A reassessment of the 1907 Sumatra "tsunami earthquake" based on macroseismic, seismological and tsunami observations and modeling

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<<< [5] Caltech Affiliation has become spurious ? >>>>

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Note that "Auvergne" is part of the new name (hence no comma) of the University, which results from the 2017 merger of Université Blaise Pascal and Université d'Auvergne. The University is in the city of Clermont-Ferrand, the capital of the Département of Puy-de-Dôme, which used to be part of the Region "Auvergne", itself merged as of 2016 with the Region "Rhône-Alpes" into "Région Auvergne-Rhône-Alpes", presumably a temporary name for lack of a better one!

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probably not apparent in 1954. In those days, they had no clue as to the relation between their M_{PAS} and tsunamis - - -

Introduction

(2)

Numerous large earthquakes have occurred off the west coast of Sumatra, Indonesia, during historical and instrumental eras [*Newcomb and McCann,* 1987], and in particular, since 2000. These events include a tsunamigenic earthquake on 4 January 1907 near the islands of Simeulue and Nias **(Figure 1)**, ~~an event~~ which *Newcomb & McCann* (1987) associated with shaking of such severity that "*people on Nias could not stand*". The tsunami that struck both islands following this earthquake was responsible for at least 2,188 fatalities (*Koloniaal Verslag*, 1907-08) and led to the disaster being embodied in myth and legend on the island of Simeulue (e.g. *McAdoo et al.*, 2006; *Syafwina*, 2014; *Rahman et al.*, 2017).

No ¶

As discussed below, the 1907 tsunami was also recorded in the far field, as far away as Réunion Island [*Berho,* 1910], which gives the event a clearly anomalous character, in the context of the comparatively low "Pasadena" magnitude ($M_{PAS} = 7.6$) assigned to the earthquake by *Gutenberg and Richter* [1954], and long the only available measure of its size.

¶ More recently, *Kanamori et al.* (2010) conducted an extensive seismological study based on a number of historical seismograms. While they did not compute a seismic moment through waveform fitting, they measured a surface wave magnitude M_S 7.8 ±0.25 and estimated a moment magnitude M_W 7.8 by scaling time domain amplitudes of body and surface waves of the 1907 earthquake to those of modern earthquakes with known moment tensors. These results were obtained in the period range 40−50s, but *Kanamori et al.* (2010) stress that the source was obviously longer, and thus the moment should be larger at longer periods suggestive of a "*tsunami earthquake*".

The term "*tsunami earthquake*" was first used by (*Kanamori*, (1972) to discuss the Source ~~seismogenesis~~ of the 1896 Meiji Sanriku and ~~the~~ 1946 Unimak (Aleutian Islands) earthquakes, both of which resulted in anomalously large tsunamis with respect to their instrument~~ed~~al magnitudes. This type of event can be distinguished based on disproportionate between surface wave magnitude (M_S) and seismic moment (M_O) relationships, [0 (zero), not letter O] from the observation of longer than expected process times (τ) despite small rupture areas (*Sykes*, 1971; *Kanamori*, 1972; *Pelayo & Wiens*, 1992; *Polet & Kanamori*, 2000), from the lack of evidence for abnormal stress drops (*Pelayo & Wiens*, 1992), and from lower than anticipated macroseismic intensities (*Kanamori*, 1972; *Fukao*, 1979; *Bourgeois et al.*, 1999). The longer process times result in red-shifting of the source spectrum, and in inconsistencies between deficient seismic magnitudes measured at short-to-moderate periods (thus relevant to macroseismic effects) and enhanced ultra-long period seismic moments (controlling the generation of tsunamis). The ruptures of tsunami earthquakes have been observed to propagate toward the trench axis (*Polet & Kanamori*, 2000) on very shallow dipping faults or on splays (*Fukao*, 1979; *Pelayo & Wiens*, 1990, 1992) located in weakly coupled regions of aseismic convergence (*Pelayo & Wiens*, 1990, 1992; *Bourgeois et al.*, 1999), or at the very top of the plate interface under conditions of sediment starvation leading to a jagged rupture (*Tanioka et al.*, 1997). Rupture velocities for such earthquakes are also less than expected for typical earthquakes owing to the low rigidity of materials proximal to the trench axis (*Fukao*, 1979; *Pelayo & Wiens*, 1990, 1992; *Heinrich et al.*, 1998; *Ihmlé et al.*, 1998).

splays are by definition less shallow dipping

Enhanced tsunami generation following moderate-to-large earthquakes can also occasionally be traced to ancillary phenomena, such as triggered landslides [*Synolakis et al.*, 2002] or volcanic processes [*Satake and Kanamori*, 1991; *Fukao et al.*, 2018]. These are generally not labelled as "tsunami earthquakes".

4.

With the exception of the 2010 Mentawai ~~tsunami earthquake~~ event (e.g. *Newman et al.*, 2011a; *Hill et al.*, 2012), no other tsunami earthquakes ha~~s~~ve been conclusively identified off the Sumatran coast during the modern or historical period. Although *Kanamori et al.* (2010) concluded in more general terms that the 1907 Sumatra earthquake held all the hallmarks of a tsunami earthquake, key aspects of this event

Original text rather confused

remain unaddressed, including (i) conclusive, quantitative evidence of its nature as a "tsunami earthquake"; and (ii) an estimate of the geometry and slip parameters of the source supporting the reported distribution of the tsunami. In addition, perplexing observations clearly in need of further study include (iii) the anomalously violent reported ground motions [*Newcomb and McCann*, 1987; *Kanamori et al.*, 2010] in comparison with other tsunami earthquakes; (iv) the lack of reported aftershocks; and (v) the lack of reported land-level changes comparable to those identified for other earthquakes in the Simeulue-Nias region (e.g. *Meltzner et al.*, (2012),

In this article, we use a multidisciplinary approach to tackle these points. Through a scrutiny of original macroseismic reports and the systematic analysis of a number of seismograms, we separate the mainshock from a previously unsuspected large aftershock, occurring only 59 minutes later. We also identify two more instrumentally recorded aftershocks, and a large event on 02 November 1908, which however is probably unrelated. Waveform analysis of digitized seismograms allow the quantification of the mainshock at mantle wave periods up to 170 s, and further demonstrate significant differences in source spectra between the mainshock and the large 06:18 aftershock. We also compile~~d~~ a set of 76 qualitative and instrumental observations of the tsunami in the Indian Ocean basin and utilise~~d~~ a subset of these to prepare a slip model for the 1907 mainshock.

5.

Reassessing Macroseismic Intensities and the Discovery of Event II.

The earliest macroseismic study of the 1907 Sumatra earthquake is a dual-zoned intensity map (**Figure 2a**; *Newcomb & McCann,* 1987) utilising the Modified Mercalli intensity scale (MMI) to distinguish between what they defined as weak (MMI I - IV) and strong shaking (MM V-VII), for what was assumed to be ~~the mainshock~~ a single event. We use the 1998 European Macroseismic Scale (EMS-98; *Grünthal et al.*, 1998) to assess intensity with our choice of macroseismic intensity assessment adhering to modern practice (e.g. *Toppozada & Real,* 1981; *Ambraseys & Douglas,* 2004). The EMS-98 scale is equivalent to MMI (Musson *et al.*, 2010); its merits and its suitability for use outside Europe are discussed by *Martin & Hough* (2015; 2016). Our re-assessment of intensity focused on annual summaries of felt earthquakes in the Dutch East Indies (e.g. *Anonymous,* 1909) by the Royal Magnetic and Meteorological Observatory (nl: *Koninklijke Magnetisch en Meteorologisch Observatorium,* KMMO) that were culled from official correspondence. These were supplemented with further accounts extracted from colonial newspapers (see *Data & Resources*) published in Indonesia and in the Netherlands. In this and subsequent sections, non-English text appears in parenthesis (bh: Bahasa Indonesia, nl: Dutch; fr: French; de: German) accompanying the modern reformed Indonesian or English spelling, e.g., tsunami (nl: *vloedgolf*).

Grünthal alone in ref. list?

Note: up to know, you have not documented several events, so let us not talk yet of separate e.q.s.

6.

accurately associate reports with individual earthquakes.

Accurate time~~-~~keeping ~~was~~ is vital to ~~isolate the effects of separate earthquakes~~. In this respect, the KMMO summary was invaluable as it printed the standard time in Jakarta (known as Batavia Time or BT), which was 7h 7m ahead of Greenwich Mean Time (GMT), ~~if a~~ for seismic disturbances ~~was recorded instrumentally~~ recorded at Jakarta (nl: *Batavia*) or Bogor (nl: *Buitenzorg*). It also published the difference in time at places outside Jakarta with respect to Batavia Time. For example, BT was 36~~-~~minutes ahead of local time ~~on~~ at Gunungsitoli (nl: *Goenoeng Sitoli*) on Nias, and 17~~-~~minutes behind local time at Pacitan in central Java. Thus, local time was GMT + 06:31 at Gunungsitoli, GMT +07:07 at Jakarta, and GMT +07:24 at Pacitan.

A surprising result of our macroseismic re-analysis (**Tables S1-S2**) was the discovery of two distinctly separate earthquakes that were felt on Nias within an hour of each other on 4 January 1907 (**Figure 2b, 2c**). A magnetograph in Bogor (*Anonymous*, 1909) recorded the first at 12:29 BT (≈05:~~19~~ 22? GMT; henceforth Event I) and a second at 13:24 BT (≈06:~~18~~ 17 GMT; henceforth Event II). TO BE FIXED SEE P.6a Be consistent Engelbertus Schröder, the civil administrator (nl: *kontrolleur*) at Gunungsitoli on Nias (**Supplementary Material**) provided the strongest evidence with which to make this distinction; he described a weak but long duration earthquake at about midday, followed by three "flood waves" (nl: *vloedgolven*) at Toyolawa, which were later followed at 12:50 pm local time (06:19 GMT) by a second violent earthquake at Gunungsitoli (*Bataviaasch Nieuwsblad*, 22 March 1907). His account appeared in several newspapers including the

6a

These times do not make sense.

All relocations show Event I occurring around 05:19:12, which I suppose is what you write as GMT in parentheses?

Propagation to Batavia is about 03:34, which gives an arrival of 05:22:46 -- which fits well the time reported in the ISC (05:22:42). It does translate to 12:29:49 BT -- so far OK (assuming the time in parentheses is the origin time).

Now, for Event II. If you give an O.T. of 06:18 GMT, and allowing for about 30 s shorter travel time (if the earthquake is close to Nias), we would get an arrival time of 06:21 GMT, or 13:28 BT, not 13:24.

---> If it is 13:24 indeed (55 minutes later than for Event I), than the O.T. of Event II is about 55.5 minutes later than Event I's, or 06:14 (or 06:15) GMT

And then on Page 7, you state that Event II was at 12:50 in Gunugs... which, accounting for a few seconds of travel time, would be reconciled with 06:19 (or 06:18 GMT as O.T. of Event II). This would suggest 13:28 BT for the recording at Bogor (operating on BT).

At any rate, assuming 12:50 p.m. at Gun... this means a difference of O.T. of 59 minutes between the two events, not 50 minutes as mentioned on Page 7, Line 124.

Then, it is not consistent with the Figure at Manila which suggests 53 mn between I and II.

We need to be consistent on this throughout the paper...

7.

Nieuws van den Daag voor Nederlands-Indië (22 March 1907) which reported the time of the second earthquake to be "12:05." We believe this is a typesetting error as the identical account printed on the same day by the *Bataviaasch Nieuwsblad* (22 March 1907) states "12:50," and importantly 12:50 is also repeated by *Schröder* (1917a) himself. The occurrence of two felt earthquakes separated by 50 minutes is also corroborated by independent observers from Gunungsitoli (*De Padanger*, 26 January 1907), Natal, and Talu (nl: *Taloe*), and discussed below based on seismograms.

Also, n Note that *Masó* [1907] interpreted the Manila records as requiring two events, but pointed out the lack of evidence for Event II at Zikawei (Shanghai) or Europe; we believe this to be incorrect, as Event II appears in the Zikawei bulletin (*de Moidrey*, 1912), and in some European bulletins (*Conrad*, 1907; *Etzold*, 1907; *Harisch*, 1907; *Rudzki*, 1908; *Schwab*, 1909). 1907 in ref. list

The weak, long duration ground motions in the near field, and the barely perceptible shaking at larger distances associated with Event I are characteristic of tsunami earthquakes (e.g. *Kanamori*, 1972). In contrast, the descriptions of shaking and damage at Gunungsitoli from Event II (**Figure 2c; Table S2**) lead us to believe it was responsible for the violent shaking on Nias that was misconstrued as the mainshock by *Newcomb & McCann* (1987) and flagged as anomalous by *Kanamori et al.* (2010). Countless houses were destroyed elsewhere on Nias and even weeks later many people were living under temporary shelters (bh: *pondoks*) put up near their dwellings (*Bataviaasch Nieuwsblad*, 22 March 1907). Unusually, neither earthquake was reported in an official summary of news and affairs for January 1907 from Aceh (nl: *Atjeh*) i.e. "Atjeh-verslag over Januari" (*Java Bode*, 16 April 1907). These earthquakes also do~~es~~ not appear in catalogues of felt earthquakes in Thailand (*Nutalaya et al.*, 1985), nor were they reported from anywhere else in south-east Asia. ✓

8.

