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Submergence and uplift associated with the giant 1833 Sumatran subduction earthquake: Evidence from coral microatolls

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Abstract. The giant Sumatran subduction earthquake of 1833 appears as a large emergence event in fossil coral microatolls on the reefs of Sumatra's outer-arc ridge. Stratigraphic analysis of these and living microatolls nearby allow us to estimate that 1833 emergence increased trenchward from about 1 to 2 m. This pattern and magnitude of uplift are consistent with about 13 m of slip on the subduction interface and suggest a magnitude (M_w) of 8.8–9.2 for the earthquake. The fossil microatolls also record rapid submergence in the decades prior to the earthquake, with rates increasing trenchward from 5 to 11 mm/yr. Living microatolls show similar rates and a similar pattern. The fossil microatolls also record at least two less extensive emergence events in the decades prior to 1833. These observations show that coral microatolls can be useful paleoseismic and paleogeodetic instruments in convergent tectonic environments.

1. Introduction

In most subduction-zone settings, geological records of vertical coseismic and interseismic displacements are obscure. The subduction interface itself is seldom directly observable, so deformation must be inferred from records of relative sea level change over the hanging wall block. Even then, data are commonly available only along coastlines far above the subduction interface and far landward from the trench or deformation front.

Above the Cascadia subduction zone, for example, paleoseismic events have been recognized in estuarine sediments about 200 km from the deformation front. Marsh and forest flora buried by intertidal sediments record the occurrence of sudden, probably seismic, submergence events. In some locales, thin tsunami sand deposits atop the marsh and forest strata demonstrate the suddenness of these events [e.g., *Atwater*, 1987; *Atwater and Yamaguchi*, 1991]. Radiocarbon analyses of buried plants constrain the dates of the events to within a few decades. Along other coastlines, workers have interpreted suites of elevated coral reef and clastic marine terraces as evidence of successive sudden uplift events [e.g., *Plafker and Rubin*, 1967; *Ota*, 1985; *Ota et al.*, 1991, 1993; *Pandolfi et al.*, 1994; *Ota and Chappell*, 1996].

Paleoseismic data obtained from these studies have placed valuable constraints on the nature of earthquake sources and their recurrence. Nonetheless, more precise and extensive records will be required to resolve several pressing issues in earthquake science. Longer and better dated paleoseismic records of multiple events are needed, for example, to determine if coseismic ruptures are similar from event to event and to understand the temporal patterns of earthquake recurrence [*Sieh*, 1996].

Toward these goals, we are developing a method for documenting sea level changes that involves the use of coral microatolls, massive solitary coral heads that record sea level fluctuations in their morphology and can be precisely dated. Work of this kind began about two decades ago [*Scoffin and Stoddart*, 1978; *Stoddart and Scoffin*, 1979; *Taylor et al.*, 1987]. These earlier studies showed that upward growth of microatolls is limited by lowest water levels. As lowest water level fluctuates with time, so does the upper limit of coral skeletal growth. The skeleton develops a distinctive morphology that reflects the sea level changes to which it has been exposed. The microatoll skeleton records sea level changes as small as a centimeter or two [*Taylor et al.*, 1987; *Zachariasen*, 1998; J. Zachariasen et al., Modern vertical deformation at the Sumatran subduction zone: Paleogeodetic insights from coral microatolls, submitted to *Bulletin of the Seismological Society of America*, 1998; hereinafter referred to as J. Zachariasen et al., submitted paper, 1998].

Thermal ionization mass spectrometric (TIMS) U-series dating methods permit unusually precise dating of corals that are a few tens to several hundred thousand years old. *Edwards et al.* [1987a, b] and *Edwards* [1988] have shown that errors of only a few years can be obtained under favorable conditions for corals between zero and 1000 years old. Thus, in tropical regions where corals flourish, there is an opportunity to obtain paleoseismic dates an order of magnitude more precise than can be achieved with radiocarbon analysis. *Edwards et al.* [1987a] and *Taylor et al.* [1990] were the first to use TIMS on microatolls to obtain paleoseismic records of recent earthquakes. They used the U-Th system to date precisely subduction earthquakes in the 18th and 19th centuries in Vanuatu. In this paper, we use this method to date past large earthquakes in Sumatra.

The Sumatran subduction zone occurs in an ideal setting for obtaining an exceptional paleoseismic record from microatolls. Microatolls are ubiquitous in its waters. Reefs on the outer-arc ridge, 80–130 km from the trench, and reefs along the coast of mainland Sumatra, 200–240 km from the trench, allow sampling across a wide transect perpendicular to the trench. Fur-

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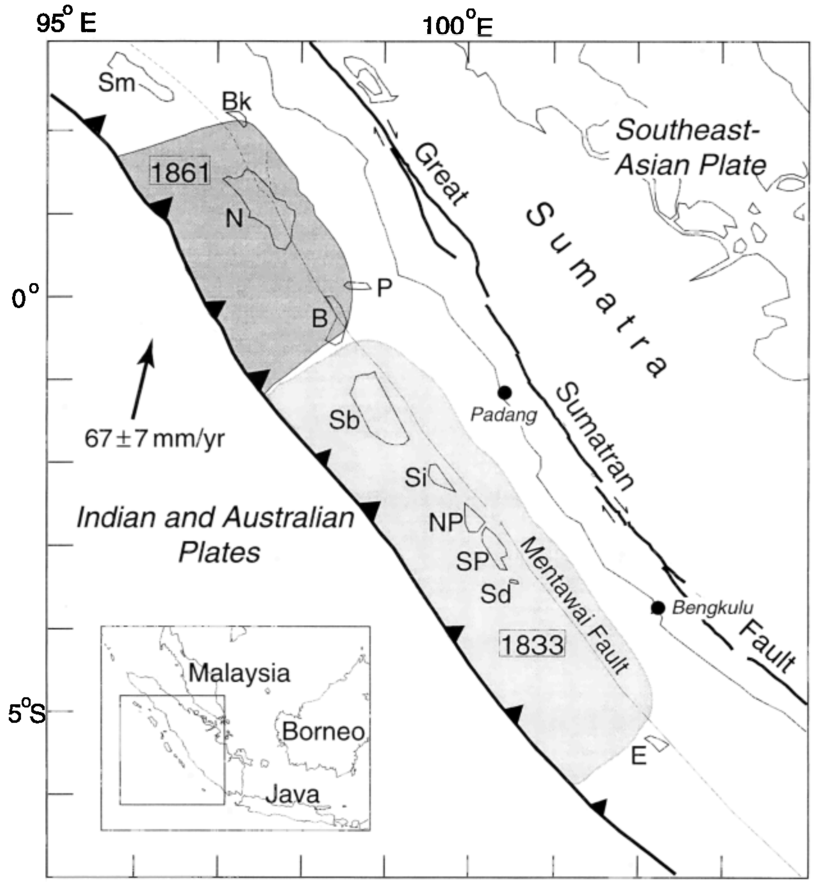


Figure 1. The Sumatran subduction zone and Great Sumatran fault separate the Indian and Australian plates from the Southeast Asian plate. Inferred rupture zones of the two most recent giant subduction zone earthquakes are shaded. Field work for this project was carried out on four islands of the Mentawai chain: Sipora (Si), South Pagai (SP), North Pagai (NP), and Sanding (Sd). Other islands and island groups in the chain include Simeuleue (Sm), Banyak (Bk), Pini (P), Nias (N), Batu (B), Siberut (Sb), and Enggano (E).

thermore, the Sumatran subduction zone has generated numerous large earthquakes that are likely to have produced observable displacements.

1.1. Geologic Setting of the Sumatran Subduction Zone

At the Sunda trench the Indian and Australian plates are subducting beneath the Southeast Asian plate (Figure 1). The relative convergence vector, measured between West Java and Christmas Island, is 67 ± 7 mm/yr, N11°E ± 4° [*Tregoning et al.*, 1994]. Sumatra and Java lie on the overriding plate, a few hundred kilometers from the trench. Convergence is nearly orthogonal to the trench axis near Java but is highly oblique near Sumatra, where strain is strongly partitioned between dip slip on the subduction zone interface and right-lateral slip on the Great Sumatran fault [*Fitch*, 1972; *McCaffrey*, 1991]. Earthquake focal mechanisms and hypocentral distributions indicate that the subducting plate dips less than 15° beneath the outer-arc ridge and steepens to about 50° below the volcanic arc [*Newcomb and McCann*, 1987; *Fauzi et al.*, 1996].

Both the subduction zone and the Great Sumatran fault are seismically active. In the 20th century, numerous magnitude (M_w) 7.5 and smaller earthquakes have occurred on both structures [*Katili and Hehuwat*, 1967; *Newcomb and McCann*, 1987; Harvard CMT catalog]. Overshadowing these are two giant earthquakes that occurred in 1833 and 1861. Using sparse Dutch accounts of shaking and tsunami extent, *Newcomb and McCann* [1987] argued convincingly that these earthquakes were produced by rupture of adjacent segments of the subduction zone (Figure 1). They estimated M_w = 8.3–8.5 for the 1861 event and 8.7–8.8 for the 1833 earthquake.

Newcomb and McCann [1987] conclude that the source of the 1833 earthquake was a 550-km-long segment of the subduction interface that extended from the Batu Islands to Enggano Island and that the rupture extended from the trench to a point below the outer-arc islands. From their estimated rupture dimensions, they suggest that slip of about 4–8 m occurred on the interface. Slip of this magnitude on a gently dipping fault should produce uplift of the order of a meter at the surface of the overriding plate.

We have, in fact, found evidence in emerged fossil microatolls of sea level changes associated with the 1833 event. In these fossil corals from the islands of Sumatra's outer-arc ridge, we also find clear indications of submergence in the decades prior to the earthquake. We interpret this history of vertical tectonic displacement to be the result of the interseismic accumulation of strain prior to the 1833 earthquake and its relief by slip along the subduction interface during the giant earthquake.

1.2. Coral Microatolls as Paleoseismometers

Coral microatolls start growth as a single polyp that attaches itself to a shallow substrate. They then grow as hemispherical

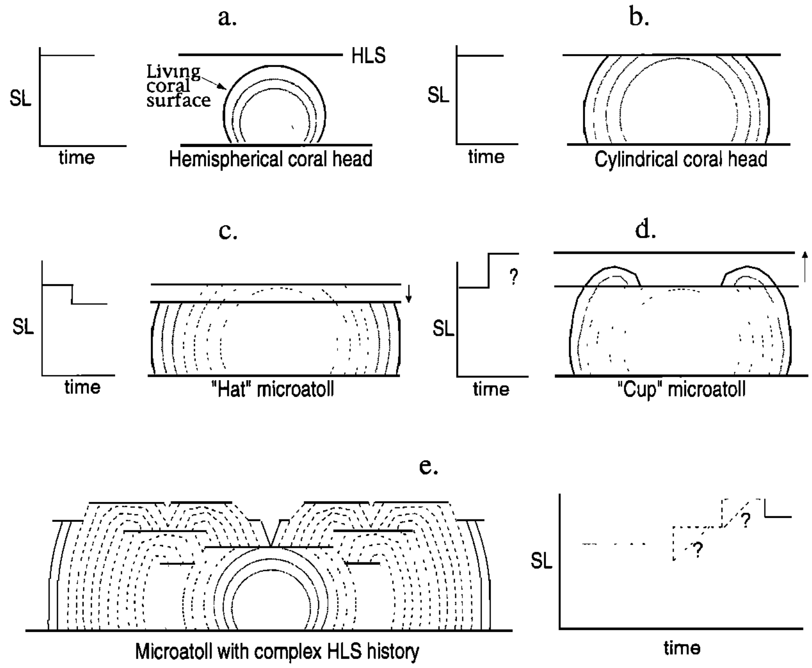


Figure 2. Cross sections that depict the effect of sea level changes on coral growth. The solid black ring is the living coral band. The solid gray rings are dead coral bands that grew under the same sea level conditions as the living coral. The dashed rings are dead bands that grew under different, prior sea level conditions [modified from *Scoffin and Stoddart*, 1978]. (a) A coral growing hemispherically below highest level of survival (HLS), with new skeleton accreting on the outside of the hemisphere. (b) The same coral head under the same sea level conditions after it has reached HLS. No further upward coral growth is possible, but horizontal growth of the part of the coral below HLS continues. (c) The "hat" morphology of a coral head that grew up to HLS but then experienced a drop in HLS. The part of the coral exposed above HLS has died. Lateral growth below the new HLS develops a lower outer rim around a higher center. The elevation difference between the two flats is a measure of the amount of emergence. (d) The "cup" morphology of a microatoll that has experienced a sea level rise after the coral had been growing at HLS. The coral grows upward toward the new HLS, constrained only by its growth rate. Upward and outward growth over the old HLS surface produces a raised outer rim, indicative of submergence. The elevation of the new HLS is not recorded by the coral until the coral grows up to it. (e) Schematic cross section through a hypothetical coral head illustrating how coral stratigraphy can be related to sea level changes. The accompanying graph is a sea level history derived from the coral head. The lines with question marks offer another equally possible sea level history, illustrating the difficulty in distinguishing between gradual and sudden sea level rise.

colonies that accrete new skeletal material in concentric shells. Variations in water temperature, rainfall, sediment influx, and other factors cause the density of the skeletal material in these coral "heads" to vary seasonally in such a way that the concentric shells appear as annual growth bands [*Scoffin and Stoddart*, 1978; *Stoddart and Scoffin*, 1979]. These annual bands are similar to tree rings in that they provide a yearly record of coral growth throughout the life of the coral [*Taylor et al.*, 1987; *Zachariasen*, 1998]. The growth bands are usually about 10 ± 5 mm thick, so heads several meters in radius represent growth over several centuries. The annual bands are often visible to the naked eye in vertical cross-sectional cuts through the coral hemisphere but are commonly more pronounced in X-radiographs of thin cross-sectional slabs [*Taylor et al.*, 1987; *Zachariasen*, 1998].

The upward growth of coral microatolls is limited by low water levels because exposure above water for prolonged periods, especially in sunlight, is fatal to the living corallites on the outer surface of the head [*Scoffin and Stoddart*, 1978]. *Taylor et al.* [1987] termed this upper limit of growth the "highest level of survival," or HLS. HLS tracks sea level, moving up and down as the water level fluctuates. More detailed discussions of microatoll growth, HLS, and their relation to sea level appear in the works by *Scoffin and Stoddart* [1978], *Taylor et al.* [1987], *Zachariasen* [1998], and J. Zachariasen et al. (submitted paper, 1998). To understand this paper, it should suffice to know that the HLS of a coral head is controlled by the elevation of annual lowest water levels.

What is important about HLS for paleoseismic studies is that the microatolls retain a record of HLS in their skeletal morphology. Microatolls grow hemispherically, outward and upward, until they reach the HLS (Figure 2a). At that point, constrained to grow below HLS, they grow only laterally, which leads to the creation of a flat upper surface at the HLS (Figure 2b). A rise or drop in relative sea level produces a commensurate change in HLS, either freeing the microatoll to grow upward again or constraining it to grow laterally at a lower level (Figures 2c and 2d). Total emergence above HLS will kill the entire coral head. This record of HLS history is retained in the morphology and stratigraphy of the coral. A vertical cross-sectional slab cut through the coral head exposes its annual growth bands and their response to changes in HLS (Figure 2e).

