km from 50-m water depths.

We treated the magnitude of these spatial variations in SLAs as an independent error and added this error in quadrature to the uncertainties due to potential coral irregularities that we previously calculated following *Meltzner et al.* [2012]. These modified errors are given in Table S2.2. Our hypothesis is that if the summed uncertainties in the displacement rate changes at each site do not overlap at the 1σ confidence level, then the spatiotemporal changes in the coral displacement rates are unlikely due to oceanographic sea-level changes, and more likely consistent with tectonic changes on the underlying megathrust.

As we can see in Table S2.2, considering both sources of error, the rate change at PBK around 1966 was $+10.6 \pm 4.5$ mm/yr (2σ). This differs significantly from the permissible rate change at site LBJ, 0.0 ± 2.2 mm/yr. The rate change at PLM is not well constrained, but if only 1σ errors are considered, it does not overlap the rate change at PBK. In 1981, the rate change at PBK, -10.2 ± 2.7 mm/yr (2σ), differs significantly from the permissible rate changes at sites LBJ, SBA, and MZL. Again, the rate changes at PLM and PUB are not well determined, but if only 1σ errors are considered, neither overlaps the rate change at PBK.

Investigating spatiotemporal changes in SLA rates

We also sought to understand the spatiotemporal changes in SLA rate in our study area. The same globally gridded SLA dataset was used. For each grid point in our study area, we computed the time series of SLA rates. In order to smooth out the contribution of seasonal and interannual oceanographic signals, we computed the daily SLA rates with a running time window of ± 8 years. This length of time window averages out interannual sea surface signals with periods of less than ~ 16 years. To illustrate the sensitivity of the results to the time window length, we have presented in Figure S2.9 the results of using time windows ranging from ± 5 to ± 10 years. Because we do not fully understand the seasonal and interannual oceanographic processes in this region, we caution against detailed interpretation of these time series of SLA rates. Furthermore, due to the spatial resolution of the dataset (and spatial interpolation), local coastal effects may not be captured in the Aviso data. Understanding such local processes awaits data from regional tide gauges or future satellite missions. Therefore, the objective of this analysis was to show only to first order the spatial differences in sea-level trends across this region.

Figure S2.9a shows the time series of SLA rates estimated with a ± 8 -year running time window, for each grid point. Similar time series for all grid points suggests that the whole region experiences coherent sea-level changes with time. To validate this, we calculated the differences in the time series of SLA rates between grid points, to quantify the spatial differences in sea-level trends across this region. It should be noted that these differences are sensitive to the length of running time window chosen. We calculated the time series of SLA rate differences between each pair of grid points – for example, grid point 1 is differenced with each of the remaining 24 grid points. The results are shown in Figure S2.9b.

The 95% confidence level of SLA rate differences is 0.9 mm/yr. This magnitude of difference is much less than the magnitude of rate changes between any two time periods (e.g., between 1966–1981, and 1981–2005), shown in Table S2.2. With a shorter running time window of ± 5 years, the results suggest an SLA rate difference of 1.4 mm/yr at the 95% confidence level, which is still less than the magnitude of rate changes between any two time periods.

Text S2.2

Corrections for long-term sea-level rise

Following the methodology of *Meltzner et al.* [2010], the RSL time series at each coral site were adjusted vertically by a constant 2 mm/yr, to account for eustatic sea-level rise in the 20th century. This correction compares well with average estimates of sea-level rise in the Indian Ocean from ~1945 to 2007, from previous studies of 20th century sea-level rise [*Beckley et al.*, 2007; *Hamlington et al.*, 2011; *Jevrejeva et al.*, 2006]. However, these studies also show that rates of sea-level rise vary with time, within a given region. For example, *Jevrejeva et al.* [2006] estimated variable rates of sea level rise in the Indian Ocean during different time periods: between 1938 to 2000, rates of sea-level rise decreased non-linearly from $\sim +3.9$ to $\sim +0.2$ mm/yr. Specifically, the rate of sea-level rise between ~1930 and 1947 was $\sim +4$ mm/yr; between ~1947 and 1958 the rate was $\sim +3$ mm/yr; between 1958 and 1965, $\sim +2$ mm/yr; between 1965 and 1980, $\sim +1$ mm/yr; and from 1980, $\sim +0.2$ mm/yr.

By analyzing tide gauge data from ~1900, some researchers have reported 60-year oscillations in global sea level, which suggest a fall in mean sea level after 1960 [*Chambers et al.*, 2012; *Woodworth et al.*, 2009]. These oscillations were assumed or argued by the respective authors to operate on the scale of entire ocean basins, leading us to expect contemporaneous changes in RSL at all sites within our study region. However, the coral data show that while sites in the Banyak Islands experienced significant RSL change from 1966-1981, similar changes are precluded at sites on the neighboring islands of Simeulue and Nias. Hence, if multi-decadal fluctuations in sea level are responsible for the coral observations in the Banyak Islands, these sea-level fluctuations must have been extremely limited in geographic extent, and they must have been distinct from the basin-wide oscillations suggested by *Woodworth et al.* [2009] and *Chambers et al.* [2012]. We

note that the nearest tide gauge records used by these authors were from sites located more than 3000 km from our study area, and as the authors point out, the spatiotemporal extent of the sea-level dataset needs to be enhanced before we can better understand the spatial dependence of these oscillations.

In this section, we consider the following questions: what might be the implications if we have overcorrected for 20th century sea-level rise? Could this overcorrection change our conclusion that spatial changes in locking patterns cannot explain the coral observations? Would we still require a slow slip event to explain the coral observations? In order to address these questions, we draw attention to the 1966–1981 uplift rates from site PBK in the Banyak Islands, since the coral at this site has the best temporal resolution and the smallest uncertainties; the other sites in the Banyak Islands (PLM and PUB) have larger uncertainties, so that any model would be less sensitive to an additional independent error of 1 to 2 mm/yr from either of these sites. If, after assuming a conservative stance of applying no sea-level rise corrections for these sites, spatial changes in coupling patterns still cannot explain the PBK uplift rate and neighbouring sites on Simeulue and Nias (and that an SSE is still required to explain the displacement rates), then we can deduce that our results are not biased by the modern sea-level rise correction that we adopted in this study.

In the case of applying no sea-level rise corrections, the alternate uplift rate at site PBK would be a lower rate of $+5.8 \pm 2.1$ mm/yr (corresponding to a 2σ confidence interval of + 3.7 to 7.9 mm/yr). At sites PLM and PUB, the displacement rates would be -0.6 ± 5.3 and 0 ± 5.3 mm/yr, respectively. At sites LBJ, SBA and MZL, the subsidence rates would be -10.2 ± 1.5 , -6.4 ± 1.9 , and -4.5 ± 3 mm/yr, respectively. If we employ models of pure backslip on the fault (where slip on the fault does not exceed the plate subduction rate),

the model with the lowest WMSSR is that shown in Figure 2.2b (also shown in Table S2.3). As mentioned in the main text, this model shows the fault locked down to 30 km depth along a 90 km section under the Banyak Islands, and creeping at the plate subduction rate at depths below 30 km. This model suggests an uplift rate of + 4.8 mm/yr at site PBK. Although this model uplift rate is within the + 3.7 mm/yr lower-end estimate at site PBK (at the 2σ level), the model would not simultaneously fit the faster subsidence rates at sites LBJ and SBA, where the model uplift rates at these two sites are -7.7 and -2.5 mm/yr, respectively. Hence, it is difficult to explain the observed displacement rates with models that only include spatially variable coupling on the megathrust, even in the case of applying no sea-level rise corrections. Our considerations here suggest that the modern sea-level rise correction that we adopt in this study does not bias our results, and our conclusion holds – that a slow slip event is required to explain the 1966–1981 displacement rates.

We note that the maximum model uplift rate at site PBK shown in Figure 2.2b is less than the maximum uplift rate suggested by trench-normal surface uplift rate profiles shown in Figure 18 of *Meltzner et al.* [2015]’s study (their figure is reproduced and modified in Figure S2.7a) . This is because in our study we have additional constraints from sites on Simeulue and Nias. As shown in Figure S2.7 (and as modeled by *Meltzner et al.* [2015]), a uniform along-strike downdip limit of locking with no downdip transition zone fails to fit displacement rates at all sites simultaneously; rather, along-strike variations in the downdip limit of locking and a downdip transition zone are needed to produce satisfactory fits at all sites. Because we incorporated these modifications in our models, the maximum model uplift rate at site PBK is less than that suggested by *Meltzner et al.* [2015].

Text S2.3

Instrumental catalogues of global seismicity

There are no known records of major earthquakes in the Sumatran region between 1956 and 1966, based on both the coral microatolls and historical records [*Newcomb and McCann, 1987*]. In addition, we searched the National Earthquake Information Center (NEIC) and International Seismological Centre (ISC) catalogs for M6+ earthquakes around Sumatra, from 1956 to 2004. Figure 2.1 shows that pre-1966 seismicity along the Sumatran subduction zone did not exceed M7, and pre-1966 earthquakes within ~500 km of the Banyak Islands all have magnitudes lower than ~6.7. On 2 April 1956, a M6.3 earthquake occurred in the vicinity of the Banyak Islands, but any postseismic deformation would likely have been small in amplitude and short-lived.

Text S2.4.

Fault geometry

The dip geometry for the fault approximates trench-normal vertical cross-sections through the Slab 1.0 model [Hayes *et al.*, 2012] along the Simeulue-Nias section of the megathrust, and is consistent with the depths of relocated regional seismicity beneath southeastern Simeulue [Tilman *et al.*, 2010] (Figure S2.10). The variable strike of the trench along this section was also incorporated into the fault geometry. The along-strike length and downdip width of the patches on the fault are 10 km and 2 km, respectively. This geometry is slightly modified from that used by Meltzner *et al.* [2015], which results in slight differences between the depths reported in this and in the earlier paper. The map-view locations of the fully locked, partially locked, and freely slipping patches should be comparable between the two papers, but the depths are only approximately comparable. The broader conclusions and implications of either paper should not be impacted by these small differences.

Coupling model parameters

For all time periods, we assigned the coupling ratio along the shallowest 22 km of the fault to 0.4. The shallow region of partial coupling is needed to explain the observed coral displacement rates in southern Simeulue and northern Nias. The location of this shallow, partially coupled zone is consistent with that of the updip seismic/aseismic transition zone reported by Tilman *et al.* [2010]. However, we note that the shallow region of the megathrust here is capable of rupturing during large earthquakes, as suggested by the M 7.6 tsunami earthquake of 1907 [Kanamori *et al.*, 2010].

In order to fit the observed subsidence rates in the Banyak Islands, we included in the models a region of partial coupling along the downdip portion of the fault (a downdip transition zone). If this downdip transition zone is not included, then either the model subsidence rates are too large to fit the data at one or more sites, or uplift, instead of subsidence, occurs at other sites (Figure S2.10). We tested a range of values for the updip limit, downdip limit and coupling ratio of the downdip transition zone. For the updip and downdip limits, we tested depths ranging between 30 and 70 km, which are consistent with previous regional and global studies of locking and transition zone depths [McCaffrey *et al.*, 2000; Tichelaar and Ruff, 1993].

In addition, models with a downdip limit of the fully coupled zone that is uniform along strike yielded poor data-model fits. However, if we incorporate a deeper downdip limit of full coupling under Simeulue Island, the data-model fits improve substantially (Figure S2.10). Hence, we estimated separate downdip limits under Simeulue and the Banyak Islands, respectively. We successively tested downdip limits ranging from 20 to 55 km, at 5 km depth increments.

To demonstrate the trade-offs between various parameters in the coupling and SSE models, Figures S2.11 and S2.13 show how the weighted mean of sum of squared residuals (WMSSR) changes as each parameter is perturbed from the preferred, best-fit values.

Text S2.5

Inverse models

We inverted the coral data to obtain an estimated slip distribution, and with this, predicted model displacements to compare with the data. Our goal here was to ensure that we had not overlooked any distribution of coupling which could explain the Banyak Islands uplift rates. In the inversion, slip was constrained to slip backwards at rates from 0 to 40 mm/yr (the full plate subduction rate). Due to the limited spatial coverage of the data (and resulting limitations in model resolution), our objective was simply to see if there is any model (even one that does not fit our expectation of smoothness) that can fit the data; if not, this helps to confirm our conclusion that backslip alone cannot explain the 1966–1981 coral displacement rates. We therefore designed a fault model with coarse subfault patches (50 km by 50 km in size), and did not apply any Laplacian smoothing constraints to the inversion.

We performed a bounds-constrained linear least squares inversion using a priori knowledge of the full plate subduction rate of 4 cm/yr. To determine coupling models for each of the three periods, where each patch was constrained to only slip backwards, we applied a lower slip bound of 0 cm/yr of backslip and an upper slip bound of 4 cm/yr of backslip (corresponding to an uncoupled, freely slipping patch and a fully coupled patch, respectively). On the other hand, to model an SSE on the fault, we allowed both forward, thrust motion slip and backwards slip. The corresponding slip bounds applied to the inversion were: a lower bound of –4 cm/yr of backslip, and an upper bound of +4 cm/yr of backslip (corresponding to forward, thrust motion slip of 8 cm/yr, and a fully coupled patch, respectively).

Figures S2.14a and S2.14b show the inversion results for the 1981–2005, and 1966–1981 coral displacement rates, respectively. On patches with relatively good resolution (see

section below), the spatial patterns of backslip in the inverse and forward models are similar to first order, although the absolute values of estimated backslip between the two models differ. Importantly, the inverse models fit the 1981–2005 coral displacement rates well (weighted mean of summed squared residuals = 0), while there are large misfits with the 1966–1981 coral displacement rates, particularly at site PBK (weighted mean of summed squared residuals = 3.1; similar to the value for the forward model in Figure 2.2b). Therefore, we argue that backslip alone cannot explain the 1966–1981 coral displacement rates.

On the other hand, if we allow for an SSE in the inverse models, i.e., we allow for thrust motion slip on the fault at rates greater than the plate subduction rate, the model displacement rates fit the 1966–1981 coral displacement rates significantly better (Figure S2.14c). However, the inverse models are sensitive to the magnitude of forward slip constraints, such that we could not confidently estimate the SSE slip rate from the inverse models.

Supporting Information Figures

Figure S2.1 (a) Cross sections of slab PBK-4, from site PBK-B.

Figure S2.1 (b) Relative sea-level history derived from slab PBK-4.

Open orange circle above triangles designate years with inferred diedowns.

Figure S2.2 (a) Cross section of slab PLM-1, from site PLM-A.

Figure S2.2 (b) Relative sea-level history derived from slab PLM-1.

Open orange circles above triangles designate years with inferred diedowns.

(a)

Figure S2.3 (a) Cross section of slab PUB-1, from site PUB-A.

Figure S2.3 (b) Relative sea-level history derived from slab PUB-1.

Open orange circles above triangles designate years with inferred diedowns.

Figure S2.4 (a) Cross section of slab SBA-1, from site SBA-A.

Figure S2.4 (b) Relative sea-level history derived from slab SBA-1. Open orange circles above triangles designate years with inferred diedowns.

Figure S2.5 (a) Cross section of slab MZL-1, from site MZL-A.

(b)

Relative Sea Level (HLS History) for Site MZL

Figure S2.5 (b) Relative sea-level history derived from slab MZL-1.

Figure S2.6 (a) Cross section of slab LBJ-1, from site LBJ-A.

Figure S2.6 (b) Relative sea-level history derived from slab LBJ-1.

Open orange circles above triangles designate years with inferred diedowns.

Figure S2.7 (a) Figure 18 from *Meltzner et al.* [2015], showing model profiles of surface uplift and subsidence, plotted as a function of distance from the trench, and for various downdip limits of locking (LD, locking depth). Superimposed are 1966 – 1981 vertical displacement rates for each site (for MZL, SBA and LBJ, the 20th century displacement rates are plotted). (b) Same model profiles shown in (a), but the 2 mm/yr modern sea level correction is removed from the 1966 – 1981 displacement rates at each site. In each case, none of the locking depth profiles can simultaneously explain displacement rates at all six sites.

