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## Persistent termini of 2004- and 2005-like ruptures of the Sunda megathrust

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[1] To gain insight into the longevity of subduction zone segmentation, we use coral microatolls to examine an 1100-year record of large earthquakes across the boundary of the great 2004 and 2005 Sunda megathrust ruptures. Simeulue, a 100-km-long island off the west coast of northern Sumatra, Indonesia, straddles this boundary: northern Simeulue was uplifted in the 2004 earthquake, whereas southern Simeulue rose in 2005. Northern Simeulue corals reveal that predecessors of the 2004 earthquake occurred in the 10th century AD, in AD 1394  $\pm$  2, and in AD 1450  $\pm$  3. Corals from southern Simeulue indicate that none of the major uplifts inferred on northern Simeulue in the past 1100 years extended to southern Simeulue. The two largest uplifts recognized at a south-central Simeulue site—around AD 1422 and in 2005—involved little or no uplift of northern Simeulue. The distribution of uplift and strong shaking during a historical earthquake in 1861 suggests the 1861 rupture area was also restricted to south of central Simeulue, as in 2005. The strikingly different histories of the two adjacent patches demonstrate that this boundary has persisted as an impediment to rupture through at least seven earthquakes in the past 1100 years. This implies that the rupture lengths, and hence sizes, of at least some future great earthquakes and tsunamis can be forecast. These microatolls also provide insight into megathrust behavior between earthquakes, revealing sudden and substantial changes in interseismic strain accumulation rates.

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### 1. Introduction

[2] Tectonic earthquakes are caused by the rupture of faults—breaks in the Earth's brittle shell along which plates and blocks move relative to one another. Earth's greatest earthquakes and tsunamis are generated by its biggest faults. These megathrusts are the gently inclined upper surfaces of oceanic plates, where they dive under other plates and into the Earth's interior.

[3] It is curious that although many of these subduction megathrusts are thousands of kilometers long, individual

seismic and tsunamigenic ruptures seldom extend along them for more than a few hundred kilometers. This is fortunate, because the principal factor that determines the size of an earthquake is the area of the rupture patch. For example, the ruptures that caused the giant 2004  $M_W$  9.2 Aceh–Andaman and 2011  $M_W$  9.0 Tōhoku earthquakes propagated 1600 km and 400 km from start to finish, respectively [Meltzner *et al.*, 2006; Ozawa *et al.*, 2011], whereas for a  $M_W$   $\sim$ 7 earthquake, propagation ceases after merely a few tens of kilometers.

[4] Although this relationship between rupture length and earthquake and tsunami size has long been known, we can say little about whether a particular, known initiation or termination point has persisted through many earthquake cycles. Thus, we are able to say little about the size (and destructive potential) of future megathrust ruptures. Specifically, if at certain locations along a fault, impediments or barriers to rupture persist over many cycles of strain accumulation and release, one could anticipate the largest earthquake that a fault can produce. If, however, the origins and endpoints of ruptures do not persist through many earthquake cycles, then forecasting the maximum size of future earthquakes will be far more difficult.

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[5] Most seismic hazard calculations assume ruptures repeat in some fashion [e.g., *Headquarters for Earthquake Research Promotion*, 2006; *Field et al.*, 2008]. The beginning and end of model ruptures are commonly assumed to occur at structural or other irregularities or discontinuities along a fault. These assumptions are not without some basis in both observation and theory. For example, modern observations show that ruptures along strike-slip faults are commonly restricted by geometrical irregularities. Ruptures terminate if the fault trace is broken by a step over of 4 km or more [*Wesnousky*, 2006].

[6] Along subduction megathrusts, recognition of barriers to rupture has been severely limited by the fault inaccessibility and the paucity of long paleoseismic records that span multiple earthquake cycles. Still, some barriers to rupture in one sequence are known to have broken through in another, whereas others have arrested rupture repeatedly [*Natawidjaja et al.*, 2006; *Konca et al.*, 2008; *Sieh et al.*, 2008; *Lay*, 2011].

[7] In some cases, most notably in Sumatra in 2004 and in Japan in 2011, ruptures sweep across hundreds of kilometers of megathrust that had experienced only far smaller ruptures in modern times. Modern history and instrumental seismology suggested that these two deadly reaches were segmented into far smaller rupture patches by closely spaced rupture barriers. Had we recognized earlier that those modern boundaries were not permanent barriers, we might have anticipated the potential for the 2004 and 2011 earthquakes.

[8] One exceptional opportunity to gain insight into the longevity of barriers exists on Simeulue island, which straddles the boundary between the  $M_W$  9.2 and  $M_W$  8.6 Sunda megathrust ruptures of 26 December 2004 and 28 March 2005. Geodetic and coral uplift data constrain the extent of the two ruptures well, and corals provide an 1100-year earthquake history across this boundary (Figure 1).

[9] Megathrust slip in 2004 died off abruptly southeastward under Simeulue [*Subarya et al.*, 2006], from values of 10–15 m [*Chlieh et al.*, 2007], even though the adjacent section of the fault to the southeast had not slipped for a century and a half and had accumulated the potential for many meters of slip. Three months later, that southeastern section did fail, but the rupture was separated from the 2004 rupture by a narrow region of low cumulative slip [*Briggs et al.*, 2006]. This provoked an important question: has this section also been a barrier to rupture in the past?

[10] To determine a long rupture history, we extracted records of relative sea level change from coral microatolls on fringing reefs directly above the termini of the 2004 and 2005 ruptures, following the methods described by *Meltzner et al.* [2010]. Coral microatolls grow at the base of the intertidal zone, and their upper surfaces record a history of local sea level [*Scoffin and Stoddart*, 1978; *Taylor et al.*, 1987; *Zachariassen et al.*, 2000; *Meltzner et al.*, 2010]. Microatoll shapes form because subaerial exposure at times of extreme low water limits the highest level to which the coral colonies can grow [*Briggs et al.*, 2006; *Meltzner et al.*, 2010]. That level is termed the highest level of survival (HLS) [*Taylor et al.*, 1987]. Flat-topped pancake-like heads record sea level stability; colonies with HLS surfaces that rise radially outward toward their perimeter reflect rising

sea level during their decades of growth. As reefs subside or rise in the course of elastic strain accumulation and release, microatoll morphologies record changes in relative sea level. Because these corals' skeletons have annual growth bands, rates of change in elevation, when gradual, can be calculated precisely.

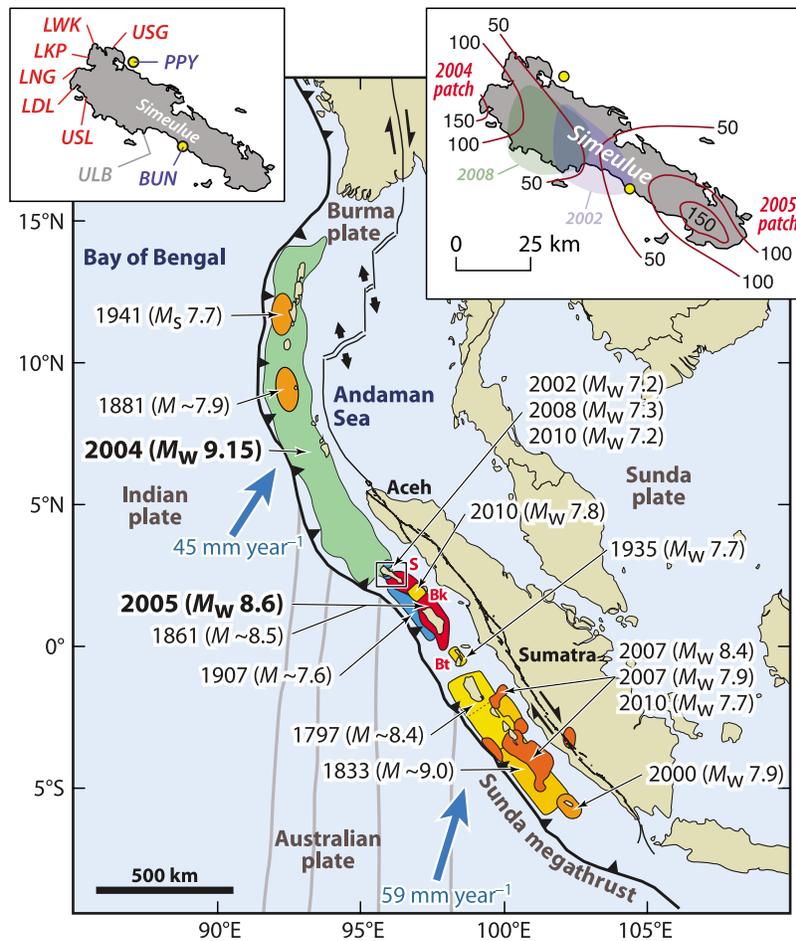
[11] The times of past uplift or subsidence events can be dated using U-Th techniques, which optimally enable determination of the age of a coral sample to within a few years [*Shen et al.*, 2002, 2008, 2010; *Frohlich et al.*, 2009]. In cases where individual samples yield insufficiently precise ages, dates from multiple samples on an individual slab can be combined, along with information about how many annual bands separate the dated samples, to provide a more precise weighted mean estimate for the dates of past uplift or subsidence events. Additionally, if we are confident that two microatolls at a site were contemporaneous, we can combine dates from the two slabs to obtain an even more precise weighted mean estimate for the dates of the events. We would be confident that two microatolls were contemporaneous if both had long records with matching diedown histories (with the slabs displaying similar elevation changes and identical intervals between the respective diedowns) and individual U-Th dates that were similar and precise enough that at least some parts of the slabs must overlap. Commonly, in cases with matching diedown histories, we can argue that the two slabs are compatible either with a complete overlap of the two records, or with no overlap, but not with a partial overlap; if the U-Th dates are close enough and precise enough, we can argue further that only a complete overlap is permitted by the data, and we can correlate individual diedowns and bands on the two slabs. Examples are discussed in detail by *Meltzner et al.* [2010].