Location should go before moment / slowness.

Location of the Mainshock (Event I)

The earliest published locations for Event I come~~s~~ from *Szirtes* (1912) and *Turner et al.* (1912), ~~and~~ with phase arrival times from over 66 stations ~~for the event were~~ listed by *Szirtes* (1912). As summarized by *Kanamori et al.* (2010), and not surprising ly for an event at the beginning of the 20th century, epicentral estimates ~~for Event I~~ are poorly constrained and significantly scattered (**Figure 4**). *Gutenberg & Richter's* (1954) solution (2°N, 94 ½°E) is on the outer rise and ~~is~~ slightly west of the solution (2°N, 95°E) by *Turner et al.* (1912), but both are very unlikely locations for a tsunami earthquake. By contrast, *Szirtes* (1912) reports a location by T.H. Staverman (2°N, 96 ¼°E) in the immediate vicinity of the trench.

New ¶

In their recent ~~re~~compilation of relocated early instrumental earthquakes, *Storchak et al.* (2013) locate the event at (1.87°N, 94.21°E), ~~which is essentially~~ near the *Turner et al.* (1912) and *Gutenberg & Richter* (1954) epicentres, and with a remarkably small error ellipse (semi-axes: 28 km and 16 km).

The most recent version (5.0) of the ISC-GEM catalog moves the epicenter North by 61 km, to (2.422°N; 94.258°E); it is noteworthy that the confidence ellipse (again very small at 20 by 17 km) does not intersect its counterpart for *Storchak et al.*'s [2013] solution (Figure 4), thus casting doubt on their reliability since the two locations are supposed to be derived from the same algo-rithm. On the other hand, a systematic grid search (of the same dataset) confirms a large scatter of possible solutions, but proposes a best fitting epicentre at 2.67°N and 95.01°E (D. Di Giacomo, *personal communication*, 2017), 125 km NE and across the plate boundary from *Storchak et al.*'s [2013].

New ¶

Kanamori et al. (2010) relocated Event I based on travel times listed in Gutenberg's personal notepads (*Goodstein et al.*, 1980), with an emphasis on S – P times, to minimize errors due to clock uncertainties. ~~The epicenter which they obtained~~ Their solution (2.48°N,

italics

96.11°E; **Figure 4**) locates approximately 60 km closer to the trench than the 2004 and 2005 epicenters, at the seaward limit of the zone of interplate seismicity (*Pesicek et al.*, 2010), this configuration being reminiscent of the 2010 Mentawai tsunami earthquake (*Newman et al.*, 2011a; *Hill et al.*, 2012). However, the 95%-confidence ellipse from *Kanamori et al.* (2010), whose parameters are listed in their Appendix A (*Kanamori et al.*, 2010; p. 371), extends ~500 km, from the outer rise, beyond the *Gutenberg & Richter* (1954) and the *Storchak et al.* (2013) locations, all the way to inland Sumatra.

We ~~also~~ performed an independent relocation ~~from~~ using the dataset listed by the ISC and ~~using~~ the interactive method of *Wysession et al.* (1991) which includes a Monte Carlo algorithm injecting Gaussian noise in the data. For an event in the 1900s, we give the noise a standard deviation σ_G = 12.5 s.

Our solution converges to (2.24°N; 94.72°E; O.T. 05:19:13 GMT), a location on the outer rise, in the vicinity of the *Storchak et al.* (2013) and *Gutenberg & Richter* (1954) epicenters, but our confidence ellipse extends across the trench and grazes the solution from *Kanamori et al.* (2010); it is essentially contained inside these authors' ellipse.

New parag. ¶

While all these relocations have very large uncertainties (we regard the *Storchak et al.* (2013) confidence ellipse ~~to be~~ as deceptively small), the general pattern emerging (**Figure 4**) is that Event I was probably significantly displaced up-dip and trenchward with respect to the area of main seismogenesis at the interplate contact, to the general area located between the ISC grid-search solution and that of Staverman, in a tectonic environment reminiscent ~~to~~ of that of the Mentawai tsunami earthquake of 25 October 2010 (*Newman et al.*, 2011a; *Hill et al.*, 2012). This is our preferred location for the epicentre of Event I (2.5°N; 95.5°E), and ~~it~~ is indicated by a red star on **Figure 4**. Incidentally, the's preferred location ~~of our preferred epicentre for Event I~~ lies within a zone of low seismic productivity between 1918 and 2007, and adjacent to, but not within, the aftershock zones of the 2004 and 2005 earthquakes (*Engdahl et al.*, 2007; *Pesicek et al.*, 2010).

10.

Event II and later aftershocks

Our study is the first to identify several aftershocks associated with the 1907 Sumatran earthquake. For the newly discovered large earthquake (Event II) at ≈6:18 GMT, we were eager to find seismological evidence with which to quantify, and possibly locate it. Waveforms were far fewer than for the mainshock, but we identified it on ~~new~~ new records from Manila (*Masó*, 1907) and Shimla (*Patterson*, 1909), and on three previous records scrutinised by *Kanamori et al.* (2010). Station bulletins from Leipzig (*Etzold*, 1907), Manila (*Masó*, 1907), and Osaka (*Anonymous*, 1931) also list it: at the latter two, it appears in addition to the mainshock. (OK)

The paucity of instrumental records prevents an instrumental location for Event II, but an epicentral distance (5,420 km) to Osaka has been previously published (*Anonymous*, 1931).

(Unfortunately, this datum is of little help, since Osaka is at essentially equal distances (within 10 km) from the preferred locations of Events I and II.)

Theoretical travel times for P and S arrivals at Leipzig determined using the TauP module in Obspy (*Beyreuther et al.*, 2010) with the IASPEI 91 velocity model (*Kennett & Engdahl*, 1991) from an event originating at ≈6:18 GMT with a source near Nias (1°N; 97.5°E; $\Delta = 86°$) ~~coincide~~ agree with published phase arrivals (*Etzold*, 1907).

The proposed locations of Events I and II result in a difference of 285 km in epicentral distance and hence only 12 s in $S-P$ times, at Leipzig; this is probably resolvable in terms of relative times given then prevailing instrumental characteristics; however, in view of significant differences in their seismic source spectra (see below), the reported Leipzig arrivals can only be regarded as supporting, rather than constraining, a difference in location between the two events.

Finally, an epicenter in the region of Nias for Event II is further supported by the damage at Gunungsitoli (**Table S2**) and the destruction of houses elsewhere on Nias (*Bataviaasch Nieuwsblad*, 22 March 1907).

Lines 251–253 deleted (add nothing)

M.

Two strongly felt aftershocks (*Anonymous*, 1909) at ≈9:58 GMT on 4 January and ≈23:44 GMT on 5 January could be corelated with sparse regional and teleseismic recordings (*Levitski*, 1907; *Pechau*, 1907; *Geiger*, 1909; *Szirtes*, 1912).

In addition to the ~~events discussed,~~ above, nineteen felt aftershocks were counted on Simeulue on 4 January (*Bataviaasch Nieuwsblad*, 12 February 1907), ~~Earthquakes~~ and more were felt daily (*Nieuws van Den Dag*, 24 January 1907), with ~~further~~ rarer aftershocks experienced after 15 January 1907 (*De Sumatra Post*, 1 February 1907). Earthquakes were also felt at Sinabang on Simeulue at 23:45 LT on 19 January 1907 (*Bataviaasch Nieuwsblad*, 12 February 1907) and on 8 July 1907 (*Het Nieuws van Den Dag*, 20 September 1907), while on Nias, aftershocks were felt continuously until 17 January (*Bataviaasch Nieuwsblad*, 29 January 1907). Aftershocks later than 5 January 1907 were ~~largely~~ mostly only reported from Nias (*Anonymous*, 1909). Well-timed felt earthquakes are almost absent from Simeulue except for a shock that was quite heavy (nl: *vrij hevig*) at Sinabang around midday (≈12:42 BT) on 4 January 1907 (*Haagsche Courant*, 12 February 1907) without causing any damage (*De Sumatra Post*, 1 February 1907). We suspect this shock could be related to Event I or a strong local aftershock, neither of which we can definitively rule out.

12.

Later activity is difficult to limit to the Simeulue-Nias region, but we did find an event on 2 November 1908 which was reported severely felt on Simeulue where it caused liquefaction (*Deli Courant*, 23 November 1908). People fled to higher ground on Simeulue fearing a "*vloedgolf*" (*Deli Courant*, 23 November 1908). This shock was also felt in Aceh and northern Sumatra (*Anonymous*, 1910).

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While this earthquake was reported at (2°N; 97°E) by *Abe and Noguchi* [1981], it is not clear what these authors' source was (the event is not listed by *Gutenberg and Richter* [1954]). Rather, modern relocations place it at (7.96°S; 90.92°E; ISC-GEM Version 5.0) and (8.04°S; 90.91°E; [*D. DiGiacomo and E.R. Engdahl,* pers. comm., 2018]); our own relocation places it at (6.01°S; 91.17°E), with a large confidence ellipse grazing those relocations, but not reaching *Abe and Noguchi*'s [1981] reported epicenter. These proposed locations are in the Wharton Basin, ~ 1300 km to the Southwest of the 1907 event, and the 1908 event is most probably unrelated to it. Rather, it falls in the general area of strong seismicity in the vicinity of Ninetyeast Ridge, originally indentified by *Stein and Okal* [1978], and expressing the diffuse boundary between the Indian and Australian plates [*Wiens et al.,* 1985]. It remains remarkable that it was felt in Aceh, at a distance of ~1500 km, suggesting a source particularly rich in high frequencies.

[1983]

I have the GTT recs. But they are difficult to interpret – NO good surface waves; NO shortperiods (missing)

Stay tuned ?

12.5

Conventional magnitude estimates

Estimates of conventional magnitudes for Event I have been offered by *Gutenberg and Richter* [1954] ($M_{PAS} = 7.6$), *Abe and Noguchi* [1983] ($M_s = 7.6$), and recomputed by the ISC [*Storchak et al.,* 2013] ($M_s = 7.8$). *Duda* [1965] proposed a figure of 7.8, the nature of his magnitude being unspecified. *Kanamori et al.* [2010] estimated M_s by comparing 20–s amplitudes on records available for 1907 to those of records simulated on historical instruments for modern earthquakes with abundant determinations of M_s, and concluded that the 1907 shock probably had $M_s = 7.8 \pm 0.25$. In addition, a review of amplitude data available from undamped Milne instruments (*Levitski,* 1909; *Turner & Milne,* 1908a, 1908b; 1909, 1910) in the manner of *Abe & Noguchi* (1983) and *Abe* (1988) yields a Milne surface-wave magnitude $M_M = 8.0 \pm 0.23$ (one–σ) which, while higher than previous estimates of M_s [*e.g., Gutenberg and Richter,* 1954; *Abe and Noguchi,* 1983], is in the same range as proposed by *Kanamori et al.* [2010].

In the case of Event II, we recomputed conventional magnitude estimates from those reports of ground amplitude for which associations could be made, using established formulæ and guidelines [*Kárník et al.,* 1962; {*Vaněk et al.,* 1962}; *Ambraseys and Douglas,* 2000]. Our best estimate of M_s is 7¼, which represents an average for Leipzig (7.5), Manila (7.0) and Osaka (7.2).

see comment p.13a
which one do we use?

In the case of the aftershocks of 04 January (09:58 GMT) and 05 January (23:44 GMT), we estimated $M_s = 5.8$ for the first one, and a "Milne magnitude" (obtained from undamped Milne records and ajusted by *Abe* [1988] to mimic M_s in the range 5.5–6.0 for both events; however, we emphasize that these values are very poorly constrained, and probably associated with large uncertainties

<<< HOW LARGE EXACTLY ? >>>

How?

As for the earthquake of 02 November 1908 near the Ninetyeast Ridge, the ISC-GEM catalog (Version 5.0) proposes an unspecified magnitude of 6.7. We were able to estimate $M_s = 6.4 \pm 0.3$ (one-σ). We note that *Abe and Noguchi* [1983] proposed a much larger value of $M_s = 7.2$; however, this is based on *Duda*'s [1965] magnitude of 7.3, which *Abe and Noguchi* [1983] have commented, has a tendency to be biased by the use of body waves of unspecified, but presumably shorter, periods. If, as suggested by enhanced felt intensities at relatively large distances, the source spectrum of the 1908 earthquake was blue-shifted towards high frequencies, Duda's magnitudes would indeed be expected to overestimate M_s.

I have the GTT records, but this will take some time!

13a

The classical reference, which I had always quoted, for the Prague formula is

Váněk, J., A. Zátopek, V. Kárník, N.V. Kondorskaya, Yu.V. Riznichenko, E.F. Savarenskiĭ, S.L. Solov'ev, and N.V. Shebalin, Standardization of magnitude scales, *Izv. Akad. Nauk SSSR, Ser. Geofiz.*, **2,** 153–158, 1962.