Figure S2.8 (a) We extracted from Aviso data the daily sea level anomalies (SLA) from 25 grid points over the Simeulue-Nias section. (b) Time series of mean SLA for all grid points, calculated by taking a running time window of ± 30 days, for each day. (c) Time series of differences in time-averaged SLA between each pair of grid points. (d) Cumulative frequency histogram of mean SLA differences between each pair of grid points, and values corresponding to the 65th and 95th percentile of the data.

Figure S2.9 (a) Time series of SLA rate for each grid point, estimated with a ± 8 -year running time window. Similar time series for all grid points suggests that on the whole, the region experiences similar regional sea-level changes with time. (b) – (e) Time series of differences in SLA rates between each pair of grid points, estimated with a running time window of (b) ± 8 years, (c) ± 5 years (d) ± 6 years and (e) ± 10 years. Note the different scales on the axes.

Figure S2.10 (a) The vertical cross-section of the fault approximates the Slab 1.0 model of *Hayes et al.* [2012] along the Simeulue-Nias section of the megathrust. Regional seismicity locations are based on *Tilmann et al.* [2010]’s results from southeastern Simeulue. (b)–(d) In the case of not including a downdip transition zone in the model (b,c), it is difficult to fit the displacement rates in the Banyak Islands within their 2σ level of uncertainties, whereas fits improve significantly with the inclusion of a downdip transition zone (d). The locking depth in each model is labeled. Data vectors in the main part of each figure correspond to 1981 – 2005 vertical displacement rates; insets show 1956 – 1966 vertical displacement rates. (b) A uniform along-strike locking depth of 35 km results in uplift, instead of subsidence, at all three Banyak Islands sites. (c) A uniform along-strike locking depth of 40 km results in too much uplift at site PUB, the site farthest from the trench. (d) Significantly better fits result from including a downdip transition zone at 30 to 60 km depths that is slipping at 50% of the plate subduction rate. However, a deeper locking depth under Simeulue Island is needed to fit the displacement rate at LBJ (as shown in our preferred best-fit model in Figure 2.2a).

Figure S2.11 Sensitivity of data-model fits to various backslip model parameters (data-model fits are quantified by the weighted mean of the sum of squared residuals), by changing parameters, one at a time, from those in the preferred best-fit backslip model in (a). (a) The preferred best-fit backslip model, with the updip limit of locking at 22 km depth, the downdip limit of the fully coupled zone at 30 km depth under the Banyak Islands and northern Nias, and the downdip limit of the fully coupled zone at 45 km depth under southern Simeulue. The

maximum depth of the downdip transition zone is at 60 km depth. (b) – (f) Example models with a single parameter perturbed in each case from the preferred model: (b) changing the updip limit of locking from 22 km to 10 km, (c) changing the downdip limit of the fully coupled zone along the northern part of the fault from 45 km to 30 km depth, (d) changing the downdip limit of the fully coupled zone along the southern part of the fault from 30 km to 45 km depth, (e) changing the maximum depth of the downdip transition zone from 60 km to 50 km depth, and (f) changing the coupling ratio of the downdip transition zone from 50% to 90% of the full plate subduction rate (where the full plate subduction rate is 40 mm/yr). These are only a subset of all the models considered; see Table S2.3 for a complete set.

Figure S2.12 Additional models of spatially variable interseismic coupling that attempt to produce best fits to the 1966 – 1981 coral displacement rates. In (a), we show the same along strike location of the freely slipping patch as our preferred SSE event model (Figure 2.2c), with the updip limit of the freely slipping patch at 30 km depth. This fails to reproduce rapid uplift at site PBK. In (b), we widen the area of the freely slipping patch under the Banyak Islands from that shown in (a). This model produces better fits than in (a), but still fails to fit the rapid uplift rate at site PBK. This model is also shown in Figure 2.2b. In (c) and (d), we made further modifications to the model, but they resulted in degraded fits. Other variations that we considered are shown in Table S2.4.

(e) extending the SSE in (a) 20 km further to the northwest, and (f) extending the SSE in (a) 20 km further to the southeast. Each of the changes in (b) through (f) results in a poorer fit of the model to the data. These are only a subset of all models considered; see Table S2.5 for a complete set.

Figure S2.14 Inverse models for (a) the 1956–1966 and 1981–2005 coral displacement rates, assuming backward slip on the fault, (b) the 1966–1981 coral displacement rates, assuming backward slip on the fault, and (c) the 1966–1981 coral displacement rates, allowing for both forward and backward slip on the fault. Note that the color scale shows real slip values (as opposed to backslip values) to be consistent with the color scale shown in the forward models, so that a slip of 0 cm/yr indicates a locked interface, 4 cm/yr indicates a freely slipping interface (slip at the full plate subduction rate of 4 cm/yr), and slip greater than 4 cm/yr indicates slow slip at a rate greater than the full plate subduction rate.

Table S2.1 Coral site locations, vertical displacement rates, and 2σ level of uncertainties in rates, for each time period. For sites PBK, PLM and PUB, average vertical displacement rates for the entire time period of available data are given for reference; these rates were not used in the modeling.

Site	Longitude	Latitude	Distance from trench (km)	Valid years (AD)	Uplift (mm/yr) [#]	Notes
PBK	97.12740	2.10935	135	1956-2005 AD	1.5± 0.9	Overall average
				1751-1861 AD	-2.2± 0.7	
				1812-1861 AD	-2.5±1.6	
				1833-1861 AD	-4.6± 2.9	
				1875-1894 AD	-3± 4.2	
				1956-1966 AD	-2.7± 3.4	
				1966-1981 AD	7.8± 2.1	
				1981-2005 AD	-2.4± 0.8	
PLM	97.20795	2.26300	153	1952-2005 AD	-0.1± 1.5	Overall average
				1952-1966 AD	-0.5± 5.7	
				1966-1981 AD	1.4± 5.3	
				1981-2005 AD	-0.5± 3.3	
PUB	97.40552	2.32384	174	1964-2005 AD	-1.1± 2.0	Overall average
				1966-1981 AD	2± 5.3	
				1981-2005 AD	-2.8± 3.3	
SBA	97.38281	1.51414	115	1962 - 2005 AD	-4.4± 1.9	
MZL	97.32974	1.47508	95	1970-1997 AD	-2.5± 3	
LBJ	96.48550	2.40224	107	1945-1997 AD	-8.2± 1.5	

[#] = 2σ level of uncertainties, calculated by the method of *Meltzner et al.* [2012].

Table S2.2 Additional uncertainties in vertical displacement rates at each site due to spatial variations in sea level, calculated by dividing the 95th percentile of mean SLA differences (17.3 mm) by the length of time over which each displacement rate was determined. The total uncertainty equals the uncertainties due to differential erosion and growth rates of the coral (calculated with the methodology of *Meltzner et al.* [2012]), plus the additional uncertainty due to spatial variations in sea level. Also shown are the magnitudes of rate changes between each time period ('1966 change' is the change in rates between 1956–1966 and 1966–1981, and '1981 change' is the change in rates between 1966–1981 and 1981–2005), and their 1 σ and 2 σ level uncertainties.

Site	Uplift (mm/yr)	Valid Years (AD)	Additional uncertainty	Total Uncertainty #	1966 Change (mm/yr)			1981 Change (mm/yr)		
					Rate	Uncertainty		Rate	Uncertainty	
						2 σ	1 σ		2 σ	1 σ
PBK	1.5 $\pm$ 0.9	1956 to 2005	$\pm$ 0.4	$\pm$ 1.0						
	-2.2 $\pm$ 0.7	1751 to 1861	$\pm$ 0.2	$\pm$ 0.7						
	-2.5 $\pm$ 1.6	1812 to 1861	$\pm$ 0.4	$\pm$ 1.7						
	-4.6 $\pm$ 2.9	1833 to 1861	$\pm$ 0.6	$\pm$ 2.9						
	-3.0 $\pm$ 4.2	1875 to 1894	$\pm$ 0.9	$\pm$ 4.3						
	-2.7 $\pm$ 3.4	1956 to 1966	$\pm$ 1.7	$\pm$ 3.8	10.6	$\pm$ 4.5	$\pm$ 2.3			
	7.8 $\pm$ 2.1	1966 to 1981	$\pm$ 1.2	$\pm$ 2.4				-10.2	$\pm$ 2.7	$\pm$ 1.4
	-2.4 $\pm$ 0.8	1981 to 2005	$\pm$ 0.7	$\pm$ 1.1						
PLM	-0.1 $\pm$ 1.5	1952 to 2005	$\pm$ 0.3	$\pm$ 1.5						
	-0.5 $\pm$ 5.7	1952 to 1966	$\pm$ 1.2	$\pm$ 5.8	1.9	$\pm$ 8.0	$\pm$ 4.0			
	1.4 $\pm$ 5.3	1966 to 1981	$\pm$ 1.2	$\pm$ 5.4				-1.9	$\pm$ 6.4	$\pm$ 3.2
	-0.5 $\pm$ 3.3	1981 to 2005	$\pm$ 0.7	$\pm$ 3.4						
PUB	-1.1 $\pm$ 2.0	1964 to 2005	$\pm$ 0.4	$\pm$ 2.0						
	2.0 $\pm$ 5.3	1966 to 1981	$\pm$ 1.2	$\pm$ 5.4				-4.8	$\pm$ 6.4	$\pm$ 3.2
	-2.8 $\pm$ 3.3	1981 to 2005	$\pm$ 0.7	$\pm$ 3.4						
SBA	-4.4 $\pm$ 1.9	1962 to 2005	$\pm$ 0.4	$\pm$ 1.9				0.0	$\pm$ 2.7	$\pm$ 1.4
MZL	-2.5 $\pm$ 3.0	1970 to 1997	$\pm$ 0.6	$\pm$ 3.0				0.0	$\pm$ 4.2	$\pm$ 2.1
LBJ	-8.2 $\pm$ 1.5	1945 to 1997	$\pm$ 0.3	$\pm$ 1.6	0.0	$\pm$ 2.2	$\pm$ 1.1	0.0	$\pm$ 2.2	$\pm$ 1.1

#: The 2σ total uncertainty equals the 2σ level of uncertainties due to differential coral erosion and growth rates (calculated with the methodology of *Meltzner et al.* [2012]), plus the additional 2σ uncertainty due to spatial variations in sea level.

Table S2.3 (a) Models of spatially variable fault coupling that yield reasonable fits to the 1956–1966 and 1981–2005 displacement rates; the preferred best-fit model shown in Figure S2.11a is highlighted in bold. (b) Sensitivity of data-model fits to various backslip model parameters, by changing the parameters, one at a time, from those in the preferred best-fit model. A subset of these models is shown in Figure S2.11. (c) Testing the downdip transition zone parameters under the Banyak Islands, with a fixed downdip limit of the fully coupled zone at 45 km depth under Simeulue, an updip limit of locking at 22 km depth, and coupling ratio of 0.4 at < 22 km depths. Models that yield good data-model fits are highlighted in bold. (d) Testing the downdip limit of the fully coupled zone under Simeulue Island, starting from the subset of models highlighted in (c). (e) Testing the updip limit of locking (with a fixed coupling ratio of 0.4 at depths shallower than the updip limit), starting from the subset of models highlighted in (c). The results in (d) and (e) affirm that the four models highlighted in (c) are models that yield good data-model fits. (LD = locking depth; BI = Banyak Islands; SS = southern Simuelue; CR = coupling ratio)

(a) Backslip models for time periods 1956 to 1966, and 1981 to 2005

LD (km)		Updip limit of locking (km)	Downdip transition zone		1956 - 1966 data-model fits		1981 - 2005 data-model fits		1956-1966 and 1981-2005 data-model fits	
BI	SS		Max. depth (km)	CR	χ^2	$\chi^2_r (n = 3)$	χ^2	$\chi^2_r (n = 6)$	χ^2	$\chi^2_r (n=9)$
30	45	22	55	0.5	0.5	0.2	8.6	1.4	9.1	1.0
30	45	22	60	0.5	0.7	0.2	7.4	1.2	8.1	0.9
30	45	22	65	0.5	1.0	0.3	7.6	1.3	8.5	0.9
30	45	22	70	0.5	1.2	0.4	8.2	1.4	9.5	1.1

(b) Perturbation of backslip model parameters from best-fit values

Changing model parameter	LD (km)		Updip limit of locking (km)	Downdip transition zone		1956-1966 and 1981-2005 data-model fits		Figure number
	BI	SS		Max. depth (km)	CR	χ^2	$\chi^2_r (n=9)$	
Preferred best-fit model	30	45	22	60	0.5	8.1	0.9	S10a
Updip limit of locking	30	45	10	60	0.5	31.8	3.5	S10b
	30	45	25	60	0.5	48.7	5.4	-
Locking depth under Simeulue	30	30	22	60	0.5	83.8	9.3	S10c

	30	50	22	60	0.5	10.0	1.1	-
Locking depth under Banyak Islands	25	45	22	60	0.5	29.3	3.3	-
	45	45	22	60	0.5	214.0	23.8	S10d
Maximum depth of transition zone	30	45	22	50	0.5	14.9	1.7	S10e
	30	45	22	70	0.5	9.5	1.1	-
Coupling ratio of transition zone	30	45	22	60	0.9	226.6	25.2	S10f
	30	45	22	60	0.3	60.3	6.7	-

(c) Testing the locking depth and downdip transition zone parameters

LD (km)		Updip limit of locking (km)	Downdip transition zone		1956-1966 and 1981-2005 data-model fits	
BI	SS		Max. depth (km)	CR	χ^2	χ^2_r (n=9)
20	45	22	30	0.3	111.8	12.4
20	45	22	30	0.5	159.4	17.7
20	45	22	30	0.7	221.1	24.6
20	45	22	35	0.3	53.2	5.9
20	45	22	35	0.5	52.1	5.8
20	45	22	35	0.7	57.6	6.4
20	45	22	40	0.3	25.6	2.8
20	45	22	40	0.5	16.9	1.9
20	45	22	40	0.7	23.6	2.6
20	45	22	45	0.3	19.0	2.1
20	45	22	45	0.5	18.5	2.1
20	45	22	45	0.7	43.5	4.8
20	45	22	50	0.3	17.0	1.9
20	45	22	50	0.5	28.2	3.1
20	45	22	50	0.7	75.2	8.4
20	45	22	55	0.3	17.4	1.9
20	45	22	55	0.5	40.1	4.5
20	45	22	55	0.7	107.5	11.9
20	45	22	60	0.3	18.5	2.1
20	45	22	60	0.5	49.4	5.5
20	45	22	60	0.7	131.0	14.6
20	45	22	65	0.3	19.7	2.2
20	45	22	65	0.5	57.1	6.3

20	45	22	65	0.7	150.0	16.7
20	45	22	70	0.3	20.7	2.3
20	45	22	70	0.5	63.1	7.0
20	45	22	70	0.7	164.4	18.3
25	45	22	30	0.3	236.2	26.2
25	45	22	30	0.5	256.3	28.5
25	45	22	30	0.7	284.1	31.6
25	45	22	35	0.3	142.8	15.9
25	45	22	35	0.5	107.5	11.9
25	45	22	35	0.7	85.8	9.5
25	45	22	40	0.3	85.1	9.5
25	45	22	40	0.5	36.7	4.1
25	45	22	40	0.7	21.8	2.4
25	45	22	45	0.3	65.2	7.2
25	45	22	45	0.5	22.4	2.5
25	45	22	45	0.7	28.3	3.1
25	45	22	50	0.3	53.4	5.9
25	45	22	50	0.5	20.4	2.3
25	45	22	50	0.7	50.2	5.6
25	45	22	55	0.3	47.3	5.3
25	45	22	55	0.5	24.5	2.7
25	45	22	55	0.7	76.0	8.4
25	45	22	60	0.3	44.6	5.0
25	45	22	60	0.5	29.3	3.3
25	45	22	60	0.7	95.7	10.6
25	45	22	65	0.3	43.1	4.8
25	45	22	65	0.5	33.9	3.8
25	45	22	65	0.7	112.0	12.4
25	45	22	70	0.3	42.4	4.7
25	45	22	70	0.5	37.8	4.2
25	45	22	70	0.7	124.7	13.9
30	45	22	35	0.3	229.7	25.5
30	45	22	35	0.5	171.2	19.0
30	45	22	35	0.7	124.8	13.9
30	45	22	40	0.3	141.4	15.7
30	45	22	40	0.5	63.7	7.1
30	45	22	40	0.7	30.1	3.3
30	45	22	45	0.3	106.3	11.8
30	45	22	45	0.5	31.4	3.5
30	45	22	45	0.7	21.5	2.4
30	45	22	50	0.3	82.4	9.2
30	45	22	50	0.5	14.9	1.7
30	45	22	50	0.7	31.2	3.5
30	45	22	55	0.3	67.9	7.5
30	45	22	55	0.5	9.1	1.0
30	45	22	55	0.7	48.5	5.4