Figure 3 illustrates how we would expect a microatoll to

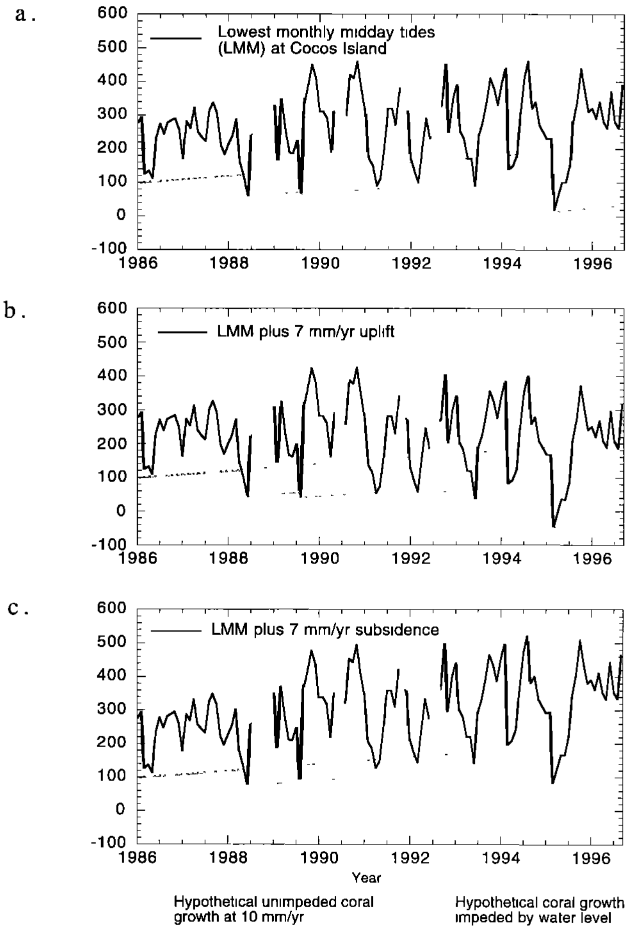


Figure 3. (a) Hypothetical coral growth based on a 10-year tidal record from the Cocos Islands, which are in a tectonically stable part of the Indian Ocean, about 1000 km SSE of Sumatra. Solid black line shows the lowest low monthly water level that occurred during midday hours, when Sun exposure was greatest. Elevations are relative to an arbitrary datum. The top of a hypothetical coral, growing unimpeded at a rate of 10 mm/yr, would rise upward as shown by the dashed line. The effect of the fluctuating lowest low water levels would have been to interrupt upward growth of the coral three times during the decade. Each impingement would have produced a downdrop in the HLS of a few tens of millimeters (solid gray line). Renewed growth upward at the coral's natural growth rate would have followed each downdrop. (b) Similar hypothetical upward coral growth, but with a steady 7 mm/yr emergence superimposed on the Cocos record. Emergence increases the number of impingements from three to six. (c) Similar hypothetical upward growth of a coral, but with a steady 7 mm/yr submergence superimposed on the Cocos record. Submergence reduces the number of impingements from three to two.

respond to normal tidal fluctuations in stable, submerging and emerging environments. On a stable coast, one could expect HLS to drop slightly below the highest living corallites of the microatoll every other year or so. On a coast where steady uplift is occurring at 7 mm/yr, HLS impingements would occur almost every year. Conversely, on a coastline that is subsiding at 7 mm/yr, diedowns due to subaerial exposure would occur only a couple times per decade. Thus a microatoll should record HLS about every one to several years, depending on the rate of submergence or emergence. If rates of submergence are equal to or greater than the upward growth rate of the coral, HLS clips might never occur, and the form of the coral head would remain hemispherical. So, if one knows the absolute age of a fossil microatoll, then its quasi-annual HLS history can be determined for the period during which the microatoll was growing. *Zachariasen* [1998] and J. Zachariasen et al. (submitted paper, 1998) discuss this matter in much more detail.

Because of the characteristics described above, and because TIMS U-Th dating methods permit dating of individual uncontaminated coral growth bands to a precision of 2–10 years in a sample a couple of centuries old [*Edwards et al.*, 1987a, b], microatolls may act as paleoseismometers and paleogeodetic instruments, documenting relative sea level changes as small as a few centimeters caused by vertical tectonic displacements [*Zachariasen*, 1998; J. Zachariasen et al., submitted paper, 1998]. A sudden uplift of more than a few centimeters will produce a drop in HLS and a consequent lower limit of upward growth (Figure 2c) [*Taylor et al.*, 1980, 1987, 1990; *Edwards et al.*, 1987a, 1988; *Edwards*, 1988]. Conversely, the coral can also document a submergence event that suddenly frees it to grow upward to a higher HLS (Figure 2d).

Zachariasen [1998] and J. Zachariasen et al. (submitted paper, 1998) examined living corals on reefs of the outer-arc islands and western coast of Sumatra to determine how well microatolls function as paleogeodetic recorders and to document vertical displacements in the upper plate over the past few decades. They studied the gross morphology of live microatolls and the stratigraphy of cross-sectional slabs cut from the coral heads. Their morphological analysis revealed that the absolute elevation of HLS for a single year on a single coral head could vary by as much as ± 40 mm (2σ), while the absolute elevation of mean HLS of coral heads within a single bay usually varied by less than ± 100 mm (2σ). We assume that the uncertainties derived from this study of living corals can be applied to the analysis of fossil corals.

2. Results

2.1. Field Methods and Observations

In July 1994 and January-February 1996, we studied emerged fossil microatolls located on the reef flats fringing several of the Mentawai Islands, the string of islands west of the Sumatran mainland that are the subaerial expression of the outer-arc ridge (Figure 1). Most of these were *Porites* sp., but a few were *Goniastrea retiformis*. We focused our attention on the islands of Sipora, North Pagai, and South Pagai, which overlie the central 180 km of the 1833 rupture zone suggested by *Newcomb and McCann* [1987] (Figures 1 and 4). We measured the elevations of our samples relative to water level at the time of collection and with respect to HLS of living corals nearby. From several microatolls, we used a chain saw to collect 100-mm-thick cross-sectional slabs for stratigraphic analysis. From other microatolls, we collected samples with a rock hammer and chisel.

Emerged fossil microatolls in nearly pristine condition were common in the bays of the Pagai and Sipora Islands. They often occurred in the same bays as living corals, but nearer shore and at higher elevations, within the intertidal zone. Their tops usually were about half a meter higher than the tops of the modern microatolls. They sat upon either an eroded reef flat composed of cemented coral fragments or on a sandy substrate.

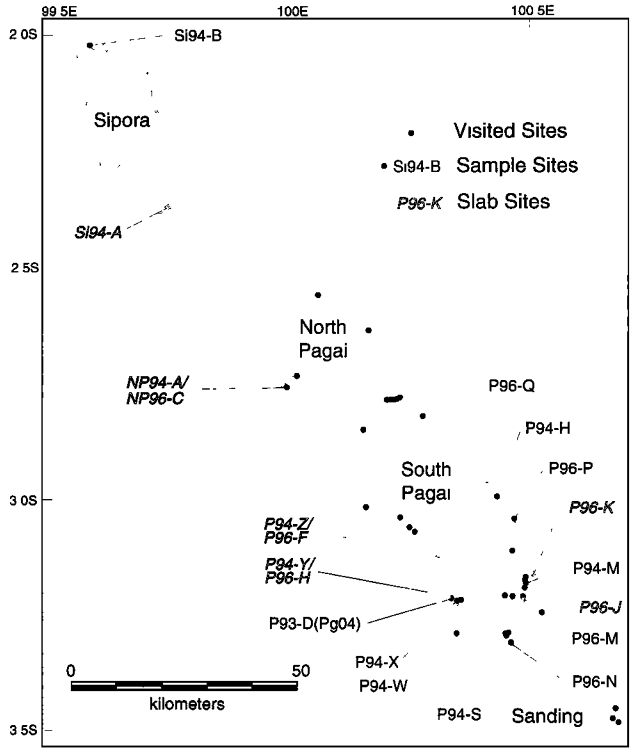


Figure 4. Map of sites on North and South Pagai, Sipora, and Sanding visited during July 1994 and January 1996. We collected samples of 19th century corals at the named sites.

Unlike fossil coral terraces along rising coasts (e.g., Barbados and New Guinea [*Bloom et al.*, 1974; *Chappell*, 1974; *Ota et al.*, 1991, 1993; *Pandolfi et al.*, 1994; *Jouannic et al.*, 1988; *Bard et al.*, 1990]), this set of fossil corals is not associated with higher raised terraces that indicate ongoing permanent uplift. The presence of mid-Holocene fossil microatolls within the intertidal zone demonstrates that long-term uplift or submergence rates here are no more than a fraction of a millimeter per year [*Zachariasen*, 1998].

The form of uneroded fossil microatolls was broadly similar throughout the islands. Their perimeters were commonly steep sided, and many had a cup-like morphology indicative of submergence during their lifetime (Figures 2 and 5). The predominant gross morphology of the fossil corals was similar to that of the live corals. Of the hundreds of fossil corals observed, only a handful had an outer, lower rim that suggested an emergence prior to death (Figures 2c and 6). These observations suggest that most of the fossil corals died by complete emergence, following several decades of submergence. The steep sides and absence of multiple outer rims imply that the emergence and resultant death were sudden.

2.2. Ages of the Fossil Corals

We analyzed 81 samples from 63 emerged fossil coral heads from numerous sites around the islands with U-Th geochronometric methods (Figure 4). We X-rayed 3- to 5-mm-thick slabs cut from these samples, in order to expose the density contrasts in the growth bands. For geochronologic analysis, we removed 1–3 g of coral from the slabs, usually from a single growth band. We avoided portions with the sugary or powdery texture that commonly indicates recrystallization. After determining the age of the sample, we determined the age of death

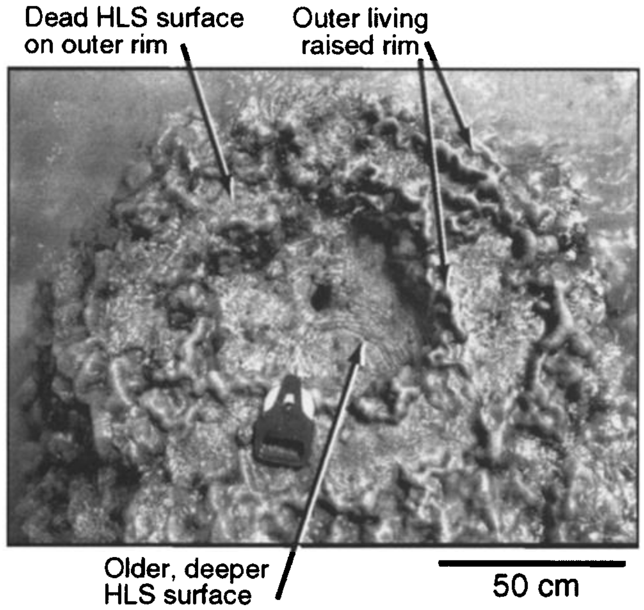


Figure 5. Photograph of a living cup-shaped microatoll. The high rim surrounding a lower dead central flat is indicative of submergence. View is obliquely downward.

of the microatoll by counting the number of annual rings between the dated ring and the outermost ring. The ^{230}Th method and our laboratory procedures are described in the appendix.

About half of the dated samples yielded ^{230}Th ages from the late 18th and early 19th centuries (Table 1). The likelihood that many of these microatolls were killed by emergence during the 1833 earthquake is high, although some of them probably represent other emergence events that occurred in prior

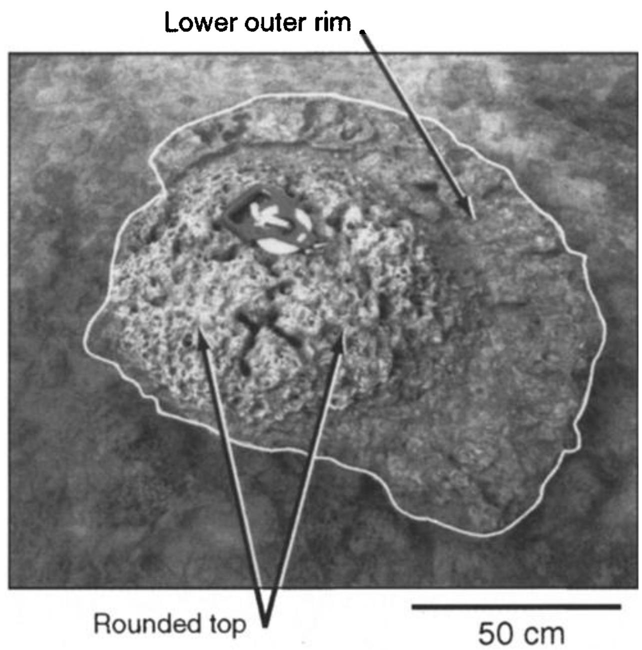


Figure 6. Photograph of dead hat-shaped microatoll. The lower outer rim that surrounds the higher, central dome indicates partial emergence. The rounded cap in the center of the coral shows that the coral had not reached HLS before its emergence. View is obliquely downward.

Table 1. Thorium 230 Dates of Death of Emerged Fossil Corals in North Pagai, South Pagai, and Sipora Islands

Sample Name	Date of Death (Error)	Corrected Date (Error)	Sample Name	Date of Death (Error)	Corrected Date (Error)
Si94-A-3.A	1778 (7)	1790 (15)	P94-M-4.A	1835 (8)	1836 (9)
Si94-A-2.A*	1759 (6)	1771 (15)	P94-S-1.B	1731 (10)	1807 (90)
Si94-A-1A.A*	1716 (4)	1724 (12)	P94-S-3B.B	1825 (4)	1834 (11)
Si94-A-5(1).A	1759 (8)	1769 (12)	P94-W-1.B	1794 (5)	1835 (47)
Si94-A-6A4.A	1827 (7)	1829 (7)	P94-X-1.A	1830 (8)	1838 (12)
Si94-A-6A3.A*	1822 (5)	1824 (5)	P94-X-3.B	1829 (4)	1834 (7)
Si94-B-1.B	1728 (25)	1789 (75)	P94-X-4.A	1836 (4)	1839 (6)
			P94-Y-3A.B	1815 (3)	1830 (18)
NP94-A-5A1(1).A	1792 (17)	1797 (18)	P94-Z-4B.B	1818 (6)	1824 (9)
NP94-A-5A2.A*	1772 (26)	1810 (51)	P96-F-1.B†	1817 (4)	1831 (17)
NP94-A-8A4(1).B	1832 (5)	1841 (11)	P96-H-1.A	1824 (3)	1836 (14)
NP94-A-8A1(2).B	1836 (7)	1842 (11)	P96-I-1.A	1835 (15)	1839 (16)
NP94-A-9A1.B	1834 (6)	1837 (7)	P96-J-1.A	1839 (10)	1842 (11)
NP96-C-1.B	1795 (6)	1807 (15)	P96-J-2.C	1831 (4)	1835 (6)
NP96-C-2.A	1820 (11)	1824 (12)	P96-K-1.B	1830 (3)	1831 (3)
NP96-C-3.A	1798 (5)	1844 (54)	P96-K-4.A	1810 (13)	1817 (16)
NP96-C-4.A	1814 (3)	1829 (18)	P96-K-5.A	1821 (6)	1827 (10)
NP96-C-5.B	1787 (3)	1799 (14)	P96-N-1.B	1860 (32)	1866 (33)
			P96-N-2.A	1632 (7)	1800 (197)
Pg04E-0.0.2(B)	1802 (11)	1826 (31)	P96-N-3.A	1662 (4)	1803 (165)
Pg04E-0.0.2(A)‡	1803 (4)	1842 (45)	P96-N-5.A	1765 (3)	1819 (63)
Pg04G-0.0.2(B)	1706 (12)	1834 (150)	P96-P-1.B	1750 (5)	1831 (55)
Pg04G-0.0.2(A)‡	1463 (29)	1834 (434)	P96-P-2.B	1777 (25)	1778 (25)
P94-H-1A.A	1766 (7)	1807 (30)	P96-Q-1.A	1789 (9)	1800 (16)
P94-H-2.A*	1809 (6)	1817 (8)			

We dated one or more pieces from each coral head with U-Th geochronometry. The uncorrected date of coral death is calculated by adding the number of annual growth bands between the sampled annual band and the outermost band of the coral head to the measured age of the sample. Uncertainties in the ages combine the 2σ analytical uncertainties and the ring counting uncertainties, which are estimated based on ring clarity. The corrected dates of death are determined in the same way, but from measured ages that have been corrected for contamination with initial thorium [*Zachariasen*, 1998]. Uncertainties in the corrected dates of death include 2σ analytical uncertainties, ring counting uncertainties, and uncertainties in the correction factor.

*Sample is from the interior of the same coral the above-listed sample comes from; first sample is usually from outer raised rim, second sample from interior flat.

†Analysis is from same region of slab as above-listed sample (e.g., both from outer rim) but from a different hand sample collected in a different year.

‡Repeat analysis of same sample as the above-listed sample.

decades. A histogram of the dates of coral death (weighted by their uncertainties, after the method of *McGill and Sieh* [1991]) has a pronounced peak at 1832 (Figure 7). A secondary peak near 1800 may also represent a distinct event. This cluster of ages around 1800 is clearer when the dates are plotted individually with their uncertainties (Figure 8). These earlier deaths may represent a single emergence event, but they could also represent several ages of death. An earthquake in 1797, which generated tsunamis on the Batu Islands and on the coast of Sumatra, northeast of Sipora [*Newcomb and McCann*, 1987] may be represented by some of the corals in this cluster.