It is very interesting that the text of the two papers is essentially the same. It was submitted as *Kárník et al.* to *Studia Geofisica et Geodaetica*, which was at the time a publication of the Czechoslovak of Sciences, with all authors listed alphabetically in the Czech (Latin) alphabet; and then to the *Izvestiya* of the USSR, with all authors listed alphabetically in Russian; this is how Váněk ends up first author, since V transcribes to B, the third letter of the Russian alphabet. Note also that the abstract in Russian p. 8 of the Kárník *et al.* paper lists the Russian authors alphabetically, then the Czech ones. Very interesting are the submission dates: 11 AUG 1961 for the Czech version, 31 OCT 1961 for the Russian one. So at face value, the Czech text pre-empts the Russian one.

I suppose one never stops learning something !!!

Reassessment of long-period seismic moment and source slowness

For Event I, *Kanamori et al.* [2010] inferred a seismic moment $M_0 \approx 6 \times 10^{27}$ dyn*cm at periods of ~50 s, through a comparison of time-domain amplitudes of body waves at Göttingen with those obtained at the nearby Black Forest Observatory (BFO) ~~station~~ during the earthquake of 02 November 2002. This method, which assumes the same focal geometry for both events (strike (ϕ) = 297°, dip (δ) = 16°, slip ~~angle~~ (λ) = 73°), is confined to the time domain and, as such, tacitly assumes that Event I had a source spectrum similar to that of the reference event, and hence followed scaling laws. However those are expected to be violated precisely by tsunami earthquakes, whose source spectrum is red-shifted towards lower frequencies with respect to a more traditional source, such as the 2002 event. For this reason, it is necessary to further explore the source spectrum of Event I in the frequency domain to constrain its seismic moment ~~value~~ at periods typical of mantle waves.

In addition to the waveforms interpreted by *Kanamori et al.* (2010), we revisited the record from Göttingen (GTT) in Germany, and located three previously unused waveforms from Pulkovo, Russia (*Golitsyn*, 1908), Shimla, India (*Patterson*, 1909), and Manila, Philippines (*Masó*, 1907). The Pulkovo record only provides a partial, single component record for Event I, but none for ~~and no~~ aftershocks (*Golitsyn*, 1908) while the Shimla seismogram clipped during Event I (*Patterson*, 1909) but recorded the aftershock (*see subsequent section*). The Manila record was written on a Vicentini mechanical seismometer with a period T = 1.4s, which makes it the only short-period recording available to us of the 1907 Sumatra earthquakes, and its crucial importance to our study is discussed in a subsequent section.

We use the first passages of Rayleigh (R_1) and Love (G_1) waves at Göttingen (GTT) for which precise metadata is available, with records digitized at a sampling rate of 1s. Our results are presented (**Figure 3**) in the mantle range of frequencies (6 – 10 mHz), expressed

15.

as a mantle magnitude corrected for focal mechansim ($M_c = \log_{10} M_0 - 20$ with M_0 in dyn*cm), in the formalism of *Okal and Talandier* [1989].

We use a slightly adapted focal mechanism ($\phi = 320°$, $\delta = 15°$; $\lambda = 90°$), rotated only 8° in the formalism of *Kagan* (1991) from *Kanamori et al.* (2010) in order to reduce the scatter between Love and Rayleigh spectral amplitudes. Note that this new mechanism is also closer to the geometry of the nearby 2005 Nias megathrust earthquake ($\phi = 333°$, $\delta = 8°$; $\lambda = 118°$). Our results **(Figure 3)** suggest an average moment $M_o \approx 2.5$ ~~x~~ 10^{28} dyn*cm ($M_C \approx 8.2$) at periods between 100~~s –~~ and 160~~s~~; at the largest resolvable period (~170~~s~~) the magnitude estimate approaches $M_C \approx 8.4$.

This value of the seismic moment, about 4 times larger than proposed by *Kanamori et al.* [2010] from an estimate of time-domain amplitudes of *S* waves around 50 s, demonstrates slowness in the source of the 1907 mainshock, in clear agreement with its nature as a "tsunami earthquake". In addition, Figure xxx documents a growth of moment with period, with a regression slope for M_C vs. frequency of −0.09 logarithmic units per mHz. This number is comparable to values obtained for ~~instrumented~~ documented tsunami earthquakes (e.g., Java, 2006: −0.11; Mentawai, 2010: −0.08; Manzanillo, 22 June 1932: −0.14; Hikurangi, 1947: −0.07 and −0.08); by contrast, traditional subduction events feature lower slopes (in absolute value) that do not exceed −0.05 (e.g., Maule, 2010: −0.05; Illapel, 2015: −0.05; Manzanillo, 03 June 1932: −0.01; see *Okal & Saloor*, 2017). This property provides a quantitative confirmation of the nature of Event I as a tsunami earthquake in conjunction with ~~the~~ its anomalously low felt intensities.

Ratio of estimated energies between Event I and II

We reproduce a copy of the Manila seismic record from *Masó* (1907; **Figure 5**). The seismogram in the top frame **(Figure 5a)** was written on a Vicentini mechanical seismometer with a period T = 1.4~~s~~, which makes it the only short-period recording available to us of the 1907 Sumatra earthquakes. The distance to Manila is $\Delta = 29°$. Even though the orientation of this horizontal seismogram (NNW−SSE or 337−157°) is pure transverse (the back-azimuth at

16.

β

Manila being b = 247°), this record clearly shows the P-waves from two earthquakes separated by ~53m, corresponding to the two shocks identified on the basis of macroseismic ~53 mn, data. We aligned the traces of the two events vertically to ease the comparison of their waveforms (**Figures 5a; 5b**). The body waves have a dominant recorded period of ~4s. The P-waves of the mainshock are both much weaker and of longer duration than those of the aftershock. Note also that the mainshock P-wave train consists of several subevents, lasting a total of about two minutes, while the aftershock waveform features two main packets, each lasting about 3s. By contrast, we show on the lower frame a record (**Figure 5b**) written on a longer-period Omori system with a natural period T = 6.4 s, oriented ENE−WSW, which this time is purely orthoradial. ~~No verb!~~ On that record, the amplitudes of the P-waves from the two events have become very comparable, and the main shock displays prominent Rayleigh waves featuring a period of ~9s at their maximum amplitude (probably reflecting the peaked response of the instrument), while the surface waves from the aftershock are much weaker.

In an attempt to quantify these observations, we enlarged the Vicentini records (**Figure 5a)** and digitized them at a time sampling δt = 0.1s. Unfortunately, the Vicentini instrument was undamped (*Masó*, 1907), which means that it would, at least theoretically, have an infinite response at its natural period. In this context, it would not be possible to formally compute an energy-to-moment ratio and a slowness parameter (Θ) as defined by *Newman & Okal* (1998). This is further compounded by the fact that the photographic magnification of the figure by *Masó* (1907) is unknown. Under the circumstances, we used the model of a very weakly damped instrument (ε e = 1.1) and were able to obtain a ratio of estimated energies between the mainshock (Event I) and the aftershock (Event II): $E_I^E / E_{II}^E \approx 0.2$.

17.

Furthermore, we have obtained an estimate of $M_s = 7¼$ for the classcial surface-wave magnitude of Event II in the previous section, which would correspond to a moment of 10^{27} dyn*cm assuming the earthquake follows scaling laws. However, the strong intensities reported may suggest that its source is blue-shifted towards higher frequencies, in which case the moment could be lower. Given the long-period moment of Event I (2.5×10^{28}), this would suggest a factor of about 200 between the energy-to-moment ratios of Events I and II, amounting to a difference of ~2.3 units in their parameters Θ. While this number is not unreasonable, it comes close to the maximum range of Θ values that we have documented (see *Okal and Saloor* [2017, Fig. 13] for an extensive dataset), and could be explained using both a very low value of $\Theta \approx -6.4$ for the mainshock, and a high one, $\Theta \approx -4.1$, for Event II. The former value is within the range of observed Θ for tsunami earthquakes (El Salvador, 2012: –6.42; Hikurangi 1947 (II, 17 May): –6.51 [*Okal and Saloor,* 2017]), while the latter would be typical of a "snappy" intraplate earthquake, such as the main normal faulting aftershock of the 2011 Tohoku earthquake (–4.12) or the great 1939 Chillán earthquake (–4.04) [*Okal and Kirby,* 2002; *Okal et al.,* 2016]. The jagged nature of the *P* waves of the mainshock at Manila would also be in line with observations during the 1992 Nicaragua event [*Polet and Kanamori,* 2000] and the resulting model by *Tanioka et al.* [2000].

I had to tweak the numbers, using 7 1/4 rather than 7 for $(M_s)_{II}$.

17a

Indian Ocean Tsunami Observations

We were able to compile 76 tsunami observations of the tsunami generated by Event I across the Indian Ocean, including xxx for which numerical constraints could be extracted (Table S3 and Figure 6). They were obtained from individual accounts [*e.g., Schröder,* 1917*a,b*], as well as from reports published in local newspapers. Most of the Dutch news reports were corroborated by later official Dutch correspondence (*Openbaar Verbaal,* 23 April 1908). First and second-hand accounts from a few locales in the near and far-field are available (e.g. *Anonymous,* 1909; *Vissier,* 1922), and some were repeated by *Solov'iev & Go* (1974). We also extracted eyewitness or second-hand accounts appearing in contemporary anthropological (e.g.,

Соловьёв
Solov'ev

specify number!

18.

Baumwoll, 2008; *Rahman et al.*, 2017) and scientific studies (*Yogaswara & Yulianto*, 2006; *Whitlow*, 2008; *Fujino et al.*, 2014).

The impact of the tsunami was ~~the~~ greatest on the southern coast of Simeulue, and on the northern, and western coast of Nias (**Figure 6a, b**). Without providing additional location details, news and official reports from Simeulue suggest the tsunami struck after Friday afternoon prayers, the sea flooding inland as much as 600-800 m~~etres~~ with isolated ~~run-ups~~ inundations of 1,200 m~~eters~~ that swept coral boulders and marine fauna inland (*Bataviaasch Nieuwsblad*, 12 February 1907; *Openbaar Verbaal*, 23 April 1908; *Yogaswara & Yulianto*, 2006). Better numerical constraint is forthcoming from Nias (**Figure 6a, 6c**), for example, from Afulu (nl: *Afoeloe*), where the sea washed inland for over a km (nl: *ongeveer een paal*†), as far as the foot of the hills where it deposited debris that included human remains and large uprooted trees [*Bataviaasch Nieuwsblad*, 22 March 1907]. However, to the South, the tsunami appears to have rapidly diminished in character at Lagundi (**Figure 6b**; *Bataviaasch Nieuwsblad*, 22 March 1907).

see 18a

The island of Simeulue Cut (nl: *Simeuloeë Tjoet* or *Simaloer Tjoet*) off the southern coast of Simeulue (**Figure 6a**) "was lost" or "disappeared" (nl: *is verdwenen*) with only the hill at the centre of the island above the water (**Figure 6b**), suggestive of extreme tsunami erosion or possibly, subsidence. West of Nias, on Pulau Wunga (**Figure 6b**) the tsunami swept the island from the west (*Schröder*, 1917a), destroying three-fourths of it including ten thousand coconut trees, with debris and victims swept into the lagoon at the centre of the island (*Bataviaasch Nieuwsblad*, 22 March 1907). The height of the tsunami is estimated at

† The old Dutch measure *"een paal"* used in colonial Indonesia is equivalent to 1507 m [*Staring*, 1871].

18a.

<<< CONFUSED >>>

Do you mean **run-up,** which would be on the order of meters (1.2 m) or **inundation** which should be on the order of hundreds of meters or even kilometers, as suggested below.

Recall that *inundation* is maximum distance of horizontal penetration, and *run-up* altitude of the corresponding point above initial sea-level.

In definition of "een paal", you need a decimal point (English usage), not a comma (French usage). Better yet use neither for a 4-digit number...

I suggest using footnote to define the "paal'. Otherwise distracts from flow of text.

between 6~~m to~~ and 15m (*Schröder* (1917a; *De Padanger*, 25 February 1907), and in 2005, modern residents told one of the authors that corpses were stranded in coconut trees in 1907. The only photographic evidence of this disaster (**Figure SX**) is from Pulau Wunga (*Schröder*, 1917b); how representative it is of damage elsewhere on the island is unknown but the similarity of the post-tsunami landscape ~~with~~ to that of Pulau Sibigau (*Hill et al.*, 2010) following the 2010 Mentawai tsunami earthquake is striking.

are you sure?

On the Sumatran mainland (**Figure 6a**), vague accounts (*Vissier*, 1931; ~~p165;~~ p. 165; *Aceh Post*, 10 June 2013) suggest that Kayu Menang and Sibolga were possibly flooded. At Guhanaga (formerly Geni) a flood in the Krueng Sabe river (*Aceh Post*, 10 June 2013) ~~related~~ was attributed to the tsunami by an eyewitness but we consider this account cautiously as the date associated with it is unreliable. The tsunami was recorded by a tide gauge in the harbour at Padang (*Anonymous*, 1909) but this record has been lost.