30	45	22	60	0.3	60.3	6.7
30	45	22	60	0.5	8.1	0.9
30	45	22	60	0.7	63.5	7.1
30	45	22	65	0.3	55.4	6.2
30	45	22	65	0.5	8.5	0.9
30	45	22	65	0.7	76.3	8.5
30	45	22	70	0.3	52.1	5.8
30	45	22	70	0.5	9.5	1.1
30	45	22	70	0.7	86.5	9.6
35	45	22	40	0.2	53.7	6.0
35	45	22	40	0.4	39.7	4.4
35	45	22	40	0.6	36.5	4.1
35	45	22	45	0.2	42.5	4.7
35	45	22	45	0.4	29.5	3.3
35	45	22	45	0.6	39.6	4.4
35	45	22	50	0.2	33.8	3.8
35	45	22	50	0.4	24.9	2.8
35	45	22	50	0.6	51.8	5.8
35	45	22	55	0.2	28.2	3.1
35	45	22	55	0.4	24.5	2.7
35	45	22	55	0.6	67.3	7.5
35	45	22	60	0.2	25.1	2.8
35	45	22	60	0.4	25.5	2.8
35	45	22	60	0.6	79.5	8.8
35	45	22	65	0.2	23.0	2.6
35	45	22	65	0.4	26.8	3.0
35	45	22	65	0.6	89.7	10.0
35	45	22	70	0.2	21.6	2.4
35	45	22	70	0.4	28.1	3.1
35	45	22	70	0.6	97.7	10.9

(d) Testing the locking depth under Simeulue Island

LD (km)		Updip limit of locking (km)	Downdip transition zone		1956-1966 and 1981-2005 data-model fits	
BI	SS		Max. depth (km)	CR	χ^2	χ^2_r (n=9)
30	30	22	55	0.5	92.4	10.3
30	35	22	55	0.5	24.3	2.7
30	40	22	55	0.5	10.7	1.2
30	45	22	55	0.5	9.1	1.0
30	50	22	55	0.5	9.6	1.1
30	55	22	55	0.5	10.6	1.2
30	30	22	60	0.5	83.8	9.3
30	35	22	60	0.5	18.6	2.1

30	40	22	60	0.5	7.7	0.9
30	45	22	60	0.5	8.1	0.9
30	50	22	60	0.5	10.0	1.1
30	55	22	60	0.5	12.1	1.3
30	30	22	65	0.5	79.0	8.8
30	35	22	65	0.5	15.7	1.7
30	40	22	65	0.5	6.8	0.8
30	45	22	65	0.5	8.5	0.9
30	50	22	65	0.5	11.5	1.3
30	55	22	65	0.5	14.4	1.6
30	30	22	70	0.5	76.4	8.5
30	35	22	70	0.5	14.4	1.6
30	40	22	70	0.5	6.8	0.8
30	45	22	70	0.5	9.5	1.1
30	50	22	70	0.5	13.1	1.5
30	55	22	70	0.5	16.6	1.8

(e) Testing the updip limit of locking

LD (km)		Updip limit of locking (km)	Downdip transition zone		1956-1966 and 1981-2005 data-model fits	
BI	SS		Max. depth (km)	CR	χ^2	χ^2_r (n=9)
30	45	10	55	0.5	38.8	4.3
30	45	22	55	0.5	9.1	1.0
30	45	25	55	0.5	42.9	4.8
30	45	10	60	0.5	31.8	3.5
30	45	22	60	0.5	8.1	0.9
30	45	25	60	0.5	48.7	5.4
30	45	10	65	0.5	28.1	3.1
30	45	22	65	0.5	8.5	0.9
30	45	25	65	0.5	53.9	6.0
30	45	10	70	0.5	26.3	2.9
30	45	22	70	0.5	9.5	1.1
30	45	25	70	0.5	58.0	6.4

Table S2.4 Models of spatially variable fault coupling that attempt to produce best fits to the 1966–1981 displacement rates. (a) Testing models with different downdip limits of the fully coupled zone under the Banyak Islands, while keeping the following parameters fixed: the downdip limit of the fully coupled zone under Simeulue at 45 km depth, an updip limit of locking at 22 km depth, a coupling ratio of 0.4 at < 22 km depths, a downdip limit of the transition zone at 60 km depth, and a coupling ratio of 0.5 in the downdip transition zone. A subset of models are shown in Figure S2.12. (b) Testing the along-strike location of the freely slipping patch under the Banyak Islands. In each of tables, the best models are highlighted in bold. (LD = locking depth; BI = Banyak Islands; SS = southern Simuelue; CR = coupling ratio)

(a) Testing the locking depth under the Banyak Islands

LD (km)		Distance of LD boundary (km)		Updip limit of locking (km)	Downdip transition zone		Depth of creeping zone under BI(km)	1966 - 1981 data-model fits		Figure number
BI	SS	NW boundary from NW edge of fault	SE boundary from SE edge of fault		Max. depth (km)	CR		χ^2	χ^2_r (n = 6)	
20	45	100	80	22	60	0.5	> 30	74.7	12.5	-
30	45	100	80	22	60	0.5	> 30	22.2	3.7	S11a
35	45	100	80	22	60	0.5	> 30	41.8	7.0	-
40	45	100	80	22	60	0.5	> 30	70.8	11.8	-
25	45	90	60	22	60	0.5	> 25	40.5	6.7	S11c
30	45	90	60	22	60	0.5	> 30	15.1	2.5	S11b
35	45	90	60	22	60	0.5	> 35	55.4	9.2	S11d
40	45	90	60	22	60	0.5	> 40	136.6	22.8	-

(b) Testing the along-strike location of locking depth change (or change in the downdip limit of the fully coupled zone) across the fault

LD (km)		Distance of LD boundary (km)		Updip limit of locking (km)	Downdip transition zone		Depth of creeping zone under BI (km)	1966-1981 data-model fits		Figure number
BI	SS	NW boundary from NW edge of fault	SE boundary from SE edge of fault		Max. depth (km)	CR		χ^2	$\chi^2, (n=6)$	
30	45	100	30	22	60	0.5	> 30	27.6	4.6	-
30	45	100	40	22	60	0.5	> 30	21.6	3.6	-
30	45	100	50	22	60	0.5	> 30	18.5	3.1	-
30	45	100	60	22	60	0.5	> 30	17.9	3.0	-
30	45	100	70	22	60	0.5	> 30	19.0	3.2	-
30	45	100	80	22	60	0.5	> 30	22.2	3.7	S11a
30	45	100	90	22	60	0.5	> 30	28.5	4.8	-
30	45	100	100	22	60	0.5	> 30	39.4	6.6	-
30	45	100	110	22	60	0.5	> 30	54.4	9.1	-
30	45	90	40	22	60	0.5	> 30	19.2	3.2	-
30	45	90	50	22	60	0.5	> 30	16.0	2.7	-
30	45	90	60	22	60	0.5	> 30	15.1	2.5	S11b
30	45	90	70	22	60	0.5	> 30	15.9	2.6	-
30	45	90	80	22	60	0.5	> 30	18.5	3.1	-
30	45	90	90	22	60	0.5	> 30	23.8	4.0	-
30	45	90	100	22	60	0.5	> 30	33.4	5.6	-
30	45	90	110	22	60	0.5	> 30	46.9	7.8	-
30	45	80	40	22	60	0.5	> 30	20.4	3.4	-
30	45	80	50	22	60	0.5	> 30	17.0	2.8	-
30	45	80	60	22	60	0.5	> 30	16.0	2.7	-
30	45	80	70	22	60	0.5	> 30	16.5	2.8	-
30	45	80	80	22	60	0.5	> 30	18.8	3.1	-
30	45	80	90	22	60	0.5	> 30	23.5	3.9	-
30	45	80	100	22	60	0.5	> 30	32.2	5.4	-
30	45	80	110	22	60	0.5	> 30	44.8	7.5	-

Table S2.5 (a) Range of SSE models that yield good fits to the 1966–1981 coral displacement rates. The preferred best-fit model is highlighted in bold. (b) Sensitivity of data-model fits to various SSE model parameters, as these parameters are deviated from the preferred best-fit SSE model highlighted in (a). A subset of these models are shown in Figure S2.13. (c) Testing SSE depth range, slip rate, location and size. Models that yield good data-model fits are highlighted in bold.

(a) SSE models for 1966 to 1981

SSE depth range (km)	SSE slip rate (mm/yr)	Along-strike length of SSE (km)	Downdip width of SSE (km)	Area (km ²)	Seismic moment (N m)	Moment magnitude	Chi-square	Reduced chi-square (n = 6)
30 - 45	61	60	52	3120	8.56E+19	7.2	8.0	1.3
30 - 50	59	60	64	3840	1.02E+20	7.3	7.2	1.2
30 - 55	56	60	76	4560	1.15E+20	7.3	7.2	1.2
30 - 60	55	60	88	5280	1.31E+20	7.3	7.7	1.3
30 - 65	55	60	98	5880	1.46E+20	7.4	7.9	1.3
30 - 70	55	60	106	6360	1.57E+20	7.4	8.0	1.3

(b) Perturbations of SSE model parameters from best-fit values

Changing model parameter	SSE depth range (km)	SSE slip rate (mm/yr)	Along-strike length of SSE (km)	Downdip width of SSE (km)	Distance of NW limit of SSE from NW edge of fault	Distance of SE limit of SSE from SE edge of fault	Area (km ²)	X ²	X ² _r (n = 6)	Figure number
Preferred best-fit model	30 - 55	56	60	76	100	80	4560	7.2	1.2	S12a
SSE updip limit	20 - 55	56	60	130	100	80	7800	41.5	6.9	-
	40 - 55	56	60	38	100	80	2280	56.5	9.4	S12b
SSE downdip limit	30 - 45	56	60	52	100	80	3120	9.0	1.5	-
	30 - 65	56	60	98	100	80	5880	8.2	1.4	S12c
SSE slip rate	30 - 55	46	60	76	100	80	4560	13.0	2.2	-
	30 - 55	66	60	76	100	80	4560	15.7	2.6	S12d
North-western extent of	30 - 55	56	80	76	80	80	6080	17.3	2.9	S12e
SSE	30 - 55	56	40	76	120	80	3040	13.6	2.3	-
South-eastern extent of	30 - 55	56	80	76	100	60	6080	12.1	2.0	S12f
SSE	30 - 55	56	40	76	100	100	3040	15.8	2.6	-

(c) Testing SSE depth range, slip rate, location and size

SSE depth range (km)	SSE slip rate (mm/yr)	Distance from NW edge of fault (km)	Distance from SE edge of fault (km)	Along-strike length of SSE (km)	Width of SSE (km)	Area (km ²)	χ^2	$\chi^2_r (n = 6)$
30 - 35	70	90	80	70	20	1400	37.9	6.3
30 - 40	55	90	80	70	38	2660	16.8	2.8
	65	90	80	70	38	2660	12.2	2.0
	65	100	80	60	38	2280	12.0	2.0
	70	100	80	60	38	2280	11.3	1.9
30 - 45	55	90	80	70	52	3640	9.3	1.5
	59	90	80	70	52	3640	9.3	1.5
	59	100	80	60	52	3120	8.1	1.3
	61	100	80	60	52	3120	8.0	1.3
	62	100	80	60	52	3120	8.1	1.3
30 - 50	55	90	80	70	64	4480	8.3	1.4
	57	100	80	60	64	3840	7.0	1.2
	59	100	80	60	64	3840	7.2	1.2
30 - 55	53	90	80	70	76	5320	8.6	1.4
	54	90	80	70	76	5320	8.9	1.5
	55	90	80	70	76	5320	9.3	1.6
	55	100	80	60	76	4560	7.2	1.2
	56	100	80	60	76	4560	7.2	1.2
	57	100	80	60	76	4560	7.4	1.2
30 - 60	53	90	80	70	88	6160	9.4	1.6
	55	90	80	70	88	6160	10.7	1.8
	55	100	80	60	88	5280	7.7	1.3
	56	100	80	60	88	5280	8.0	1.3
30 - 65	52	90	80	70	98	6860	9.3	1.5
	53	90	80	70	98	6860	9.7	1.6
	55	90	80	70	98	6860	11.1	1.9
	55	100	80	60	98	5880	7.9	1.3
	56	100	80	60	98	5880	8.2	1.4
30 - 70	51	90	80	70	106	7420	9.1	1.5
	52	90	80	70	106	7420	9.4	1.6
	55	100	80	60	106	6360	8.0	1.3
	56	100	80	60	106	6360	8.4	1.4
35 - 40	50	90	80	70	18	1260	48.6	8.1
	100	90	80	70	18	1260	12.1	2.0
	110	90	80	70	18	1260	11.4	1.9
	120	90	80	70	18	1260	12.9	2.2
	60	100	80	60	18	1080	40.5	6.8
35 - 45	60	90	80	70	32	2240	20.0	3.3

	90	90	80	70	32	2240	16.6	2.8
	90	100	80	60	32	1920	13.2	2.2
35 - 50	70	90	80	70	44	3080	14.9	2.5
	90	90	80	70	44	3080	32.0	5.3
	80	100	80	60	44	2640	15.8	2.6
	90	100	80	60	44	2640	23.2	3.9

Table S2.6 List of SSEs at various subduction zones, compiled by *Peng and Gomberg [2010]* and *Schwartz and Rokosky [2007]*.