[12] In this study, we present observations and analyses from two sites above the northwestern limit of the 2005 rupture. In an earlier paper [*Meltzner et al.*, 2010], we presented results from northern Simeulue sites above the southeastern end of the 2004 rupture. The stark contrast in the rupture histories of the 2004 and 2005 patches leads us to conclude that the 2004–2005 rupture boundary has been a persistent barrier to rupture over at least the past 1100 years.

## 2. Northern Simeulue Rupture History

[13] Using coral microatolls, *Meltzner et al.* [2010] uncovered a remarkably consistent and precise 14th–15th century earthquake history at several northwestern Simeulue sites (red locations in Figure 1 inset). At sites that rose more than 40 cm in 2004, earlier large uplifts occurred within 2 or 3 years of AD 1394 and 1450, and a smaller uplift occurred about AD 1430 (Figure 2).

[14] In AD 1394, corals from sites LDL, LNG, LKP, and possibly LWK rose [*Meltzner et al.*, 2010]. Uplift was half a meter both at LKP in  $AD\ 1394 \pm 2$  ( $2\sigma$  uncertainty, hereafter) and at LDL in  $AD\ 1393 \pm 3$  (Figures 3 and 4). Thirty-six years later, about AD 1430, microatolls at LKP recorded a 12-cm uplift. The spatial extent of this small uplift is not well resolved, but corals to the northeast at LWK show it did not occur there (Figure 3) [*Meltzner et al.*, 2010].



**Figure 1.** Regional map of the Sunda megathrust and recent large ruptures. S, Simeulue Island; Bk, Banyak Islands; Bt, Batu Islands. Rupture locations and magnitudes are from Briggs *et al.* [2006], Konca *et al.* [2008], Meltzner *et al.* [2010], and E. M. Hill *et al.* (The 2010  $M_W$  7.8 Mentawai earthquake: Very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS, submitted to *Journal of Geophysical Research*, 2012), and references therein; the 1907 location is speculative. Relative plate motions from Shearer and Bürgmann [2010]. Black lines are faults; gray lines are fracture zones. Left inset shows northern (red) and central (blue) Simeulue sites; auxiliary site ULB (gray) is also discussed in the text. Right inset shows 2004 and 2005 uplift contours in centimeters, updated from Briggs *et al.* [2006], along with the 2002 and 2008 uplift patches; these patches circumscribe areas where uplift was  $\sim 20$  cm or more. (See sections S1 and S2 of the auxiliary material and Meltzner *et al.* [2010] for a discussion of revised 2004 and 2005 uplift estimates and the determination of 2002 and 2008 uplifts.) Location of insets shown by box on the main map.

[15] Then, in  $AD\ 1450 \pm 3$ , all the corals on the reef flats above the 2004 rupture patch of northern Simeulue died suddenly. The most straightforward explanation for this complete mortality is that a large uplift raised the shallow reef flats sufficiently to expose all corals at extreme low water and kill them. Meltzner *et al.* [2010] argued that at LKP, the uplift must have been at least 1.2 m to accomplish this and could have been 2.5 m or more. In contrast, the site rose only 1.0 m in 2004. Following 1450, these reef flats remained devoid of living corals until the early 20th century.

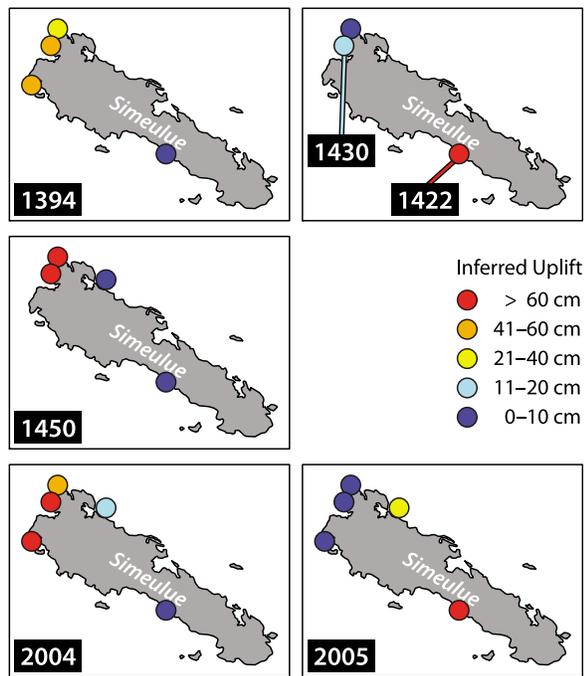
[16] Far less is known about earlier events. At USL, which experienced uplift of 1.3 m in 2004, a population of micro-atolls died in  $AD\ 956 \pm 16$  [Meltzner *et al.*, 2010]. If this death was related to uplift, then the uplift must have been at

least 38 cm, the height of the coral's outer living perimeter prior to uplift.

### 3. Southern–Central Simeulue Rupture History

[17] Here we present results from two sites farther south-east on Simeulue (blue labels in the Figure 1 inset; details of observations and analyses appear in sections S1 and S2, Figures S1–S18, and Tables S1–S4 in the auxiliary material).<sup>1</sup> These sites were affected more by the 2005 than the 2004 rupture, and their long-term uplift histories differ markedly from those just described to the north and west.

<sup>1</sup>Auxiliary materials are available in the HTML. doi:10.1029/2011JB008888.



**Figure 2.** Maps showing locations of uplift in 1394, 1422–1430, 1450, 2004, and 2005. For 2004 and 2005, only selected sites are shown to facilitate comparison with the earlier uplifts.

This contrast provides tight constraints on the boundaries of past ruptures.

### 3.1. The Bunon (BUN) Site

[18] The Bunon site sits on a broad promontory along the southwest coast of Simeulue,  $\sim 10$  km south of the center of the island, near Bunon village (BUN, Figure 1). The Bunon site consists of two subsites: primary site BUN-A and subsidiary site BUN-B. Site BUN-A rose  $\sim 80$  cm during the 2005 earthquake, while BUN-B,  $\sim 1.8$  km to the west-northwest, rose  $\sim 65$  cm; both subsites experienced little vertical change in 2004 (Figures 1 and 2, and section S1 of the auxiliary material). Thus, during at least the 2004–2005 sequence, Bunon acted in concert with the southern Simeulue patch and was independent of northern Simeulue. In addition, the Bunon sites rose  $\sim 20$  cm during an  $M_w$  7.2 earthquake on 2 November 2002 and up to 10 cm during an  $M_w$  7.3 earthquake on 20 February 2008.

[19] Both Bunon subsites have abundant modern heads (i.e., coral heads that were living at the time of the 2004 and 2005 earthquakes), although none of the modern heads had records of relative sea level extending back more than  $\sim 25$  years. In addition, the BUN-A site has multiple generations of large fossil microatolls (i.e., microatolls that died long before 2004, possibly in prior uplift events) from the 9th–11th and 14th–16th centuries AD. A total of three modern and seven fossil coral microatolls were sampled from the BUN sites; all but one modern head originated from site BUN-A (Table S1 of the auxiliary material).

[20] The sampled microatolls provide three discrete histories of relative sea level at BUN-A, spanning the mid-9th to early 11th centuries AD, the early 14th to late 16th centuries, and AD 1982 to 2005. The mid-9th to early 11th century

record is one of remarkably steady relative sea level, with no sudden uplift or subsidence apparent during the 170-year period preceding the BUN-9 microatoll's death. The death of BUN-9 around AD 1024 hints at a moderate or large uplift at that time, although the solitary microatoll conceivably could have died from another cause. The BUN-A record picks up again three centuries later around AD 1311 as the subjacent megathrust was slowly accumulating strain, with a submergence rate of 2.2 mm/yr (Figures 3 and 4). Submergence accelerated around 1340 and remained at 6.6 mm/yr until the site suddenly rose 66–77 cm around AD 1422. This large uplift may have been followed by  $\sim 10$  cm of postseismic subsidence. Bunon was then nearly stable from  $\sim 1433$  to  $\sim 1466$ . Submergence began again in the 1470s. The site continued to submerge at an average rate of 6.0 mm/yr from  $\sim 1481$  until at least 1576, although it may have been as fast as 10–11 mm/yr from  $\sim 1516$  to  $\sim 1538$ . Shortly after 1576, the remaining microatolls at the site died, probably due to coseismic uplift. The modern record reveals an interseismic tectonic subsidence rate of 5.3 mm/yr from 1986 to 1995, followed by the uplifts in 2002, 2005, and 2008.