On the Indian subcontinent and its dependencies, tide gauges operated by the Survey of India at Port Blair (Andaman Islands), Apollo Bandar (in Mumbai), Chennai (Madras) and Karachi provided the only reported instrumental readings from this tsunami [*Erskine,* 1908, 1909] (Figure 6d), but maregrams were unavailable to us. Similar intruments at Kidderpore (Calcutta), Yangon (Rangoon, Myanmar) and Aden (Yemen) did not record the disturbance [*Erskine,* 1908, 1909].

In the Southern Indian Ocean (Figure 6f), the tsunami was observed at Mauritius [*Erskine,* 1909], on Rodrigues [*The West Australian,* 5 January 1907], on La Réunion [*Bertho,* 1910; *Sahal et al.,* 2011], and reportedly in the Seychelles, and at the French islands of Amsterdam and St. Paul [*Bertho,* 1910]. However, the remoteness of those two islands, whose settlement at the time is not proven beyond doubt, lead us to question the veracity of the corresponding reports.

20.

Rodrigues: interesting: O.T. 1200 + 4 hrs est. = 1600 GMT +7 = ~~[illegible]~~ 1700 GMT +8

Time to send telegram to Perth, still makes it to the 5 Jan 07 edition of the newspaper!

Curiously this event is missing from the French national tsunami database (*Lambert & Terrier*, 2011).

As for Madagascar, ~~H~~historical tide gauge data available from *Service hydrographique et océanographique de la Marine* (SHOM) for ~~Madagascar~~ (SHOM Shelfmark: SHD R 7JJ2384) have no entries ~~does not include data~~ for 3 – 6 January 1907 ~~at~~ Diego-Suarez (~~*Isabelle*~~ I. *Garnier-Loussaut*, personal communication, 2014). Observations of high and low water levels are missing for January 1907 from Tamatave, and diurnal visual observations made with a tide staff are only available for 15 February until 2 March 1907 from Baie du Courrier (*Yann Ferret*, personal communication, 2014).

No reports could be found in the several French-language newspapers published in Madagascar at the time.

21.

Tsunami Modelling

Methodology

We chose a subset of 12 locations with ~~known~~ reliable geographic co-ordinates ~~for~~ at which either run-up, wave heights, or inundation distances could be determined **(Table S3)** on the islands of Simeulue and Nias (**Figure 6a, 7b**).

These sites extend approximately parallel to the Northern Sunda Trench, along strike from Simeulue to Nias, for a total distance of about 220 km. Their spatial distribution and their proximity to the Sunda Trench (60–80 km), along with the effect of directivity [*e.g., Miyoshi,* 1955; *Kajiura,* 1972] then allow a causal relationship between field observations and the parameters of the rupture, with each site serving as a control point to estimate and constrain slip on a corresponding segment of a model of the parent seismic source.

Our tsunami simulations use COMCOT (Cornell Multi-Grid Coupled Tsunami Model), a fully validated finite-difference algorithm solving the non-linear shallow-water approximation of hydrodynamics [*Liu et al.,* 1998], which has been successfully applied to the investigation of both historical and modern tsunamis [*e.g., Li et al.,* 2015; *Wang and Liu,* 2006]. However, the extremely limited nature and quality of the available seismological data precludes the use of comonplace modern methods [*e.g., Kikuchi and Kanamori,* 1982, 1986, 1991] to constrain the parameters (fault length and width, seismic slip) necessary to compute a field of initial conditions for the simulation of the tsunami.

In a recent study, *Ebel and Chambers* [2016] have documented that seismicity occurring in the vicinity of past large ruptures in Eastern North America and California shows a tendency to concentrate its largest events at the edges of rupture. They went on to show ~~the~~ that this geographical trend can be used to map the rupture area of historical events predating the development of modern (and especially digital) data necessary to conduct detailed source tomography. In this context, and assuming that *Ebel and Chambers'* [2016] paradigm can be extended to the case of large events at subduction interfaces, we use our preferred epicentral location, and catalogues of relocated modern seismicity before and after the 2004 Sumatra-Andaman and 2005 Nias earthquakes [*Engdahl et al.,* 2007; *Pesicek et al.,* 2010], to define a zone of presently low seismic productivity in the uppermost part of the plate contact, extending roughly 250 km between 94.7 and 97°N. In this general area, we then envision a number of scenarios for seismic rupture along the shallow up-dip plate interface of the Sunda Trench, which we divide into sixteen elementary sub-faults, themselves regrouped into four segments labeled A to D, this methodology being detailed in the *Supplementary Material.* The 16 patches, each measuring 30 km × 40 km, are distributed in updip–downdip pairs (Figure 7a), their exact geometry following Model Slab1.0 [*Hayes et al.,* 2012].

Because of the slowness we have documented in the source of the 1907 earthquake, we use slip estimates departing from the classical scaling laws applicable to subduction events [*Geller,* 1976; *Blaser et al.,* 2010] and consider models with variable rigidity μ. We start by assuming that the rupture was limited to the shallowest portion of the interface; conventional crustal material would feature a typical value $\mu = 3 \times 10^{11}$ dyn/cm^2, which using the moment $M_0 = 2.5 \times 10^{28}$ dyn*cm obtained from mantle waves, would result in a slip $\Delta u = 7$ m on each of the eight up-dip sub-faults. We then model a tsunami earthquake by varying the rigidity μ on each segment between 0.7 and 2.0×10^{11} dyn/cm^2, which increases the slip to between 9 and 24 m (Figures 7a, and S1–7), in 3–m increments. These values of μ fall in the range proposed in the context of tsunami earthquakes by a number of investigators [*Ide et al.,* 1993; *Satake,* 1995; *Bilek and Lay,* 1999; *Geist and Bilek,* 2001; *Hill et al.,* 2012; *Li et al.,* 2015].

For each of these models, we then compare the simulated tsunami amplitudes to those estimated from documentary material (Figure 6).

Lines 429-437

The tsunami impacts documented on Simeulue (**Figure 6b**) were best replicated in association with more than 21m of slip on segment B of the rupture. For example, at Lakubang, the inundation distance is thought to have approached 3km (*Yogaswara & Yulianto*, 2006) despite this location being relatively sheltered by offshore islands (**Figure 7b**). At this location, despite slip greater then 21m on segment B, the inundation ~~is~~ remains underpredicted and we suspect the inundation distance reported by *Yogaswara & Yulianto* (2006) is strongly controlled by local drainage that is known to allow tsunamis to travel much farther inland following the channel of natural and man-made drainage features (e.g. *Mori et al.*, 2011). In the region of Nias to the south (**Figure 7c**), we find that a slip of 15m is

required on Segment D to reproduce the wave heights reported from Pulau Wunga (**Figure 7c**). Our constraints on the amount of slip on Segments A and C (**Figure 6**) are weaker due to the lack of quantitative records from northern Simeulue and the Banyak islands respectively (**Figure 2c**). However, we infer that the noticeable tsunami impacts on the west coast of Sumatra (**Figure 2c**) such as the flooding at Susoh (*Bataviaasch Nieuwsblad*, 12 February 1907), Tapaktuan (*Anonymous,* 1909), and possibly Kayu Menang (*Aceh Post,* 10 June 2013) are probably indicative of slip of 6-9m on those segments.

24

In conclusion, our simulation experiments indicate that the values of run-up and inundation estimated from the documentary material require seismic slips varying from 6 to 9 m on Segments A and C to as much as 15 m on Segment D, and probably more than 21 m on Segment B. These values are greater, by factors ranging from 1.3 to 5, than expected under scaling laws [*e.g., Geller,* 1976] for an earthquake of moment $M_0 = 2.5 \times 10^{28}$ dyn*cm. This provides an additional argument for the anomalous properties of the 1907 event, and quantifies its characcter as a "tsunami earthquake", as defined in *Kanamori*'s [1972] landmark study.

Our preferred rupture model satisfactorily predicts tsunami heights at all but one location, Afulu on Nias island (Figure xx), where our model considerably underestimates

<<< run-up (xxx m as opposed to the reported xxx m) and/or inundation (xxx m as opposed to xxx m). >>> please insert numbers!

We note that the modern village is located near a semi-circular bay (Telok Afulu), opening to the South and facing a small island to the West, at 1.25°N; 97.24°E; further, a review of topographic data (see *Data and Resources*) reveals steep relief (≥ 20 m) close to the shore, and a drainage channel extending inland for ≈ 1 km from the northwestern shore of Telok Afulu. Such features can cause highly localized amplification of tsunami waves [*e.g., Shimozono et al.,* 2014], while a small island can also amplify tsunami waves in its lee, a scenario known as "Babi effect" after it was first observed and modeled at Babi Island† during the 1992 Flores tsunami [*Yeh et al.,* 1994; *Briggs et al.,* 1995]. Localized extreme inundation could also be due to an underwater landslide (possibly triggered by the snappier Event II), as in the case at Riangkroko during the Flores tsunami [*Tsuji et al.,* 1995; *Plafker,* 1997]. We also cannot rule out the possibility of a separate, local tsunami caused by Event II, especially given the lack of arrival times in the accounts of the tsunami at Afulu.

Far field?

† a small island on the Northern shore of Flores island (8.43°S; 122.51°E), not to be confused with Babi Island (2.09°N; 96.65°E), ~30 km SE of Simeulue

Discussion and Conclusion

This study sheds a considerably enhanced vision on the 1907 Sumatra earthquake, resolving apparently contradictory properties through the documentation of Event II, reassessing its long-period moment and quantifying its character as a slow "tsunami earthquake", and proposing a model of seismic rupture which provides an acceptable fit to a new dataset of tsunami run-up and inundation values. In the following sections, we provide a perspective on our most important results.

The value of non-instrumental archives

Our study research underscores the importance of meticulously collated and analysed non-instrumental evidence that ~~supplements~~ allows the study of historical ~~and early-instrumental~~ earthquakes for which instrumental data is lacking or severely limited. ~~such as the 1907 Sumatra tsunami earthquake.~~ The discovery of Event II was an unexpected outcome borne out of our careful scrutiny of macroseismic data that led us to look for seismological evidence to support it. The absence of this event from earthquake catalogues (e.g. *Gutenberg & Richter*, 1954; *Storchak et al.*, 2013) is very conspicuous but not unusual, as with other recently discovered large early-instrumental earthquakes (*Hough et al.*, 2005); it results from the limited number of seismograms at hand, and the inadequacy (or simple inexistence) of proper algorithms for association of reported phase times, not to mention the adjustment of local times before the implementation of standard time zones. The previously untapped historical and scientific sources located by us have also formed the basis of early 20th century descriptions of the disaster (e.g. *de Meyier*, 1907; *Rappard*, 1909; *van der Linde*, 1920). In stark contrast with other natural disasters in the Dutch East Indies such as the 1883 eruption of Krakatau ~~in Sunda Strait~~ for which detailed official reports were written (e.g. *Veerbeck*, 1886), locating an official report for the 1907 disaster was futile despite exhaustive efforts on our part.

26

Szirtes (1912; pp.5) carries cursory mention of a "detailed study" (de: *eingehende Untersuchung*) by T.H. Staverman, including possibly a study of its epicentral location, but without a complete citation.

The scientific bibliography in The Dutch East Indies between 1907 and 1912 [*e.g., Verbeek,* 1912; *Visser,* 1922] has no record of this document, and its whereabouts remain in mystery, including at Strasbourg, where Siegmund Szirtes was based [*L. Rivera,* pers. comm., 2017].

The 1907 epicenter in the local tectonic context

Our preferred location of the rupture from Event I is trench ward i.e. up-dip, in a north-east facing re-entrant on the Sunda Megathrust (*Franke et al.*, 2008) between the ruptures of the 2004 (*Chieh et al.*, 2007) and 2005 (*Konca et al.*, 2008) earthquakes. This is in agreement with ~~supported by~~ our conclusion that Event I was a tsunami earthquake, and with the documented observations of the near-field tsunami. Near the western terminus of our rupture, seismic reflection surveys (*Franke et al.*, 2008) have identified a ridge of oceanic basement (**Figure 1b**) that projects into the Wharton Basin coinciding with a mapped fracture zone (*Singh et al.*, 2011; *Jacob et al.*, 2014). On Simeulue, this coincides with the Simeulue Saddle (*Sieh et al.*, 2005) which has served as a persistent barrier to rupture in the past (*Meltzner et al.*, 2012). This feature, or barrier, along with another to the south in the Batu Islands (*Natawidjaja et al.*, 2006), demarcates an important segment boundary on the Sunda megathrust (*Meltzner et al.*, 2015). Alt Though our best-fit epicentral location lies slightly to the west of this feature, ~~the~~ its uncertainty ellipse ~~associated with it~~ overlaps this basement ridge. Incidentally, the largest inferred slip in 1907 on the interface locates ~~occurred~~ in the region of this structural feature, a pattern ~~which is~~ very similar to that ~~studies~~ of other tsunami earthquakes, that associate both event nucleation and the largest slip with the subduction of subsurface topography (e.g. *Tanioka et al.*, 1997; *Abercrombie et al.*, 2001; *Newman et al*, 2011b; *Bell et al.*, 2014); this remark is also supported by the jagged nature of the Omori waveform discussed previously. The weaker

see enclosed comment

6 in ref list

difficult to say which one supports the other tsunami amplitudes on Southern Nias suggest that the 1907 rupture may not have extended to he southern segment barrier.