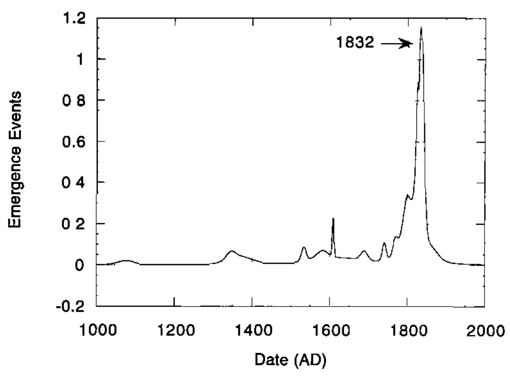


Figure 7. Plot of late Holocene dates of coral death shows dominance of deaths near 1833. For each coral date, we produced a Gaussian curve, centered on the corrected ^{230}Th age and with a standard deviation equal to half the age uncertainty. This plot is the sum of all the individual curves.

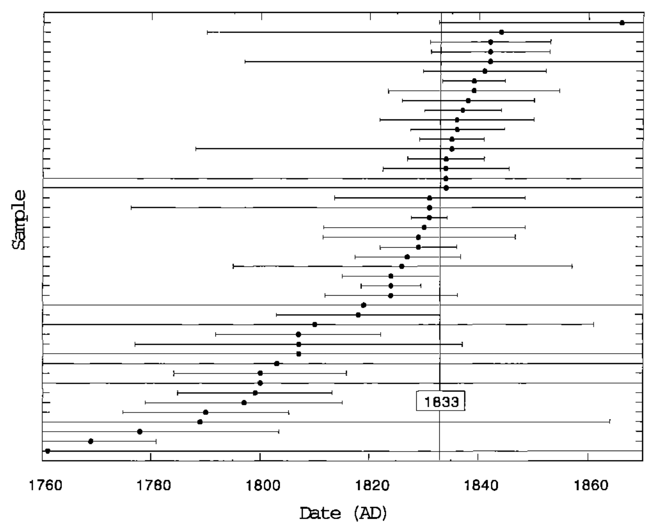


Figure 8. Corrected ^{230}Th dates of individual coral deaths between the mid-18th and mid-19th centuries and their uncertainties.

Figure 8 shows that many of the emerged microatolls died suddenly between about 1820 and 1845. The coincidence of this cluster of coral deaths with 1833, the year of the giant earthquake, suggests strongly that these corals died from exposure due to emergence during that event. Because the death dates of these corals are consistent with an 1833 death, we will assume 1833 in the following discussions to be the date of death. Hence discussion of the dates of older rings and HLS impingements recorded in those corals will be referenced to an 1833 death.

2.3. Interpretation of the Coral Slabs

We collected large cross-sectional slabs from seven corals whose ages are consistent with an 1833 death. The slabs came from sites on or near all three islands: one from Sipora, two from North Pagai, and four from South Pagai (Figure 4). X-radiographs of the slabs reveal patterns of annual density bands that reflect multiple old HLS surfaces. Using the X-radiographs, we measured the changes in elevation of successive HLS surfaces and documented the timing of the HLS impingements by counting the annual bands. From these stratigraphic analyses, we obtained annual records of relative sea level change that affected the corals, up to and including their deaths by emergence.

Here we describe the seven slabs and their vertical-displacement histories. We use one sample to illustrate the method of analyzing the X-radiographs, and then we present the results of similar analyses of the remaining samples. We also compare the stratigraphic records from the slabbed corals with the morphological signature of other microatolls of similar age on the same reef as a test of consistency. All contemporaneous corals in one bay should have consistent records of HLS change, since each has been exposed to the same relative sea level history. Finally, we use the displacement histories as a basis for discussion of the tectonic and paleoseismic history of the decades prior to and including the 1833 earthquake.

2.3.1. Two records of the 1833 event from a site on the west coast of North Pagai. On the west side of North Pagai, at site NP94-A/NP96-C, we surveyed 16 emerged coral heads. These dead corals fall into two groups, according to the elevation of their tops (Figure 9). Dead corals D1, D2, D6, D7, D8, and possibly D3, D4, and D5 belong to the high group, and corals D9, D12, D13, D14, D15, D16, and NP94-A-9 belong to the low group. Corals D11 and NP94-A-6 are low-topped but are highly eroded.

The elevation difference between the high and low populations of fossil corals at this site is as much as 550 mm, significantly larger than the ±100 mm variation in HLS elevation usually observed between living corals at a single site [*Zachariasen*, 1998; J. Zachariasen et al., submitted paper, 1998]. The ages of the two populations of coral also vary. The low corals D9 (sample NP94-A-8), D13 (sample NP96-C-2), D14 (sample NP96-C-3), D16 (sample NP96-C-4), and NP94-A-9 are all possible 1833 fatalities (Table 1; Figure 9b). Three higher corals, D1 (source of sample NP96-C-5), D7 (sample NP96-C-1), and high D10 (sample NP94-A-5) died in the late 1700s. These are too old to be 1833 fatalities and may have died in a distinct emergence event around the turn of the century. The correlation of older ages with higher elevations, and 1833 ages with lower elevations, suggests that the two populations represent different generations.

Two slabs, both from apparent 1833 casualties, were collected from corals at this site. Sample NP94-A-9 came from a small *Goniastrea retiformis* coral (Figures 9 and 10). It died in 1837 ± 7. The X-radiograph of the coral slab reveals a series of concentric dark and light bands, each pair representing an annual cycle of skeletal growth (Figure 11a). From the X-radiograph, we identified the annual bands and former HLS surfaces and unconformities and generated a tracing of these primary features (Figure 11b). We assigned each band an age by counting from the dated ring. We measured the elevations of the tops of each of the annual bands relative to the top of the coral, the level at which the coral head was growing just prior to final emergence, to determine how relative HLS height changed during the final decades of the microatoll's life. Rings that were clipped and impeded from upward growth form surfaces that represent the HLS. For complete rings, upward growth was limited by the coral's natural growth rate, not by exposure, and therefore the tops of these rings do not represent HLS. They do, however, indicate that the coral did not grow up as high as it could have and therefore that HLS was above the level of the top of those rings. Thus we can use these elevations of the tops of those unclipped rings only as a lower bound to HLS elevation during that year.

The rounded top of the coral head and the uniform thickness of the annual ring at the top of the coral indicate that this coral head was still below HLS in the year before the final emergence event. Slight HLS impingements, or clips, created when HLS dipped below the top of the coral, appear to have occurred in 1826 and 1828. A larger impingement produced a small flat surface in 1819. Prior to that, there is evidence in the left lobe of an old HLS clip about 260 mm from the top. The flat is eroded, but the growth of younger rings toward and over the eroded flat surface suggests that it formed in the years just prior to 1811. The amount of time represented by the flat is unclear because bands that formed before about 1805 are absent. Nonetheless, the younger part of the flat appears to have formed between 1806 and 1810. The right lobe also has an eroded flat about 220 mm from the top, which may correlate with the oldest HLS surface recorded in the left lobe. However, the ambiguity of the rings that formed before 1817 preclude precise dating of the HLS flat.

The slab stratigraphy reveals that HLS and relative sea level were rising in the two decades prior to coral death. The coral submerged at least 180 mm between 1819 and 1833 and then suddenly and completely emerged and died in 1833. After emergence, HLS must have been below the base of the coral because the entire head died. Thus the amount of emergence must have been at least 290 mm, the vertical distance between the lowest and highest parts of the outermost ring.

A slab cut from a microatoll only about 50 m seaward of the head we have just discussed contains a similar HLS history. Slab NP94-A-8 came from Coral D9, a *Porites* sp. microatoll 1.65 m in diameter (Figure 9). The slab was cut radially inward, to a point slightly beyond the center of the head (Figure 12). We dated two pieces from this sample, one from the interior of the central flat and one from the raised outer rim. The corrected ^{230}Th date of death was 1841 ± 11, counting out from the dated exterior piece, and 1842 ± 11 from the interior piece.

Four HLS surfaces are apparent in the slab morphology and two additional HLS impingements are buried within the body of the microatoll (Figure 13a). Though the rings at the center of the microatoll are difficult to discern, it appears that the coral grew upward and outward for 30–50 years before first reaching HLS in the 1780s. This first interaction with HLS occurred at or near the beginning of an emergence event. The a.

b.

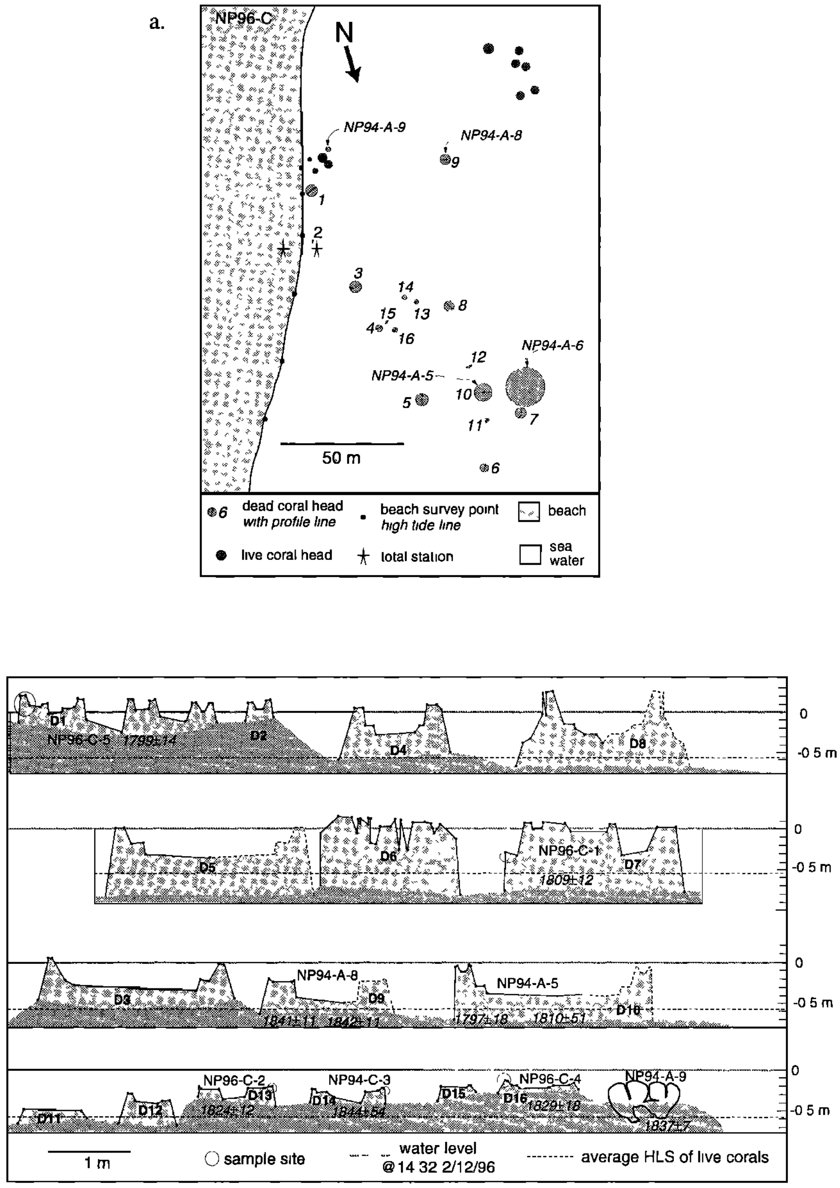


Figure 9. (a) Map and (b) cross-sectional profiles of dead coral microatolls at site NP96-C (also NP94-A). The numbered dead corals on the map correspond to those in the profiles. The profiles reveal at least two populations of microatolls. The higher corals have a more complex morphology, whereas the lower corals are smaller and have a simpler morphology. Several corals from the lower population appear to have died in the 1833 emergence event. The higher corals may have died earlier, near the end of the 18th century. The morphology of both populations suggests they experienced preseismic submergence.

precise date of the emergence is uncertain because the rings near the top of the hemisphere are truncated and may be slightly eroded. The oldest ring exposed at the surface grew sometime between 1782 and 1788. If this reflects the first year of HLS impingement and emergence, emergence continued over several years, until about 1793, when HLS was 140 mm below the top of the hemisphere. An alternative interpretation is that the emergence occurred more rapidly, in about 1793, and was followed by slight erosion of the central hemisphere's surface. A third possibility is that the emergence was sudden, sometime between 1782 and 1788, but the microatoll did not respond to the emergence for several years. Regardless of which of these three interpretations is correct, the hemispherical core surrounded by the raised outer rim indicates that the coral had been growing an indeterminate depth below HLS until an emergence in the late 1780s or early 1790s brought HLS about 140 mm below the top of the hemisphere.

Soon after this emergence, HLS must have risen for about a decade at an average rate greater than the rate of upward coral growth because the rim grew upward unimpeded until about 1800. Upward growth was again obstructed briefly by HLS in about 1808. Four consecutive years of HLS impingement occurred between about 1813 and 1816, after which the coral began to grow upward again, unimpeded by HLS, until about 1830. The two horns at the top of the outermost rim indicate that the coral had once again resumed unimpeded upward growth in the year or two before its death in 1833.

So, like NP94-A-9, this coral also experienced significant

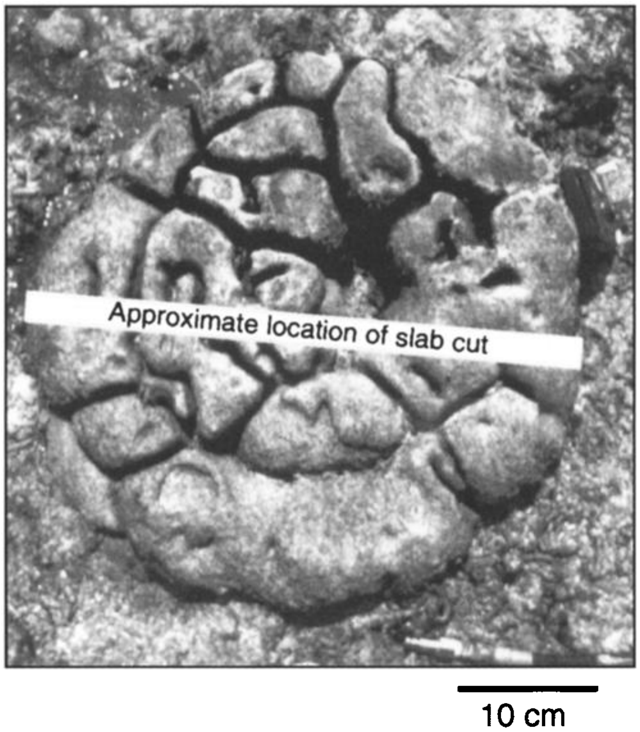


Figure 10. Photograph of the top of microatoll NP94-A-9, looking downward. Form is typical of *Goniastrea* sp. corals.

submergence prior to death: HLS rose over 400 mm after the emergence of the 1790s. Final emergence was sufficient to raise the living portions of this coral completely above HLS, causing the head to die. Unfortunately, the outermost ring on our slab does not continue downward more than about 130 mm from the top, so the sample provides a minimum bound to the magnitude of emergence of just 130 mm. The two slabs from this site reveal a history of rapid submergence prior to final emergence and death, probably in 1833. Final emergence was sudden, and vertical displacement was at least 290 mm.

The morphological characteristics of the other corals that died in about 1833 are consistent with the HLS history of the two slabs. They possess a raised outer rim around a central flat, the signature of rising water level. Some of these rims are 200 mm high, similar to the height of the rim in NP94-A-8 and the elevation of the top of NP94-A-9 above its HLS surface of 1819. The similarity of the stratigraphic and morphological signatures of all these emerged microatolls, their large numbers, the suddenness and size of their emergence, and the coincidence in time with a known earthquake all suggest that coseismic uplift in the 1833 earthquake followed several decades of rapid submergence.

2.3.2. Earlier emergence(s) on the west coast of North Pagai. In addition to the final emergence event in 1833, the stratigraphic record of NP94-A-8 also reveals the occurrence of an earlier emergence around 1790. Because the coral head had not reached HLS just prior to that emergence, the magnitude of the emergence is poorly constrained. It is at least 140 mm, the elevation difference between the top of the central hemisphere and the thin bridge from hemisphere to rim.