[21] Because the BUN-A record spans the 14th–15th century, we can assess any changes there during the large northern Simeulue uplifts of that period. Microatolls show that Bunon subsided steadily from  $\sim 1350$  until  $\sim 1420$  and no diedown larger than  $\sim 5$  cm occurred in AD 1394 (Figure 3 and section S1 of the auxiliary material). Similarly, Bunon experienced no uplifts larger than  $\sim 10$  cm between  $\sim 1425$  and  $\sim 1465$ , although a diedown of  $\sim 10$  cm around 1450 may correspond to the 1450 uplift of northern Simeulue. Even allowing for all dating uncertainties, the corals at Bunon constrain maximum uplift there in 1394 and 1450 to  $\sim 5$  cm and  $\sim 10$  cm, respectively, thus showing that these events did not propagate southeastward beneath Bunon (Figure 2).

[22] The small, 12-cm  $\sim$ AD 1430 uplift at LKP may, however, be coeval with the 66–77 cm  $\sim$ AD 1422 uplift at BUN-A. If these two uplifts are contemporaneous, the much smaller amplitude of the uplift at LKP indicates the rupture died out northwestward. If the two uplifts are not contemporaneous, any uplift at LKP in  $\sim 1422$  was substantially less than 12 cm. This would be even stronger evidence that the 1422 event did not propagate beneath northern Simeulue.

[23] The much older flat-topped BUN-9 microatoll exhibits a continuous history of relative sea level stability from around AD 875 (or even 850) until AD 1020 (Figure 5). This record precludes extension of the northern Simeulue rupture of  $\sim$ AD 956 as far southeast as Bunon. Although the head's ultimate death some time after 1020 hints at a moderate or large uplift at that time, perhaps similar to the uplifts of  $\sim 1422$  and 2005, the head yielded no evidence for any earlier large uplift or subsidence events over the course of its 170-year history. Thus it appears that the 956 event—like the northern Simeulue uplifts in 1394, 1450, and 2004—did not extend beneath Bunon.

### 3.2. The Pulau Peny (PPY) Site

[24] We now turn to the other site that limits southeastern propagation of ruptures from northern Simeulue. Pulau Peny is a tiny islet 3 km off the northern northeast coast of Simeulue (PPY, Figure 1). Although Pulau Peny is closer to the northern Simeulue locus of uplift, which was associated with

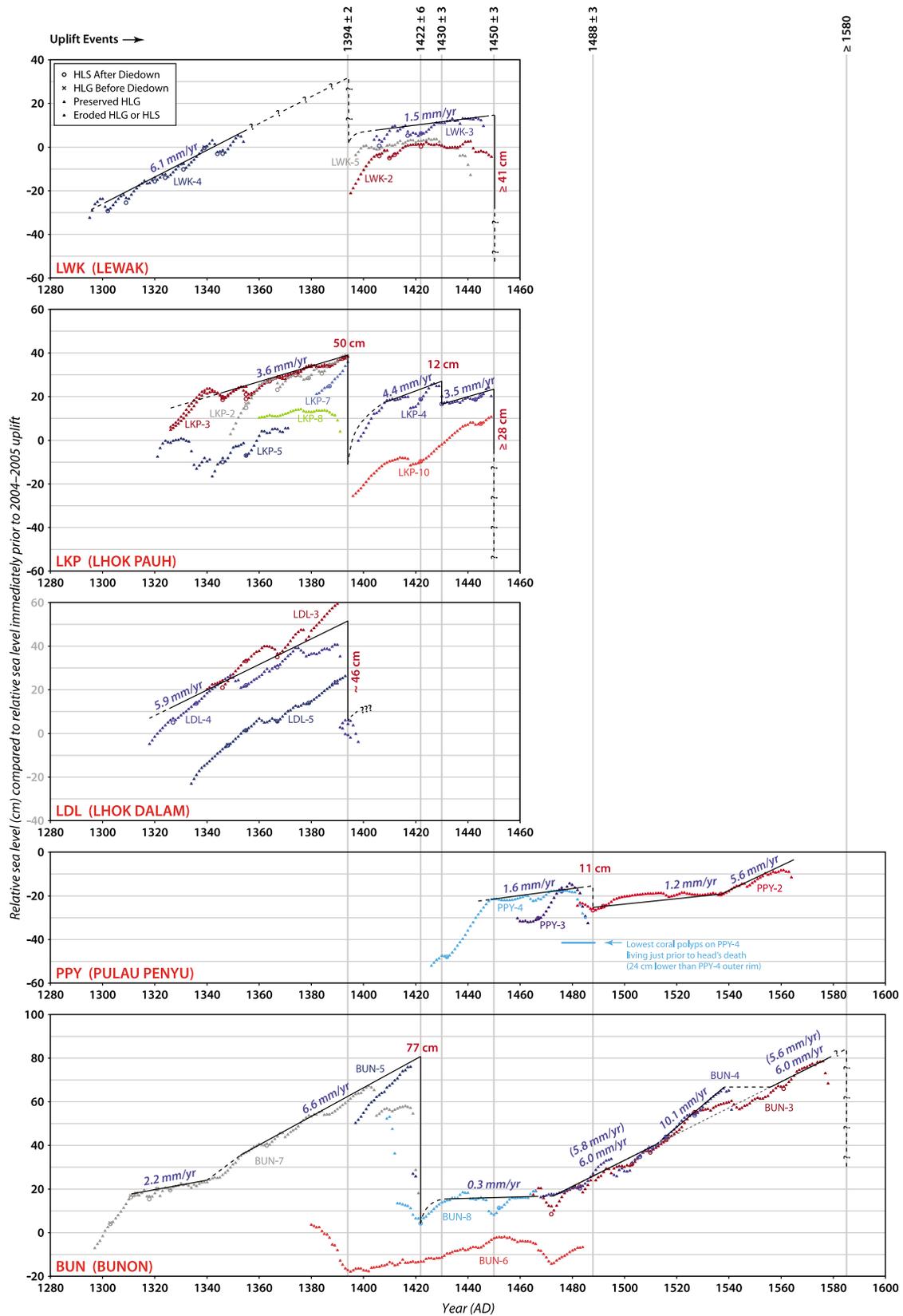


Figure 3

the 2004 earthquake, the majority (if not all) of the uplift at Pulau Penyu occurred in 2005. Uplift at Pulau Penyu totaled 36–38 cm in 2004–2005; no more than 14 cm of this occurred in 2004. Thus, like Bunon, Pulau Penyu mostly acted as part of the southern Simeulue patch for the 2004–2005 sequence and was largely independent of the 2004 patch. In addition, Pulau Penyu rose  $\sim 10$  cm and  $\sim 11$  cm, respectively, during the moderate earthquakes of 2002 and 2008.

[25] The Pulau Penyu site, PPY-A, occupies the northern and eastern sides of this islet. There are abundant modern heads and at least two generations of fossil microatolls that span the 15th–16th centuries AD. A total of one modern and three fossil coral microatolls were sampled from the PPY-A site (Table S1 of the auxiliary material).

[26] The sampled microatolls provide two discrete continuous histories of relative sea level at the PPY-A site, spanning the 15th–16th centuries and AD 1928 to 2005. The Pulau Penyu record begins around AD 1430 with the initial diedown of the oldest sampled microatoll (Figures 3 and 4). The amplitude of this diedown is unknown and may not have been large, but the occurrence of the diedown suggests the head was not too far below HLS thereafter. This initial diedown was followed by  $\sim 26$  cm of submergence, which may have occurred either suddenly (as during an earthquake) or over as much as a decade. Slow and gradual submergence from the 1450s ended in a sudden uplift of 10–24 cm around AD 1488. This uplift may have been followed by  $\sim 14$  cm of postseismic subsidence. The site then submerged gradually but not uniformly until at least the 1560s. In the 1560s or 1570s, the population of microatolls on the shallow reef flats of PPY-A died, possibly due to coseismic uplift. The modern record indicates the site submerged steadily at 5.9 mm/yr from 1932 to 1980, but the submergence slowed to 1.7 mm/yr from 1980 through at least 1995. The submergence rate deceleration might be partly due to a decrease in the rate of sea level rise in the Indian Ocean basin [Jevrejeva *et al.*, 2006]. This was followed by coseismic uplifts in 2002, 2005, 2008, and possibly 2004.