Since our rupture model is only constrained by tsunami observations, further research would also be required to refine the western limit of the 1907 rupture in the region of the subsurface ridge identified by *Franke et al.* (2008), in order to determine whether the rupture stopped at, or propagated through, this segement boundary.

The value of maximum slip in our rupture model is ~~constrained on a~~ obtained in a model of simple planar rupture, but we also recognise the possibility of a complex rupture involving splay faults within the overlying accretionary wedge. Seismic reflection profiles indicate the presence of frontal folds and pop-up structures south (*Franke et al.*, 2008) and west (*Singh et al.*, 2008) of Simeulue. The relation between splay faults and unexpectedly large sea floor displacements has been numerically proven (e.g. *Wendt et al.*, 2009) and ~~observed in~~ proposed for other tsunami earthquakes (e.g. *Fan et al.*, 2017). However, given the uncertainties in the near field associated with both the quantifiable numeric observations, i.e., run-up and inundation distances, as well as the largely unknown arrival times of the tsunami locally, we ~~choose a~~ restricted this study to a simple planar model.

28.

The next sentence doubts that they are first hand.

The lack of geodetic signals

Our study has ~~accounted for~~ resolved many unaddressed ~~aspects about~~ aspects of the 1907 Sumatra earthquake sequence but leaves unanswered the question of whether any subsidence was associated with this sequence? .

Our preferred rupture model would be expected to generate subsidence on the order of xxx m at Simeulue and uplift of xxx at xxx.

crucial, fill in!

In this context, we note *Hodgon*'s [1934] remark that "the south coast was partially submerged by an earthquake", which could be interpretd either as coseismic subsidence [*Meltzner et al.*, 2015], or as tsunami inundation under the semantics of the 1930s.

Reports from the southern coast of Simeulue are limited and none of the available Dutch accounts ~~report~~ mention land level changes associated with the 1907 earthquake, except in brief descriptions (see *Supplementary Material*) from Simeulue Cut (*Utrechts Nieuwsblad*, 14 February 1907; **Figure 5b**) and the Hinako Islands (*Bataviaasch Nieuwsblad*, 22 March 1907; **Figure 5b**).

(need to give some estimate ~~and~~ of the relevant amplitudes).

These accounts do not state whether the writer actually visited the island, viewed it from a distance, interviewed direct witnesses, or simply ~~derived~~ reported this information ~~from local accounts.~~ second-hand. If taken at face value, the description from Simeulue Cut might be inferred as coseismic subsidence or extreme tsunami erosion. The other perplexing detail comes from the Hinako islands (**Figure 5x**) where the seabed was reportedly exposed (uplifted?) for up to 2 4 hours (*Bataviaasch Nieuwsblad*, 22 March 1907). Coral microatolls have been well studied on Simeulue and

29.

contributed have to our understanding of the rupture extents of medieval great earthquakes (*Meltzner et al.*, 2015) as they are natural long-term indicators of changes in relative sea-level, but these are often subject to preservation issues that can undermine temporal completeness (*Meltzner et al.*, 2012). In this regard, and pertinent to our study, *Meltzner et al.* (2012) investigated a coral microatoll at Ujung Lambajo (LBJ; **Figure 1b**) at a distance of ~6km north-west of Simeulue Cut but found that the paleo-record at this site only extended to 1955. Much further south, *Meltzner et al.* (2015) records the death of a microatoll (LAG-3B; **Figure 1b**) near Lagundri on Nias and suggest this could either be related to the 1907 earthquake or had “nothing to do with tectonics or relative sea level (RSL) change.” Unlike the locations of many tsunami earthquakes that are devoid of offshore islands, Simeulue and Nias lie within 100km of the trench axis and therefore increase the likelihood that the coseismic signature of land level changes from tsunami earthquakes could be recorded in the paleo-record of microatolls. It is also entirely plausible that the inferred land level changes on Simeulue Cut and in the Hinako Islands are indicative of a rupture that extended further downdip than in our first-order model. Also, they could have been, both or independently, associated with the aftershock, or with post-seismic after-slip. Bearing in mind the uncertainties in all of the above, we do not incorporate these in our rupture model but suggest that further work is required in this unique physiographic environment to uncover the paleo-record of previous tsunami earthquakes near Simeulue.

The 1907 event as a Tsunami Earthquake: A Global Perspective

The main conclusion of our study is the confirmation and quantification of the slowness of the 1907 mainshock (Event I), which we achieved from a broad range of seismological observations. By computing its seismic moment from the spectral amplitude of surface waves measured at mantle periods (up to 170 s), we obtain a value significantly greater (by a factor of about 4) than previously derived around 50 s by *Kanamori et al.* [2010]. Although it must remain qualitative in the absence of instrumental metadata, the comparison of amplitudes and duration of body and surface waves at Manila (Figure xx) clearly documents the difference in source properties between Events I and II, and hence the slowness of the mainshock. The identification of Event II, occurring only 59 minutes after the mainshock, clearly resolves the paradox of an event holding many hallmarks of a "tsunami earthquake", but being felt at surprisingly high intensities.

adjust as necessary

Because of its size and location, the identification of the 1907 North Sumatra event as a tsunami earthquake has a number of implications which must be discussed in a global perspective. Table xxx compiles characteristics for 21 known such events, including estimates of seismic moments and, when available, of the slowness parameter Θ introduced by *Newman and Okal* [1998]. At $M_0 = 2.5 \times 10^{28}$ dyn*cm, the 1907 earthquake ranks as the third largest documented "tsunami earthquake", after the 2004 Sumatra-Andaman earthquake (which can be considered a tsunami earthquake, in view of the clear slowness in its source [*Choy and Boatwright,* 2007; *Okal,* 2011]) and the 1946 Aleutian one. Note that the list in Table xx may not be exhaustive since the origin of some historical destructive waves remains debated, *e.g.,* on 26 February 1902 in El Salvador [*Cruz and Wyss,* 1983; *Ambraseys and Adams,* 1996]. Some evidence would also characterize the 1700 Cascadia meagathrust event as a tsunami earthquake [*Obermeier and Dickenson,* 2000; *Okal,* 2011].

As summarized for example by *Okal* [2008], tsunami earthquakes are generally thought to occur either through rupture in mechanically deficient media, such as accretionary prisms *e.g.,* 31,

along splay faults propagating through sedimentary structures [*Fukao,* 1979], or under conditions of sediment starvation leading to seismic failure along the uppermost section of the plate interface, involving a jagged, and hence disproportionally long, rupture [*Tanioka et al.,* 1997; *Polet and Kanamori,* 2000]. In addition, tsunami earthquakes can occur as mainshocks, which *Okal and Saloor* [2017] qualified as "Primary Tsunami Earthquakes" (PTEs) or as "Aftershock Tsunami Earthquakes" (ATEs), following a larger, regular megathrust event. The exceptional tsunami of the 2011 Tohoku event may be the result of the combination of a regular megathrust event, and of a lower-frequency rupture of the shallowest portion of the interface, which might have qualified as an ATE, but for its occurrence only 3 minutes after the initial nucleation [*Satake et al.,* 2013].

Because they are felt at deceptively low levels not conducive to the perception of an appropriate level of threat, tsunami earthquakes form a particularly vicious kind of natural hazard, and their mitigation constitutes one of the biggest challenges facing today's tsunami community. Paramount among remaining questions of primary importance is the relationship between regular megathrust, PTEs and ATEs:

(i) *"Does the distribution of PTEs bear a regional signal, in other words does the occurrence of a PTE constitute a harbinger of more to come along the same subduction zone?"*

A systematic study by *Okal and Newman* [2001] in the regions of the three 1990s PTEs (Nicaragua 1992; Java, 1994; and Northern Peru, 1996) found no systematic trend for slowness in background seismicity, but identified as PTE an older (1960) event in Northern Peru, 340 km away from the 1996 Chimbote source, thus suggesting a lateral coherence in the updip properties of the interface governing the occurrence of PTEs. Posterior to their study, the 2006 Java earthquake provided a spectacular confirmation of this trend, when it took place 580 km from the 1994 epicenter. Similarly, the 2012 El Salvador PTE occurred only 175 km from the 1992 Nicaragua event; should the 1902 event (which combined weak

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inland destruction and a wave killing about 170 people) be confirmed as a PTE, then it would earmark a 350–km section of the Central American subduction zone as prone to PTEs.

In this context, the 1907 event provides some limited new insight. It was indeed followed, 98 years later, by a slow megathrust earthquake, but the origin of their slow character may not be comparable: based on its location near the trench, the 1907 event's slowness was rooted in its shallow source, whereas in 2004, it emanated from a relatively slow and jagged propagation along its exceptionally long fault length [*Ishii et al.,* 2005; *de Groot-Hedlin,* 2005; *Tolstoy and Bohnenstiehl,* 2005].

More comparable are the events of 1907 and 2010, which both took place along the shallowest, probably structurally similar, portions of the plate interface, and were separated laterally by 900 km. However the 2010 Mentawai earthquake was clearly an ATE of the 2007 Bengkulu megathrust event, while the 1907 earthquake was Primary: there is no evidence anywhere in historical events of an immediately preceding mainshock which would have been significantly larger [*Newcomb and McCann,* 1987].

(ii) *"Does the occurrence of PTEs along a subduction zone rule out that of regular (and possibly larger) megathrusts?"*

The question is particularly critical along Java where two PTEs featuring very similar source characteristics took place in 1994 and 2006, and where no truly gigantic earthquake is known in the admittedly short historical record [*Newcomb and McCann,* 1987]. It would also apply to the Hikurangi Trough, off the North Island of New Zealand, where the only large known interplate thrust earthquakes are the 1947 tsunami earthquakes [*Okal and Saloor,* 2017] (and the historical record is even shorter), and conceivably to Cascadia if indeed the 1700 event had a slow source.

In this context, the proximity of the 1907 and 2005 epicenters would suggest a negative

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answer, namely that PTEs and regular megathrust events may occur along neighboring subduction segments (or even along the same one at different depths along the interface). Note that this conclusion might be supported by the case of the 1946 Aleutian PTE, in comparison to the nearby 1957 and 1938 events which did not exhibit source slowness, the latter even generating a deceptively small tsunami [*Stover and Coffman,* 1993]; however, this analogy remains tentative in the presence of the Shumagin Gap between the 1946 and 1938 ruptures, where no large events are known since at least 1787 [*Davies et al.,* 1981], which leaves the question of its potential for a PTE wide open until the next large earthquake fills it.

(iii) Finally, the ultimate question is *"Can any subduction zone be considered immune to tsuanmi earthquakes, either PTEs or ATEs?"*

While tsunami earthquakes have not been documented (at least yet) in a number of subduction zones (*e.g.,* Southern Peru, Northern Chile), the small scale study in Vanutu by *Okal and Saloor* [2017] documents a lateral, along-strike, variation of the slowness parameter Θ, in conjuction with the subduction of submarine structures, on a scale of a only a few hundred km. Given the largely random nature of the fragmentation of rupture at subduction zones [*Ando,* 1975], and the gross undersampling of tectonic cycles by the modern science of seismology, which resulted in unsuspectedly large events (2004 Sumatra; 2011 Tohoku) having led to the demise of otherwise promising paradigms [*Stein and Okal,* 2007; *McCaffrey,* 2007], the only prudent answer to this question must remain negative.

The 1907 event provides no new insight in this respect.

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The future: Bracing for the next one

In the absence of definitive evidence of aseismic slip, tremor, or slow-slip earthquakes up-dip (*Feng et al.*, 2015; *Tsang et al.*, 2015), and based on newly modelled estimates of convergence of 46 mm/yr at (1°N; 96°E) [*Bradley et al.*, 2017], we believe that the rupture patch identified in this study, having now accumulated 5.1 m of locked slip since 1907, could host a future $M_w \geq 7.7$ event, which may be a tsunami earthquake, and which could also be enhanced by Coulomb stress transfer from the 2004 event [*McCloskey et al.*, 2005]. With dimensions and physical properties similar to our model (Figure 7a), this future event could conceivably feature deceptive ground accelerations insufficient to elicit self-evacuation.