This emergence may correlate with the emergence that killed some or all of the corals with higher elevations. Two high corals (D1 and D7) and one intermediate-depth coral (D10) died in 1799 ± 14, 1809 ± 12, and 1797 ± 18, respectively. Both this older set of microatolls and the set that died in 1833 were alive in the late 1700s and thus must have experienced the same fluctuations in sea level during that period. The slight emergence of NP94-A-8 and complete emergence of these other, higher heads may constrain emergence during this event to about 800 mm (Figure 9). For more discussion of the evidence of this earlier emergence event, see *Zachariasen* [1998].

2.3.3. Evidence for emergence in 1833 and 1810 from a site on Sipora. Site Si94-A is on Siruamata, a small island near the southern tip of Sipora (Figure 4). Several corals at this site attest to two large emergence events that occurred in rapid succession in the early 1800s (Figure 14a and 14b). The latter probably occurred during the 1833 earthquake, and an earlier large event occurred about 1810.

Slab Si94-A-6 was taken from Coral 1 (Figure 14). This microatoll was one of the few that had a lower outer rim. The rim indicates that the microatoll survived a large but partial emergence a couple decades prior to final emergence. The coral head was 3.2 m in diameter at the base and stood about 1 m above a rocky substrate and the sand that partially buried its base. The outer rim was 200–250 mm high, 250–350 mm wide, and about 700 mm above the outermost, lower rim.

The slab from this coral was the most intact of all slabs collected in 1994 and 1996 and possessed the least ambiguous annual banding (Figure 13b). We dated two samples from two separate rings, which yield consistent ages for the death. The outermost sample had a corrected date of 1797 ± 7, which yielded a date of death of 1829 ± 7. The inner sample had a corrected date of 1770 ± 5, which yielded a death date of 1824 ± 5, which is slightly older than 1833. So, perhaps we have underestimated the error in the age of death.

The slab cross section displays a history of submergence followed by two large emergence events. Growth from about 1764 to 1789 was unimpeded except for a brief HLS impingement in 1779. In about 1789 an emergence of 80–100 mm occurred. Subsequent submergence was rapid enough to prevent HLS impingements between 1789 and 1810. The coral had not again reached HLS when, in 1810, HLS dropped at least 700 mm, and all but the bottom 250 mm of the outer living band died.

Lateral growth of the lowest 250 mm of the head continued after 1810 and produced the lower outer flange. The narrow growth bands from 1810 to about 1824 demonstrate that postemergence growth occurred at a reduced rate. The average thickness of these rings is about half that of the pre-1810 bands. Perhaps shallower water over the reef flat following the emergence led to conditions less favorable to growth (for example, more rapid warming of the water by sunlight or poorer circulation of nutrients during low tide).

Erosion has removed the top of the lower rim, so it is unclear exactly how the coral responded to HLS in the decade or two following the emergence. However, its shape and elevation suggest HLS was stable or rose slightly in the decades after 1810. The submergence was no more than about 80 mm during the next 23 years, and probably less. Emergence in 1833 was at least 270 mm, the height of the outer rim above the substrate.

If uplift caused Coral 1 to emerge fully and die in 1833, other corals living at the site at that time should also have been uplifted. The degree to which the HLS records of these corals are similar or dissimilar to that of Coral 1 is an indication of the precision of the method.

Of the four corals examined on this reef, two had a lower outer rim, indicative of partial emergence (Figure 14b). The

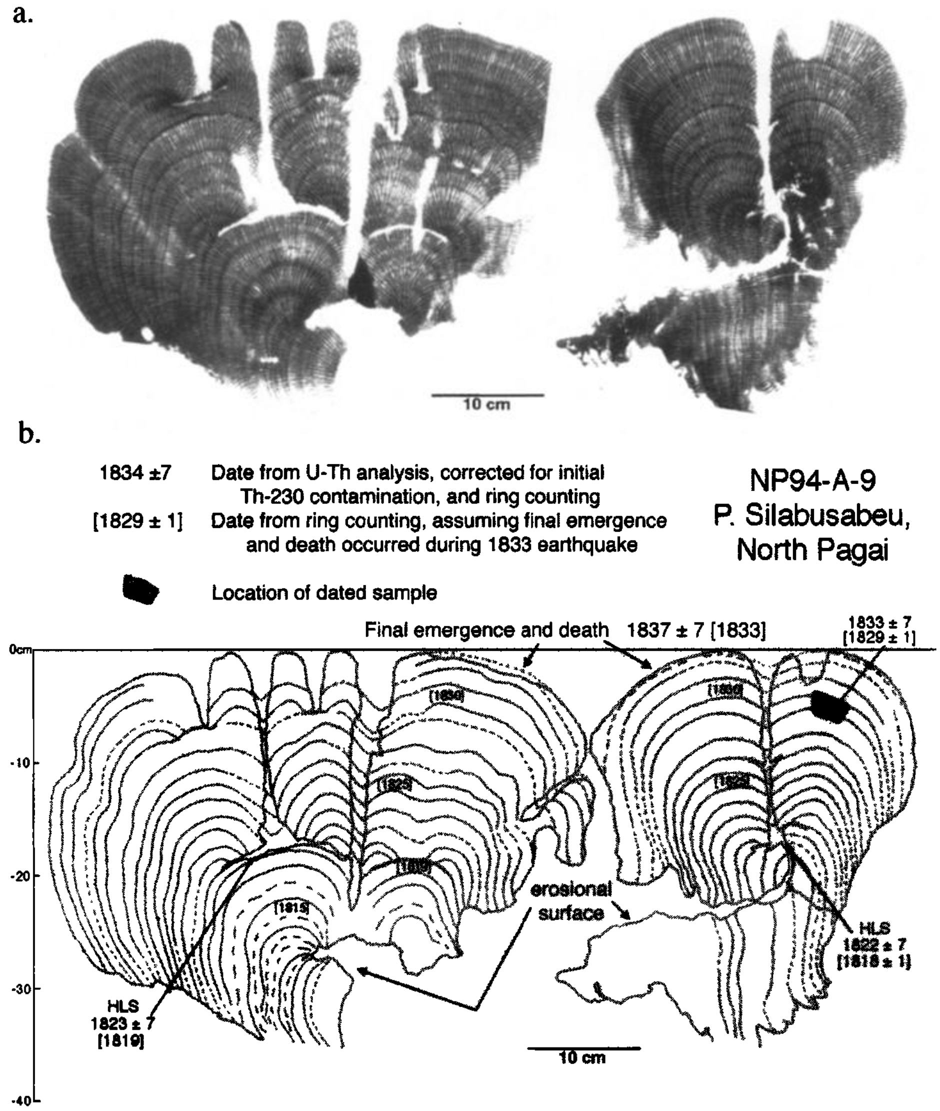


Figure 11. (a) X-radiograph of cross-sectional slab of coral head NP94-A-9. (b) Drawing of annual bands and HLS surfaces interpreted from the X-radiograph. A series of HLS impingements indicates rapid submergence in the two decades prior to 1833. The magnitude of the final emergence in about 1833 is at least 290 mm, the vertical distance from the top to the bottom of the outermost ring. Coseismic emergence could have been greater, since the coral head rose completely above HLS and died.

other two did not. The rims on both Coral 1 and Coral 3 are about 250 mm wide and 700 mm below the highest part of the coral. We did not date Coral 3, but its morphology suggests that it was a contemporary of Coral 1 and experienced a similar HLS history.

2.3.4. An event in the middle to late 1700s at the same site on Sipora. Corals 2 and 4 at site Si94-A have no lower ring. But this might be expected because their bases sit at approximately the same elevation as the top of the outer lower rings in Corals 1 and 3 (Figure 14b). Thus the morphology and base elevations of Corals 2 and 3 suggest that they could have emerged fully and died during the 1810 emergence event that is so clear in the slab from Coral 1. However, their ^{230}Th dates show that these corals died decades earlier, in 1790 ± 15 and 1769 ± 12, respectively.

Our ^{230}Th dates show that Coral 2 could not have emerged and died any later than 1805 or any earlier than 1775. Within this period, Coral 1 grew upward unimpeded, except for a minor HLS clip in about 1779 and HLS clips and a modest emergence between about 1786 and about 1789. One can argue that neither the minor clip of the 1779 HLS nor the clip of the 1786 HLS below the top of Coral 1 is likely to have been contemporaneous with the death of Coral 2 because the base of Coral 2 extends about 400 mm below the 1779 and 1786 HLS's of Coral 1. This scenario is plausible only if the lower outside perimeter of Coral 2 was already dead or was buried by sand, below the 1779 or 1789 HLS. If so, an emergence of about 350 mm could have killed Coral 2 and produced either the 1779 or 1786 mm HLS clips in Coral 1. Collection and analysis of a slab from Coral 2 would allow us to test these ideas.

The morphology of Coral 4 is also difficult to reconcile with the history of Coral 1. The date of the outermost part of the raised rim of Coral 4 is 1769 ± 12. This should be the date of its death by emergence. An emergence sufficient to raise the entire coral out of the water would have been at least 500 mm.

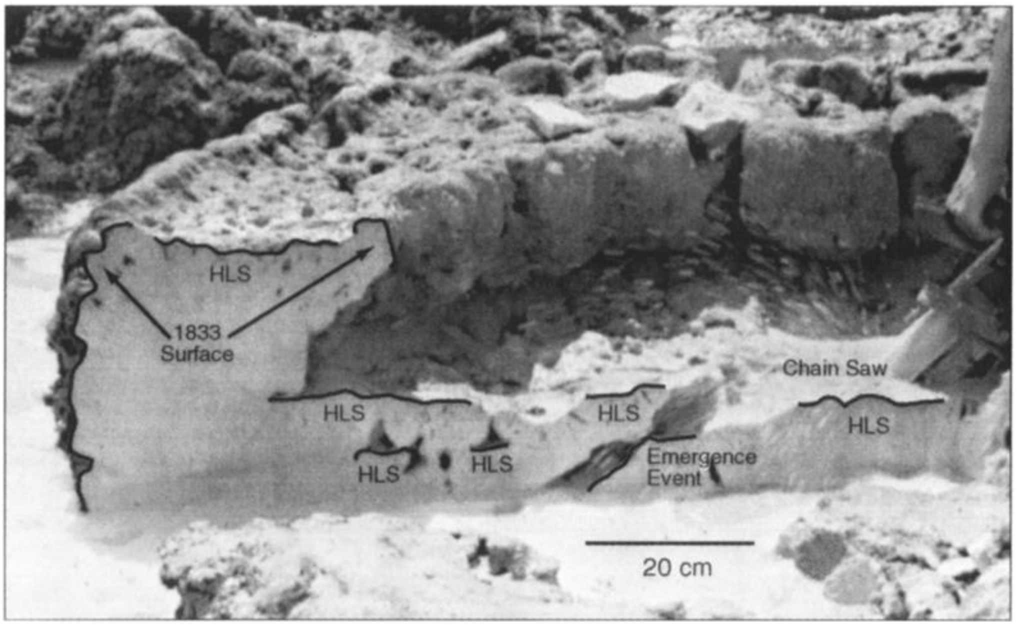


Figure 12. Photograph of coral head NP94-A-8 during removal of a slab. Numerous HLS surfaces within the head indicate a predominance of submergence in the decades prior to final emergence and death. The base of the coral and the sandy substrate are under water.

Such an emergence is plausible if the date of the emergence is in the 1750s or early 1760s because we have no record of that period in the slab from Coral 1. It is less plausible if the date of emergence is in the late 1760s or the 1770s because the annual bands in Coral 1 show nearly unimpeded growth during this period. As in the case of Coral 2, we could hypothesize that only the upper parts of the outer perimeter of Coral 4 were alive during this final emergence. If this were so, a drop of HLS to the level seen in the minor clip in Coral 1 could have killed Coral 4.

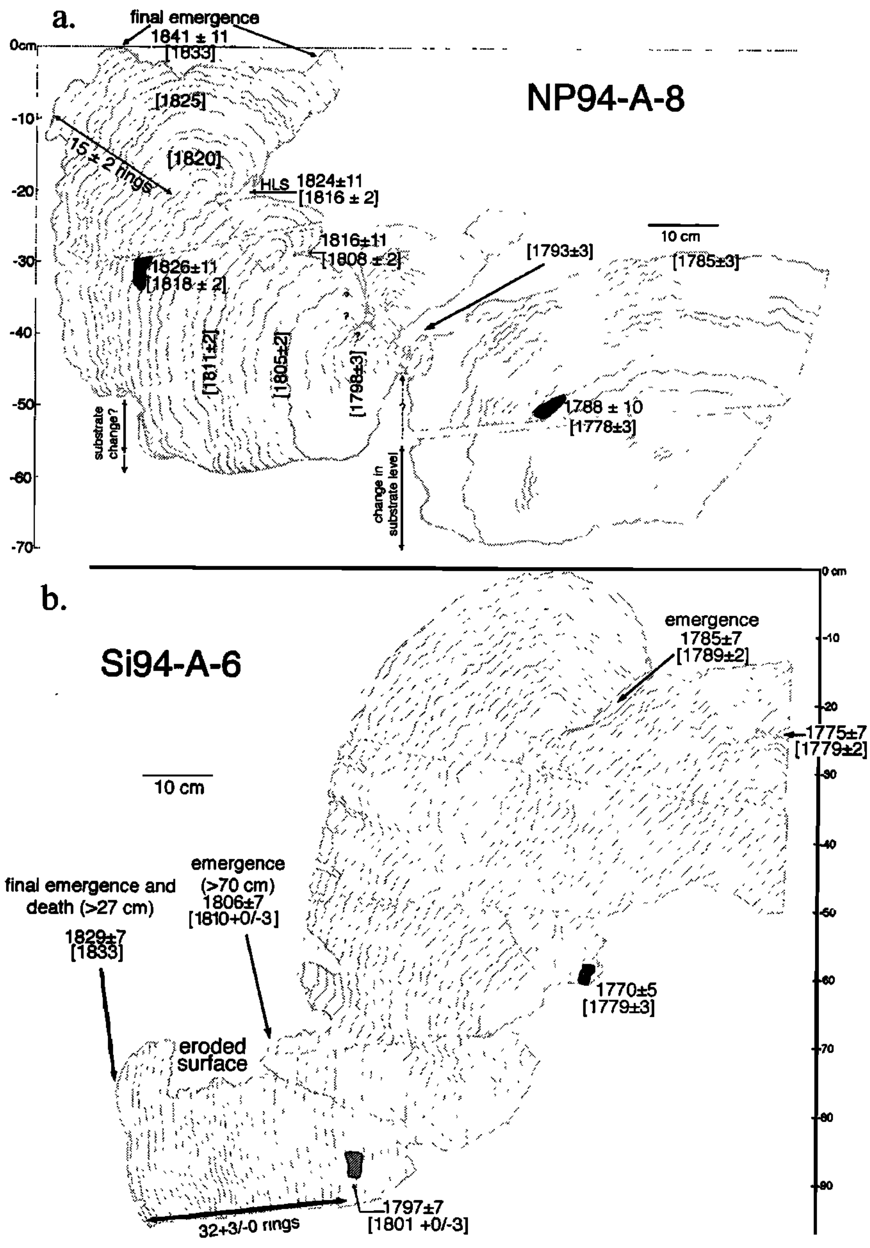
The four corals from the site near the southern tip of Sipora illustrate the possibilities and difficulties inherent in using fossil microatolls to document past changes in sea level. The HLS record derived from slab Si94-A-6 is very clear. Three emergence events of about 100, 700, and >250 mm are clearly documented. The times between the events are also constrained to within a couple of years, since this slab has such clear growth rings. The questions raised by the ages and morphology of the three unslabbed corals show that morphological analysis is insufficient and must be coupled with stratigraphic analysis of cut slabs to achieve well-constrained sea level histories.

2.3.5. The 1833 emergence on the west coast of south Pagai. P96-H (also P94-Y) is a site within an elongate embayment on the western side of South Pagai (Figure 4). Sample P96-H-1 came from a large cup-shaped *Porites* microatoll, 2.35 m wide and 1.26 m tall. The raised outer rim was two-tiered, with the higher outer rim about 300 mm above the central flat (Figure 13c). The perimeter of the coral dipped about 50–60° outward, except near the base of the coral, where it was nearly vertical.