[27] The oldest microatoll sampled at PPY-A overlaps with a large inferred uplift on the 2004 patch of northern Simeulue in AD  $1450 \pm 3$  [Meltzner *et al.*, 2010], allowing us to compare the behavior of the 2004 patch with that of the PPY-A site at the time of that earlier rupture. What are the constraints on uplift at PPY-A during the 1450 northern Simeulue earthquake? The morphology of PPY-A microatolls alive at the time precludes a diedown exceeding 10–15 cm (Figure 3). An uplift in excess of 15 cm is possible only if one postulates an inordinately large amount of subsidence (considerably more than 26 cm) following the  $\sim 1430$  diedown (a detailed discussion appears in section S2 of the auxiliary material).

[28] The small maximum limits on uplift ( $\sim 15$  cm) at PPY-A in 2004 and in 1450 imply that rupture did not

propagate far eastward or southeastward past Pulau Penyu during either event. The largest uplift documented at PPY-A reached 22–38 cm, in 2005. Although Pulau Penyu is closer to the peak uplift of the island in 2004 than to that in 2005 (Figure 1), there is a much steeper gradient in uplift separating Pulau Penyu from the former than from the latter.

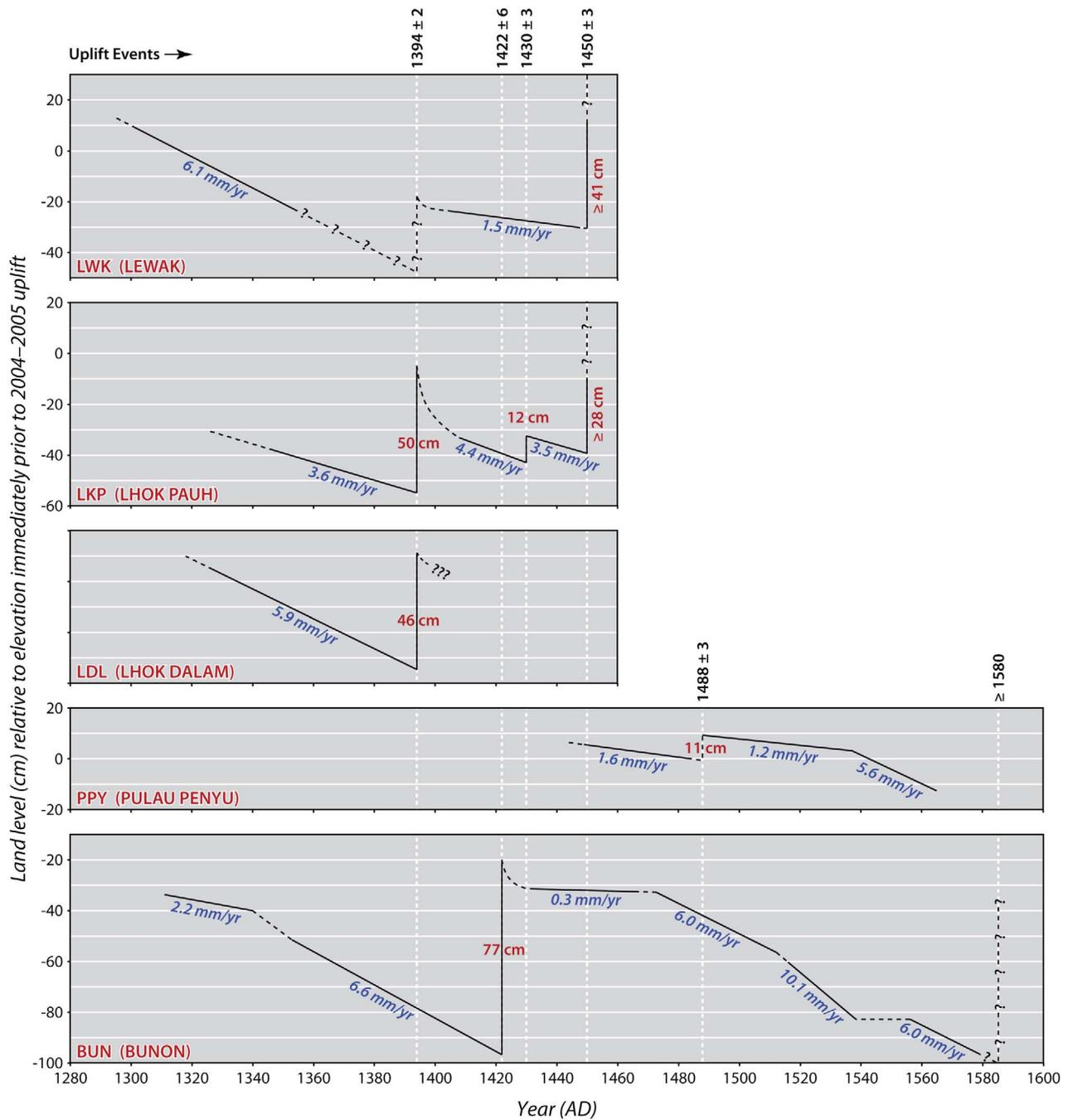
#### 4. Interseismic Subsidence

[29] In addition to their demonstration of the persistence of the central Simeulue rupture boundary, the coral microatolls suggest that interseismic subsidence rates at individual sites may have been markedly non-uniform. Until very recently, Earth scientists generally believed that interseismic motions are more or less constant, punctuated only by sudden earthquakes and postseismic deformation that follows the earthquakes and decays predictably with time. Over the past decade, however, new technologies and monitoring investments have resulted in the discovery and exploration of diurnal to monthly slow slip events in Japan, Cascadia, and elsewhere [Beroza and Ide, 2009; Gombert *et al.*, 2010]. These recent discoveries underscore our incomplete understanding of plate-boundary processes. Still, many recent studies assume that geodetic deformation rates determined over one or several decades approximate or represent deformation rates throughout the interseismic cycle [e.g., Bock *et al.*, 2003; Subarya *et al.*, 2006; Chlieh *et al.*, 2008; Sieh *et al.*, 2008].

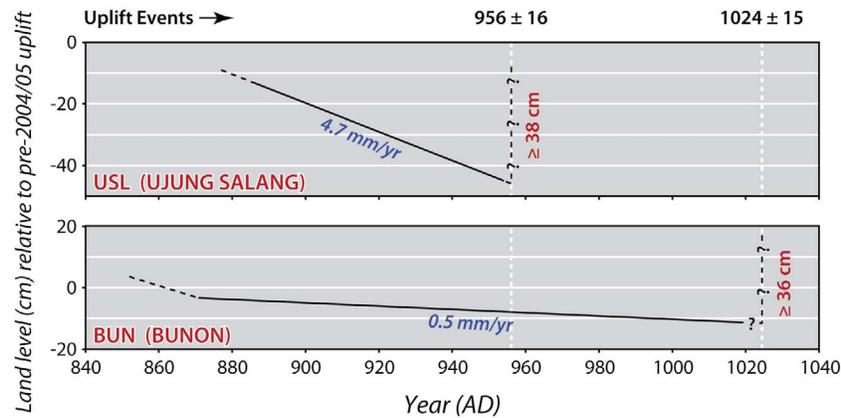
[30] Our data show that submergence rates are not stationary over decades to centuries. The submergence rates at Bunon, Pulau Penyu, and some northern Simeulue sites have varied considerably through individual seismic cycles and from one seismic cycle to the next. BUN-A, for example, submerged at a remarkably steady rate of 0.5 mm/yr between about AD 875 and 1020 or later (Figure 5 and Figure S12 of the auxiliary material). By contrast, the site submerged at 2.2 mm/yr from 1311 to 1340, at 6.6 mm/yr from 1353 to 1422, at merely 0.3 mm/yr from 1433 to 1466, and at an average rate of 6.0 mm/yr between 1481 and 1576, with an exceptionally fast interlude (at 10–11 mm/yr) between 1516 and 1538 (Figures 3 and 4). Details are in section S1 of the auxiliary material.

[31] Pulau Penyu also experienced substantial changes in submergence rates. From shortly prior to 1500 until  $\sim 1537$ , PPY-A submerged at 1.2 mm/yr or less, but then suddenly began submerging at 5.6 mm/yr until at least 1560 (Figures 3 and 4). The modern microatoll at PPY-A records submergence of 5.9 mm/yr from 1932 to 1980, followed by slower submergence of 1.7 mm/yr from 1980 through at least 1995. Even considering the temporally variable rates of sea level rise in the Indian Ocean over the 20th century estimated by Jevrejeva *et al.* [2006], a change in the subsidence rates is apparent: the submergence rates determined from the PPY-A microatoll imply PPY-A subsided tectonically at

**Figure 3.** The 14th–16th century relative sea level history at northern Simeulue sites LWK, LKP, and LDL [Meltzner *et al.*, 2010], and southern Simeulue sites PPY and BUN. This figure reflects Scenario 1 at BUN (Figure S11a of the auxiliary material) and Scenario A at PPY (Figure S18a of the auxiliary material); for a discussion of slightly different scenarios permitted by the data, see sections S1 and S2 of the auxiliary material. Data constrain solid parts of the curves well; dashed portions are inferred, and queried portions are conjectural. Dotted line at BUN indicates the AD 1481–1576 average. Diedowns (in centimeters) are red. Submergence rates (in millimeters per year, defined in a relative sense as the rate at which the coral descends below the surface of the water) are blue. Vertical gray lines mark dates of uplifts. The zero elevation datum at each site is the HLG just prior to the 2004–2005 uplift; 14th-century elevations at LDL are not known relative to 2004 elevations because none of the 14th-century heads there were in place.