We recall that on Simeulue Island, the local legend of *s'mong* associates strong earthquakes with tsunami hazard, and elicits self-evacuation to high ground upon feeling strong shaking [*McAdoo et al.,* 2006; *Baumwoll,* 2008; *Syafwina,* 2014]. It was particularly effective during the 2004 Sumatra-Andaman disaster, where an estimated 800 people were saved by self-evacuating from several villages which were later totally eradicated by the tsunami [*McAdoo et al.,* 2006]. It has been proposed [*e.g., Syafwina,* 2014] that *s'mong* is based on memories of the 1907 event, preserved in verse or in stories passed down by older generations (we note, however that *Yogaswara and Yulianto* [2066] record versions of the legend, both with and without severe shaking). If so, the exact role played, in the *s'mong* legend, by the weak shaking due to Event I, and the presumably stronger one from Event II, remains unclear; at any rate, shaking from Event II, estimated to lag Event I by 59 minutes, may not have occurred before the tsunami attacked Simeulue, where it obliterated many villages and resulted in hundreds of casualties; this is further suggested by E. Schröder's account of the second shock being felt on Nias (located farther from Event I than Simeulue) *after* flooding by the tsunami [*Bataviaasch Nieuwsblad,* 22 March 1907].

adjust as necessary

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While the *s'mong* tradition may have saved many lives in 2004 and 2005, it is clear that it is not adapted to the particular challenge of tsunami earthquakes, as illustrated by the case of the 2010 Mentawai event, 600 km Southeast of Simeulue (Figure xxx) and a typical ATE following the 2007 Bengkulu megathrust earthquake. As detailed by *Hill et al.* [2012], the earthquake was felt only "gently" or even not felt at all at some locations on the Pagai islands. Yet, the tsunami reached run-up heights of 17 m and killed more than 500 people. Over-reliance on the legend of the *s'mong* prevented local villagers from evacuating immediately, since the shaking was weaker than during the seismologically much larger 2007 Bengkulu mainshock [*Borrero et al.,* 2009] and during the 2009 intraplate Padang earthquake (which took place deeper in the slab, and under the Island of Sumatra [*McCloskey et al.,* 2010]), both of which did not produce a significant tsunami in the Mentawai Islands.

In this context, some interviewed witnesses reported that the 2010 shaking on the Mentawai Islands was both weak and long, lasting as much as "several minutes" [*Hill et al.,* 2012, p. 4]. This is an important observation, since source slowness is expected to both decrease high-frequency ground motion (and hence the level of shaking), and increase the duration of seismic wavetrains; the comparison of these two properties has been earmarked by *Convers and Newman* [2013] and *Okal* [2013] as a means of identifying tsunami earthquakes from the analysis of first-arriving *P* waveshapes. Indeed, during the 2010 Mentawai event, *Newman et al.* [2011] had iden- a? tified the slow character of the event only 17 minutes after the origin time using a prototype of b? their method, but unfortunately, there existed no means to relay this information in due time to the populations at risk.

Accordingly, we end this study by reiterating the recommendation made by *Hill et al.* [2012, p. 19] to include the duration of felt shaking as a warning for spontaneous evacutation, in a sense to implement a variant of *Convers and Newman*'s [2013] algorithm based on human perception. This constitutes a formidable challenge, since our experience gathered from a large number of post-tsunami surveys [*e.g., Synolakis and Okal,* 2005] is that the perception of relatively short time durations (seconds, minutes) by lay individuals in a situation of emergency and panic is even less accurate than that of distance.

TABLE xx. List of Documented Tsunami Earthquakes

Date D M (J) Y	Region	Type *	Moment (10^{27} dyn*cm)	Θ	Reference	Landslide *	Notes
15 JUN (167) 1896	Sanriku, Japan	PTE	12		a		Charter Event [*Kanamori,* 1972]
04 JAN (004) 1907	**Northern Sumatra**	**PTE**	**25**	**−6.5 ?**			**This Study**
13 APR (103) 1923	Kamchatka	ATE	1.2		b	L	
22 JUN (174) 1932	Manzanillo, Mexico	ATE	5.2	−6.18	c		
21 JUL (202) 1934	Santa Cruz	ATE	10	−6.10	d		
01 APR (091) 1946	Aleutian	PTE	85	−7.03	e	L	Charter Event [*Kanamori,* 1972]
25 MAR (084) 1947	Hikurangi, N.Z.	PTE	4	−5.94	d		
17 MAY (137) 1947	Hikurangi, N.Z.	†	3	−6.51	d		
20 NOV (325) 1960	Northern Peru	PTE	2.7	−6.13	f		
20 OCT (293) 1963	Kuril Is.	ATE	7.5	−6.42	g		
13 AUG (225) 1965	Vanuatu	ATE	2	−5.88	d		
10 JUN (161) 1975	Nemuro-Oki.	ATE	0.8	−6.43	g		
19 DEC (353) 1982	Tonga	PTE	2.0	−5.76	h		
02 SEP (246) 1992	Nicaragua	PTE	3.4	−6.30	h		
02 JUN (153) 1994	Java	PTE	5.3	−6.01	h		
21 FEB (052) 1996	Chimbote, Peru	PTE	2.2	−5.94	h		
26 DEC (361) 2004	Sumatra-Andaman	PTE	1200	−6.40	h		
17 JUL (198) 2006	Java	PTE	4.6	−6.13	j		
25 OCT (298) 2010	Mentawai	ATE	6.8	−6.22	j		
27 AUG (240) 2012	El Salvador	PTE	1.3	−6.42	j		
06 FEB (037) 2013	Santa Cruz	PTE	9.4	−5.94	d		

* PTE: Primary Tsunami Earthquake; ATE: Aftershock Tsunami Earthquake [*Okal and Saloor,* 2017]; L: Tsunami enhanced by ancillary landslide.

† While the event is smaller than its predecessor two months earlier, the moment ratio (3/4) is too close to qualify the second event as an aftershock of the first one.

References: a: *Tanioka and Satake* [1996]; b: *Salaree and Okal* [2018]; c: *Okal and Borrero,* [2011]; d: *Okal and Saloor* [2017]; e: *López and Okal,* [2007]; f: *Okal and Newman* [2001]; g: *Fukao* [1979], *Okal et al.* [2003]; h: *Newman and Okal* [1998]; i: *Stein and Okal* [2007], *Choy and Boatwright,* 2007]; j: *Saloor and Okal* [2018].

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Data and Resources

Digital Dutch newspaper reports from the Netherlands and colonial Indonesia are available from Delpher (http://www.delpher.nl). Figures were created using freely available Generic Mapping Tools (*Wessel and Smith,* 1998) and QGIS software (http://download.qgis.org/). Bathymetry was digitised for Nias and Simeulue from AMS T503 Series maps (1:250,000; sheets NA 47-5, NA 47-8, NA 47-9, NA 47-13, NA 47-14, NA 46-8) available via the Perry-Castañeda Map Collection at the University of Texas at Austin (http://www.lib.utexas.edu/maps/ams/indonesia/, last accessed July 2017). We also use the General Bathymetric Chart of the Oceans (GEBCO) digital bathymetry dataset (http://www.gebco.net/, last accessed in October 2014). Digital topographic contours (1:5000) are available from Badan Informasi Geospasial (http://tanahair.indonesia.go.id; last accessed 27 November 2017). The topographic data used in the tsunami simulations is bare-ground Shuttle Radar Topography Mission (SRTM) data downloaded from https://data.bris.ac.uk/data/dataset/10tv0p32gizt01nh9edcjzd6wa (last accessed, April 2018). For some key locations including Lakubang, Latak Ayah, Lukon, and Simeulue Cut, we replace the SRTM data with 4-m resolution digital terrain data purchased from NTT Data Corporation. Topographic maps (1:50,000) prepared by the 653rd Topographic Engineering Battalion of the United States Air Force for Simeulue (Series HIND 605 Sheet 5; NLA shelf mark: MAP G8082.S5 s50) and by the British War Office for Nias (Series HIND 614 Sheet 9; NLA shelf mark: MAP G8082.N5 s50) were procured from the National Library of Australia (NLA).

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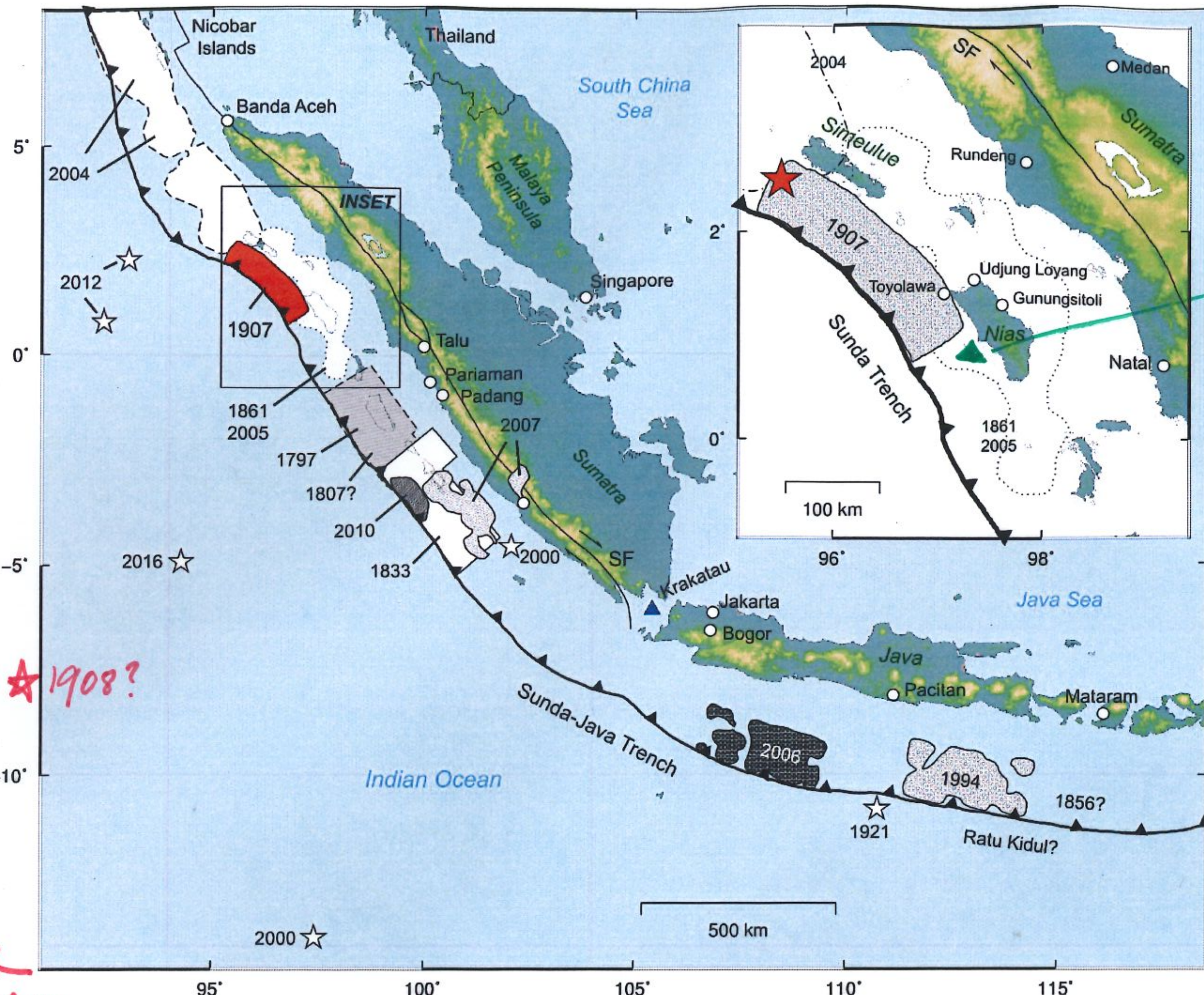


Event II

*1908?

extend to 90° to add 1908?

Figure 1: Generalized tectonic map of the Sunda-Java trench in Indonesia. First-order rupture of the 1907 earthquake (this study) indicated by a filled red polygon. Rupture areas for the 1797, 1833, 1861, 1994, 2004, 2006, 2007, and 2010 earthquakes are also shown (*Bilek & Engdahl*, 2007; *Chlieh et al.*, 20XX; *Konca et al.*, 2006; *Bilek & Engdahl*, 2007; *Konca et al.*, 2008; *Hill et al.*, 2010). Locations of possible historical tsunami earthquakes in Ratu Kidul, 1807, and 1856 discussed in the text are labelled and appended by question marks. Stars represent selected earthquakes including recent earthquakes in the Wharton basin. **BASEMENT RIDGES TO BE ADDED TO FIGURE**

do not clutter!

green triangle in inset indicates preferred location of Event II.

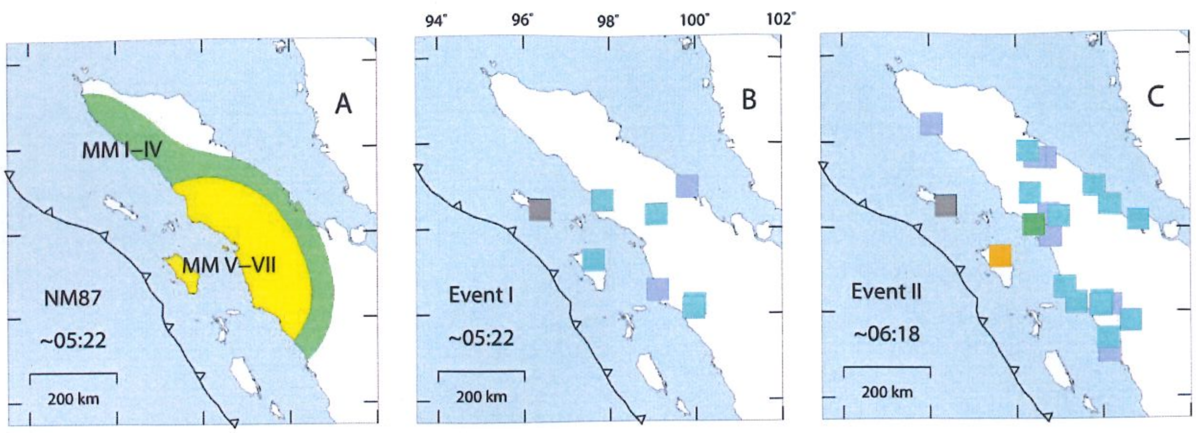


Figure 2: Intensity map (a) from *Newcomb & McCann* (1987) in comparison with intensities determined by our study for the mainshock at 05:22 GMT (b), and the largest aftershock at 06:18 (c). Grey coloured boxes indicate that an earthquake was felt but that macroseismic data were insufficient ~~with which~~ to assign an intensity.