P96-H-1 records several decades of submergence punctuated by a few small emergences and two periods of HLS stability. Most of the central core is missing from the sample, but the innermost part of the sample is from the central flat of the coral. Correlation of the rings of the separate innermost piece of the central flat with those of the main part of the slab is highly uncertain. Our tentative correlation suggests that upward growth of the coral was impeded first between about 1786 and 1789. The shape of the rings suggests that an emergence of about 60–80 mm in 1791 was followed by unimpeded upward growth until a small emergence in 1797. The coral was impeded at a higher elevation between about 1808 or 1809 and 1815. Between 1815 and 1833 the coral grew upward about 20 cm. The 1833 ring does not continue over the top of the head, so the microatoll was clipped by HLS just prior to emergence in 1833. Although the coral stands more than a meter above the reef flat, the absence of the outermost ring on most of the outer edge of the slab means that we can be certain of a drop in HLS of only 100 mm in 1833. Another steep-sided *Porites* microatoll (P94-Y-3) at this site was dated but not slabbed. That coral died in 1830 ± 18, and its morphology was similar to that of P96-H-1.

Site P96-F (also P94-Z) is on an island off the central west coast of South Pagai (Figure 4). P96-F-1 is a slab from a cup-shaped *Goniastrea retiformis* microatoll (Figure 13d). Tilted about 5° seaward, with half its rim broken off, the microatoll stood about 600–650 mm above the substrate and was 1800 mm wide. Its raised outer rim was 450 mm wide and contained a deep hollow between its inner and outer halves. The microatoll died in 1831 ± 17.

The slab from this coral also records a history of submergence punctuated by brief periods of stability or emergence (Figure 13d). As with the sample just discussed, the interior pieces of the slab are detached from the outer, so we cannot match well the rings of the central flat with those of the raised outer rim. The tops of the smaller detached pieces to the right and below the outer rim were part of the central flat surface of the microatoll. We tentatively estimate that the flat represents about 5 years of relative HLS stability, from 1801 to 1806. The innermost fragment of the coral has a small HLS impingement of about 2 years' duration, about 15 years and 130 mm below



Legend: Location of dated sample

1809±15 Date from U-Th analysis, corrected for Th-230 contamination, plus ring counting

[1824±2] Date ring counting, assuming final emergence and death occurred in 1833 the top of the fragment. If the top of the fragment has a date of about 1803, the HLS impingement occurred in 1789.

The main part of the slab shows that three multiyear HLS impingements interrupted upward growth of the head between about 1800 and final emergence in 1833. The first of these occurred before 1808. The next occurred at a higher level about 10 years later. This HLS flat is eroded, so the age of first impingement is uncertain by a couple of years, and the elevation of the first impingement is uncertain by a couple of centimeters. We estimate that HLS impeded growth between about 1815 and 1819. Growth was impeded by a still-higher HLS from 1831 to 1833. Emergence of at least several tens of millimeters killed the head in 1833.

2.3.6. The 1833 emergence on the southeast coast of South Pagai. Sample P96-J-2 was collected from the east coast of Siatanusa Island, one of the small islands of the archipelago off the southeast coast of South Pagai (Figure 4). The coral sat on an eroded reef, upon which only a few small microatolls were preserved, near the modern storm berm. Most of these were only about 400–600 mm wide and tall. Their subdued rims suggested that they had experienced minimal preseismic submergence and/or some erosion. Two fossil corals from this site were dated. They yielded dates of death of 1842 ± 11 and 1835 ± 6. The latter date came from P96-J-2, the largest fossil microatoll observed at this site, about 1.10 m in diameter (Figure 13e).

P96-J-2 experienced a couple decades of unimpeded upward growth that were terminated by at least 6 years of clipping by a stable HLS between about 1810 and 1815 (Figure 13e). HLS dropped about 50 mm in 1816. Upward growth then resumed unimpeded, and by 1829 the microatoll had created a raised rim nearly 100 mm above the flat of 1810–1815. After a year or two of impingements near this elevation, HLS rose again, enabling the microatoll to add the small horns on top of the rim. In 1833, however, the entire coral emerged above HLS and died. The vertical distance from the top of the outermost ring to its lowest extent, 220 mm, is a minimum measure of the emergence in 1833.

Six kilometers north of the site just described is site P96-K, on the east side of a large island in the archipelago (Figure 4). Numerous living and dead microatolls exist in several bays around this site. Dates of death of three of the corals from this site, K-1, K-4, and K-5, are near 1833. We collected a slab from P96-K-4, a small *Porites* head about 500–600 mm wide. Because the coral was small, the P96-K-4 slab records only a short HLS history (Figure 13f). HLS impingement first restricted upward growth in about 1811. Shortly thereafter, HLS dropped, and the coral emerged about 30 mm. Further unimpeded upward growth followed until it emerged fully in 1833 and died. The minimum amount of final emergence was about 270 mm.

2.3.7. Other sites with 1833 or late 19th century corals. We found many other microatolls on the reefs of North and South Pagai that died around 1833 (Table 1; Figure 15). We did not slab these, but we did describe their morphologies and measure their elevations. All had the morphological signature of sudden emergence following decades of submergence. Their occurrence throughout the Pagai Islands helps constrain the geographic extent of the coseismic uplift in 1833.

3. Synthesis and Interpretation

What do these diverse records of sea level fluctuation contribute toward an understanding of the earthquake geology of the Sumatran subduction zone? In this section, we attempt to construct a coherent interpretation of the giant 1833 earthquake and related phenomena, using the data we have presented above.

First, we discuss the nature of pre-1833 submergence that we see in all of the microatolls. Second, we consider the evidence for emergence events in the decades prior to 1833. Third, we attempt to reconstruct the magnitude of coseismic emergence in 1833. Fourth, we use this and a simple elastic dislocation model to estimate the amount of slip on the subduction interface in 1833. Fifth, we use our estimate of coseismic slip and the geographic extent of rupture to constrain the magnitude of the 1833 event. And finally, we discuss the broader implications of this work in understanding the nature of the earthquake cycle in Sumatra.

Figure 13. (opposite) Drawings from X-radiographs of NP94-A-8 and five other slabs from various sites in the Pagais and Sipora. (a) NP94-A-8. Compare with Figure 12. The interior of this head was at or below HLS until it emerged partially above HLS in the late 1700s. An increase in substrate level coincident or slightly after this emergence may reflect partial burial by sand. Submergence beginning shortly after the mid-1790s allowed growth of the outer raised rim. Several HLS clips occurred during this lengthy period of submergence. The microatolls died during a large emergence event, in about 1833. The magnitude of that event was at least 130 mm, but probably more than the full height of the coral head. (b) Si94-A-6. The outer portion of this microatoll reveals a complex history of submergence and emergence. Final emergence probably occurred in 1833, but a 700-mm emergence event occurred in about 1810. This slab also shows an early history of submergence. (c) P96-H-1. Slab from 1.2-m-high *Porites* sp. microatoll. The microatoll died by emergence in about 1833. In the decades prior to emergence, the head was submerging, except during two multiyear periods of HLS stability. The lack of a lower outer rim suggests that emergence in 1833 was greater than the 1.2-m height of the coral head. (d) P96-F-1. Outer rim and part of the central flat of a *Goniastrea* sp. coral. A period of HLS stability represented by the central flat was followed by about 300 mm of submergence, punctuated by at least one multiyear episode of HLS impingement. The magnitude of final coseismic emergence was at least 200 mm. Photographs show the top of the coral before slabbing and the slab after sawing. The shape of the thin X-rayed slab is not identical to that in the photo of the slab because the thin slab was cut from the interior of the larger slab. (e) P96-J-2. Outer rims and portion of the interior of a *Porites* sp. microatoll from one of the many islands southeast of the South Pagai peninsula. The presence of both a central flat and a flat near the top of the outer rim suggests two short periods of sea level stability and of HLS impingement during several decades of overall submergence. The horns at the top of the rim suggest that submergence was occurring in the few years before final emergence. The final emergence in about 1833 was at least 220 mm. (f) P96-K-4. Outer rim and portion of the interior flat of a *Porites* sp. microatoll. Only one episode of HLS stability appears in the few decades prior to death by emergence. Submergence dominates the history of this head in the two decades prior to its death. The magnitude of emergence was at least 270 mm.

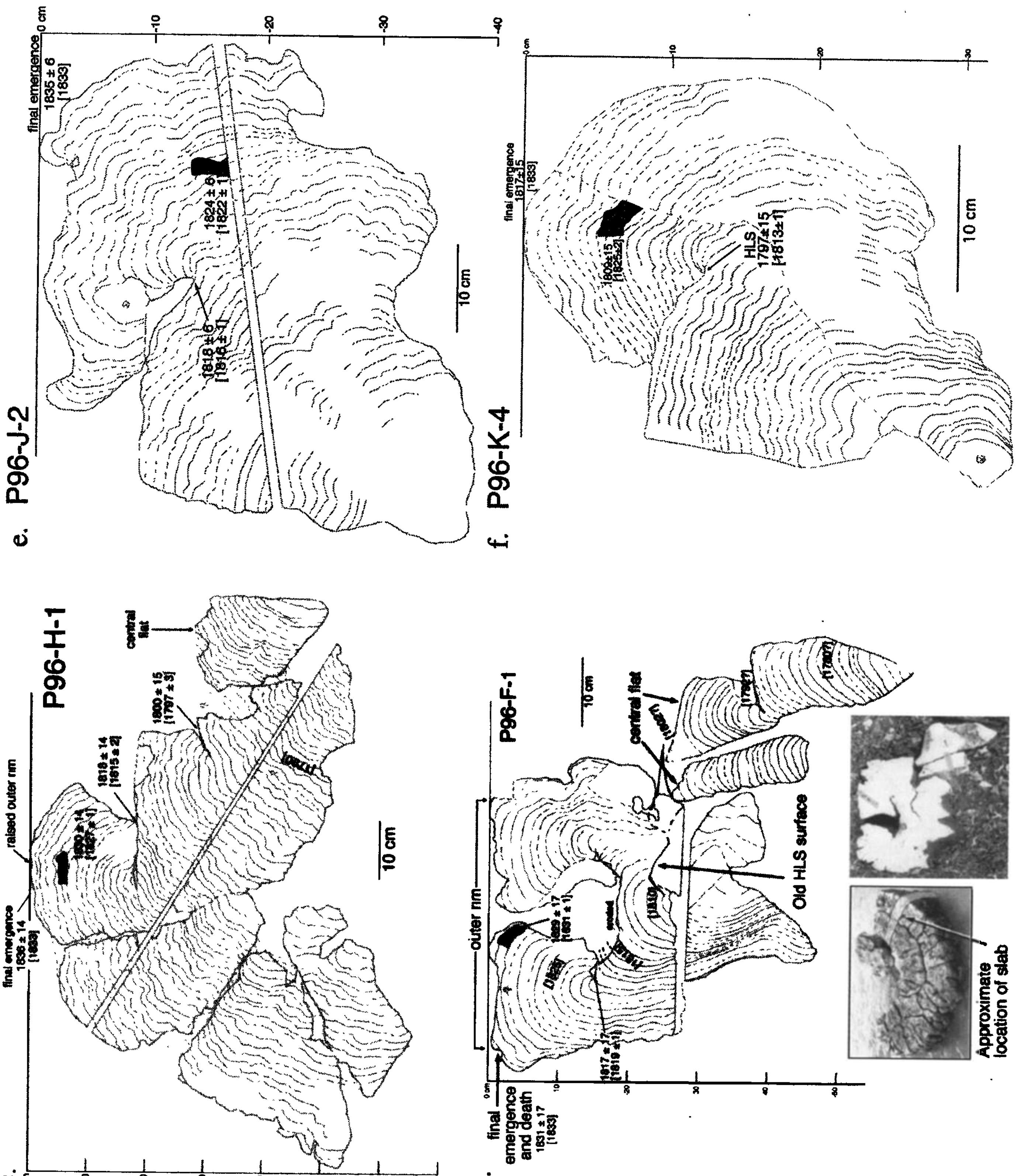


Figure 13. (continued)

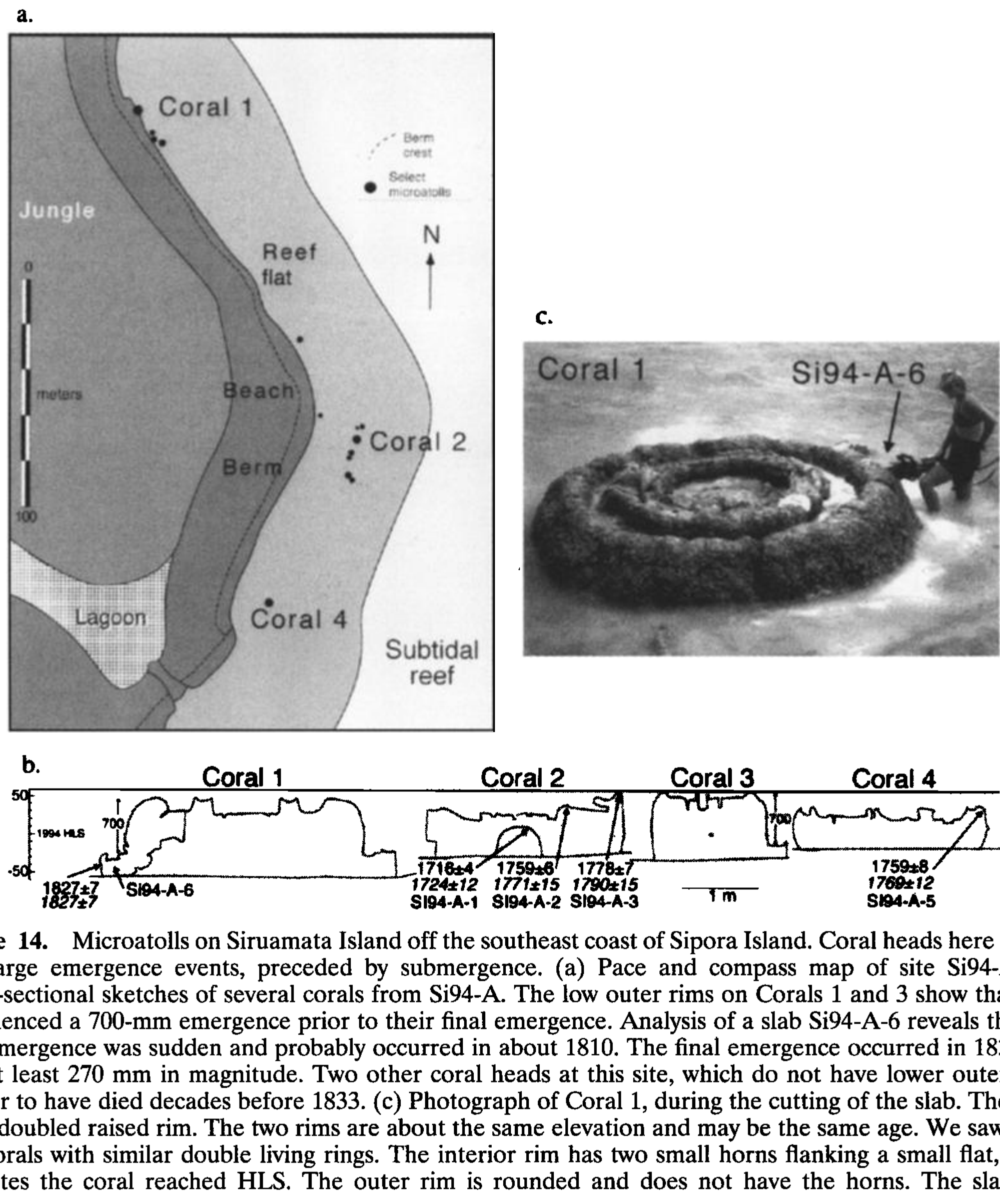


Figure 14. Microatolls on Siruamata Island off the southeast coast of Sipora Island. Coral heads here record two large emergence events, preceded by submergence. (a) Pace and compass map of site Si94-A. (b) Cross-sectional sketches of several corals from Si94-A. The low outer rims on Corals 1 and 3 show that they experienced a 700-mm emergence prior to their final emergence. Analysis of a slab Si94-A-6 reveals that the first emergence was sudden and probably occurred in about 1810. The final emergence occurred in 1833 and was at least 270 mm in magnitude. Two other coral heads at this site, which do not have lower outer rims, appear to have died decades before 1833. (c) Photograph of Coral 1, during the cutting of the slab. The coral has a doubled raised rim. The two rims are about the same elevation and may be the same age. We saw many live corals with similar double living rings. The interior rim has two small horns flanking a small flat, which indicates the coral reached HLS. The outer rim is rounded and does not have the horns. The slab only included the outer rim and a piece of the flat surface between the two rims.