**Figure 4.** Histories of interseismic subsidence and coseismic uplift through the 14th–16th centuries at the LWK, LKP, LDL, PPY, and BUN sites. The rates and elevations shown have been inverted from the corresponding relative sea level histories (Figure 3), and the time series has been shifted vertically to account for eustatic sea level rise since the 20th century, following the methodology of *Meltzner et al.* [2010]. Data constrain solid parts of the curves well; dashed portions are inferred, and queried portions are conjectural. Uplift amounts (in centimeters) are red. Interseismic subsidence rates (in millimeters per year, defined in an absolute sense as the geodetic rate at which the land moves downward) are blue. Vertical dotted white lines mark dates of uplifts. The zero elevation datum at each site is the site’s elevation immediately prior to the 2004–2005 uplift; 14th-century elevations at LDL are not known relative to 2004 elevations because none of the 14th-century heads at the site were in place.



**Figure 5.** Histories of interseismic subsidence and coseismic uplift through the 9th–11th centuries at the USL and BUN sites. See Figure 4 for explanation.

roughly 3–4 mm/yr from 1932 to 1980, but at little more than 1 mm/yr from 1980 to 1995. Details are in section S2 of the auxiliary material.

[32] Although the submergence rate variations we observe in the coral records are sometimes dramatic, it is important to be mindful of the fact that many factors besides tectonic subsidence influence interseismic submergence rates, and one must assess the tectonic significance of any submergence rate variations recorded by the microatolls. We divide these non-tectonic influences into two categories: (1) those that affect individual microatolls and are not coherent from one coral head to the next, and (2) those that operate over regional to global scales and have broad coherence. We will consider each of these categories separately. Only in cases where we can preclude influences of both types can we consider submergence rate changes to be tectonically significant.

[33] Factors that can influence the relative sea level history recorded by an individual microatoll include the coral's growth rate, which can vary from one head to another at a given site, or even from one part of a head to another part of the same head; and erosion, which might leave one part of a head better preserved than another part of the same head, or better preserved than a contemporaneous part of a nearby head. In order to better appreciate the extent to which such factors can affect the apparent relative sea level history recorded by a coral microatoll, we explore two cases in which we have slabs from microatolls that were living contemporaneously at the same site.

[34] The first case involves a comparison of coeval heads BUN-3 and BUN-4, which grew 200 m apart at the BUN-A site (Figures S1, S9, S10, and S11, and section S1 of the auxiliary material), and explores the potential consequences of under-appreciated erosion. Both of these microatolls recorded the site's relative sea level history from ~1470 to ~1540. Analysis of either head suggests a relative sea level rise (i.e., submergence) of about 5.8 or 5.9 mm/yr through ~1516, followed by a rise of ~10 mm/yr or more beginning in ~1516; details differ, however, on the two heads beyond 1516. At first glance BUN-3 suggests relative sea level rose by 11.7 mm/yr from 1516 to at least 1527, was below the long-term average until ~1545, and then returned to 5.6 mm/yr from 1545 until at least 1576 (Figure S9b of the auxiliary material). BUN-4, on the other hand, suggests an

average submergence rate of 10.1 mm/yr from 1516 until at least 1538 (Figure S10b of the auxiliary material). Closer inspection of the two microatoll slabs reveals that the apparent discrepancies arise from erosion of the upper part of the 1527 through 1545 bands on BUN-3, but the two slabs provide otherwise consistent relative sea level histories. In particular, the biased low rate estimated from BUN-3 for 1527 through 1538 (or perhaps through 1545) is based only on two (or three) diedowns (Figure S9 of the auxiliary material). This observation leads us to envision a method to test the significance of any submergence rates (or rate changes) determined from any individual microatoll.

[35] We propose that only submergence rates based upon four or more HLG points (e.g., orange squares in Figures S9b and S10b of the auxiliary material) spanning three or more diedowns (e.g., blue or pink dotted lines in Figures S9a and S10a of the auxiliary material) can be considered significant. For the BUN-3 and BUN-4 heads, this proposition would imply that the 5.8–5.9 mm/yr rate from ~1481 until 1516 (as determined on either head) is significant, as are the 10.1 mm/yr rate from 1516 to 1538 (determined from BUN-4) and the 5.6 mm/yr rate from 1545 to 1576 (determined from BUN-3); the average rate of relative sea level rise of 6.0 mm/yr from 1481 to 1576, determined from BUN-3, would likewise be significant. In contrast, the shorter-term apparent rates we originally inferred from BUN-3 for the respective periods 1516–1527 and 1527–1545 would not be considered significant, because of the brevity of those intervals and the potential for bias that could result from preservation peculiarities of a particular microatoll.

[36] The second case in which we can compare contemporaneous microatolls involves two modern microatolls sampled from the Ujung Lambajo (ULB-A) site on the southwest coast of Simeulue (ULB, Figure 1), 18 km west-northwest of Bunon. With few exceptions, modern microatolls at a particular site all tend to have consistent morphologies: although some might be taller, some might have started growing earlier, and some might have sustained more erosion, coeval microatolls at a site tend to show the same diedowns once they first reach HLS. At ULB-A, however, we observed two modern microatolls with seemingly inconsistent diedown histories, and we slabbed both because they were so distinct. These microatolls, which grew 26 m apart, both died because of

uplift in the 2004 earthquake, and both recorded relative sea level history at the site from 1982 or earlier through 2004. For whatever reason, ULB-1 (Figure S19 of the auxiliary material) grew upward faster than ULB-2 (Figure S20 of the auxiliary material), and as a result of the faster growth rate, ULB-1 experienced minor diedowns in late 1994 and late 1995 not seen on ULB-2, and ULB-1 experienced a substantially larger diedown than ULB-2 in late 1997.

[37] Indeed, out of 26 sites near Simeulue or Nias where we slabbed microatolls, and among many additional sites we have surveyed but not sampled, the pair of slabbed corals at ULB-A stands out as an extreme example of morphologically dissimilar coeval modern microatolls at a single site. Hence, we argue that it is appropriate and justifiable to use the ULB-A pair to illustrate a worst-case scenario for irregularities that might bias a single coral head's relative sea level history. Analyses of the microatolls individually yield submergence rates of 10.9 mm/yr over 1982–1997 from ULB-1 (Figure S19b of the auxiliary material) and 6.0 mm/yr over 1971–1997 from ULB-2 (Figure S20b of the auxiliary material), whereas analyzing the data from both heads jointly suggests submergence at 9.4 mm/yr over 1971–1997 (Figure S21 of the auxiliary material). Although, by the methodology proposed earlier, all three submergence rates are potentially significant, the lack of congruence of the three estimates raises the issue of uncertainties or errors in the estimated submergence rates.

[38] We can use the BUN-3 and BUN-4 pair, and the ULB-1 and ULB-2 pair, to estimate reasonable (albeit conservative) errors for the various submergence rates. Comparing the relative sea level time series recorded by BUN-3 and BUN-4, we note that the maximum differential erosion between the heads is 7 cm (Figure 3). Similarly, the maximum difference in any year's HLG on ULB-1 and ULB-2 is 8 cm (Figure S21 of the auxiliary material). As argued earlier, we will consider the worst-case scenario to be an 8-cm error in the apparent elevation gain recorded by a microatoll slab due to differential erosion of one part of the head compared to another part, or due to deficient upward growth. A simple calculation reveals that, for a rate averaged over 16 years, the associated error is 5 mm/yr, whereas for a rate averaged over 40 years, the error drops to 2 mm/yr. Assigning errors in this manner, we obtain submergence rates of  $10.9 \pm 5.3$  mm/yr over 1982–1997 from ULB-1,  $6.0 \pm 3.1$  mm/yr over 1971–1997 from ULB-2, and  $9.4 \pm 3.1$  mm/yr over 1971–1997 from the joint analysis. With these errors, the rates are indistinguishable, and the differences in the rates are not significant. Although microatolls with substantially more erosion can easily be found on any reef, there comes a limit beyond which it is difficult to recognize diedowns or concentric rings, and we generally do not employ such heads to estimate interseismic rates.

[39] Returning to the time series at BUN-A, we consider the following submergence rates to be significant:  $2.2 \pm 2.8$  mm/yr over 1311–1340 (from BUN-7);  $6.6 \pm 1.2$  mm/yr over 1353–1422 (from BUN-7);  $0.3 \pm 2.4$  mm/yr over 1433–1466 (from BUN-8);  $5.8 \pm 2.3$  mm/yr over 1481–1516 (from BUN-3);  $10.1 \pm 3.6$  mm/yr over 1516–1538 (from BUN-4);  $5.6 \pm 2.6$  mm/yr over 1545–1576 (from BUN-3). Accordingly, the difference in rates before and after the 1422 earthquake is significant, as are the difference in the 1311–1340 and 1353–1422 rates and the difference in the 1433–

1466 and 1481–1516 rates. The apparent rate changes in 1516 and around 1540, while perhaps real, cannot be considered significant based on available data.