Note that reloc. put it at 05:19 !

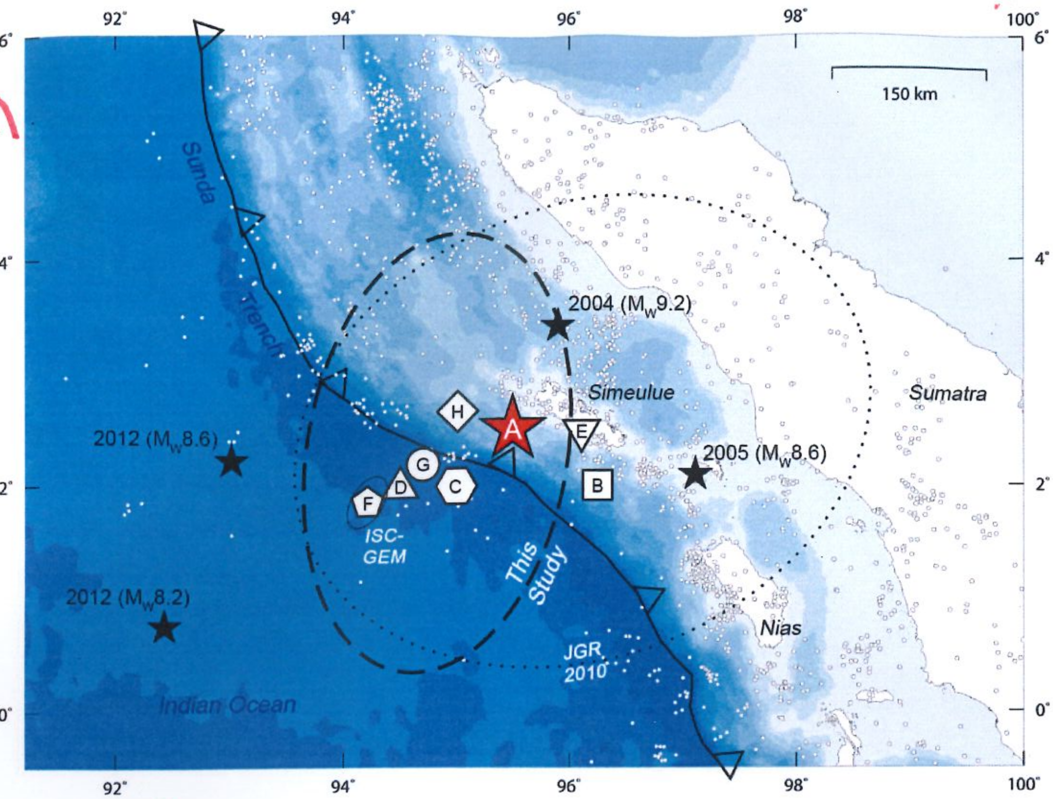
Scale of palet for squares ?

Note: grey not sufficiently different from light purple (itself too close to sea background).

Suggest for those open squares.

62.

Need Mentawai 2010 (and, B 2007)



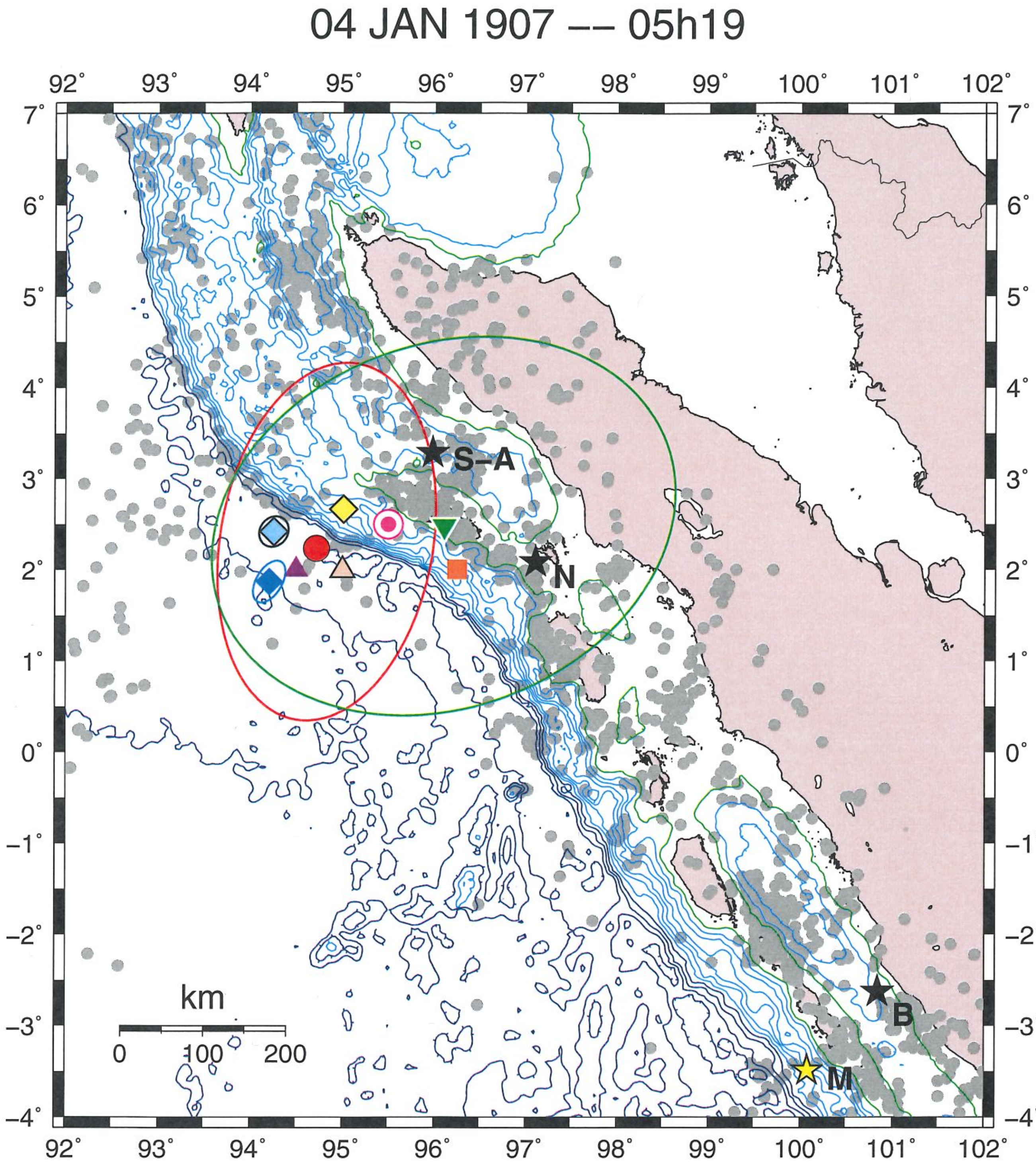
change order

Figure 4: The red star shows our relocation solution with associated Monte Carlo ellipse (dashed line). Epicentres by b) Staverman from Szirtes (1912), c) Gutenberg & Richter (1954), d) e) *Kanamori et al.* (1990) along with the confidence ellipse described in their Appendix (dotted ellipse). The brown upward triangle is Gutenberg and Richter's [1954] epicenter, and the orange square T. Staveman's [Szirtes, 1912]. The blue diamond (with minuscule ellipse) shows the ISC GEM epicenter, and the yellow one is the best estimate from a grid search [D. Di Giacomo, pers. comm., 2017]. The black stars show the epicenters of of 2004 Sumatra-Andaman, 2005 Nias, and the 2012 Wharton basin earthqakes. Background seismicity (1918−2015; h < 100 km) shown as light grey dots.

You seem to have been using my old caption after redrafting it totally. No longer matches

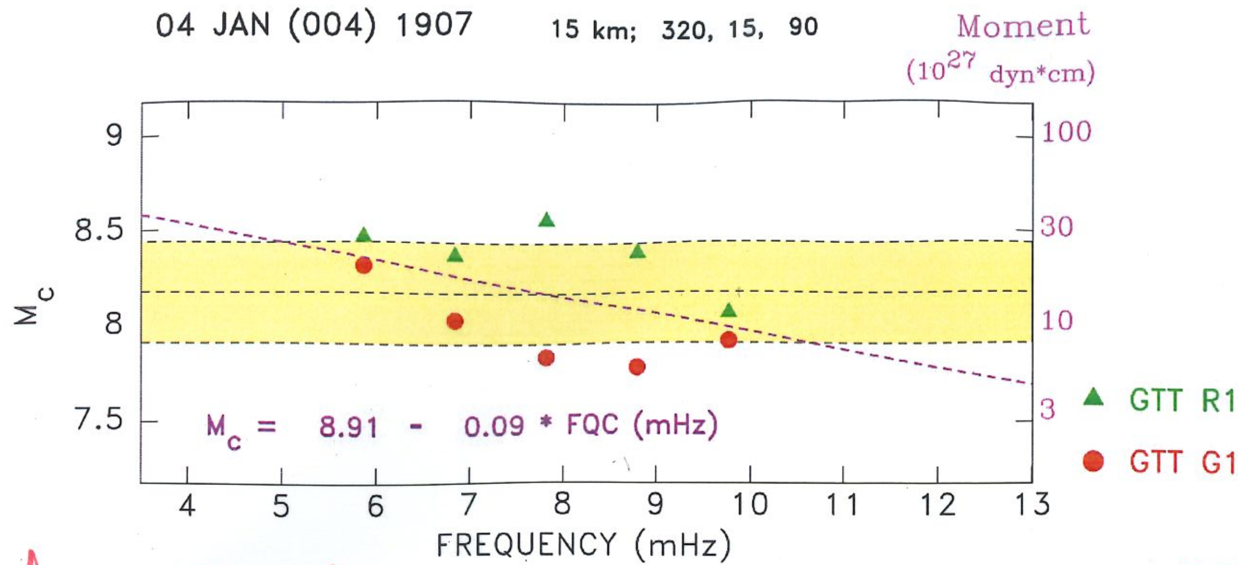
at which magnitude?

See "my" new version attached which I still prefer!



Added to Figure 4xx.: ... The blue diamonds (with respective confidence ellipses) are the ISC solution by *Storchak et al.* [2013] (solid, bright), and from the ISC–GEM (5.0) catalog (outlined, light); note that the two ellipses do not intersect, suggesting that they are deceptively small. The light brown triangle is *Turner et al.*'s [1912] location.

64



change order !

Figure 3: Spectral amplitudes of Rayleigh and Love waves at Go¨ttingen, interpreted as corrected mantle magnitudes Mc (*Okal & Talndier*, 1989), computed in the geometry f = 320°, d = 10°; l = 90° in the freaquency band 6–10 mHz. The black dashed lines and yellow band show the average value and standard deviation of the full dataset, while the purple dashed line shows its best linear regression. Note the strong increase of moment with period.