Location	Date	Moment (N m)	Moment (N m)(2)	M _w	M _w	Duration (seconds)	Reference
New Zealand							
Kapiti	2003	9.50E+18	-	-	-	3.15E+07	Peng and Gomberg [2010]
Manawatu	2004	1.72E+19	-	-	-	2.94E+07	Peng and Gomberg [2010]
Manawatu	2004	3.07E+19	-	-	-	1.64E+07	Peng and Gomberg [2010]
Manawatu	2004	3.98E+19	-	7.0	-	4.74E+07	Schwartz and Rokosky [2007]
Gisborne	2002	1.00E+19	-	-	-	8.64E+05	Peng and Gomberg [2010]
Gisborne	2004	7.70E+18	-	-	-	1.47E+06	Peng and Gomberg [2010]
Hastings	2006	6.20E+18	-	-	-	7.78E+05	Peng and Gomberg [2010]
Gisborne	2006	5.00E+18	-	-	-	1.04E+06	Peng and Gomberg [2010]
Hastings	2006	1.33E+19	-	-	-	1.56E+06	Peng and Gomberg [2010]
Southwest Japan							
Shikoku	04/01/2001	9.30E+17	-	-	-	6.91E+05	Peng and Gomberg [2010]
Shikoku	08/16/2001	1.01E+18	-	-	-	1.73E+05	Peng and Gomberg [2010]
Shikoku	08/18/2001	8.10E+17	-	-	-	1.73E+05	Peng and Gomberg [2010]
Shikoku	02/10/2002	9.60E+17	-	-	-	4.32E+05	Peng and Gomberg [2010]
Shikoku	2/15/2002	1.14E+18	-	-	-	2.59E+05	Peng and Gomberg [2010]
Shikoku	08/06/2002	6.40E+17	-	-	-	2.59E+05	Peng and Gomberg [2010]
Shikoku	08/09/2002	5.20E+17	-	-	-	3.46E+05	Peng and Gomberg [2010]
Shikoku	4/17/2003	9.00E+17	-	-	-	3.46E+05	Peng and Gomberg [2010]
Shikoku	8/27/2003	7.60E+17	-	-	-	2.59E+05	Peng and Gomberg [2010]
Shikoku	8/20/2003	1.15E+18	-	-	-	4.32E+05	Peng and Gomberg [2010]
Shikoku	11/07/2003	2.81E+18	-	-	-	5.18E+05	Peng and Gomberg [2010]
Shikoku	11/19/2003	1.05E+18	-	-	-	5.18E+05	Peng and Gomberg [2010]
Shikoku	02/10/2004	1.11E+18	-	-	-	2.59E+05	Peng and Gomberg [2010]
Shikoku	4/19/2004	1.35E+18	-	-	-	6.05E+05	Peng and Gomberg [2010]
Shikoku	12/27/2004	9.30E+17	-	-	-	4.32E+05	Peng and Gomberg [2010]
Shikoku	05/12/2005	1.21E+18	-	-	-	1.73E+05	Peng and Gomberg [2010]

Shikoku	5/14/2005	5.40E+17	-	-	-	3.46E+05	Peng and Gomborg [2010]
Shikoku	10/23/2005	5.10E+17	-	-	-	3.46E+05	Peng and Gomborg [2010]
Shikoku	4/15/2006	1.18E+18	-	-	-	4.32E+05	Peng and Gomborg [2010]
Shikoku	09/08/2006	2.89E+18	-	-	-	7.78E+05	Peng and Gomborg [2010]
Shikoku	3/13/2007	4.70E+17	-	-	-	3.46E+05	Peng and Gomborg [2010]
Shikoku	8/27/2007	5.50E+17	-	-	-	6.05E+05	Peng and Gomborg [2010]
Shikoku	09/09/2007	5.20E+17	-	-	-	2.59E+05	Peng and Gomborg [2010]
Shikoku	12/19/2007	1.99E+18	-	-	-	5.18E+05	Peng and Gomborg [2010]
Shikoku	3/13/2008	8.20E+17	-	-	-	5.18E+05	Peng and Gomborg [2010]
Shikoku	11/08/2006	9.89E+17	-	-	-	2.59E+05	Peng and Gomborg [2010]
Shikoku	02/13/2008	1.35E+18	-	-	-	4.32E+05	Peng and Gomborg [2010]
Bungo channel	1997	1.00E+19	7.94E+19	6.6	7.2	2.59E+07	Schwartz and Rokosky [2007]
Bungo channel	2003	2.00E+19	3.98E+19	6.8	7.0	7.78E+06	Schwartz and Rokosky [2007]
Bungo channel	2002	1.26E+18	-	6.0	-	4.32E+05	Schwartz and Rokosky [2007]
Bungo channel	2003	1.78E+18	-	6.1	-	8.64E+05	Schwartz and Rokosky [2007]
Bungo channel	2003	2.51E+18	-	6.2	-	6.05E+05	Schwartz and Rokosky [2007]
Bungo channel	2004	1.26E+18	-	6.0	-	3.46E+05	Schwartz and Rokosky [2007]
Bungo channel	2004	6.31E+17	-	5.8	-	4.32E+05	Schwartz and Rokosky [2007]
Tokai	2000	1.00E+19	3.98E+19	6.6	>7.0	1.73E+08	Schwartz and Rokosky [2007]
Tokai	2004	6.31E+17	-	5.8	-	4.32E+05	Schwartz and Rokosky [2007]
Tokai	2005	4.47E+17	-	5.7	-	2.59E+05	Schwartz and Rokosky [2007]
Boso	1996	5.01E+18	7.08E+18	6.4	6.5	6.05E+05	Schwartz and Rokosky [2007]
Boso	2002	1.00E+19	-	6.6	-	4.32E+06	Schwartz and Rokosky [2007]
Alaska-Aleutian subduction zone							
South central Alaska	1998	7.94E+19	-	7.2	-	8.64E+07	Schwartz and Rokosky [2007]
Mexico/Guerrero subduction zone							
Guerrero	2001	2.00E+19	-	6.8	7.6	1.73E+07	Schwartz and Rokosky [2007]
Cascadia subduction zone							
Cascadia	1999	1.41E+19	2.82E+19	6.7	6.9	4.32E+06	Schwartz and Rokosky [2007]
Cascadia	2003	1.00E+19	-	6.6	-	3.46E+06	Schwartz and Rokosky [2007]
Cascadia	2004	2.00E+19	-	6.8	-	4.32E+06	Schwartz and Rokosky [2007]

Chapter 3

Afterslip following the 2007 M_w 8.4 Bengkulu earthquake located adjacent to the rupture zone of the 2010 M_w 7.8 Mentawai earthquake

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Abstract

We model the spatial and temporal distribution of afterslip following the 12 September 2007 M_w 8.4 Bengkulu earthquake with Sumatran GPS Array (SuGAR) time series spanning ~ 6.3 years after the earthquake, up until the end of 2013. Our inversion results suggest moderate amounts of cumulative afterslip concentrated updip and downdip of the 2007 mainshock rupture area, and shallow afterslip that borders and overlaps the 2010 M_w 7.8 Mentawai earthquake rupture zone. The spatial proximity between the shallow afterslip and the 2010 rupture zone suggests that afterslip may play an important role in redistributing stresses in the shallow region of the megathrust. In particular, the cumulative contribution of stress changes due to the mainshock event and the ensuing afterslip likely increased strain rates in the shallow portion of the megathrust adjacent to the 2010 Mentawai earthquake rupture area, potentially advancing in time the rupture of this region in 2010.

3.1. Introduction

The 2007 M_w 8.4 Bengkulu earthquake in Sumatra marks the first in a modern series of large earthquakes that occurred on the Mentawai section of the Sunda megathrust – a section that has been forecast to rupture in the coming decades [Sieh *et al.*, 2008]. The Mentawai section last ruptured in 1797 and 1833, producing earthquakes of M_w 8.6–8.8 and M_w 8.8–8.9, respectively [Philibosian *et al.*, 2014; Sieh, 2007]. The moment released by the 2007 event represents only a fraction of that released in 1833 along this portion of the Mentawai section, as well as of the moment deficit accumulated since 1833, suggesting that the potential for a great earthquake along this section remains high [Konca *et al.*, 2008].

In the months and years following the Bengkulu earthquake, Sumatran GPS Array (SuGAR) stations located close to the mainshock rupture area recorded substantial amounts of postseismic deformation. One relevant question that stems from this observation is: to what extent has the remainder of the accumulated moment deficit since 1833 been released by afterslip? In addition, an unusual earthquake occurred ~ 3.1 years after the Bengkulu earthquake – on 25 October, 2010, a M_w 7.8 “tsunami earthquake” [Hill *et al.*, 2012] occurred in the region updip of the Bengkulu earthquake rupture zone, during ongoing postseismic deformation following the 2007 event. The spatial proximity between these two earthquakes begs further questions: what is the role of afterslip in redistributing stress on this part of the megathrust? Did afterslip advance in time the rupture of the shallow region of the megathrust in 2010? What implications does this have for the seismic hazard of other shallow regions along the megathrust, which are often inferred to be aseismic?

Generally, afterslip plays an important role in releasing and redistributing stress on megathrusts. Some researchers propose that aftershocks are driven by afterslip, based on observations of similar temporal patterns of afterslip and aftershocks [*Hsu et al.*, 2006; *Perfettini and Avouac*, 2004; 2007; *Perfettini et al.*, 2005]. In addition, the shallow regions of megathrusts have been proposed to consist of conditionally stable patches that usually behave aseismically, but may fail coseismically under high strain loads [*Bilek and Lay*, 2002; *Bilek et al.*, 2004; *Lay and Bilek*, 2007; *Lay et al.*, 2012; *Scholz*, 1998]. These hypotheses have been proposed based on far-field teleseismic data and inferences of the rigidity of material in the shallow region of subduction zones. However, support for these hypotheses requires geodetic studies of cases where “tsunami earthquakes” occur within a period of continued afterslip following coseismic rupture of the deeper part of the megathrust. The case of the 2010 Mentawai “tsunami earthquake” occurring within the time frame of continued afterslip following the 2007 Bengkulu earthquake hence presents an excellent opportunity to study the role of afterslip in redistributing stress along the shallow portion of the megathrust.

To address the questions above, we need to study postseismic time series of displacements that span a long time period, and that at least span the period between the two earthquakes. Even though previous researchers have studied the afterslip following this earthquake, the maximum time period spans 15 months following the earthquake [*Lubis et al.*, 2012]. To our knowledge, there have not been any studies on postseismic deformation spanning longer time periods after the earthquake. In our study, we model the entire post-Bengkulu earthquake time series of displacements recorded by SuGAR stations – a time period that spans an ~6.3-year-long period until the end of 2013 and which includes the period up to and beyond the 2010 Mentawai earthquake.

Modeling this longer time series of displacements requires taking into account the contribution of viscoelastic relaxation of the mantle to postseismic time series. In this study, we present two suites of afterslip models. The first suite of models is based on assuming that the majority of recorded postseismic deformation is due to afterslip on the megathrust. The second suite of models additionally includes finite volume blocks located at asthenospheric depths that accommodate ductile deformation due to viscous flow. We define these blocks as ductile shear zones. We then jointly invert for strains associated with viscoelastic deformation in each ductile shear zone, in addition to afterslip on the megathrust. We compare the afterslip distributions between the two suites of models and discuss the spatial relationship between the post-Bengkulu earthquake afterslip and the 2010 Mentawai earthquake.

In addition, we re-estimate the coseismic slip distribution of the Bengkulu earthquake using coral uplift measurements from *Konca et al.* [2008], and improved coseismic SuGAR offsets estimated by *Feng et al.* [2015b], which allows us to fully understand the regions of the fault that slipped, and the spatio-temporal distribution of coseismic stresses relaxed as afterslip. Previously published studies about this earthquake modeled slip on a planar fault, but our new results are based on more realistic fault geometry.

3.2. GPS data

We use GPS data from 17 SuGAR stations and 1 IGS station (NTUS) which were operating at the time of the 2007 Bengkulu earthquake, and which recorded afterslip following the event (Figure 3.1). We processed the GPS data using the GPS-Inferred Positioning System and Orbit Analysis Simulation Software (GIPSY-OASIS) version 6.2 [Zumberge *et al.*, 1997]. Full details of the data processing are described in *Feng et al.* [2015b]. We transformed the daily positions in the ITRF2008 (International Terrestrial

Reference Frame 2008) into daily positions relative to the Sunda plate, using the ITRF2008-Sunda transformation [Altamimi *et al.*, 2012].

Feng et al. [2015b] estimated GPS coseismic offsets and postseismic decays, together with long-term rates, and annual and semi-annual seasonal signals, on all three components of each station using a nonlinear least squares optimization procedure. The resulting coseismic offsets estimated from their optimization procedure represent more accurate and spatially coherent estimates compared to previous studies. Using their estimates, we obtained the postseismic GPS time series for the 2007 event by removing long-term rates, seasonal signals, and coseismic offsets and postseismic decays associated with other earthquakes that occurred during the 2007 postseismic period.

Our GPS postseismic time series span ~ 6.3 years, starting from one day after the earthquake (13 September 2007) until the end of 2013 (31 December 2013). We note that two aftershocks of M_w 7.9 and 7.0 occurred ~ 12 and ~ 16 hours after the mainshock, respectively; the coseismic offsets associated with these events were estimated together with the M_w 8.4 mainshock. It is likely that a sizeable amount of afterslip occurred in the hours following the earthquake, so the total amount of afterslip in our models likely represents an underestimate. Studies of coseismic slip and afterslip with high rate GPS data allow separation of the early afterslip deformation from the mainshock deformation (e.g. *Miyazaki and Larson* [2008], *Fukuda et al.* [2009]), and in the case of the 2012 Nicoya earthquake, estimated afterslip in the first 3 hours after the earthquake amounts to $\sim 35\%$ of the cumulative afterslip estimated 70 days following the earthquake [*Malservisi et al.*, 2015].

In Table 3.1, we list the GPS coseismic displacements, and the cumulative GPS postseismic displacements over the ~ 6.3 -year-long time period after the Bengkulu earthquake. The cumulative postseismic displacements range from $+15.7 \pm 1.8$ (1σ uncertainties) to -73.5 ± 1.5 mm on the east component, $+6.3 \pm 0.8$ to -102.3 ± 9.7 mm on the north component, and -58.9 ± 8.5 to -15.3 ± 2.3 mm on the vertical component. The largest cumulative displacements were observed at stations BSAT and PKRT (Figure 3.1), adjacent to the parts of the fault that experienced peak coseismic slip during the M_w 8.4 mainshock and M_w 7.9 aftershock.

For our models of coseismic slip, we additionally included 18 coral uplift measurements reported by *Konca et al.* [2008] (Table 3.2). The coral uplift measurements were taken up to a month after the earthquake, so the measurements include a portion of postseismic deformation. We note that the GPS and coral data record both the M_w 8.4 mainshock and the M_w 7.9 and M_w 7.0 aftershocks, so our inverted coseismic slip distribution (described in the next section) includes slip from all three events.

Figure 3.1 Map of the 17 Sumatran GPS Array (SuGAR) stations and 1 IGS station (NTUS) used to invert for coseismic slip and afterslip, and cumulative ~6.3-year postseismic horizontal (arrows) and vertical (bars) displacements. Coloured slip represents the 2007 Bengkulu earthquake mainshock and aftershock source models from *Konca et al.* [2008] (outlined in black), and the 2010 Mentawai "tsunami earthquake" source model from *Hill et al.* [2012] (outlined in magenta). Stars: Hypocenters of each earthquake are from the ANSS catalog.

3.3. Inversion methods

3.3.1. Modeling the coseismic slip distribution

We inverted for the coseismic slip distribution of the M_w 8.4, 7.9 and 7.0 earthquakes using a linear least squares inversion technique with non-negativity constraints. We inverted for coseismic slip on 20 km by 20 km rectangular subfault patches. Due to the limited resolution, we constrained the rake direction on each patch to vary within $\pm 45^\circ$ of a rake direction of 114° reported by the GCMT (Global Centroid Moment Tensor) catalog. The model fault spans the length of the Mentawai section from the Batu Islands in the northwest to Enggano Island in the southeast. The dip profile of the fault approximates the Slab 1.0 model [Hayes *et al.*, 2012] along this section of the megathrust (Figure 3.2a), and extends to 100 km depth. We regularized our solution by minimizing the second order derivative of the fault slip (i.e., applying Laplacian smoothing), tuned by a Laplacian smoothing weight. We tested weights that varied by several orders of magnitude in range to determine its influence on the model solution. We selected an optimal Laplacian weight that preserves the robust spatial characteristics of the model solution. We placed slip penalization constraints on subfault patches located at the base and NW and SE edges of the fault model.

3.3.2. Modeling the spatio-temporal distribution of afterslip with PCAIM

We inverted for the spatial and temporal distribution of afterslip on our model fault with the principal component analysis-based inversion method (PCAIM) [Kositsky and Avouac, 2010]. PCAIM uses Okada's solutions for deformation due to dislocations embedded in an elastic half-space to calculate the displacements at each GPS station [Okada, 1985]. The advantage of the PCAIM method is its ability to deal with time periods with missing data, and the method effectively filters the data time series for the main afterslip signal by representing the time series by an optimum number of principal

components. To select the optimum number of principal components to represent the postseismic time series, we examined the cumulative amount of variance in the data explained by incrementally adding principal components to represent the time series. Using one principal component to represent the time series explains ~74% of the variance in the data, while adding a second component explains ~94% of the data variance (Figure S3.2). Adding further principal components explains only marginally more of the cumulative amount of variance in the data. Theoretically, the higher order principal components represent other signals in the time series, such as seasonal signals, but we note that researchers should be wary of directly interpreting the spatial patterns suggested by each principal component to a particular tectonic or physical process. We chose to use one principal component to represent the time series, since this adequately fits the shape of the postseismic time series. Even though using two principal components yields a similar slip distribution (see Figure S3.3), higher order principal components can lead to model solutions that are not realistic or tectonically feasible.