[40] Finally, we address the issue of non-tectonic influences that operate over regional to global scales and have broad coherence. Even if a rate change recorded by a microatoll is significant, it is not necessarily tectonically significant (in other words, it does not necessarily reflect a change in underlying tectonic parameters or processes). In particular, changes in relative sea level at a given site depend upon changes in the eustatic sea level or regional changes in sea level that arise from nonuniform ocean warming, salinity variations, gravitational effects, and changes in ocean circulation; upon isostatic adjustments of the land due to loading or unloading of the lithosphere by glaciers or glacial meltwater; and upon tectonic adjustments of the land over the earthquake cycle. We wish to isolate the third contribution to relative sea level changes at any site, but we cannot blindly ignore the other contributions. For modern microatolls—particularly for time series from recent decades—we can utilize independent records of sea level change. For fossil microatolls, however, no reliable sea level data exist for this region [Meltzner *et al.*, 2010], which presents a challenge. For depiction purposes in Figures 4 and 5, we assume that pre-20th century tectonic subsidence rates equal the submergence rates recorded by microatolls, but this assumption has not been validated by independent data.

[41] Fortunately, we can use coeval paleogeodetic data from different sites on Simeulue to our advantage. Any changes in eustatic or regional sea level should have a nearly uniform effect at sites over distances of several hundred kilometers. Nicholls and Cazenave [2010] show decadal scale variations in sea surface height of up to  $\pm 20$  mm/yr, but the largest and most spatially variable deviations all occur in specific locations, associated with major currents like the Kuroshio Current, the Gulf Stream, and the Southern Ocean; since 1992, sea surface heights in the eastern tropical Indian Ocean have been rising more uniformly and closer to the global average. Similarly, isostatic adjustments in the tropics should be roughly uniform over distances of several hundred kilometers, such that any spatial differences in isostatic responses since the mid-Holocene would be negligible at sites within 200–300 km of one another [Mitrovica and Milne, 2002]. Because Simeulue is only  $\sim 100$  km long, any eustatic or isostatic changes should be seen synchronously at all sites on Simeulue. Any significant change seen on one part of Simeulue but precluded by coeval corals on another part of the island is considered to be tectonically significant.

[42] As it turns out, corals at Lewak on northern Simeulue record interseismic rate changes that are similar in amplitude to, but uncorrelated with, the variations at Bunon. Between 1308 and 1352, a period of slow submergence at Bunon, microatoll LWK-4 at Lewak recorded steady, rapid submergence of  $6.1 \pm 1.8$  mm/yr. A century later, coeval LWK-2 and LWK-3 suggest rates of  $1.4 \pm 2.4$  mm/yr over 1408–1441 and  $1.5 \pm 3.5$  mm/yr over 1411–1434, respectively. Each of these rates is significant, and the difference between the LWK-4 rate and the LWK-2 rate represents a significant change at the site. Furthermore, this 15th century record overlaps with both rapid and slow submergence at Bunon (Figures 3 and 4), indicating the changes at the two sites are asynchronous and

uncorrelated. We therefore infer that these rate changes reflect real changes in tectonic processes under the respective sites.

[43] The modern microatoll at Lewak, LWK-1, records submergence of  $5.3 \pm 1.8$  mm/yr on average from 1953 to 1997, which would correspond to tectonic subsidence of  $3.3 \pm 1.8$  mm/yr if an average eustatic sea level rise of 2 mm/yr is assumed [Meltzner *et al.*, 2010]. However, the submergence recorded by LWK-1 was slightly faster before the 1970s than after. As discussed for the observations at PPY-A, this difference could be largely explained by temporal variations in the rate of sea level rise in the Indian Ocean, which is estimated to have been 1–3 mm/yr faster before the 1970s than since [Jevrejeva *et al.*, 2006]. In this case, the similarity and synchronicity of the changes at Lewak and Pulau Penyu (and at Lhok Pau; see Meltzner *et al.* [2010]) suggest these changes may be largely or entirely due to regional changes in sea level; nonetheless, independent sea level records suggest that at least some of the apparent change at Pulau Penyu may be tectonic.

## 5. Discussion

[44] Coral microatolls above Simeulue island provide an 1100-year paleoseismic and paleogeodetic history across the boundary of the great 2004 and 2005 Sunda megathrust ruptures, and they support two important conclusions. First, the contrasting earthquake histories across this modern boundary are clear evidence for a persistent barrier to rupture. And second, strain accumulation during the interseismic period is not uniform over time, and not all the variations can be explained by postseismic processes.

### 5.1. A Barrier to Rupture

[45] The records from Bunon provide robust evidence that none of the major uplifts known or inferred on northern Simeulue in the past 1100 years (~AD 956, 1394, 1450, or 2004) involved significant uplift or subsidence at Bunon. Moreover, during the two northern Simeulue events for which there is a record at Pulau Penyu (1450 and 2004), the uplifts there were small. Likewise, the largest uplifts at Bunon—the 66–77 cm uplift around AD 1422 and the ~80 cm uplift in 2005—coincided with little or no vertical deformation on northern Simeulue. Historical intensities [Newcomb and McCann, 1987] and corals at other sites [Meltzner *et al.*, 2009] suggest the large megathrust earthquake of 1861 was very similar to that of 2005, in that it involved rupture below Nias island and southeastern Simeulue but not below northwestern Simeulue. Thus central Simeulue was above a barrier to rupture from the northwest in 956, 1394, 1450, and 2004 and from the southeast in 1422, 1861, and 2005. Although there are gaps in the paleoseismic record, there is no evidence for throughgoing rupture under central Simeulue at any time in the past 1100 years.

[46] Moderate ( $M_W$  7.2–7.3) megathrust ruptures in 2002 and 2008 illuminate a narrow patch of low cumulative slip between the 2004 and 2005 ruptures [DeShon *et al.*, 2005; Tilmann *et al.*, 2010], and an additional poorly located  $M_W$  7.2 earthquake on 9 May 2010 may also lie along this patch. If this narrow section is characterized over the long term by exclusively moderate and smaller ruptures, then it would support the hypothesis that central Simeulue overlies a narrow region of the megathrust with fundamentally different properties than elsewhere along strike.

[47] The Batu Islands section of the megathrust, between about  $0.5^\circ\text{S}$  and the Equator, is another persistent barrier to rupture (Figure 1) [Natawidjaja *et al.*, 2006]. Great ruptures from the southeast in 1797 and 1833 and from the northwest in 1861 and 2005 did not propagate through. At Badgugu (BDG), a site in the Batu Islands, one long-lived coral recorded small diedowns of ~20, ~10, and ~5 cm, respectively, during the great earthquakes of 1797, 1833, and 1861 [Natawidjaja *et al.*, 2006]. Continuous GPS observations nearby (at sites within 60 km, also in the Batu Islands) suggest uplift in 2005 was at most a few centimeters there [Briggs *et al.*, 2006]. The only large uplift in the 260-year record at Badgugu was ~70 cm, during a moderate ( $M_W$  7.7) earthquake in 1935, wholly within the Batu Islands patch [Natawidjaja *et al.*, 2004, 2006]. Although the Batu Islands paleoseismic record is shorter than the Simeulue record, it demonstrates that, like the central Simeulue patch, the Batu Islands patch has been a persistent barrier to throughgoing rupture during several large earthquakes.

[48] What causes the central Simeulue and Batu Islands patches to persistently arrest rupture? Understanding why some barriers to megathrust rupture appear to persist whereas others are more ephemeral has been an elusive goal. On strike-slip faults, both structural and rheological irregularities have the potential to be persistent barriers to rupture: step overs of 4 km or more have been shown to be reliable terminators of coseismic rupture [Wesnowsky, 2006], and the creeping section of the San Andreas fault in California, which slips more or less steadily at nearly the long-term rate and accumulates little or no strain [Rolandone *et al.*, 2008], has been assumed to be incapable of sustaining large throughgoing ruptures. Along subduction zones, numerous relationships between structures and rupture segmentation have been proposed [e.g., Kodaira *et al.*, 2000; Cummins *et al.*, 2002; Bilek, 2010; Loveless *et al.*, 2010; Wang and Bilek, 2011], but identifying a consistent relationship has been challenging [Loveless *et al.*, 2010]; perhaps this is partly because it has been difficult to distinguish truly persistent barriers from more ephemeral ones, where records are short.