3.1. Pre-1833 Submergence

Figure 16 displays the history of HLS fluctuation for each of the seven coral slabs. The graphs appear roughly in geographic order, north to south and west to east (compare with Figure 4). In each of the seven plots, HLS elevations are shown relative to the top of the coral in 1833. Each point represents 1 year. A dot indicates that an HLS impingement occurred that year. An arrow indicates that upward growth was not impeded by HLS that year. The base of the arrow represents the elevation of the top of the year's annual band. So, HLS for a year shown with an arrow would have been an indeterminate amount above the base of the arrow.

All seven curves show a predominance of submergence in the decades prior to 1833. Average rates of submergence range from about 5 to 11 mm/yr. The rates shown were derived by least squares fit of the HLS clip points (Table 2) [*Zachariasen*, 1998]. The two sites farthest from the trench, off the southeast coast of South Pagai, are at the lower limit of this range in pre-1833 rates. Other, unslabbed corals from southeastern South Pagai have more subdued raised rims than corals farther north and west. This supports the contention that southeastern South Pagai was submerging more slowly than elsewhere prior to 1833. The two curves from the west coast of North Pagai are at the upper end of the range of submergence rates. Thus the rates appear to vary regionally. Furthermore, all values are higher than 2 mm/yr, the current global average rate of sea level rise due to hydro-isostasy [*Peltier and Tushingham*, 1991].

3.1.1. Gradual or episodic? In all of the records, episodes of HLS impingement 1–8 years long punctuate longer periods of unimpeded upward growth. During most of the multiyear impingements, HLS does not vary more than a few tens of millimeters. Two- to 8-year periods of nearly stable HLS, separated by unimpeded upward growth, warrant the conclusion that there is episodicity in the submergence. The magnitude of the submergence episodes is 50–200 mm. We see this in the records of submerging modern microatolls as well [*Zachariasen*, 1998; J. Zachariasen et al., submitted paper, 1998]. Thus

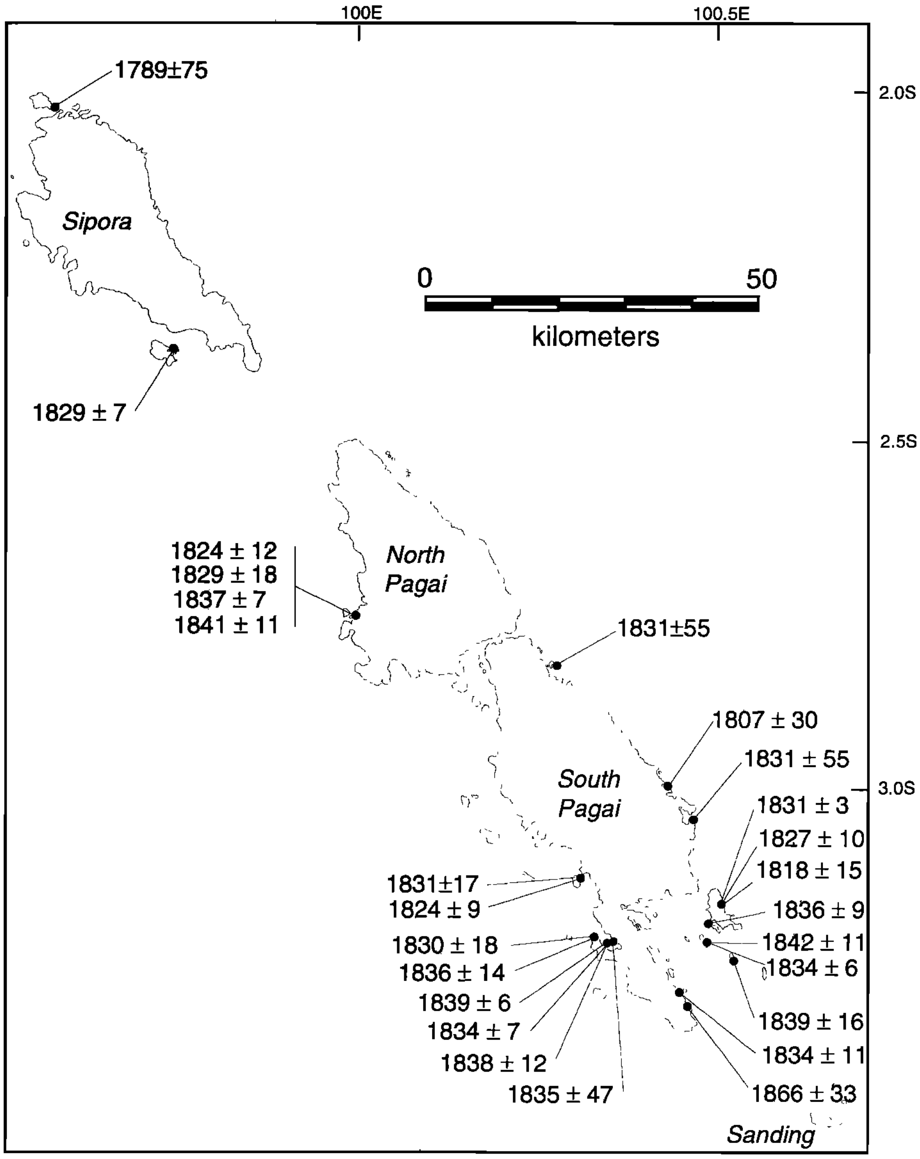


Figure 15. Map of sites with dated corals that died in about 1833. The dates of death have been corrected for initial ^{230}Th contamination. The geographical distribution of these samples suggests that the 180 km of outer-arc ridge represented by the Pagai Islands and Sipora Islands experienced substantial uplift during the 1833 earthquake.

it appears that periods of HLS stability are interrupted by periods of submergence.

But what is the nature of these long periods that lack HLS clips? One possibility is that these represent periods during which the rate of submergence was constant and just rapid enough to keep annual lowest tide levels from dropping below the top of the microatoll. At the other extreme would be sudden submergence events of several tens to hundreds of millimeters at the beginning of each period. In this case, many years of upward growth would be required before the top of the microatoll reached the level of the new HLS and recorded the next HLS clip. Discriminating between these two possibilities is important because it is vital to know if strain accumulation and relief above the subduction interface is continuous, episodic, or some combination of the two. Unfortunately, as we will see, it is difficult to argue strongly from our data for either case.

Gradual submergence appears plausible, given the high average rates of submergence. Recall from Figure 3c that a constant rate of submergence of 7 mm/yr limited HLS impingements for the Cocos Island record to just two in 10 years. Thus one might argue that the numerous long periods that lack HLS clips in Figure 16 are the result of constant or nearly constant submergence. Further support for an hypothesis of gradual submergence comes from the one site at which HLS clips are closely spaced. At site NP94-A the combined record from the two slabs shows HLS clips in 1810, 1814–1816, 1819, 1826, 1829, and 1830–1833 (Figure 16). Six years is the longest period without an HLS clip.

Weak evidence against gradual submergence events is the long duration of many of the periods that are free of HLS clips. In several cases, these are a decade or longer. The typical decadal record of lowest low water would have to be different from the hypothetical record in Figure 3c to produce such long periods without HLS clips.

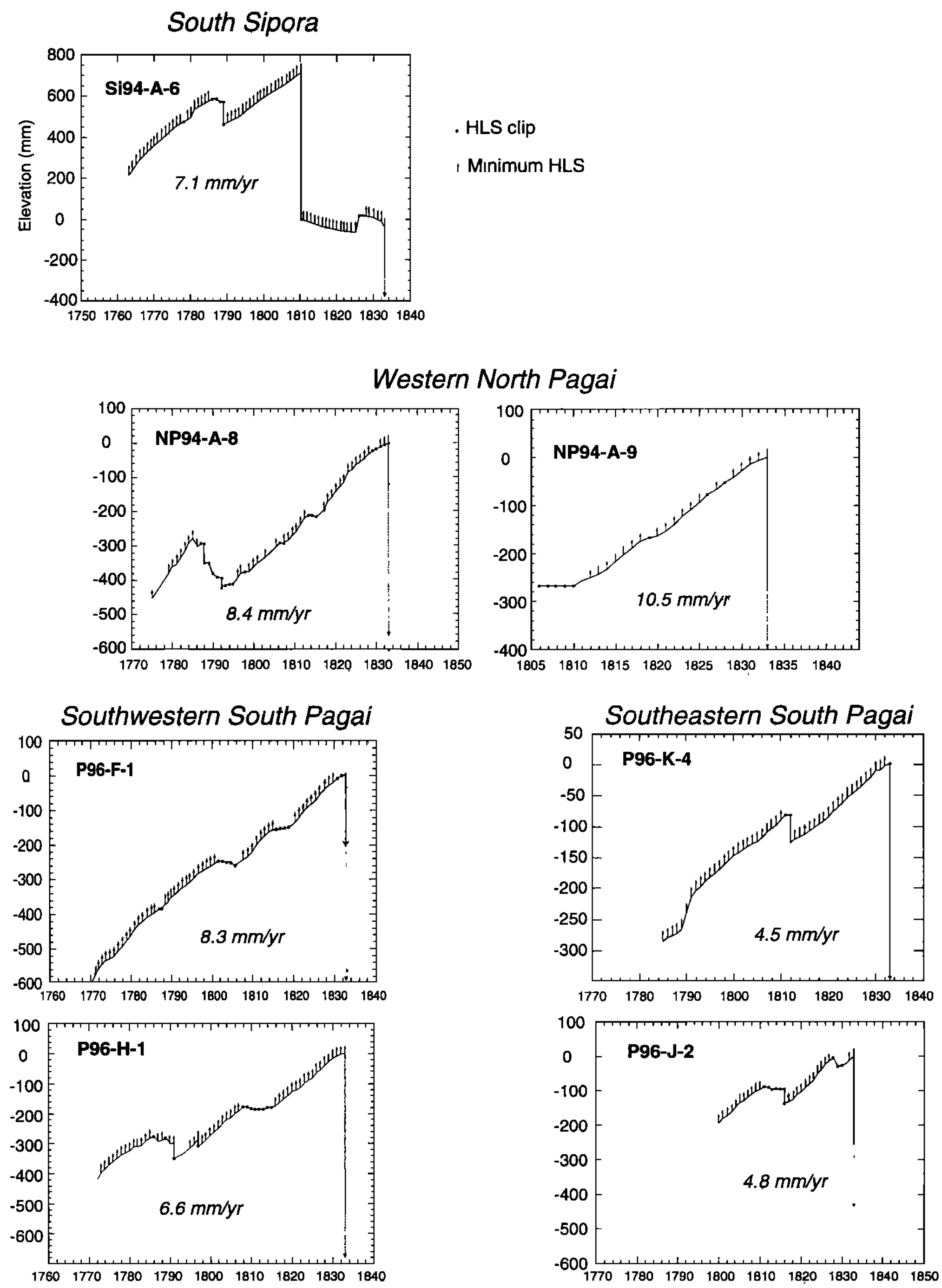


Figure 16. Graphs of HLS elevation versus time for the seven corals that died from emergence in about 1833. Dots mark the elevations of HLS surfaces or "clips." Arrows indicate lowest limit for HLS in years in which the rings were unimpeded by HLS. In these years, upward growth was limited by their natural growth rate. Thus HLS must have been somewhere above the tops of the rings. Downward pointing arrows at the end of each curve indicate that the measured total coseismic uplift is greater than or equal to the vertical extent of the youngest annual band, since the coral head emerged completely. We have assumed an 1833 death date for the corals.

Our data give us no strong preference for either gradual or sudden submergence producing these long periods of unimpeded upward growth. Since the response of microatolls to large emergence events is more rapid than their response to submergence events, this question may need to be resolved in an environment of interseismic emergence rather than submergence.

3.1.2. Poor correlation of HLS clips between slabs and sites. If the various episodes of HLS stability and submergence were caused by tectonic processes, one might expect they would correlate from site to site. In fact, the correlation is poor even among slabs from the same site.

Figure 17 facilitates comparison of the seven records by displaying them with a common time axis. The numerals at

Table 2. Preseismic Submergence Rates and Magnitude of Coseismic Emergence Derived From Coral HLS Records of Seven Slabbed Microatolls From Six Sites on the Coasts of the Sipora and Pagai Islands

Sample	Distance From Trench, km	Preseismic Submergence Rate, mm/yr	Minimum Coseismic Emergence, mm	Coral Height, mm (Minimum)	Last 2 Decades Preseismic Submergence Rate (and Year Beginning)
P96-J-2	122.6	4.8	200	400	0.81 (1816)
P96-K-4	128.2	4.5	50	330	0.60 (1812)
P96-H-1	109.4	6.6	100	1260	1.12 (1815)
P96-F-1	113.4	8.3	170	650	1.13 (1819)
NP94-A-9	112.1	10.5	330	370	1.18 (1816)
NP94-A-8	112.1	8.4	130	700	1.41 (1818)
Si94-A-6	117.3	7.1*	700	970	1.21 (1789)

All seven corals probably died as a result of emergence in the 1833 giant earthquake. Table indicates distance from each site to the trench, the preseismic submergence rate calculated from the slope of the best fit line through HLS-elevation points, minimum coseismic emergence determined from the slabs, and height above the substrate of the emerged corals.
*For the period prior to the 700-mm emergence event in about 1810.

various points below each curve are estimates of the uncertainty in age due solely to ring counting ambiguity. For example, in slab P96-H-1 the uncertainty in counting rings from 1833 to about 1810 is about 2 years and to about 1790 is about 3 years. The bar and query at about 1790 indicate that the older part of the slab is not physically connected with the younger part of the slab, and so the ages of its rings can be assigned only tentatively and could be in error by many years.

A comparison of the two records from NP94-A shows that even for microatolls only 50 m apart (Figure 9a), HLS impingements do not occur typically in the same year. This suggests that nontectonic processes have produced the second-order details superimposed upon the overall record of submergence. These processes must be very site-specific; perhaps they include such factors as differing response time to low water levels and differing vulnerability to low water due to variety in species and varying exposure to wave action and sunlight. Because of these puzzling nontectonic, local influences on the timing of small HLS clips, it appears that in our search for tectonic phenomena we should focus primarily on the larger, more prominent deviations in the microatoll histories.

3.2. Pre-1833 Emergence Events

The vertical lines in Figure 17 indicate the dates of the two largest emergence events before 1833. These emergences occurred in about 1790 and about 1810. We have described each of these above, in the presentation of individual slabs. Now, we consider possible correlation of these events between slabs.

3.2.1. Emergence in about 1790. Slab Si94-A-6 from southern Sipora displays clear evidence of an emergence episode within a couple years of 1790. Uncertainties in ring-counting back to this event from 1833 are only about 2 years. The fact that this event was preceded by several years of stable HLS limits the size of the emergence event at this site to about 100 mm (Figure 17). That is to say, this HLS clip cannot represent a larger emergence event that brought the top of the head 100 mm above HLS. For this reason, we cannot correlate this 100-mm emergence event with the emergence event that killed Corals 2 and 4 on the same reef. The difficulty with this hypothesis is that HLS prior to the death of these corals was near the top of corals 2 and 4, about 400 mm above the stable HLS documented in Si94-A-6 in the 4 years prior to its emergence event. More plausible, but still very tenuous, would be the suggestion that the event that killed the higher corals involved a 400-mm drop in HLS from the top of Coral 2 to the 1779 or 1786 HLS of Si94-A-6.

NP94-A-8 is from a reef about 60 km southeast of Si94-A-6, on the western coast of North Pagai. It also shows a prominent emergence event in the late 1780s or early 1790s. As we explained in the preceding section, this emergence could have been sudden, sometime in the early 1790s or gradual over several years in the late 1780s and 1790s. Recall that the record in NP94-A-8 may reflect the lower 140 mm of a 400-mm emergence that is dated in three unslabbed heads from the same site. It is tempting to associate this event or these events on North Pagai with the possible 400-mm event in 1779 or 1786 on southern Sipora or the 100-mm emergence there in about 1790.