PCAIM inverts for slip on each fault patch with a linear least squares inversion method with non-negativity constraints. We constrained the rake direction on each patch to $\pm 45^\circ$ of thrust directed motion (90°). Similar to our coseismic inversions, we applied slip penalization constraints on subfault patches located at the edges of our fault model, and Laplacian smoothing constraints. We tested Laplacian smoothing weights that varied by several orders of magnitude in range, and selected a weight that yields a solution at the corner of the L-curve of residuals.

3.3.3. Joint inversions of afterslip and viscoelastic deformation with PCAIM

Due to release of fluids into the mantle wedge from the subducting oceanic lithosphere, the viscosity of the mantle prism is one to two orders of magnitude lower than the oceanic lithosphere [Karato *et al.*, 2010]. Viscous flow in the mantle prism occurs due to

coseismic stress changes imparted by the mainshock rupture. The timescale of this deformation depends on the viscosity of the mantle prism. Previous researchers investigating postseismic deformation following the Sumatran earthquakes have estimated mantle viscosities in this region between 10^{17} to 10^{19} Pa s [Lubis *et al.*, 2012; Panet *et al.*, 2007; Pollitz *et al.*, 2006]. The value of the mantle viscosity can depend on the time window of geodetic observations examined; examining geodetic observations late in the earthquake cycle yield mantle viscosities one order of magnitude higher than geodetic observations early in the earthquake cycle [Meade *et al.*, 2013]. In addition, some studies of early stage postseismic deformation assume predominantly viscoelastic deformation and do not take into account afterslip – employing instead transient rheologies to explain different time-dependent decay patterns of geodetic time series (e.g. Panet *et al.* [2007]). As we are modeling a multi-year-long time series of postseismic displacements, we need to account for the relative contribution of both afterslip and viscoelastic relaxation of the mantle to the postseismic time series. Further, Feng *et al.* [2015b] found that the postseismic time series are represented by a logarithmic dependence of amplitude decay with time, and that a wide range of logarithmic decay times are needed to fit the time series. They propose that this wide range of temporal decay times indicates time series that represent the cumulative effects of several different postseismic relaxation mechanisms.

Traditionally, geodetic studies remove the contribution of viscoelastic deformation from postseismic time series by forward modeling viscoelastic deformation based on reasonable assumptions of the coseismic slip distribution and the Earth model (e.g., Suito and Freymueller [2009]), and invert the residual postseismic time series for afterslip distributions. This method assumes that viscoelastic deformation is driven only by coseismic stress changes, but neglects the contribution of viscoelastic deformation driven

by time-dependent afterslip, which may be a significant contribution if afterslip occurs at predominantly deeper depths along the megathrust. Also, this method suffers from our insufficient knowledge of the Earth model.

We therefore investigated an alternative method of determining the relative contributions of afterslip and viscoelastic deformation to the postseismic time series. We discretized the mantle volume into nine blocks of finite volume that can accommodate ductile deformation below the downdip end of the model fault, at asthenospheric depths. We define these blocks as ductile shear zones. We expect that viscous flow occurs predominantly in the asthenosphere, and as the lithosphere-asthenosphere boundary is located at ~ 100 km depths in Sumatra [Kumar *et al.*, 2007], we placed the ductile shear zones at ~ 100 – 180 km depths. The lengths of these zones extend the Bengkulu segment (Figure S3.2a), and are approximately aligned with the subducting slab (from Slab 1.0) at these depths.

We define a system of reference e_1 , e_2 and e_3 , where e_1 and e_2 are the trench perpendicular and trench parallel directions, and e_3 the depth direction. We then linearly invert for each of the six independent strain components in each of the nine ductile shear zones. As viscous flow is deviatoric, this implies that $e_{11} + e_{22} + e_{33} = 0$. The addition of nine ductile shear zones with six components adds 54 unknown strain parameters into our inversion. The inversion recovers afterslip and strain, which have different units. We use units of centimeters and microstrains to build the afterslip and strain kernels, respectively. The afterslip field is smoothed using a 2D Laplacian. The Green's functions for distributed inelastic strain in finite volumes were derived by *S. Barbot (personal communication)*. In our study, the Green's functions for each strain component were calculated with the Relax software [Barbot and Fialko, 2010a; b; Barbot *et al.*, 2009]. We first designed a

matrix of Green's functions for both afterslip and strains, as well as the matrix of Laplacian smoothing constraints. We then loaded into PCAIM the Green's functions and Laplacian smoothing constraints, and modified the software's functions to invert for both afterslip and strain components. We experimented with a range of Laplacian smoothing weights for afterslip and strains.

Figure 3.2 (a) Dip profile of our fault model approximates the Slab 1.0 model [Hayes *et al.*, 2012] along this portion of the megathrust. Grey dots define the fault geometry of Slab 1.0 from a suite of depth cross-sections along this section of the megathrust. Oblique view (b) and cross-sectional view (c) showing relative location and geometry of the fault model at < 100 km depths, and nine ductile shear zones located at ~100–180 km asthenospheric depths.

3.4. Model resolution

We conducted checkerboard tests of different sized checks to assess the resolution of slip across our fault model. The synthetic GPS displacements from the checkerboard tests were inverted for slip on the fault employing the same Laplacian smoothing weight as employed in the inversions for the models which do not take into account viscoelastic effects.

Figure 3.3a–c show the input checkerboard of slip across our fault model with 60, 80 and 120 km sized square checks, respectively. The 120 km sized slip checks are well recovered under and NW of the Pagai Islands, except in regions closest to the trench (Figure 3.3f). The estimated slip checks are smoothed across neighbouring fault patches in the region SE of the Pagai Islands. For the 60 and 80 km sized checks, slip is only reasonably recovered under the islands, and slip is smoothed across neighbouring fault patches. Similarly, the 60 and 80 km sized square checks are poorly recovered in regions close to the trench, and over the majority of the region SE of the Pagai Islands (Figures 3.3d and 3.3e), which suggests that afterslip is poorly resolved in these regions. We cannot exclude the possibility that afterslip in these parts of the fault is an artefact of our model.

In contrast, some slip checks are recovered at the Pagai Islands and along the Sumatran coast, downdip of the 2007 mainshock rupture area, albeit with some smoothing of slip across neighbouring fault patches. This suggests that afterslip under and west of the Pagai Islands is somewhat well resolved. Also, slip checks are reasonably well recovered in the region of the megathrust between Sipora and the Pagai Islands.

3.4.1. Testing various slip penalization constraints due to limited model resolution in the 2007 mainshock rupture area

Unfortunately, the model resolution southeast of the Pagai Islands is poor, since there are no overlying forearc islands along this portion of the megathrust. We therefore tested inversions based on placing constraints to penalize slip on different parts of the fault, and evaluated the influence of the different constraints on the data-model fits. In this way, we could assess for which regions of the fault afterslip needs to occur in order to yield the best fits to the data. The four different constraints that we tested were: 1) allowing afterslip in all regions of the fault, 2) penalizing afterslip on patches closest to the trench, 3) penalizing afterslip on all patches that slipped > 4 m coseismically, and 4) penalizing afterslip on patches that slipped > 5 m coseismically. The second case of penalizing afterslip on patches closest to the trench allows us to determine the spatial relationship between any shallow afterslip and the 2010 earthquake rupture area. The latter two cases of slip penalization constraints on patches that slipped coseismically are based on geodetic studies and rate-and-state friction models that suggest areas of coseismic slip and afterslip may be complementary to each other (e.g., Miyazaki *et al.* [2004]).

We used the reduced chi-square to evaluate quantitatively the fit of the data and the model displacements. The reduced chi-square of data-model fits, χ_r^2 , is defined as:

$$\chi_r^2 = \frac{\sum_k^{Sta} \sum_j^{Comp} \sum_t^{Time} \left(\frac{data_t - model_t}{\sigma_t} \right)_{jk}^2}{DOF}$$

The numerator represents the chi-squared statistic, which is summed over: 1) the number of time epochs, 2) each of the three components (east, north, up), and 3) the number of stations. The number of degrees of freedom (DOF) is defined as the product of the number of station components and the total number of time epochs.

Figure 3.3 Checkerboard tests with various sizes of input checks: (a) 60 x 60 km, (b) 80 x 80 km, and (c) 120 x 120 km, and recovery of slip in various parts of the fault model (d-f). Data and model displacements are shown in panels (g) to (i). Green triangles and circles: locations of the 17 SuGAR stations and 1 IGS station used in the inversion.

3.5. Results

3.5.1. Coseismic slip distribution

Figure 3.4a shows the results of our inverted coseismic slip distribution for the M_w 8.4, 7.9 and 7.0 earthquakes (Figure S3.1 shows the results of inverting only GPS data). Our model yields a reduced chi-square of 1.2, and a geodetic moment of 4.18×10^{21} N m. The spatial pattern of our coseismic model is very similar to that of *Konca et al.* [2008], although the locus of peak coseismic slip is located further downdip (Figure 3.4b). This may be related to the fact that we used a curved fault geometry, rather than a planar fault with a constant dip. The maximum coseismic slip in our model is ~ 7 m, comparable to the maximum slip value in *Konca et al.* [2008]’s model. We note that although the value of peak slip is poorly resolved and depends on the choice of Laplacian smoothing, varying the smoothing over three orders of magnitude results in a small difference in estimated seismic moments (Table S3.1). The seismic moment is therefore robustly estimated.

Figure 3.4 (a) Coseismic slip distribution of the 2007 Bengkulu earthquake M_w 8.4 mainshock and M_w 7.9 and 7.0 aftershocks. Our model is similar to the coseismic model of *Konca et al.* [2008] (1,3,5,7 m slip contours plotted in purple), although the locus of peak coseismic slip is located further downdip. The maximum coseismic slip is our model is ~ 7 m. (b) Data and model GPS displacements at 17 SuGAR stations and 1 IGS station (main), and 18 coral sites (inset).

3.5.2. Suite 1: Spatio-temporal distribution of afterslip on the megathrust

Figure 3.5 shows our model of cumulative afterslip in an ~6.3-year period following the Bengkulu earthquake, and Figure 3.6 shows the time series of data-model fits. Figure S3.4 shows the temporal evolution of afterslip. Our model yields a reduced chi-square of 18.4 and a geodetic moment of 3.55×10^{21} N m, equivalent to a moment magnitude of $\sim M_w$ 8.3. The cumulative afterslip moment amounts to ~85% of the coseismic geodetic moment. Our results suggest that peak afterslip is concentrated in regions updip and downdip of the 2007 mainshock area, with some afterslip located within the mainshock area (Figure 3.5b). We note that horizontal data-model fits are poor at ABGS, NTUS and MLKN, and the vertical data-model fits are poor at MKMK, LNNG and LAIS (Figure 3.6). This may be because the postseismic displacements on these components record deformation due to viscoelastic relaxation of the mantle.

Figure 3.7 shows the spatial distribution of afterslip in the case of placing slip penalization constraints on various parts of the fault. Similar data-model fits are obtained in both cases of allowing and penalizing afterslip close to the trench (compare Figures 3.5a and 3.7a), indicating that the data are not sensitive to slip in regions close to the trench. The reduced chi-square of data-model fits increases slightly in the case of penalizing afterslip in the 2007 mainshock area, suggesting that afterslip may have occurred within the mainshock area, but this inference is probably not robust due to poor model resolution in this part of the fault.

Figures 3.5a and 3.7a therefore represent two ‘best-fit’ afterslip models. Both models suggest peak cumulative afterslip of ~2 m immediately downdip of the 2007 mainshock area (region A, Figure 3.5a), at ~30–75 km depths on the megathrust. These depths coincide with the range of depths corresponding to the downdip aseismic-seismic

transition [Tichelaar and Ruff, 1993], and the downdip limit of interseismic coupling, at $\sim 40\text{--}60$ km depths on the megathrust [Chlieh *et al.*, 2008]. Afterslip located downdip of the mainshock rupture area were also found for the 2004 M_w 9.1 Sumatra-Andaman [Chlieh *et al.*, 2007], 2005 M_w 8.6 Nias [Prawirodirdjo *et al.*, 2010], 1994 M_w 7.7 Sanriku [Melbourne *et al.*, 2002] and 1964 M_w 9.2 Alaska earthquakes [Zweck *et al.*, 2002]. In addition, a moderate amount of cumulative afterslip ($\sim 1\text{--}2$ m) occurred within the 2007 mainshock area (region D, Figure 3.5a), and close to the 2007 M_w 7.9 aftershock area (region C, Figure 3.5a). However, details of peak afterslip in the shallow region of the fault differ between the two models: one model (Figure 3.5a) suggests ~ 2 m of cumulative shallow afterslip concentrated updip of the 2007 mainshock area at < 12 km depths on the megathrust, adjacent to the 2010 rupture area (region B, Figure 3.5a), and another model (Figure 3.7a) suggests ~ 1 m of cumulative shallow afterslip located at $\sim 12\text{--}20$ km depths, immediately downdip from the 2010 earthquake rupture area. We note again that afterslip values depend on the amount of smoothing in the model.

Figure 3.5 (a) Spatial distribution of cumulative afterslip in an ~6.3-year period following the 2007 mainshock, for models that allow afterslip in all regions of the fault, and relationship between the afterslip and the 2007 and 2010 earthquake rupture areas. Purple: 1, 3, 5 and 7 m slip contours of the 2007 Bengkulu mainshock and aftershock (from *Konca et al.* [2008]). Blue: 1, 3, 5, and 7 m slip contours of the 2010 earthquake estimated by *Hill et al.* [2012]. Regions A to D are discussed in the text. Purple circles: Aftershocks of the 2007 event from December 2007 to October 2008, from *Collings et al.* [2012]. (b) Comparison of data and model displacements.

(c) Verticals (cm)

Figure 3.6 Time series of data and model displacements at each station for the model shown in Figure 3.5, on east (a), north (b) and vertical (c) components. Grey: Data displacements and 1σ uncertainties. Colors: Model displacements, colored by station.

Figure 3.7 Afterslip distribution in the case of placing slip penalization constraints on various parts of the fault: (a) penalizing afterslip close to the trench. The reduced chi-square is similar to the case of allowing afterslip everywhere on the fault, indicating that slip is poorly resolved in regions close to the trench. (b) penalizing afterslip on all patches that slipped > 4 m coseismically, and (c) penalizing afterslip on patches that slipped > 5 m coseismically. In (b) and (c), patches with slip penalization constraints are outlined in white. Data and model displacements are shown in each figure.

3.5.3. Suite 2: Spatio-temporal distribution of afterslip on the megathrust, and strains within ductile shear zones

Figure 3.8 shows our joint inversion results of afterslip on the megathrust and strains in ductile shear zones. Figure 3.9 shows the time series of data-model fits, and Figure S3.5 the spatio-temporal evolution of afterslip. Compared to the models in suite 1 that do not take into account viscoelastic effects (Figure 3.5a and 3.7a), this model suggests a lesser amount of cumulative afterslip in the region downdip of the 2007 mainshock and aftershock areas (region A). This suggests that the estimated afterslip in the downdip region in Figure 3.5a actually represents a portion of the viscoelastic deformation mapped onto the megathrust. The spatial distribution of afterslip is quite different in the shallow region of the fault. The highest amount of cumulative afterslip in the shallow region (~ 1.4 - 1.6 m) is now located in the 2010 earthquake rupture area, rather than updip of the 2007 mainshock area (region B). Also, significantly less afterslip is located in the mainshock rupture area (region D). This model yields better data-model fits (a lower reduced chi-square of 12.6), and a geodetic moment of 3.24×10^{21} N m, equivalent to a moment magnitude of $\sim M_w$ 8.3. The cumulative afterslip moment amounts to $\sim 77\%$ of the coseismic geodetic moment. The main conclusion from comparing these two suites of models is that the models that take into account viscoelastic effects yield better data-model fits than those that do not, suggesting that it is important to factor in deformation due to viscoelastic relaxation of the mantle. However, we may not yet have optimized our estimates of the viscoelastic deformation, because poor data-model fits are still evident on the vertical time series of stations LNNG, LAIS and MKMK, which are located downdip of the 2007 mainshock.