[49] Under Simeulue and the Batu Islands, structural or geometrical impediments to rupture are strong possibilities. Franke *et al.* [2008] used multichannel reflection and wide-angle/refraction seismic data to identify a rise in the oceanic basement that is elongated north-northeast to south-southwest off the southwest coast of Simeulue. This apparent ridge is masked by sedimentary cover in the trench, and neither spreading ridges nor fracture zones are evident on high-resolution bathymetry along strike immediately south of the trench; however, farther south, where the sedimentary cover thins, several fracture zones are imaged on gravity, magnetic [Liu *et al.*, 1983; Cande *et al.*, 1989], and satellite altimetry [Smith and Sandwell, 1997] data. The width (30–50 km) and relief (~500–2000 m) of those fracture zones are of the same order of magnitude as the rise on the subducting plate observed in the seismic data just south of Simeulue, and one of those fracture zones projects almost exactly into the imaged rise [Franke *et al.*, 2008]. Continuing farther to the north-northeast, this fracture zone, now deeply buried beneath trench sediment, would closely align with the persistent rupture boundary we have documented on Simeulue.

[50] Franke *et al.* [2008] further suggest that a ramp or tear along the eastern flank of the subducting fracture zone

beneath Simeulue, which appears to have 3 km of relief (down to the east), might reinforce the fracture zone itself as a barrier to rupture propagation. Some of this relief across the fracture zone could be a function of the juxtaposition of crust of significantly different ages. The general age of the oceanic crust, however, is Eocene, and, assuming symmetrical spreading, there is an age difference of  $\sim 2$  Ma [Cande *et al.*, 1989]. The resulting seafloor depth difference should, therefore, be only of the order of 100–200 m [Turcotte and Schubert, 2002; Franke *et al.*, 2008], a difference that cannot account for the overall relief observed across the fracture zone. Alternatively, Franke *et al.* [2008] suggest that a fault or tear at the eastern flank of the proposed fracture zone could explain the observed depth difference of 3 km. The N–S to NNE–SSW striking fracture zones on the oceanic plate southwest of Sumatra have been reactivated as left-lateral strike-slip faults [Deplus *et al.*, 1998], one of which ruptured in an  $M_W$  7.2 earthquake on 10 January 2012. Near the Sunda trench, these reactivated fracture zones appear to pick up a significant normal component of slip, due to flexural bending of the oceanic plate as it descends into the subduction zone [Graindorge *et al.*, 2008]. Franke *et al.* [2008] proposed that a tear in the slab could have formed along the fracture zone south of Simeulue in this manner.

[51] Focusing on the upper plate, Simeulue also displays structural discontinuities above the barrier. A broad anticline in the northwestern part of the island does not connect with anticlines and synclines in the southeast [Endharto and Sukido, 1994]. Furthermore, the above-sea level northwestern part of the island is twice as wide as its southeastern counterpart, and although this exact relationship would not hold with lower sea levels, there would continue to be a pronounced trenchward step (from southeast to northwest) along the southwest coast of Simeulue, even if sea level were 50 m lower. Farther northeast along strike of the rupture barrier, however, multichannel seismic reflection profiles across the Simeulue forearc basin reveal no rooted faults cutting Miocene and younger sediments, except for the trench-parallel West Andaman fault [Berglar *et al.*, 2008]; this suggests that any discontinuities in the upper plate are not pervasive. Any upper-plate discontinuities may simply be a reflection of the complexity already described along the subjacent megathrust.

[52] The Batu Islands are even more clearly associated with structural complexity. The Investigator fracture zone, which comprises four individual parallel ridges of up to 1900 m relief and  $\sim 120$  km total width, and which is visible in the seafloor bathymetry, projects beneath the Batu Islands (Figure 1) and correlates with irregularities in the trend of the deformation front and in the geomorphology of the upper plate [Kopp *et al.*, 2008]. This fracture zone appears to influence seismicity over a wide depth interval. A well-defined band of intense seismicity extending from 80 to 200 km depth lies along the prolongation of the Investigator fracture zone, all the way from the Batu Islands to Toba caldera [Fauzi *et al.*, 1996; Lange *et al.*, 2010]. The age contrast across the Investigator fracture zone is  $\sim 15$  Ma [Cande *et al.*, 1989; Kopp *et al.*, 2008], which should lead to a difference of several hundred meters or more in the depth of the subducting slab across the fracture zone. Sieh and Natawidjaja [2000] note that the orientations of faults in

the upper plate just west of the northward projection of the Investigator fracture zone are predominantly north–south, parallel to the topographic and structural grain of the Investigator fracture zone. They hypothesize that the topographic heterogeneity of the Investigator fracture zone has led to disruption of the forearc and outer-arc regions.

[53] There are cases, however, in which a clear structural break did not halt megathrust rupture. Notably, the 1861 and 2005 earthquakes ruptured through structural complexities in the Banyak Islands [Briggs *et al.*, 2006; Meltzner *et al.*, 2009]. In the Solomon Islands in 2007, megathrust rupture traversed the subducting Simbo ridge transform and thus broke through a triple junction [Taylor *et al.*, 2008].

[54] As an alternative (or perhaps complement) to direct structural or geometrical controls, along-strike variations in rheology, interseismic coupling, or fault friction might be a cause of these two barriers [Kaneko *et al.*, 2010]. Near the boundary region between the 2004 and 2005 ruptures, Tilmann *et al.* [2010] examined precisely located aftershocks and demonstrated that the vast majority of aftershocks in the study region occurred on the plate interface within a narrow band that they inferred marks the transition between the seismogenic zone and stable sliding. Although this tight band of aftershocks is roughly parallel to both the trench and the axis of Simeulue, and though it tends to underlie the 500 m bathymetric contour for most of its length, there is an abrupt and marked  $\sim 25$  km landward shift of the updip edge of this band of aftershocks in the vicinity of central Simeulue, immediately trenchward of the Bunon site.

[55] Although this gap in the aftershocks could simply be a reflection of the coseismic slip deficit between the 2004 and 2005 ruptures, it could also indicate a deeper onset of unstable frictional conditions and a reduced width of the seismogenic zone [Tilmann *et al.*, 2010]. The location of the aftershock gap corresponds fairly well with the eastern edge of the ridge imaged by Franke *et al.* [2008], which implies it could also overlie the tear in the slab proposed by Franke *et al.* [2008], although Tilmann *et al.* [2010] were unable to resolve any significant shift in the depths of the aftershocks from one end of Simeulue to the other. Tilmann *et al.* [2010] suggest that the fracture zone could be a locus of enhanced fluid release into the megathrust, raising fluid pressure there and reducing the effective normal stress, which could in turn deepen the onset of seismogenic behavior [Moore and Saffer, 2001; Scholz, 2002]. This mechanism might hold even if the fracture zone itself is buried under sediments. Whatever the exact reason, a localized  $\sim 25$  km reduction in the width of the seismogenic zone, Tilmann *et al.* [2010] point out, might be sufficient to act as a barrier to rupture propagation in giant earthquakes along the megathrust.

[56] Modeling of coral and GPS geodetic data over the period 1962–2006 from around the Equator suggests the Batu Islands barrier has also been a poorly coupled segment of the megathrust [Chlieh *et al.*, 2008]. If that low coupling is persistent over centuries, perhaps it is due to locally enhanced fluid release into the megathrust, as suggested by Tilmann *et al.* [2010], or it might be due to weak rocks such as serpentine associated with the fracture zone itself. However, whether a 45-year geodetic record is sufficient to determine the long-term coupling of a portion of a fault

remains an open question, particularly in light of dramatic variations in interseismic subsidence rates documented in this study at timescales of decades to centuries. Indeed, *Prawirodirdjo et al.* [2010] argue that the Batu Islands section of the megathrust was more strongly coupled before 2001 than from 2001 until March 2005.

[57] Since the Batu Islands and central Simeulue patches have been persistent impediments to several great earthquake ruptures, we conclude that the intervening 1861 and 2005 ruptures are the longest one should expect from this reach of the megathrust. Earthquakes larger than those in 1861 or 2005 ( $M_W > 8.6$ ) could occur only if the width of the rupture zone was greater (e.g., if slip extended closer to the trench), or if the displacements were larger. Paleoseismological identification of other “permanent” barriers to rupture should also lead to identification of the maximum size earthquakes plausible for other sections of the world’s megathrusts. Without such evidence from the geological record, it would seem prudent to conclude that a  $M_W$  9 rupture can occur on any sufficiently long section of a megathrust.

## 5.2. Interseismic Rate Switching

[58] Certain microatolls found on Simeulue appear to reflect sudden changes in submergence rates. At any given site, the submergence rate might remain fixed for decades until switching, over a decade or so, to a new rate; submergence might remain fixed at this new rate for decades more. There are ample examples at Bunon and Lewak where significant rate changes demonstrably occur (see section 4), and even in some cases where we have not shown that a rate change is significant, the observations are nonetheless best explained by sudden rate changes. These observed rate variations are not consistent with steadily decaying postseismic deformation [*Perfettini et al.*, 2005]: not only do the rates appear to be linear (except possibly in the ~5–15 years following each uplift) and the changes abrupt (over shorter time spans than we can resolve), but we have examples of both slow then faster submergence prior to an uplift (e.g., Bunon, 1311–1422; Bunon, 1433–1580; Figure 4) and fast then slower submergence prior to an uplift (e.g., Pulau Penyu, 1932–2004; Figure S14 of the auxiliary material), as well as uniformly slow submergence for decades to centuries prior to an uplift (e.g., Bunon, 875–1024; Lewak, 1408–1450; Figures 4 and 5).