A 50-mm emergence event in P96-H-1, about 60 km still farther southeast on the southwestern coast of South Pagai, might be contemporaneous with these two larger events. The existence of this event is less certain, though, because the coral slab has a discontinuity in the vicinity of the annual bands of the late 1700s. Also, any regional emergence of the islands off the southeast coast of South Pagai would have to be very small to avoid being expressed in slab P96-F-1.

We are not confident that any of the emergence events less than or equal to about 100 mm are contemporaneous or that their cause is tectonic. Emergences of this magnitude can certainly be caused by fluctuations in lowest tides (see, for example, the 1995 diedown in our hypothetical coral in Figure 3). Nonetheless, the records do suggest the possibility of an event. The 400-mm events on southern Pagai and western North Pagai are likely tectonic in origin, but our lack of details about these late 18th century events precludes further analysis or interpretation.

Newcomb and McCann's [1987] catalog of historical earthquakes includes a large earthquake in 1797, probably near the Pagais. This event generated tsunamis in the Batu Islands and on the west coast of Sumatra just north of Padang, across from Sipora (Figure 1). This could be the 1790s event recorded at NP94-A and Si94-A, but only if our counting of annual bands is incorrect. In both cases there appear to be more than 36 annual growth bands between the 1780s/1790s disturbance and the outer (1833) edge. Therefore the event (or events) probably occurred several years before 1797. If this event(s) did not occur in 1797, then the 1797 earthquake was not associated with a significant emergence in the Pagai and Sipora Islands.

3.2.2. Emergence in about 1810. Corals at site Si94-A off the southern coast of Sipora document a sudden 700-mm emergence in about 1810 (Figure 17). Both the magnitude and the date of the emergence are well constrained. Only one microatoll (P96-K-4) has an emergence that is temporally close—a 50-mm event in about 1813. One or two multiyear HLS flats occur in all of the other slabbed microatolls within a couple years of 1810. Association of the beginning of the multiyear HLS impingements with the large emergence is unlikely. It would be quite fortuitous for an emergence event to bring several microatoll tops up precisely to HLS, rather than to overshoot and cause a larger die down. Association of the end of a multiyear HLS impingement would also be difficult to explain.

The lack of disturbance in any of the slabs from central North Pagai to southern South Pagai suggests that the 1810 event was restricted to an area to the north of these sites. Furthermore, the lack of a large earthquake or tsunami in the historical record of about 1810 suggests that the emergence occurred aseismically. The collection and analysis of slabs from sites near southern Sipora could constrain the dimensions of the uplifted region and lead to a better understanding of this mysterious event.

3.3. Giant Earthquake of 1833

3.3.1. Magnitude of emergence in 1833. Thus far, we have collected slabs from microatolls whose tops were at or near HLS just before the 1833 earthquake. The vertical range spanned by the 1833 growth band on the perimeter of the seven slabs ranges up to 290 mm (Figures 11, 13, and 17). This is a minimum measure of the magnitude of 1833 emergence. None of these slabs have low post-1833 skirts around their perimeter. And so we conclude that all were raised so far into the intertidal zone that even their bases were brought completely above the HLS. Thus none record the full magnitude of the coseseismic step. The fact, however, that some of the microatolls that died in 1833 are more than a meter tall suggests that the magnitude of emergence was a meter or more, at least locally. Even so, precise measurement of the coseismic step must await collection and analysis of microatolls that rose only partially above their HLS in 1833 (such as the one shown in Figure 6).

In lieu of direct measurements of the complete 1833 coseismic step, we can construct reasonable estimates by cobbling together the HLS records from the seven 1833 vintage microatolls and those of nearby living corals. Figure 18 displays the relevant modern records alongside the 1833 records. We have both modern and 1833 records from only two sites, P96-J and P96-K, off the southeastern coast of South Pagai. Note that the records of pre-1833 submergence and modern submergence are very similar in style and rate.

In Figure 19 the modern records and the older records appear together, with their elevations plotted in relation to 1996 HLS. For sites P96-J and K, we estimate the 1833 coseismic step by extrapolating the average modern rate of submergence linearly back to 1833. The estimated magnitudes of coseismic uplift at the two sites are about 1.1 and 1.0 m. Uncertainty in the modern rate [*Zachariasen*, 1998] leads to an uncertainty of about 0.2 m in these values.

For the other five 1833 records, we have no modern records at the same site. The nearest modern record is 20–70 km away, at site P96-B-2, at the western entrance to the strait between

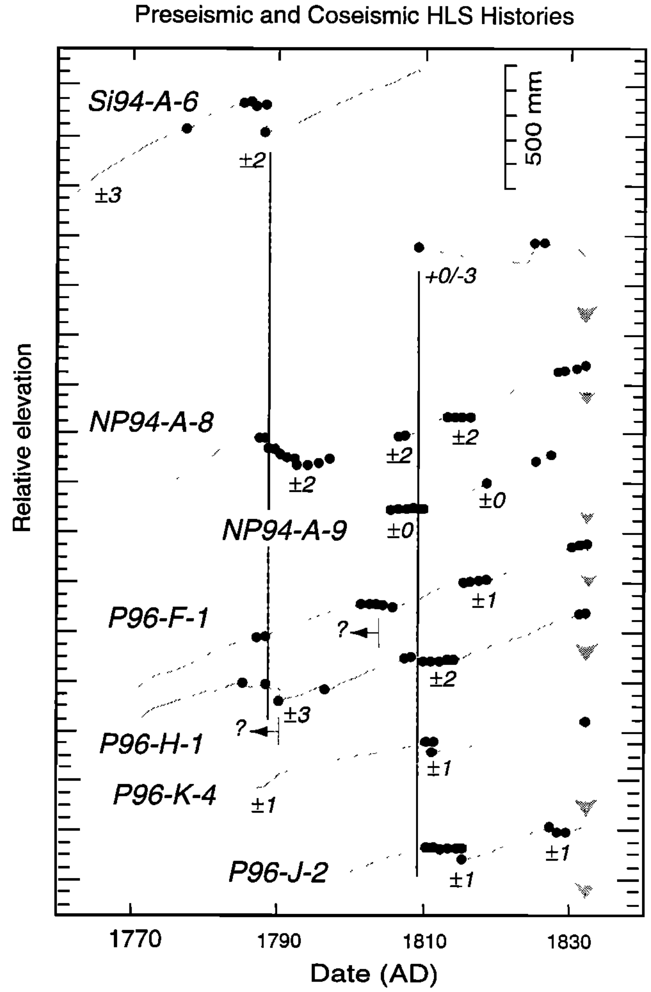


Figure 17. All seven graphs from Figure 16 plotted with a common time axis. Dots indicate HLS clips seen in the annual growth-band patterns, and gray lines pass through minimum HLS elevations. Vertical lines represent possible correlations of pre-1833 events.

North and South Pagai (Figure 4). Despite the great separation of this modern record from the five other 1833 records, its distance from the trench is similar to that of these other sites (Table 2), and its rate of submergence is similar. Since the pre-1833 and modern rates are nearly identical at P96-K and J, where we have colocated data, and since the modern rate from P96-B-2 is similar to the rates from the four other pre-1833 sites, we assume the modern rate is similar to the pre-1833 rate.

Thus for Si94-A and P96-H we use a rate of 7 ± 2 mm/yr. And for P96F and NP94-A we use 7.5 ± 2.5 mm/yr. The estimated coseismic steps in 1833 at these other sites range between about 1.6 and 1.9 m. Since we have used uncertainties of about 2–2.5 mm/yr for the 163 years since the event, the uncertainties in the estimated coseismic steps based on this extrapolation are about 0.35 m. Adding the uncertainties in HLS elevations in the microatolls increases the uncertainty in the coseismic step by only about 40 mm [*Zachariasen*, 1998; J. Zachariasen et al., submitted paper, 1998].

The coseismic emergence values that we have calculated are only rough estimates of coseismic emergence. It is quite likely that rates of submergence were not constant in the 163 years

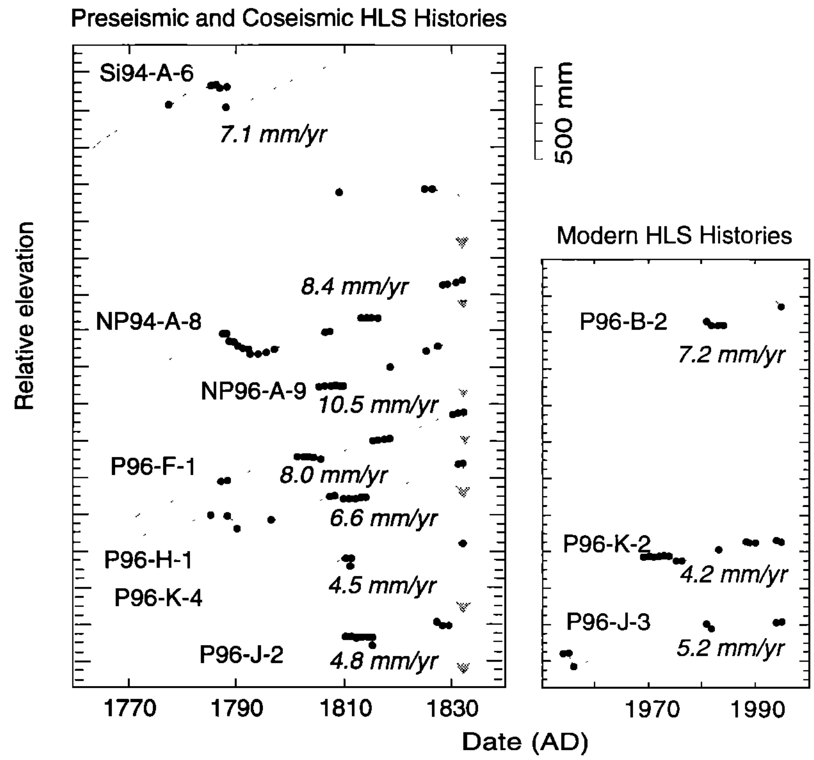


Figure 18. Comparison of modern and early 19th century HLS histories. Only sites P96-K and J have records from both the older and younger periods. For the remaining older sites, comparisons must be made with sites tens of kilometers away but at a similar distance from the trench. Site P96-B-2 is on a small island at the western entrance to the strait between North and South Pagai.

between 1833 and 1996. In fact, rates of submergence in the two decades prior to 1833 appear to have been higher than rates averaged over the several decades prior to the earthquake (see last column in Table 2). *Savage and Thatcher* [1992] found the contribution of the transient postseismic deformation phase in Japan was about one fourth of the total interseismic deformation through an earthquake cycle. If a similar postseismic phase occurred here too, the coseismic uplifts might be 250–450 mm higher than our linear extrapolation of the modern submergence rates has yielded.

3.3.2. Magnitude of slip on the subduction interface in 1833. We have used the estimates of coseismic emergence to approximate the magnitude of coseismic slip along the subduction interface. Following *Savage*'s [1983] method, we constructed simple elastic dislocation models that were appropriate for the geometry of the Sumatran subduction zone. The dip of the subduction interface beneath the islands is constrained to less than 15° by earthquake hypocenters and moment-tensor solutions [*Newcomb and McCann*, 1987; *Fauzi et al.*, 1996; Harvard CMT catalog]. The distance from the sites to the trench axis ranges between 109 and 128 km (Table 2 and Figure 1).

Figure 20 shows a profile of uplift associated with a model that fits our data well. In this case, we impose reverse slip on a 12°-dipping plane that intersects Earth's surface at the trench. Slip is a uniform 13 m from the trench to a point 175 km downdip. This downdip limit would be at a depth of about 40 km beneath the forearc basin, about where the eastern limit of the 1833 rupture plane is drawn on Figure 1. The subduction interface would be about 25 km beneath our sites.

The magnitude of emergence and its landward decrease (Figure 20) place important constraints on the magnitude of slip on the interface and its downdip extent. Greater amounts of slip would produce a steeper curve, and moving the downdip limit of slip would move the landward dipping portion of the curve either trenchward or landward. The uncertainties in the emergence estimates do not warrant more complicated models.

3.3.3. Magnitude of the 1833 earthquake. Our data indicate that uplift was associated with the 1833 Sumatran earthquake along at least a 180-km length of the outer-arc ridge, encompassing the islands of Sipora, North Pagai, and South Pagai. A dislocation on the subduction interface with a downdip width of 175 km and 13 m of slip fits the emergence data well. From these source parameters, we can calculate a provisional minimum moment magnitude for the earthquake, according to the equations

$$M_0 = \mu A D \tag{1}$$

$$M_w = (\log M_0 - 16.05)/1.5 \tag{2}$$

where M_0 is the seismic moment, $\mu = 5 \times 10^{11}$ dyn cm, A is the area of the fault plane, D is the average slip on the fault plane, and M_w is the moment magnitude. The minimum moment magnitude (M_w) thus derived is 8.8. If we assume that the extent of the source was actually 550 km, the distance *Newcomb and McCann* [1987] inferred from their analysis of historical reports and interpretation of geologic structure, the moment magnitude (M_w) rises to 9.2. The reason that this estimate is significantly greater than the M_w = 8.7–8.8 estimated by *Newcomb and McCann* [1987] is that our data suggest a downdip width that is 75 km longer and an average slip about 2 times greater than they used. A more reliable estimate of

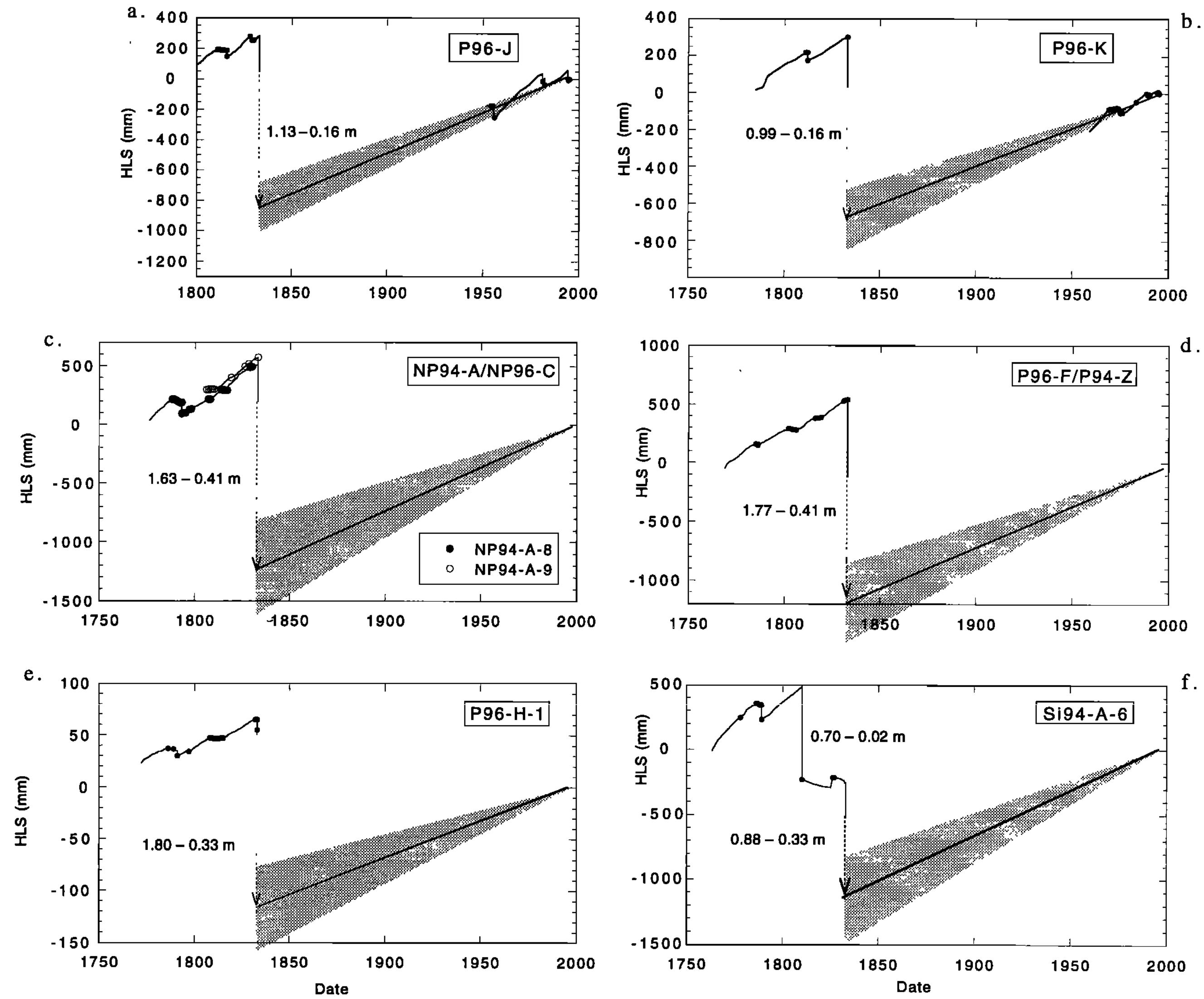


Figure 19. We have estimated the 1833 coseismic emergence by linear extrapolation of the modern submergence rates back to 1833. These estimates vary from about 1 to 2 m among the six sites. At sites P96-J and P96-K, we have slabs from both living and dead microatolls, so records of both are plotted with respect to 1996 HLS. At the other four sites, we have no modern slabs, so the modern submergence rate is assumed to be about the same as the averaged pre-1833 rate and modern values from sites more than 20 km away [*Zachariasen*, 1998; J. Zachariasen et al., submitted paper, 1998]. If one assumes a postseismic transient similar to those that have been observed after some recent subduction earthquakes, coseismic emergence would be higher.

magnitude will be possible if future work results in discovery of microatolls that emerged during the 1833 earthquake farther to the northwest and southeast.