We highlight that further work is needed to better understand and constrain these results. Firstly, we need to understand the distribution of strain suggested by these models. In

Figure 3.8b, the strain distribution in the ductile shear zones is not very smooth, and contrary to our expectation that the largest strains should be located closer to the downdip end of the fault, where coseismic stress changes are largest. Secondly, we need to better understand the tradeoffs between the changes in the afterslip distribution and strains in the ductile shear zones. Thirdly, we may need to investigate allowing viscoelastic deformation in the mantle wedge. Following the 2004 Sumatra-Andaman earthquake, researchers have found that postseismic GPS data can be explained by an elastic layer, that ranges from 62–90 km depths, overlying a viscoelastic layer [Paul *et al.*, 2012; Pollitz *et al.*, 2008], where a thicker elastic layer explains better the near-field GPS data. These elastic layer depths are shallower than that allowed for in our model parameterization, and warrants the need to investigate allowing for viscoelastic deformation in the mantle wedge i.e. at depths shallower than 100 km. Notably, if viscoelastic deformation occurs at these shallower depths, it would have a greater impact on observed surface velocity fields. In our present model, we speculate that the high strains within the shallowest ductile shear zones may represent a combination of viscoelastic deformation from the mantle wedge as well as from the asthenosphere.

Figure 3.8 Joint inversion of afterslip on the megathrust (a), deviatoric strains within ductile shear zones (b), and GPS data and model displacements (c). Note that panel (b) shows the cross-sectional ends of the shear zones, with the top row being the shallowest row of boxes. Compared to the model shown in Figure 3.5, the magnitude of afterslip in region A in this model is less.

(b) North (cm)

(c) Verticals (cm)

Figure 3.9 Time series of data and model displacements at each station for the model shown in Figure 3.7, on east (a), north (b) and vertical (c) components. See Figure 3.6 for similar captions.

3.6. Discussion

Our model results suggest that reasonably well-resolved areas of afterslip are concentrated downdip of the 2007 mainshock rupture area, and updip of the portion of the megathrust under the Pagai Islands. Our results suggest that taking into account viscoelastic effects improves data-model fits. In models that do not account for viscoelastic effects, some amount of afterslip is concentrated downdip of the mainshock rupture zone, but which may actually represent viscoelastic deformation mapped onto the megathrust. After incorporating ductile shear zones in a joint inversion of afterslip on the megathrust and strains in shear zones, less afterslip is concentrated downdip of the mainshock rupture area. Afterslip in the downdip region occurs at depths $\sim 30\text{--}75$ km on the megathrust. Geodetic studies of fault coupling along the Sumatra subduction zone suggest that the downdip limit of the seismogenic zone in this region varies between 30 to 55 km, with predominantly creep at depths >55 km [Chlieh *et al.*, 2008; Simoes *et al.*, 2004], marking the depth of the seismic-aseismic downdip transition zone, where the onset of stable sliding and velocity-strengthening behavior may occur.

Our model suggests a greater amount of afterslip located downdip of the 2007 mainshock rupture area than the model of Lubis *et al.* [2012]. One possible explanation for this difference is that the vertical deformation reported by Lubis *et al.* [2012] was accidentally reversed [Feng *et al.*, 2015b]. As mentioned in the previous section, further work is needed to better understand and constrain our models that take into account viscoelastic effects, especially to understand the influence of the depth of the ductile shear zones on the solution, and the tradeoffs between the spatial distribution of afterslip and strains in ductile shear zones.

The models that do not account for viscoelastic effects suggest peak afterslip in the shallow region of the megathrust located updip of the 2007 mainshock area, adjacent to the 2010 earthquake rupture area (Figures 3.5a and 3.8). A similar data-model fit is obtained in the case of penalizing afterslip on patches closest to the trench (Figure 3.7a). This model suggests peak afterslip located updip of the portion of the megathrust that experienced > 4 m of coseismic slip, west of the Pagai Islands. On the contrary, the models that take into account viscoelastic effects suggest that peak afterslip in the shallow region overlaps the 2010 earthquake rupture zone. Due to poor model resolution in this region of the fault, all three shallow afterslip scenarios are possible. In future work, we will investigate alternative afterslip models by testing constraints that force complementarity between coseismic and afterslip stress changes, so that we jointly invert for coseismic slip and afterslip with constraints inspired by the theory of rate-and-state friction. Nevertheless, these scenarios indicate that the spatial distribution of afterslip in the shallow region of the fault overlaps and borders the 2010 earthquake rupture area.

The shallow afterslip in our models is located at similar depths to shallow afterslip observed after the 2005 Nias earthquake [Hsu *et al.*, 2006]. Our results are consistent with the model described by Marone *et al.* [1991], in which coseismic rupture of the seismogenic zone causes stress perturbations in the updip zone of velocity-strengthening behavior, which are subsequently relaxed as earthquake afterslip. Researchers commonly characterize subduction zone trench regions as aseismic, based primarily on two inferences: 1) that the updip limit of seismicity, defining a seismic front, is located at ~ 5 -10 km depths on subduction zone megathrusts [Byrne *et al.*, 1988; Lay and Bilek, 2007; Lay *et al.*, 2012; Tichelaar and Ruff, 1993]; 2) that there is an abundance of materials at trench regions that exhibit velocity-strengthening behavior, such as unconsolidated sediments and smectite [Byrne *et al.*, 1988; Moore and Saffer, 2001; Vrolijk, 1990]. In

addition, geodetic studies regularly characterize the shallowest portions of megathrusts as aseismic, often as a result of unavailable model resolution in this region [Chlieh *et al.*, 2008]. However, it is clear that the seismic behavior of the shallow portion of the megathrust appears to be more complex than previously thought. Specifically, the shallow region of megathrusts can both participate in large earthquakes, as exemplified by the 2011 M_w 9.0 Tohoku-Oki earthquake, as well as host “tsunami earthquakes”, as exemplified by the 1314 and 2010 M_w 7.8 Mentawai 1992 M_w 7.6 Nicaragua, 1994 M_w 7.8 Java and 1996 M_w 7.5 Peru earthquakes [Hill *et al.*, 2012; Philiposian *et al.*, 2012; Polet and Kanamori, 2000]. It is intriguing that some subduction zones host more “tsunami earthquakes” than others. For example, based on historical records, Nicaragua hosts more “tsunami earthquakes” than Costa Rica (Dixon, *personal communication*). The question that needs to be addressed is therefore: what physical conditions and properties control the spatial and temporal pattern of shallow seismic ruptures?

A number of proposed models attempt to explain the spatial patterns of “tsunami earthquake” occurrence in the shallow portions of megathrusts. For example, Polet and Kanamori [2000] proposed that the subduction of rough, faulted oceanic plate increases the potential for a “tsunami earthquake”. Lay and Bilek [2007] highlighted the importance of considering temporally varying physical factors such as seafloor structures, pore fluid pressures and strain rates. They proposed a model in which the region just downdip of the zone of stable sliding close to the trench consists of conditionally stable patches. Their model implies that the shallow region commonly slips aseismically, but may occasionally fail coseismically in spatially distinct patches, if loaded at high strain rates [Bilek and Lay, 2002; Bilek *et al.*, 2004; Lay *et al.*, 2012; Scholz, 1998]. Possible processes that can cause high strain rates include afterslip and slow slip events. For example, along the weakly coupled section of the megathrust offshore Peru, a slow slip event and seismic

sequence in 2009 preceded shallow rupture of the megathrust; the preceding events were proposed to promote eventual seismic rupture of conditionally stable patches in the shallow trench region [Villegas-Lanza *et al.*, 2016]. In the context of Lay and Bilek [2007]’s model, the 2010 Mentawai earthquake represents coseismic rupture of a spatially distinct patch of conditional stability on the fault. The fact that the region closest to the trench ruptured in this earthquake implies that these conditionally stable patches can exist all the way to the trench. We propose that shallow afterslip regions in our model may have caused localized increases in strain rates adjacent to and within the Mentawai earthquake rupture area, perhaps in concert with increased coseismic stresses in this region. The increased strain rates may have brought forward in time its coseismic failure in 2010.

Other researchers have highlighted the importance of considering the role of décollements and structures in the accretionary wedge that facilitate coseismic rupture of the shallow regions of subduction zones [Hubbard *et al.*, 2015]. Bathymetric maps and seismic reflection results from the MEGATERA (Mentawai-Gap Tsunami Earthquake Risk Assessment) cruise in the 2010 Mentawai earthquake rupture area reveal paired thrust faults associated with fold and thrust belts at the toe of the accretionary wedge that were likely generated by coseismic rupture in 2010, and played an important role in tsunamigenesis [Singh *et al.*, 2011a]. Geodetic models of afterslip should ideally take into account these mapped structures in trench regions. Such efforts are currently limited by the lack of model resolution in trench regions, but with the advent of seafloor geodetic networks installed at an increasing number of subduction zones worldwide, geodesists may begin to explore these models.

The geodetic moment of our coseismic slip distribution model is 4.18×10^{21} N m, which is similar to a value of 5×10^{21} N m obtained by *Konca et al.* [2008]. Our suites of afterslip models suggest cumulative afterslip geodetic moments ranging from 3.24 to 3.55×10^{21} N m, equivalent to a moment magnitude of M_w 8.3, and which amounts to ~ 77 to 85% of the coseismic geodetic moment. Collectively, the coseismic and afterslip geodetic moment amounts to $\sim 7.42\text{--}7.73 \times 10^{21}$ N m, representing $\sim 28\%$ of the moment released during the 1833 earthquake (M_w 8.8–8.9, [*Philibosian et al.*, 2014]). We note that these estimates are of the first order, because we are not taking into account along-strike variations in moment contributions from both the 1797 and 1833 earthquakes. Still, our results suggest that the collective moment released by the mainshock and ensuing afterslip represent only a fraction of the moment released during the 1797 and 1833 earthquakes, and that the potential for another large earthquake along this portion of the megathrust remains large.

3.7. Conclusions

Our inversion model results suggest that afterslip occurs both updip and downdip of the 2007 mainshock area, likely in regions that overlap and flank the seismogenic zone. One of our most important results is that the shallow patches of afterslip border and overlap the 2010 Mentawai earthquake rupture zone. We propose that the localized increases in strain rates caused by coseismic stress changes and ensuing shallow afterslip may have brought forward in time coseismic failure of the shallow portion of the megathrust in 2010.

Table 3.1 17 SuGAR stations and 1 IGS station (NTUS) recorded coseismic and postseismic displacements from the 2007 M_w 8.4 Bengkulu earthquake, including possible contributions from the M_w 7.9 and M_w 7.0 aftershocks.

Station	Longitude	Latitude	Coseismic displacements (mm)						Postseismic displacements - cumulative, over ~ 7.3 years (mm)					
			E	1 σ	N	1 σ	U	1 σ	E	1 σ	N	1 σ	U	1 σ
ABGS	99.388	0.221	-1.1	0.3	-22.1	0.4	-5.2	0.8	-3.3	0.8	-13.1	1.9	11.4	2.3
BSAT	100.285	-3.077	-947.8	2.8	-1082.9	3.5	747.3	3.6	-73.5	1.5	-93.4	1.6	-15.3	2.3
BTET	98.644	-1.282	-7.3	0.3	7.1	1.0	-7.9	1.3	0.0	0.0	6.3	0.8	-3.4	1.3
JMBI	103.520	-1.616	-64.9	0.8	-50.3	0.8	1.9	1.4	-20.7	1.3	-18.1	1.0	2.1	1.3
LAIS	102.034	-3.505	-624.9	1.7	-364.2	1.5	-118.6	2.3	-51.9	1.2	-49.9	1.2	-3.2	1.6
LNNG	101.156	-2.285	-402.5	1.1	-485.1	1.6	-123.1	2.1	-55.6	1.9	-90.1	2.4	-3.0	2.4
MKMK	101.091	-2.543	-491.8	2.1	-599.1	2.9	-191.3	3.1	-50.0	1.0	-73.5	1.2	-9.1	2.0
MLKN	102.276	-5.317	21.2	0.4	-18.4	0.4	-23.7	1.2	15.7	1.8	-16.9	1.9	-2.6	2.6
MNNA	102.890	-4.500	-66.2	0.5	-13.3	0.3	-18.4	0.8	-25.4	1.5	-12.2	0.9	5.4	1.7
MSAI	99.089	-1.326	-22.5	1.1	14.2	0.5	-7.4	1.6	-23.2	3.1	-6.7	1.3	7.2	3.7
NGNG	99.268	-1.800	-169.9	1.0	-30.1	0.9	61.2	2.1	-30.2	1.1	-26.4	1.0	18.6	1.9
NTUS	103.680	1.346	-15.0	1.4	-19.9	2.7	-4.3	1.9	-2.7	0.5	-5.9	0.5	-0.6	0.9
PBJO	98.516	-0.637	-4.7	0.3	0.6	0.3	-4.4	0.8	-	-	-	-	-	-
PKRT	99.543	-2.151	-354.9	3.6	-494.8	4.7	236.3	6.3	-69.4	6.2	-102.3	9.7	58.9	8.5
PPNJ	99.604	-1.994	-439.6	1.1	-560.2	0.9	238.3	3.2	-59.1	2.3	-54.1	2.2	15.7	2.1
PRKB	100.400	-2.967	-984.6	4.6	-1474	3.6	331.8	0.0	-70.6	1.6	-57.1	1.5	6.8	0.0
PSKI	100.353	-1.125	-128.5	0.5	-234.1	0.6	-55.3	1.2	-39.8	0.6	-70.9	1.0	7.6	1.2
SLBU	100.009	-2.766	-215.2	1.4	-107.3	2.3	107.9	3.9	-27.6	13.2	-68.5	22	46.2	15.3
TIKU	99.944	-0.396	-11.5	0.9	-76.7	1.9	-13.5	2.1	-5.2	0.7	-14.9	1.2	1.5	1.9

Table 3.2 Coral uplift measurements reported at 18 sites on the Pagai Islands, Sipora and Mega Island (from *Konca et al.* [2008]).

Site	Longitude	Latitude	Coseismic displacements		Measurement date
			Uplift (mm)	2 σ (mm)	
MEG07-A	101.034	-4.007	1350	140	10/6/2007
MEG07-B	101.039	-4.017	1250	60	10/6/2007
STP07-A	100.683	-3.454	200	120	10/6/2007
SDG07-A	100.637	-3.486	930	230	10/7/2007
TBO07-A	100.467	-3.346	910	140	10/7/2007
LMS07-A	100.331	-3.209	990	230	10/7/2007
TNP07-A	100.506	-3.164	450	120	10/8/2007
TNK07-A	100.406	-3.186	780	230	10/8/2007
LBT07-A	100.229	-3.116	830	60	10/8/2007
PJS07-A	100.158	-3.016	350	120	10/9/2007
SPG07-A	100.177	-2.886	-30	140	10/9/2007
BSG07-A	100.184	-2.838	-190	140	10/9/2007
TNG07-A	100.280	-2.821	-30	180	10/9/2007
MBL07-A	100.013	-2.516	0	140	10/9/2007
TMS07-A	99.611	-2.019	50	120	10/10/2007
SGS07-A	99.651	-2.047	100	60	10/10/2007
RKT07-A	99.708	-2.117	200	140	10/10/2007
SMY07-A	100.111	-2.604	0	180	9/30/2007

3.8. Supporting Information

Figure S3.1 Coseismic slip distribution estimated by inverting GPS coseismic offsets only. The slip distribution is very similar to that inverted with both GPS and coral data (Figure S3.4).

Figure S3.2 (a) Singular values of each principal component, and (b) percentage of variance in the data explained by each principal component.

Figure S3.3 Comparison of the spatial distribution of afterslip in the case of approximating the data time series using (a) 1 principal component, and (b) 2 principal components.

Figure S3.4 Spatio-temporal evolution of afterslip for the model that does not take into account viscoelastic effects (Figure 3.5).