[59] Furthermore, although submergence rates recorded by microatolls reflect the combined effect of tectonic land-level changes, isostatic adjustments, and eustasy [*Meltzner et al.*, 2010], the variations at Bunon can only be explained by tectonic processes. Since isostasy and eustasy operate over regional to global scales, isostatic or eustatic changes recorded at Bunon should be recorded at all sites on Simeulue. Quite to the contrary, the notable fluctuations at Bunon between 1311 and 1450 are not correlated with fluctuations in the northern Simeulue corals (Figure 4), leaving only tectonics to explain them.

[60] Observations such as these are not limited to Simeulue. Work in the Batu and Mentawai Islands has uncovered similar variations there [*Natawidjaja et al.*, 2004, 2006, 2007; *Sieh et al.*, 2008]. Along the Japan trench north of the 2011 rupture patch, primarily between Honshu and Hokkaido, *Nishimura et al.* [2004] documented apparent changes in coupling on annual time scales between 1995

and 2002 that cannot be explained entirely by postseismic deformation following the 1994 Sanriku earthquake. And abrupt changes in interseismic rates have also been observed with GPS data in Alaska [*Frey Mueller*, 2010]: in lower Cook Inlet, near Homer, Alaska, part of the subduction interface that was creeping prior to 2004 (since at least the mid-1990s) suddenly locked, and has remained locked. *Frey Mueller* [2010] suggested that the mid-1990s through 2004 involved a long slow slip event, or that the frictional behavior of the interface at the downdip end of the locked zone is very sensitive to small stress changes.

[61] We speculate that sudden interseismic “rate switching,” as observed in the Sumatran corals, could be a common phenomenon along subduction zones. If so, any one local geodetic network will likely need to be in place for many decades to observe it. Furthermore, this phenomenon might be detectable only at stations above the locked portion of the fault or near its downdip limit; in many cases, the locked portion of the fault is entirely offshore, and only the most trenchward points on land are near the downdip limit. These challenges could explain why rate switching has not (yet) been identified as a widespread occurrence.

[62] Geological observations along exhumed paleo-subduction zones can be interpreted to support our inferences of variations in coupling over the earthquake cycle. *Bachmann et al.* [2009] presented evidence from an exhumed subduction zone in the central Alps of Europe for fluids circulating along the plate interface and for transient changes in pore pressure; they argue that these changes may give rise to variations in coupling over the seismic cycle. At an exhumed subduction thrust on Kodiak Island, Alaska, *Rowe et al.* [2011] observed mutually crosscutting fault rocks that distinctly record three general rates of slip: seismic slip, recorded by pseudotachylite bearing rocks; solution creep on discrete foliated cataclasites, representing interseismic creep; and an intermediate strain rate texture in non-foliated cataclasites. This intermediate strain rate could represent anything from afterslip to slow slip events, to a delocalization of the shear surface during or immediately after seismic slip.

[63] The changes in interseismic behavior at various sites on the Sumatran outer arc islands might arise from small heterogeneities in the frictional properties along the fault, from decadal-scale variations in fluid flow conditions along the fault, or from long-duration slow-slip events. The periods of faster interseismic subsidence imply increased coupling under the site and/or intervals during which the locked zone extended farther downdip; conversely, slower subsidence suggests decreased coupling or a narrower locked zone. Alternatively, as *Frey Mueller* [2010] suggested, the abrupt changes in “interseismic” rates might actually reflect the initiation or termination of slow slip events. *Meade and Loveless* [2009] use scaling arguments to suggest that the slip velocity of large slow events may be near the plate convergence rate, making them indistinguishable from apparent partial coupling. Elastic dislocation models of geodetic measurements above subduction zones have led to the identification of  $M_W \approx 6.0$ – $7.2$  slow slip events that release elastic strain over periods of days to months, but great ( $M_W \geq 8$ ) slow slip events have remained unidentified.

*Meade and Loveless* [2009] showed that slip velocity in slow slip events decreases with event magnitude; they predict that, if  $M_W \geq 8$  slow slip events exist, the slip velocities should be  $\leq 50$  mm/yr and their durations should be  $>10$  years. *Meade and Loveless* [2009] suggest that great slow slip events may release a fraction of the accumulating strain on the plate interface, creating the appearance of a partially coupled subduction zone that is actually a snapshot in the time evolution of an ongoing  $M_W \geq 8$  slow slip event. The largest slow slip events could last decades to centuries. Our observations of substantial multidecadal variations in the interseismic subsidence rate at individual sites is consistent with their hypothesis. It is important to point out, however, that we have not attempted to replicate the observed rate changes with physical or numerical models, and doing so will be an important next step in understanding and explaining our observations.

### 5.3. A Note on the “Missing” Northern Simeulue Corals, AD 1450–1930

[64] Last, our observations at Bunon and Pulau Peny provide supporting context for a particular argument we advanced during the course of our earlier work on northern Simeulue. Despite extensively searching much of the coast of northern Simeulue, we were unable to locate a single microatoll that had been alive between AD 1450 and the early 20th century anywhere between the USL and USG sites (Figure 1) [Meltzner *et al.*, 2010]. Abundant older (and younger) microatolls at most of the sites imply that the problem is not one of preservation; instead, it must be that no corals grew on the reef flats of northwestern Simeulue between 1450 and  $\sim 1930$ . We also did not find fossil microatolls anywhere along the stretch of coast for 30 km southeast from USL, but in these locations the complete lack of fossil microatolls suggests preservation problems could be a factor.

[65] In our earlier work, we explored the question of what might have prevented corals from colonizing and living on the reef flats of northwestern Simeulue for nearly five centuries. Although we considered biological factors, we argued that the most plausible explanation for this absence is that those reef flats were sitting above the subtidal zone for most of that period, a result of tectonic uplift [Meltzner *et al.*, 2010]. Observations at Bunon and Pulau Peny support such a conclusion and add an intriguing detail.

[66] When the microatoll records from Bunon and Pulau Peny are considered along with the northwestern Simeulue records, it becomes apparent that the region of northwestern Simeulue characterized by the 480-year absence of corals encompasses but does not extend beyond what eventually became the 2004 rupture patch. We observe that each of the sites with a 1450–1930 gap in the coral record was uplifted  $\sim 25$  cm or more in 2004 but had little or no change in 2005. In contrast, the Bunon and Pulau Peny reefs, only 10–20 km away, which rose mostly or entirely in 2005, had abundant corals living through at least the 1560s, and microatolls grew at several other southern Simeulue sites between the mid-18th century and 1861. Thus, on Simeulue, there is a one-to-one correspondence between the area lacking corals from 1450 until  $\sim 1930$  and the area that was uplifted in 2004. The perfect coincidence between these two regions argues that the two are interrelated, that the 480-year absence of corals is evidence that the reef flats of northern Simeulue must have been elevated above the subtidal zone for much of that

period, and that the spatial extent of these elevated reefs was restricted to the area west-northwest of the 2004–2005 rupture barrier.

## 6. Conclusions

[67] In this paper, we show that the coral microatolls of Simeulue document a barrier to rupture that has persisted under central Simeulue island for at least the past 1100 years, through seven major ruptures. We also show that abrupt and enduring changes in rates of tectonic subsidence are common during this period. Both of these discoveries have important implications for forecasting large Sumatran earthquakes and for understanding subduction megathrust processes elsewhere on Earth.

[68] In our earlier paper [Meltzner *et al.*, 2010], we concluded that microatoll sites on northwestern Simeulue that rose in 2004 consistently record previous large uplifts around AD 1394 and 1450. Additionally, one northwestern Simeulue site records an earlier uplift about AD 956. In marked contrast to this history, we show in this paper that microatoll sites only tens of kilometers away on central and southeastern Simeulue coasts (which rose in 2005) experienced none of these northwestern Simeulue uplifts. Conversely (as in 2005) large uplifts of southeastern Simeulue around AD 1422 and in 1861 appear to have had little or no effect at the northwestern Simeulue sites. Hence, terminations of large ruptures under central Simeulue have occurred at least seven times in the past 1100 years. Given the long paleoseismic records we have obtained on either side of central Simeulue, the absence of any throughgoing rupture shows that the megathrust beneath central Simeulue is truly a persistent barrier to rupture. This barrier might owe its persistence to a north-trending fracture zone in the downgoing slab and a related step or tear in the megathrust itself.

[69] Simeulue island’s microatolls also record gradual submergence during long interseismic intervals, due primarily to tectonic subsidence, although changes in sea level could also contribute. Microatolls at several sites record abrupt and sustained changes in interseismic submergence rates. The fact that these changes are in some cases not synchronous from one site to another implies that at least some of these changes are due to a change in tectonic conditions. Changes in the degree or geometry of plate coupling, or the onset or termination of long-duration slow-slip events, are the best candidates for causing these changes.

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