3.4. Earthquake Cycle

3.4.1. Average recurrence time for nominal 1833 earthquakes. Lacking a more complete history of large events, it is common practice to estimate the time between large earthquakes by assuming that the most recent large earthquake is typical. This mean time between nominal, characteristic earthquakes is useful in evaluating seismic hazard [cf. *Yeats et al.*, 1997, chap. 13]. If one knows the amount of slip during the hypothetically characteristic event and the long-term slip rate of the fault, one can calculate a nominal recurrence time between large events.

Our estimate of slip in 1833 is consistent with recurrence of similar earthquakes about every 265 years. We derive this interval by dividing the rate of subduction into our estimate of slip in 1833. To estimate the rate of subduction, we begin with the relative plate velocity (67 mm/yr, N11°E) determined by Global Positioning System (GPS) surveys [*Tregoning et al.*, 1994]. From this, we must remove the slip vector of the Great Sumatran fault (Figure 21). This is about 11 mm/yr, S35°E [*Sieh et al.*, 1991]. The resultant vector for the subduction zone is 60 mm/yr, N18°E. The component of this vector that is perpendicular to the subduction zone is 49 mm/yr, N54°E. At a rate of accumulation of 49 mm/yr, 13 m of dip slip would occur, on average, about every 265 years. This average interval would be longer, of course, if a part of the accumulating strain were relieved by aseismic slip. Data from Sumatran GPS sur-

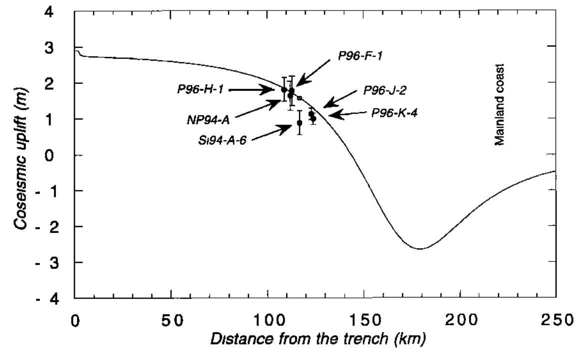


Figure 20. Slip of 13 m on the subduction interface from the trench to 175 km down dip produces a profile of coseismic uplift that is consistent with our estimates of coseismic emergence in 1833. The dip of the subduction interface in this model is 12°. Slip of this magnitude is consistent with an average recurrence of about 265 years for nominal 1833 earthquakes. Coseismic uplift estimates appear as solid circles. Open circle indicates where Si94-A-6 would plot if 700-mm emergence from 1810 event were included.

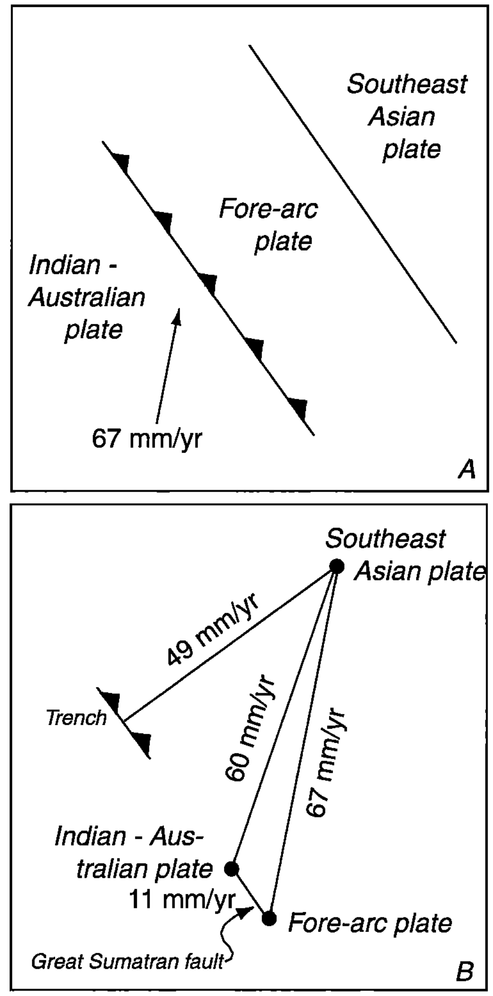


Figure 21. Calculations of the rate of slip on the subduction interface require knowledge of the rate of slip on the Great Sumatran fault, the rate of convergence between the Southeast Asian and Indian/Australian plates, and their geometries. (a) Map of the geometric relationships of the three tectonic plates and their bounding structures. (b) Vector diagram of the motions between the three plates. Convergence across the subduction zone is 60 mm/yr, N18°E. The component of this vector that is perpendicular to the trench axis is 49 mm/yr, N54°E.

veys suggest, however, that aseismic strain relief is not occurring [*Prawirodirdjo et al.*, 1997].

This hypothetical average interval of about 265 years may be close to the actual period of time between giant earthquakes along this portion of the Sumatran subduction zone. We did find a few fossil microatolls that emerged from below HLS in the mid-14th century and early 17th century [*Zachariasen*, 1998]. The two intervals separating these two events and the 1833 earthquake are both about 230 years. A paper in preparation will discuss the nature of these earlier events and will compare their preemergence record of submergence with both the pre-1833 and modern records.

3.4.2. Balance of interseismic submergence and coseismic emergence. *Zachariasen* [1998] used mid-Holocene microatolls within the intertidal zone to calculate long-term average rates of vertical deformation of the outer-arc islands. For the past several thousand years, these average rates of submergence and emergence have been between about 0.1 and 0.3 mm/yr, more than an order of magnitude lower than the rates of coseismic and interseismic deformation we have documented here.

The near stability of the outer-arc islands over the past several thousand years implies that coseismic and interseismic vertical deformation must be nearly equal but opposite in sign, at least when summed over many earthquake cycles. Thus interseismic submergence and coseismic emergence of the Sumatran outer-arc islands primarily result from the elastic accumulation and relief of strain in the hanging wall block of the subduction zone. A record of coseismic and interseismic deformations that spans several earthquake cycles would allow us to determine whether or not this balance is achieved over just one cycle. Fragments of such a lengthy record will be documented and discussed in a manuscript in preparation (J. Zachariasen et al., manuscript in preparation).

4. Conclusions

We have found that coral microatolls from the reefs of the outer-arc ridge of Sumatra contain a paleoseismic record of uplift associated with the giant subduction-zone earthquake of 1833. Furthermore, the microatolls contain a paleogeodetic record of rapid submergence during the several decades prior to the earthquake. Rates of preseismic submergence range from about 5 to 11 mm/yr and increase trenchward. Living microatolls from the same region have been submerging for the past few decades at similar rates and show a similar increase trenchward. These pre-1833 and modern records reflect patterns of strain accumulation above the portion of the subduction interface that failed during the 1833 earthquake and is currently locked. The similarity of the pre-1833 and modern rates and patterns suggests that processes of strain accumulation and relief are broadly similar through many earthquake cycles.

By plotting the pre-1833 and modern histories on the same graph and extrapolating the modern rates of submergence back to 1833, we have estimated the magnitude of uplift in 1833 at six sites above a 125-km length of the subduction zone. Values range from about 1 to 2 m. These may well underestimate the coseismic uplift by a few tens of percent because we have not included any postseismic relaxation in our estimates. Nevertheless, our estimates of coseismic uplift increase systematically trenchward and allow crude estimates of slip on the subduction interface during the 1833 event.

The pattern and magnitude of coseismic emergence is mimicked well by a uniform 13 m of reverse slip on a fault dipping 12° landward to a point 175 km from the trench, about 40 km beneath the forearc basin. Our estimate of the minimum moment magnitude (M_w) of this event is 8.8, assuming that only the portion of the subduction interface below our sites ruptured. A more likely moment magnitude, calculated using a greater along-strike length of the rupture, is 9.2. Recurrence of a nominal 1833 event about every 265 years would accommodate all of the 49 mm/yr of convergence across the subduction interface. This simple recurrence model may, in fact, be a good first approximation of reality: interpretation of data from the Sumatran GPS network suggests that the portion of the subduction interface that produced the 1833 earthquake is currently locked and has no component of aseismic slip. Furthermore, sparse evidence from older microatolls hints that each of the most recent two recurrence intervals have been about 230 years.

Coral microatolls are not perfect paleoseismic and paleogeodetic instruments. Attribution of emergence or submergence events smaller than about 100 mm to tectonic causes is tenuous because nontectonic causes can also produce effects of this magnitude. Identification of sudden submergence events is also difficult because of the time required for a microatoll to grow upward to the new level of HLS impingement. Furthermore, we have been unable to explain well the poor correlation of secondary features between individual heads and reefs.

Even though it is less than the imprecision in radiocarbon dating, imprecision of our ^{230}Th dates is large enough to create ambiguities in the correlation of events only a decade apart. For example, sites off the southwestern coast of Sipora Island and off the central western coast of North Pagai Island have microatolls that died or partially emerged in the late 1700s. The ^{230}Th dates of these microatolls are precise enough to allow us to say that none of the emergences coincide with the large historical, tsunamigenic earthquake of 1797. But the dates are too imprecise to enable unambiguous correlations between heads at the same site or between sites. Perhaps chemical fingerprinting of annual bands or work with more slabs will enable resolution of this problem.

Long, precise, and complete histories of sequential fault rupture are impossible to obtain. Exceptional histories, such as the record of great subduction events along the Nankai trough, Japan, and the southern coast of Chile, provide very precise dates of large earthquakes but leave large ambiguities in source parameters. Long paleoseismic records from sequences of clastic deposits, such as those along the San Andreas fault, are also problematic. Dating precision is seldom better than several decades, determination of magnitudes of slip is possible only at exceptional sites and requires time-consuming three-dimensional excavations. Long records of subduction-zone events from clastic deposits, such as those along the Cascadian coast, are also limited by the imprecision of radiocarbon dating and by the wide vertical ranges of the fossils and facies that are used to determine the magnitude of vertical deformation.

Microatolls appear to be exceptional natural instruments for extending the record of earthquakes and earthquake cycles into the preinstrumental past. In addition to allowing absolute dating of vertical deformations to within a decade or better, relative dating within one head can be precise to within a couple years or better. Microatolls contain records that enable dating and characterization of not only seismic events, but interseismic vertical deformation as well. In fact, the eventual construction of a record of vertical deformation through several cycles of strain accumulation and relief may well be possible using the microatolls above the Sumatran subduction zone.

Appendix: U-Th Disequilibrium Dating Methods

We prepared and analyzed samples according to *Chen and Wasserburg* [1981], *Chen et al.* [1986], *Edwards et al.* [1987b], *Edwards* [1988], and *Gallup* [1997]. The sample was dissolved in HNO_3, and U and Th spikes were added. The U spike was a double ^{233}U-^{236}U spike with $^{233}U/^{236}U \sim 1$, and a ^{233}U concentration of about 9.3 pmol/g. It was added so that the $^{235}U/^{233}U$ ratio was about 10. The Th spike was a single ^{229}Th spike, with a concentration of about 0.2 pmol/g. We usually added about 0.5–1 g of spike per gram of sample. The concentration of the uranium spike was periodically calculated by measuring mixtures of spike and gravimetric uranium standard (NBL 112a). The ^{229}Th spike was mixed with a gravimetric standard of Ames Th to determine concentration.

We used iron precipitation to extract the U and Th from the coral sample. The elution scheme to remove the U and Th from the iron precipitate was patterned after the methods of *Edwards* [1988] and *Gallup* [1997]. We loaded the samples onto zone-refined rhenium filaments. We loaded about 1/5 of the U fraction onto a rhenium filament and analyzed the uranium with the double-filament technique. Two-thirds of the Th fraction was loaded onto a rhenium filament, and either topped with or sandwiched between layers of colloidal graphite after the method of *Chen and Wasserburg* [1981]. Th was run using the single-filament tachnique.

The samples were analyzed on the Minnesota Isotope Laboratory's Finnigan-MAT 262-RPQ thermal ionization mass spectrometer. Both uranium and thorium were measured using the electron multiplier in pulse counting mode. Uranium data were gathered in peak-jumping mode, by sequentially measuring masses 233, 233.5, 234, 234.5, 235, 235.5, and 236. An exponential fractionation correction was applied to all ratios using the measured ^{233}U-^{236}U ratio in each scan. Mass 238 was not measured directly because the signal from ^{238}U was too large for the ion counter. We measured mass 235 and calculated 238 values by multiplying the measured 235 values by the natural $^{238}U/^{235}U$ ratio of 137.88 [*Garner et al.*, 1971]. We continued the analysis until we had obtained 2σ counting statistics of 1‰ on mass 234. The peak-jumping sequence for thorium runs was mass 229, 230, 230, 232. Thorium runs continued until the sample was gone and/or the filament burned through. Because most of the samples were young (<500 years), counting statistics for ^{230}Th generally yielded uncertainties of about 1%.

We used the measured isotopic ratios to calculate the concentrations of ^{234}U, ^{238}U, ^{230}Th, and ^{232}Th. All raw data were fully propagated through corrections for spike impurities, abundance sensitivity, dark noise, and filament and analytical blanks. The isotopic concentrations were in turn used to determine the age of the sample and its current and initial δ^{234}U values [*Kaufman and Broecker*, 1965; *Chen and Wasserburg*, 1981; *Edwards et al.*, 1987b]. The isotopic measurements obtained for all the analyzed samples appear in Table 3.1 of *Zachariasen* [1998].

In some cases, the sampled corals contain appreciable quantities of ^{232}Th. Since ^{232}Th is not a decay product, its presence indicates that the samples had significant initial thorium. Where we found significant quantities of ^{232}Th, we assumed that a component of the measured ^{230}Th also was incorporated initially and did not form from in situ decay in the coral. The ^{230}Th age calculation assumes that the initial ^{230}Th concentration was zero. Use of this equation will yield ^{230}Th ages that are older than the true age if significant amounts of ^{230}Th were, in fact, initially present. We corrected the ages for this initial ^{230}Th by multiplying the measured ^{232}Th concentration by the initial ^{230}Th/^{232}Th ratio. We estimated the initial ^{230}Th/^{232}Th ratio by (1) analyzing corals of known age and calculating the initial ^{230}Th/^{232}Th ratio from the measured uranium and thorium isotopic compositions and (2) analyzing subsamples of the same coral of the same age or known age difference, but with different ^{232}Th/^{238}U ratios. The initial ^{230}Th/^{232}Th ratio can be calculated from analyses of two subsamples given the assumption that both have the same initial ratio. Initial ^{230}Th/^{232}Th values for 20 such determinations ranged from 0 to 13×10^{-6}, with an average of 6×10^{-6}. We therefore use a value of initial ^{230}Th/^{232}Th = $6 +7/-6 \times 10^{-6}$ to correct for initial ^{230}Th. This value is slightly higher than the 4×10^{-6} value commonly used for such corrections. The commonly used value is calculated based on the assumption that the thorium is derived from material of average crustal composition at secular equilibrium [see *Stein et al.*, 1991; *Eisenhauer et al.*, 1993]. The value we use is more appropriate for our particular samples because our value was measured directly on local corals. Details about the correction are discussed in chapter 3 of *Zachariasen* [1998].

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