Figure S3.5 Spatio-temporal evolution of afterslip for the model that takes into account viscoelastic effects (Figure 3.8).

Table S3.1 Sensitivity of estimated coseismic moment to degree of Laplacian smoothing

Laplacian weight	Seismic moment ($\times 10^{21}$ N m)
0.03	3.70
0.3	3.52
3	3.73
30	4.22
300	4.32

Concluding thoughts

At subduction zones, slip processes such as long-term SSEs, afterslip and viscoelastic relaxation of the mantle can span timescales longer than the operating time of current GPS networks. Understanding these processes requires examining long-term geodetic records. In this thesis, I have examined long geodetic time series at the Sumatran subduction zone to better characterize the spatio-temporal spectrum of slip behavior on the Sunda megathrust.

The first chapter of my thesis addresses the question: From a long geodetic record of interseismic deformation, what is the character of interseismic deformation? Does it vary in time, and why or why not? A paleogeodetic record of interseismic deformation that spans the past two earthquake cycles at Simeulue Island reveals that interseismic deformation rates are not uniform in time. One way to explain these rate variations is temporal changes in the downdip limit of locking under the island. In addition, we found that the pattern of interseismic coupling preceding earthquakes between two seismic cycles may not necessarily be the same, even though the earthquakes themselves have a similar rupture pattern. Our results highlight the need for researchers to consider GPS derived interseismic coupling maps as only a snapshot of conditions on the fault that can vary over time. The temporal changes in fault coupling may be the manifestation of longer-term tectonic processes such as viscoelastic relaxation of the mantle – there is scope for further work on this subject.

Chapter 2 addresses the question: Do slow slip events occur in the Sumatran subduction zone? The answer is yes. From coral data, we have discovered a 15-year-long slow slip event that occurred between 1966 and 1981 on the Banyak Islands section of the

megathrust. This event is the longest lasting slow slip event reported at any subduction zone, and the first to be observed in Sumatra. Our findings complement that of *Feng et al.* [2015a], who found no slow slip events in the ~12-year-long SuGAR time series. This finding led them to consider whether the absence of SSEs in the past decade are either because they are too small to be detected, related to the current state of the earthquake cycle, or that similarly long-lasting SSEs does not enable us to yet characterize such events from the GPS time series.

Lastly, chapter 3 addresses the question: How does seismic, postseismic and interseismic deformation history affect the timing and spatial pattern of seismic ruptures? We focused particularly on the interesting tale of two earthquakes that occurred in the southern portion of the Mentawai section of the megathrust in 2007 and 2010. From examining a ~6.3-year-long GPS time series, we discovered that afterslip following the 2007 Bengkulu earthquake borders and overlaps the rupture zone of the 2010 earthquake. The collective contribution of afterslip and coseismic stress changes likely caused localized increases in strain rates that may have brought forward in time rupture of the shallow region of the megathrust in 2010.

These studies illuminate new characteristics of slip behavior on the Sunda megathrust, especially in understanding the timescale at which these slip processes occur (Figure 4.1). Understanding the spatio-temporal variability of these slip processes is important for gaining a comprehensive understanding of the earthquake cycle at the Sumatran subduction zone.

Figure 4.1 Timescale of earthquake cycle slip behavior in subduction zones (black arrows), and timescales at which we can probe this slip behavior with GPS and paleogeodesy (red arrows). The results of my thesis extends our knowledge of the timescale of slip behavior in the Sumatran subduction zone (magenta arrows): slow slip events can continue for more than a decade, and interseismic coupling changes can persist for a few tens of decades.

Bibliography

- Altamimi, Z., L. Métivier, and X. Collilieux (2012), ITRF2008 plate motion model, *Journal of Geophysical Research: Solid Earth*, 117(B7), B07402.
- Barbot, S. (2015), Unified cycle of earthquakes: Unicycle documentation, edited.
- Barbot, S., and Y. Fialko (2010a), A unified continuum representation of post-seismic relaxation mechanisms: semi-analytic models of afterslip, poroelastic rebound and viscoelastic flow, *Geophysical Journal International*, 182(3), 1124-1140.
- Barbot, S., and Y. Fialko (2010b), Fourier-domain Green's function for an elastic semi-infinite solid under gravity, with applications to earthquake and volcano deformation, *Geophysical Journal International*, 182(2), 568-582.
- Barbot, S., Y. Fialko, and Y. Bock (2009), Postseismic deformation due to the Mw 6.0 2004 Parkfield earthquake: Stress-driven creep on a fault with spatially variable rate-and-state friction parameters, *Journal of Geophysical Research: Solid Earth*, 114(B7), B07405.
- Beckley, B., F. Lemoine, S. Luthcke, R. Ray, and N. Zelensky (2007), A reassessment of global and regional mean sea level trends from TOPEX and Jason-1 altimetry based on revised reference frame and orbits, *Geophysical Research Letters*, 34(14).
- Bilek, S. L., and T. Lay (2002), Tsunami earthquakes possibly widespread manifestations of frictional conditional stability, *Geophysical Research Letters*, 29(14).
- Bilek, S. L., T. Lay, and L. J. Ruff (2004), Radiated seismic energy and earthquake source duration variations from teleseismic source time functions for shallow subduction zone thrust earthquakes, *Journal of Geophysical Research: Solid Earth*, 109(B9), B09308.
- Briggs, R. W., et al. (2006), Deformation and Slip Along the Sunda Megathrust in the Great 2005 Nias-Simeulue Earthquake, *Science*, 311(5769), 1897-1901.
- Byrne, D. E., D. M. Davis, and L. R. Sykes (1988), Loci and maximum size of thrust earthquakes and the mechanics of the shallow region of subduction zones, *Tectonics*, 7(4), 833-857.
- Chambers, D. P., M. A. Merrifield, and R. S. Nerem (2012), Is there a 60-year oscillation in global mean sea level?, *Geophysical Research Letters*, 39(18), L18607.
- Chlieh, J. P. Avouac, K. Sieh, D. H. Natawidjaja, and J. Galetzka (2008), Heterogeneous coupling of the Sumatran megathrust constrained by geodetic and paleogeodetic measurements, *Journal of Geophysical Research: Solid Earth*, 113(B5), B05305.
- Chlieh, et al. (2007), Coseismic Slip and Afterslip of the Great Mw 9.15 Sumatra-Andaman Earthquake of 2004, *Bulletin of the Seismological Society of America*, 97(1A), S152-S173.

- Collings, R., D. Lange, A. Rietbrock, F. Tilmann, D. Natawidjaja, B. Suwargadi, M. Miller, and J. Saul (2012), Structure and seismogenic properties of the Mentawai segment of the Sumatra subduction zone revealed by local earthquake traveltime tomography, *Journal of Geophysical Research: Solid Earth*, 117(B1), B01312.
- Douglas, A., J. Beavan, L. Wallace, and J. Townend (2005), Slow slip on the northern Hikurangi subduction interface, New Zealand, *Geophysical Research Letters*, 32(16), L16305.
- Feng, L., E. M. Hill, P. Elósegui, Q. Qiu, I. Hermawan, P. Banerjee, and K. Sieh (2015a), Hunt for slow slip events along the Sumatran subduction zone in a decade of continuous GPS data, *Journal of Geophysical Research: Solid Earth*, 120(12), 8623-8632.
- Feng, L., E. M. Hill, P. Banerjee, I. Hermawan, L. L. H. Tsang, D. H. Natawidjaja, B. W. Suwargadi, and K. Sieh (2015b), A unified GPS-based earthquake catalog for the Sumatran plate boundary between 2002 and 2013 *Journal of Geophysical Research: Solid Earth*.
- Franke, D., M. Schnabel, S. Ladage, D. R. Tappin, S. Neben, Y. S. Djajadihardja, C. Müller, H. Kopp, and C. Gaedicke (2008), The great Sumatra–Andaman earthquakes — Imaging the boundary between the ruptures of the great 2004 and 2005 earthquakes, *Earth and Planetary Science Letters*, 269(1–2), 118-130.
- Freymueller, J. T., S. C. Cohen, and H. J. Fletcher (2000), Spatial variations in present-day deformation, Kenai Peninsula, Alaska, and their implications, *Journal of Geophysical Research: Solid Earth*, 105(B4), 8079-8101.
- Fu, Y., and J. T. Freymueller (2013), Repeated large Slow Slip Events at the southcentral Alaska subduction zone, *Earth and Planetary Science Letters*, 375(0), 303-311.
- Fukuda, J. i., K. M. Johnson, K. M. Larson, and S. i. Miyazaki (2009), Fault friction parameters inferred from the early stages of afterslip following the 2003 Tokachi-oki earthquake, *Journal of Geophysical Research: Solid Earth*, 114(B4), n/a-n/a.
- Hamlington, B., R. Leben, R. Nerem, W. Han, and K. Y. Kim (2011), Reconstructing sea level using cyclostationary empirical orthogonal functions, *Journal of Geophysical Research: Oceans (1978–2012)*, 116(C12).
- Hayes, G. P., D. J. Wald, and R. L. Johnson (2012), Slab1.0: A three-dimensional model of global subduction zone geometries, *Journal of Geophysical Research: Solid Earth*, 117(B1), B01302.
- Hill, E. M., et al. (2012), The 2010 Mw 7.8 Mentawai earthquake: Very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS data, *Journal of Geophysical Research: Solid Earth*, 117(B6), B06402.
- Hirose, H., K. Hirahara, F. Kimata, N. Fujii, and S. i. Miyazaki (1999), A slow thrust slip event following the two 1996 Hyuganada Earthquakes beneath the Bungo Channel, southwest Japan, *Geophysical Research Letters*, 26(21), 3237-3240.
- Hsu, Y.-J., M. Simons, J.-P. Avouac, J. Galetzka, K. Sieh, M. Chlieh, D. Natawidjaja, L. Prawirodirdjo, and Y. Bock (2006), Frictional Afterslip Following the 2005 Nias-Simeulue Earthquake, Sumatra, *Science*, 312(5782), 1921-1926.

- Hubbard, J., S. Barbot, E. M. Hill, and P. Tapponnier (2015), Coseismic slip on shallow décollement megathrusts: implications for seismic and tsunami hazard, *Earth-Science Reviews*, 141, 45-55.
- Ide, S., G. C. Beroza, D. R. Shelly, and T. Uchide (2007), A scaling law for slow earthquakes, *Nature*, 447(7140), 76-79.
- Jacob, J., J. Dymment, and V. Yatheesh (2014), Revisiting the structure, age, and evolution of the Wharton Basin to better understand subduction under Indonesia, *Journal of Geophysical Research: Solid Earth*, 119(1), 2013JB010285.
- Jevrejeva, S., A. Grinsted, J. Moore, and S. Holgate (2006), Nonlinear trends and multiyear cycles in sea level records, *Journal of Geophysical Research: Oceans* (1978–2012), 111(C9).
- Kanamori, H., L. Rivera, and W. H. K. Lee (2010), Historical seismograms for unravelling a mysterious earthquake: The 1907 Sumatra Earthquake, *Geophysical Journal International*, 183(1), 358-374.
- Kaneko, Y., J.-P. Avouac, and N. Lapusta (2010), Towards inferring earthquake patterns from geodetic observations of interseismic coupling, *Nature Geosci*, 3(5), 363-369.
- Karato, S.-i. (2010), Rheology of the deep upper mantle and its implications for the preservation of the continental roots: A review, *Tectonophysics*, 481(1–4), 82-98.
- Konca, A. O., V. Hjorleifsdottir, T.-R. A. Song, J.-P. Avouac, D. V. Helmberger, C. Ji, K. Sieh, R. Briggs, and A. Meltzner (2007), Rupture Kinematics of the 2005 Mw 8.6 Nias–Simeulue Earthquake from the Joint Inversion of Seismic and Geodetic Data, *Bulletin of the Seismological Society of America*, 97(1A), S307-S322.
- Konca, A. O., et al. (2008), Partial rupture of a locked patch of the Sumatra megathrust during the 2007 earthquake sequence, *Nature*, 456(7222), 631-635.
- Kositsky, A. P., and J. P. Avouac (2010), Inverting geodetic time series with a principal component analysis-based inversion method, *Journal of Geophysical Research: Solid Earth*, 115(B3), B03401.
- Kumar, P., X. Yuan, M. R. Kumar, R. Kind, X. Li, and R. K. Chadha (2007), The rapid drift of the Indian tectonic plate, *Nature*, 449(7164), 894-897.
- Lay, T., and S. Bilek (2007), Anomalous Earthquake Ruptures at Shallow Depths on Subduction Zone Megathrusts, in *The Seismogenic Zone of Subduction Thrust Faults*, edited by T. Dixon and J. C. Moore, pp. 476 - 511, Columbia University Press.
- Lay, T., H. Kanamori, C. J. Ammon, K. D. Koper, A. R. Hutko, L. Ye, H. Yue, and T. M. Rushing (2012), Depth-varying rupture properties of subduction zone megathrust faults, *Journal of Geophysical Research: Solid Earth*, 117(B4), B04311.
- Li, S., J. T. Freymueller, and R. McCaffrey (2014), Time-dependent variations of slow slip events in Lower Cook Inlet of the Alaska-Aleutian subduction zone, paper presented at American Geophysical Union Fall Meeting, 2014, San Francisco.
- Lienkaemper, J. J., F. S. McFarland, R. W. Simpson, R. G. Bilham, D. A. Ponce, J. J. Boatwright, and S. J. Caskey (2012), Long-Term Creep Rates on the Hayward Fault:

- Evidence for Controls on the Size and Frequency of Large Earthquakes, *Bulletin of the Seismological Society of America*, 102(1), 31-41.
- Lorito, S., F. Romano, S. Atzori, X. Tong, A. Avallone, J. McCloskey, M. Cocco, E. Boschi, and A. Piatanesi (2011), Limited overlap between the seismic gap and coseismic slip of the great 2010 Chile earthquake, *Nature Geosci*, 4(3), 173-177.
- Loveless, J. P., and B. J. Meade (2011), Spatial correlation of interseismic coupling and coseismic rupture extent of the 2011 MW = 9.0 Tohoku-oki earthquake, *Geophysical Research Letters*, 38(17), L17306.
- Lubis, A. M., A. Hashima, and T. Sato (2012), Analysis of afterslip distribution following the 2007 September 12 southern Sumatra earthquake using poroelastic and viscoelastic media, *Geophysical Journal International*, 192(1), 18-37.
- Malservisi, R., et al. (2015), Multiscale postseismic behavior on a megathrust: The 2012 Nicoya earthquake, Costa Rica, *Geochemistry, Geophysics, Geosystems*, 16(6), 1848-1864.
- Marone, C. J., C. H. Scholtz, and R. Bilham (1991), On the mechanics of earthquake afterslip, *Journal of Geophysical Research: Solid Earth*, 96(B5), 8441-8452.
- McCaffrey, R., P. C. Zwick, Y. Bock, L. Prawirodirdjo, J. F. Genrich, C. W. Stevens, S. S. O. Puntodewo, and C. Subarya (2000), Strain partitioning during oblique plate convergence in northern Sumatra: Geodetic and seismologic constraints and numerical modeling, *Journal of Geophysical Research: Solid Earth*, 105(B12), 28363-28376.
- McNeill, L. C., and T. J. Henstock (2014), Forearc structure and morphology along the Sumatra-Andaman subduction zone, *Tectonics*, 33(2), 2012TC003264.
- Meade, B. J., and J. P. Loveless (2009), Predicting the geodetic signature of MW $\geq$ 8 slow slip events, *Geophysical Research Letters*, 36(1), L01306.
- Meade, B. J., Y. Klinger, and E. A. Hetland (2013), Inference of Multiple Earthquake-Cycle Relaxation Timescales from Irregular Geodetic Sampling of Interseismic Deformation, *Bulletin of the Seismological Society of America*, 103(5), 2824-2835.
- Melbourne, T. I., F. H. Webb, J. M. Stock, and C. Reigber (2002), Rapid postseismic transients in subduction zones from continuous GPS, *Journal of Geophysical Research: Solid Earth*, 107(B10), ETG 10-11-ETG 10-10.
- Meltzner, and C. D. Woodroffe (2014), Coral microatolls, in *Handbook of Sea-level Research*, edited, Wiley Chichester.
- Meltzner, K. Sieh, H.-W. Chiang, C.-C. Shen, B. W. Suwargadi, D. H. Natawidjaja, B. Philibosian, and R. W. Briggs (2012), Persistent termini of 2004- and 2005-like ruptures of the Sunda megathrust, *Journal of Geophysical Research: Solid Earth*, 117(B4), B04405.
- Meltzner, K. Sieh, H.-W. Chiang, C.-C. Shen, B. W. Suwargadi, D. H. Natawidjaja, B. E. Philibosian, R. W. Briggs, and J. Galetzka (2010), Coral evidence for earthquake recurrence and an A.D. 1390–1455 cluster at the south end of the 2004 Aceh–Andaman rupture, *Journal of Geophysical Research: Solid Earth*, 115(B10), B10402.