65

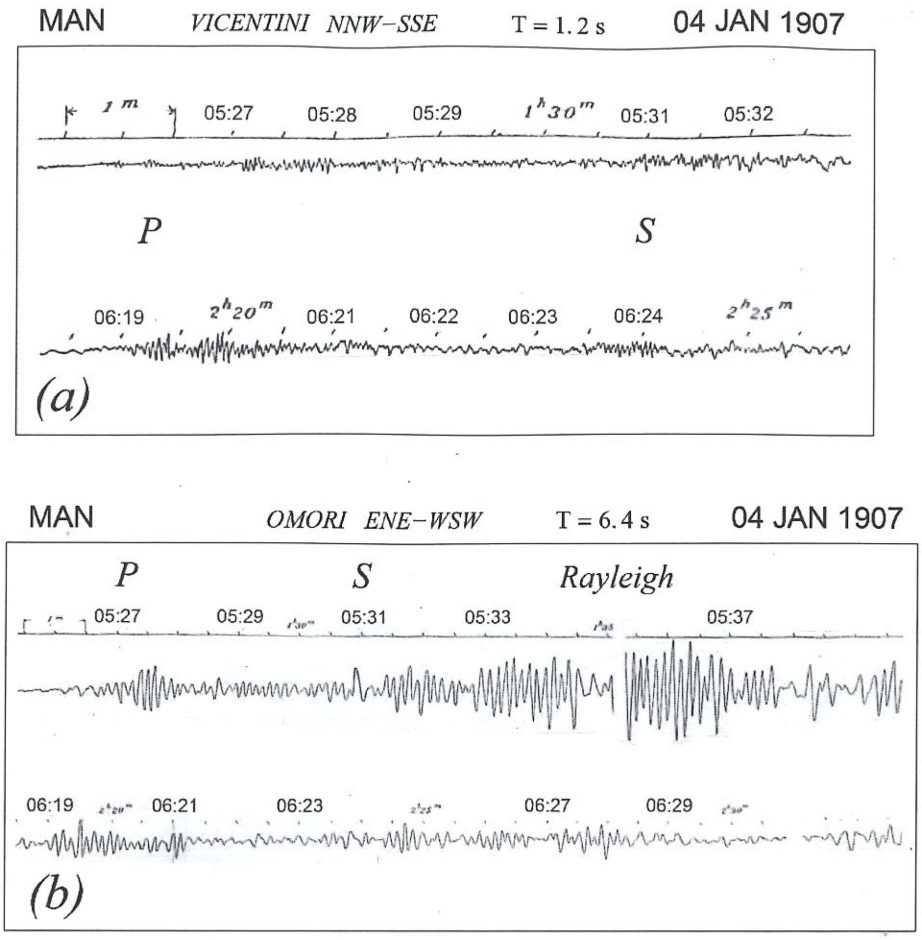


Figure 5: Seismic recordings at Manila (distance 29°; back-azimuth 247°), reproduced from *Maso* (1907), clearly showing the two events, separated by ~53 minutes. (a) Short-period Vicentini seismograph; (b) Omori seismograph. In both instances, the records have offset to align corresponding phases vertically. Time marks are at 30–s intervals, with original times given as local p.m. times (GMT +8). Standard times (GMT) are overprinted in dark blue. These records illustrate dramatically the different characteristics of the two events.

66.

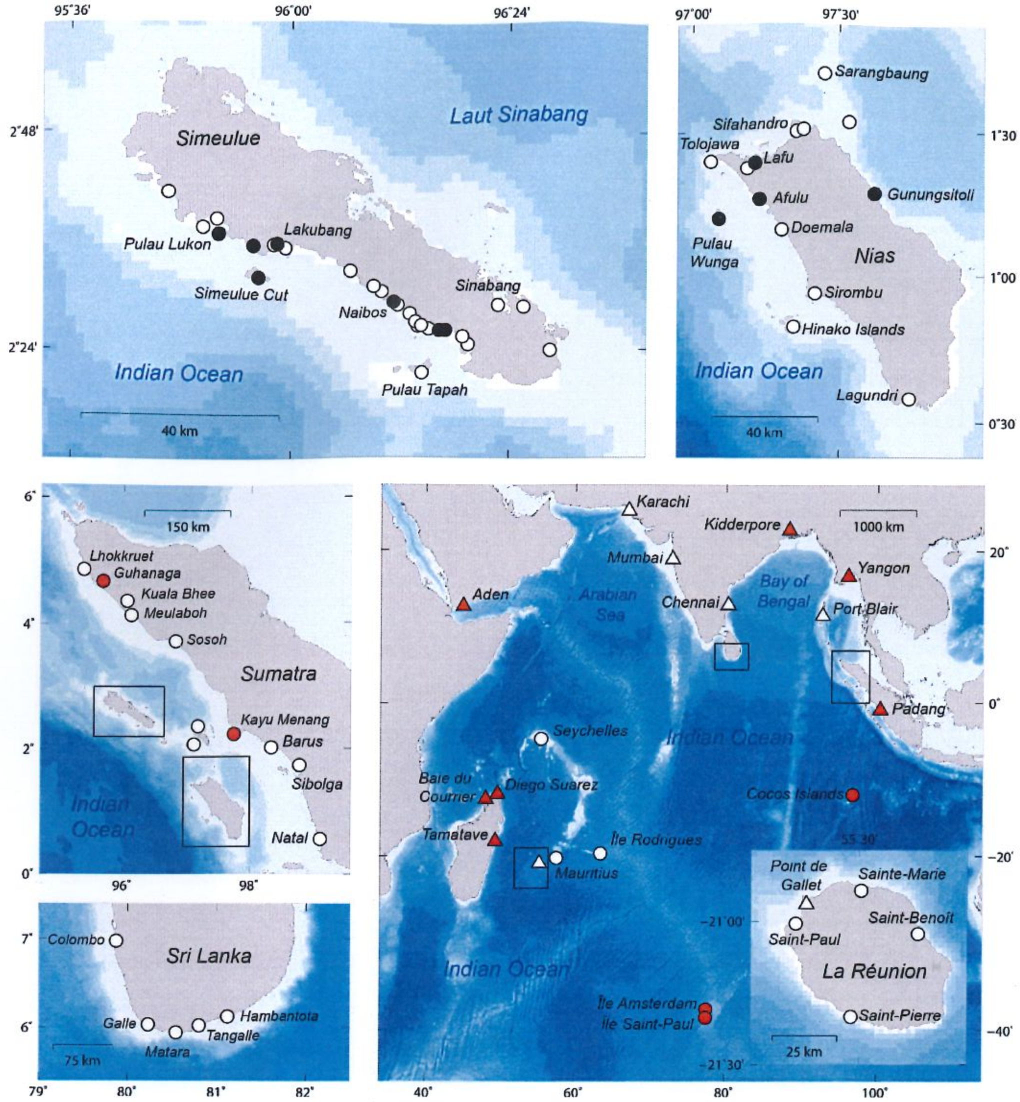


Figure 6: Tsunami observations on Simeulue (a), Nias (b), northern Sumatra (c), and in the Indian Ocean basin (d) including Sri Lanka (e) and La Réunion (f). Red circles mark questionable or false records. White triangles show tide gauges where readings were available, and red triangles show tide gauges in operation in 1907 where data was either unavailable, or the tsunami was unrecorded, or records were incomplete.

Difference between solid and open circles?

68

Figure X : Pulau Wunga (Schroeder, 1917b)

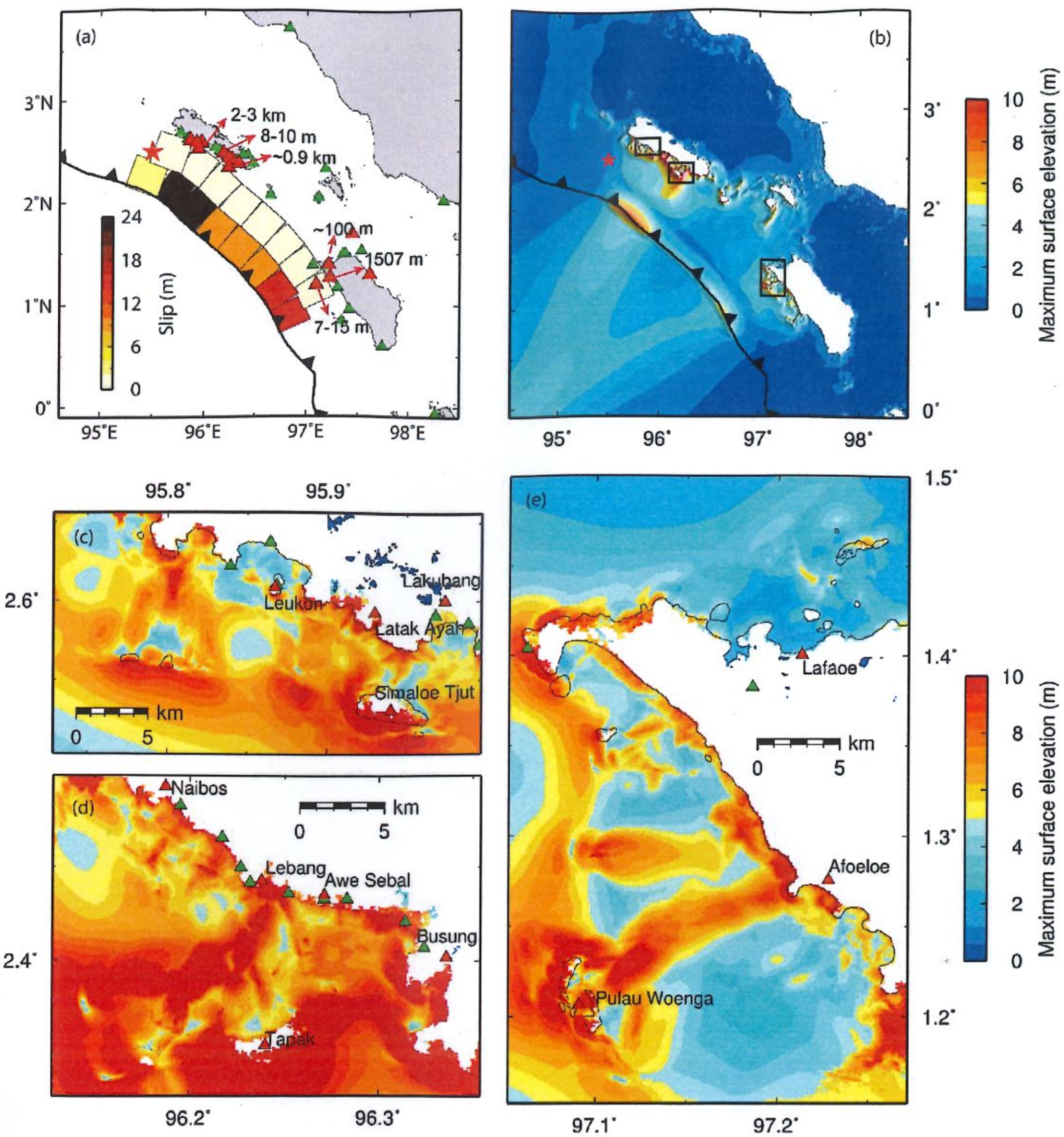


Figure 7: Hypothetical rupture geometry and slip parameters for the 1907 mainshock (a) estimated in this study (7A) off Simeulue (5A) and Nias (5B). Labels with red arrows in 7A indicate locations with estimates of wave weights, inundation distances, or flow depths.

70.

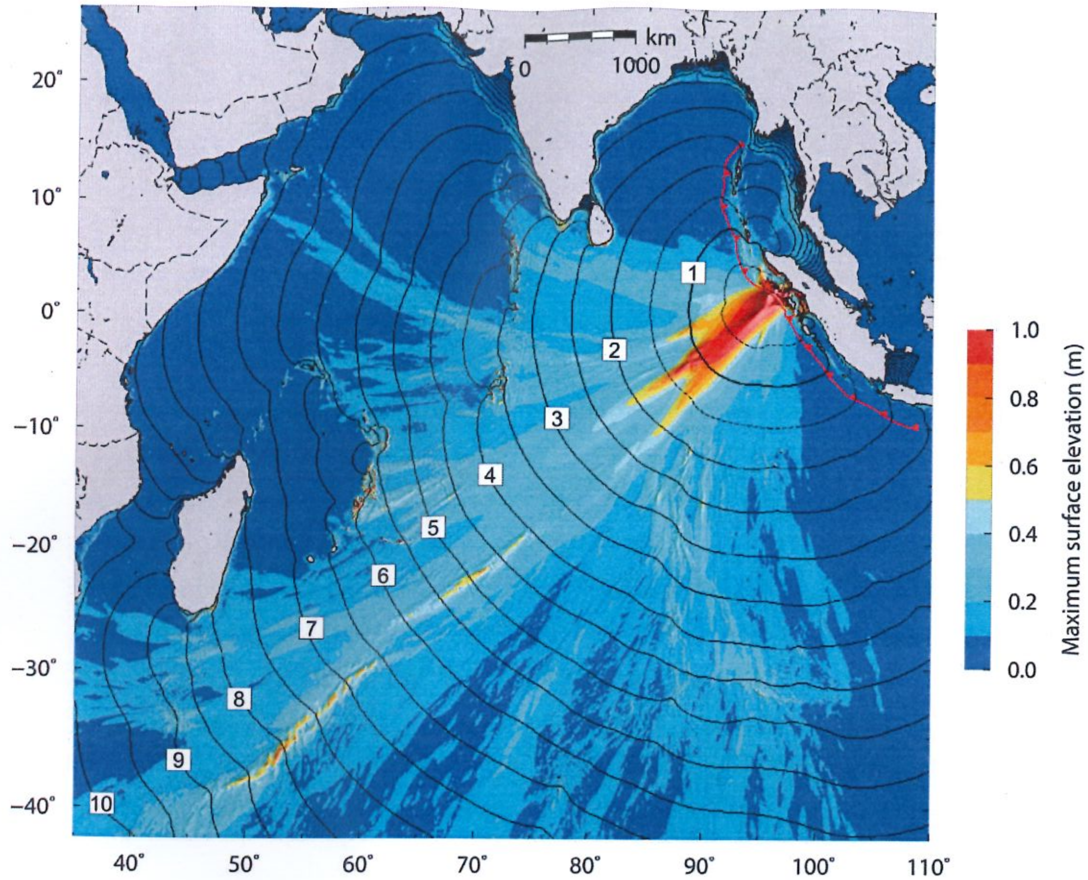


Figure 8: Tsunami propagation in the Indian Ocean basin and predicted arrival time contours.