- Meltzner, et al. (2015), Time-varying interseismic strain rates and similar seismic ruptures on the Nias-Simeulue patch of the Sunda megathrust, *Quaternary Science Reviews*.
- Métois, M., A. Socquet, and C. Vigny (2012), Interseismic coupling, segmentation and mechanical behavior of the central Chile subduction zone, *Journal of Geophysical Research: Solid Earth*, 117(B3), B03406.
- Miura, S., Y. Suwa, A. Hasegawa, and T. Nishimura (2004), The 2003 M8.0 Tokachi-Oki earthquake – How much has the great event paid back slip debts?, *Geophysical Research Letters*, 31(5), L05613.
- Miyazaki, and K. M. Larson (2008), Coseismic and early postseismic slip for the 2003 Tokachi-oki earthquake sequence inferred from GPS data, *Geophysical Research Letters*, 35(4), n/a-n/a.
- Miyazaki, S., P. Segall, J. Fukuda, and T. Kato (2004), Space time distribution of afterslip following the 2003 Tokachi-oki earthquake: Implications for variations in fault zone frictional properties, *Geophysical Research Letters*, 31(6), n/a-n/a.
- Miyazaki, S., P. Segall, J. J. McGuire, T. Kato, and Y. Hatanaka (2006), Spatial and temporal evolution of stress and slip rate during the 2000 Tokai slow earthquake, *Journal of Geophysical Research: Solid Earth*, 111(B3), B03409.
- Moore, J. C., and D. Saffer (2001), Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: An effect of diagenetic to low-grade metamorphic processes and increasing effective stress, *Geology*, 29(2), 183-186.
- Moreno, M., M. Rosenau, and O. Oncken (2010), 2010 Maule earthquake slip correlates with pre-seismic locking of Andean subduction zone, *Nature*, 467(7312), 198-202.
- Moreno, M., et al. (2011), Heterogeneous plate locking in the South–Central Chile subduction zone: Building up the next great earthquake, *Earth and Planetary Science Letters*, 305(3–4), 413-424.
- Nalbant, S., J. McCloskey, S. Steacy, M. NicBhloscaidh, and S. Murphy (2013), Interseismic coupling, stress evolution, and earthquake slip on the Sunda megathrust, *Geophysical Research Letters*, 40(16), 4204-4208.
- Natawidjaja, D. H., K. Sieh, S. N. Ward, H. Cheng, R. L. Edwards, J. Galetzka, and B. W. Suwargadi (2004), Paleogeodetic records of seismic and aseismic subduction from central Sumatran microatolls, Indonesia, *Journal of Geophysical Research: Solid Earth*, 109(B4), B04306.
- Natawidjaja, D. H., K. Sieh, J. Galetzka, B. W. Suwargadi, H. Cheng, R. L. Edwards, and M. Chlieh (2007), Interseismic deformation above the Sunda Megathrust recorded in coral microatolls of the Mentawai islands, West Sumatra, *Journal of Geophysical Research: Solid Earth*, 112(B2), B02404.
- Natawidjaja, D. H., K. Sieh, M. Chlieh, J. Galetzka, B. W. Suwargadi, H. Cheng, R. L. Edwards, J.-P. Avouac, and S. N. Ward (2006), Source parameters of the great Sumatran megathrust earthquakes of 1797 and 1833 inferred from coral microatolls, *Journal of Geophysical Research: Solid Earth*, 111(B6), B06403.

- Newcomb, K. R., and W. R. McCann (1987), Seismic history and seismotectonics of the Sunda Arc, *Journal of Geophysical Research: Solid Earth*, 92(B1), 421-439.
- Nishimura, T., T. Hirasawa, S. i. Miyazaki, T. Sagiya, T. Tada, S. Miura, and K. Tanaka (2004), Temporal change of interplate coupling in northeastern Japan during 1995–2002 estimated from continuous GPS observations, *Geophysical Journal International*, 157(2), 901-916.
- Noda, H., and T. Hori (2014), Under what circumstances does a seismogenic patch produce aseismic transients in the later interseismic period?, *Geophysical Research Letters*, 41(21), 7477-7484.
- Ochi, T., and T. Kato (2013), Depth extent of the long-term slow slip event in the Tokai district, central Japan: A new insight, *Journal of Geophysical Research: Solid Earth*, 118(9), 4847-4860.
- Ohta, Y., F. Kimata, and T. Sagiya (2004), Reexamination of the interplate coupling in the Tokai region, central Japan, based on the GPS data in 1997–2002, *Geophysical Research Letters*, 31(24), L24604.
- Ohta, Y., J. T. Freymueller, S. Hreinsdóttir, and H. Suito (2006), A large slow slip event and the depth of the seismogenic zone in the south central Alaska subduction zone, *Earth and Planetary Science Letters*, 247(1–2), 108-116.
- Okada, Y. (1985), Surface deformation due to shear and tensile faults in a half-space, *Bulletin of the Seismological Society of America*, 75(4), 1135-1154.
- Ozawa, S., M. Murakami, M. Kaidzu, T. Tada, T. Sagiya, Y. Hatanaka, H. Yarai, and T. Nishimura (2002), Detection and Monitoring of Ongoing Aseismic Slip in the Tokai Region, Central Japan, *Science*, 298(5595), 1009-1012.
- Panet, I., V. Mikhailov, M. Diament, F. Pollitz, G. King, O. De Viron, M. Holschneider, R. Biancale, and J.-M. Lemoine (2007), Coseismic and post-seismic signatures of the Sumatra 2004 December and 2005 March earthquakes in GRACE satellite gravity, *Geophysical Journal International*, 171(1), 177-190.
- Paul, J., C. P. Rajendran, A. R. Lowry, V. Andrade, and K. Rajendran (2012), Andaman Postseismic Deformation Observations: Still Slipping after All These Years?, *Bulletin of the Seismological Society of America*, 102(1), 343-351.
- Peng, Z., and J. Gomberg (2010), An integrated perspective of the continuum between earthquakes and slow-slip phenomena, *Nature Geosci*, 3(9), 599-607.
- Perfettini, H., and J. P. Avouac (2004), Postseismic relaxation driven by brittle creep: A possible mechanism to reconcile geodetic measurements and the decay rate of aftershocks, application to the Chi-Chi earthquake, Taiwan, *Journal of Geophysical Research: Solid Earth*, 109(B2), B02304.
- Perfettini, H., and J. P. Avouac (2007), Modeling afterslip and aftershocks following the 1992 Landers earthquake, *Journal of Geophysical Research: Solid Earth*, 112(B7), B07409.
- Perfettini, H., J. P. Avouac, and J. C. Ruegg (2005), Geodetic displacements and aftershocks following the 2001 Mw = 8.4 Peru earthquake: Implications for the

mechanics of the earthquake cycle along subduction zones, *Journal of Geophysical Research: Solid Earth*, 110(B9), B09404.

Philibosian, B., K. Sieh, D. H. Natawidjaja, H.-W. Chiang, C.-C. Shen, B. W. Suwargadi, E. M. Hill, and R. L. Edwards (2012), An ancient shallow slip event on the Mentawai segment of the Sunda megathrust, Sumatra, *Journal of Geophysical Research: Solid Earth*, 117(B5), B05401.

Philibosian, B., K. Sieh, J.-P. Avouac, D. H. Natawidjaja, H.-W. Chiang, C.-C. Wu, H. Perfettini, C.-C. Shen, M. R. Daryono, and B. W. Suwargadi (2014), Rupture and variable coupling behavior of the Mentawai segment of the Sunda megathrust during the supercycle culmination of 1797 to 1833, *Journal of Geophysical Research: Solid Earth*, 119(9), 2014JB011200.

Polet, J., and H. Kanamori (2000), Shallow subduction zone earthquakes and their tsunamigenic potential, *Geophysical Journal International*, 142(3), 684-702.

Pollitz, R. Bürgmann, and P. Banerjee (2006), Post-seismic relaxation following the great 2004 Sumatra-Andaman earthquake on a compressible self-gravitating Earth, *Geophysical Journal International*, 167(1), 397-420.

Pollitz, P. Banerjee, K. Grijalva, B. Nagarajan, and R. Bürgmann (2008), Effect of 3-D viscoelastic structure on post-seismic relaxation from the 2004 M= 9.2 Sumatra earthquake, *Geophysical Journal International*, 173(1), 189-204.

Ponte, R. M., and F. Lyard (2002), Effects of unresolved high-frequency signals in altimeter records inferred from tide gauge data, *Journal of Atmospheric and Oceanic Technology*, 19(4), 534-539.

Prawirodirdjo, L., R. McCaffrey, C. D. Chadwell, Y. Bock, and C. Subarya (2010), Geodetic observations of an earthquake cycle at the Sumatra subduction zone: Role of interseismic strain segmentation, *Journal of Geophysical Research: Solid Earth*, 115(B3), B03414.

Reid, H. F. (1910), *The Mechanics of the Earthquake, The California Earthquake of April 18, 1906, Report of the State Investigation Commission*, 16-28 pp., Carnegie Institution of Washington, Washington, D.C.

Romano, F., A. Piatanesi, S. Lorito, N. D'Agostino, K. Hirata, S. Atzori, Y. Yamazaki, and M. Cocco (2012), Clues from joint inversion of tsunami and geodetic data of the 2011 Tohoku-oki earthquake, *Scientific Reports*, 2, 385.

Savage, J. C. (1983), A dislocation model of strain accumulation and release at a subduction zone, *Journal of Geophysical Research: Solid Earth*, 88(B6), 4984-4996.

Scholz, C. H. (1998), Earthquakes and friction laws, *Nature*, 391(6662), 37-42.

Schwartz, S. Y., and J. M. Rokosky (2007), Slow slip events and seismic tremor at circum-Pacific subduction zones, *Reviews of Geophysics*, 45(3), RG3004.

Scoffin, T. P., D. R. Stoddart, and B. R. Rosen (1978), *The Nature and Significance of Microatolls*, 99-122 pp.

- Shelly, D. R., G. C. Beroza, S. Ide, and S. Nakamura (2006), Low-frequency earthquakes in Shikoku, Japan, and their relationship to episodic tremor and slip, *Nature*, 442(7099), 188-191.
- Sieh, K. (2007), The Sunda Megathrust - Past, Present and Future, *Journal of Earthquake and Tsunami*, 01(01), 1-19.
- Sieh, K., D. H. Natawidjaja, A. J. Meltzner, C.-C. Shen, H. Cheng, K.-S. Li, B. W. Suwargadi, J. Galetzka, B. Philipposian, and R. L. Edwards (2008), Earthquake Supercycles Inferred from Sea-Level Changes Recorded in the Corals of West Sumatra, *Science*, 322(5908), 1674-1678.
- Simoes, M., J. P. Avouac, R. Cattin, and P. Henry (2004), The Sumatra subduction zone: A case for a locked fault zone extending into the mantle, *Journal of Geophysical Research: Solid Earth*, 109(B10), B10402.
- Simons, W. J. F., et al. (2007), A decade of GPS in Southeast Asia: Resolving Sundaland motion and boundaries, *Journal of Geophysical Research: Solid Earth*, 112(B6), n/a-n/a.
- Singh, S. C., N. Hananto, M. Mukti, H. Permana, Y. Djajadihardja, and H. Harjono (2011a), Seismic images of the megathrust rupture during the 25th October 2010 Pagai earthquake, SW Sumatra: Frontal rupture and large tsunami, *Geophysical Research Letters*, 38(16), L16313.
- Singh, S. C., H. Carton, A. S. Chauhan, S. Androvandi, A. Davaille, J. Dymant, M. Cannat, and N. D. Hananto (2011b), Extremely thin crust in the Indian Ocean possibly resulting from Plume—Ridge Interaction, *Geophysical Journal International*, 184(1), 29-42.
- Suito, H., and J. T. Freymueller (2009), A viscoelastic and afterslip postseismic deformation model for the 1964 Alaska earthquake, *Journal of Geophysical Research: Solid Earth*, 114(B11), B11404.
- Taylor, F. W., C. Frohlich, J. Lecolle, and M. Strecker (1987), Analysis of partially emerged corals and reef terraces in the central Vanuatu Arc: Comparison of contemporary coseismic and nonseismic with quaternary vertical movements, *Journal of Geophysical Research: Solid Earth*, 92(B6), 4905-4933.
- Thirumalai, K., F. W. Taylor, C.-C. Shen, L. L. Lavie, C. Frohlich, L. M. Wallace, C.-C. Wu, H. Sun, and A. K. Papabatu (2015), Variable Holocene deformation above a shallow subduction zone extremely close to the trench, *Nat Commun*, 6.
- Tichelaar, B. W., and L. J. Ruff (1993), Depth of seismic coupling along subduction zones, *Journal of Geophysical Research: Solid Earth*, 98(B2), 2017-2037.
- Tilmann, F. J., T. J. Craig, I. Grevemeyer, B. Suwargadi, H. Kopp, and E. Flueh (2010), The updip seismic/aseismic transition of the Sumatra megathrust illuminated by aftershocks of the 2004 Aceh-Andaman and 2005 Nias events, *Geophysical Journal International*, 181(3), 1261-1274.
- Tsang, L. L. H., A. J. Meltzner, B. Philipposian, E. M. Hill, J. T. Freymueller, and K. Sieh (2015), A 15-year slow slip event on the Sunda megathrust offshore Sumatra, *Geophysical Research Letters*, 42(16), 6630-6638.

- Villegas-Lanza, J.-M. Nocquet, F. Rolandone, M. Vallée, H. Tavera, F. Bondoux, T. Tran, X. Martin, and M. Chlieh (2016), A mixed seismic–aseismic stress release episode in the Andean subduction zone, *Nature Geoscience*, 9, 150-154.
- Vrolijk, P. (1990), On the mechanical role of smectite in subduction zones, *Geology*, 18(8), 703-707.
- Wesson, R. L., D. Melnick, M. Cisternas, M. Moreno, and L. L. Ely (2015), Vertical deformation through a complete seismic cycle at Isla Santa Maria, Chile, *Nature Geosci*, 8(7), 547-551.
- Woodworth, P. L., N. J. White, S. Jevrejeva, S. J. Holgate, J. A. Church, and W. R. Gehrels (2009), Evidence for the accelerations of sea level on multi-decade and century timescales, *International Journal of Climatology*, 29(6), 777-789.
- Zachariasen, J., K. Sieh, F. W. Taylor, and W. S. Hantoro (2000), Modern Vertical Deformation above the Sumatran Subduction Zone: Paleogeodetic Insights from Coral Microatolls, *Bulletin of the Seismological Society of America*, 90(4), 897-913.
- Zumberge, J. F., M. B. Heflin, D. C. Jefferson, M. M. Watkins, and F. H. Webb (1997), Precise point positioning for the efficient and robust analysis of GPS data from large networks, *Journal of Geophysical Research: Solid Earth*, 102(B3), 5005-5017.
- Zweck, C., J. T. Freymueller, and S. C. Cohen (2002), Three-dimensional elastic dislocation modeling of the postseismic response to the 1964 Alaska earthquake, *Journal of Geophysical Research: Solid Earth*, 107(B4), ECV 1-1-ECV 1-11.