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Discovery of possible mega-thrust earthquake along the Seram Trough from records of 1629 tsunami in eastern Indonesian region

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Abstract Arthur Wichmann's "Earthquakes of the Indian Archipelago" documents several large earthquakes and tsunami throughout the Banda Arc region that can be interpreted as mega-thrust events. However, the source regions of these events are not known. One of the largest and well-documented events in the catalog is the great earthquake and tsunami affecting the Banda Islands on August 1, 1629. It caused severe damage from a 15-m tsunami that arrived at the Banda Islands about a half hour after violent shaking stopped. The earthquake was also recorded 230 km away in Ambon, but no tsunami is mentioned. This event was followed by at least 9 years of uncommonly frequent seismic activity in the region that tapered off with time, which can be interpreted as aftershocks. The combination of these observations indicates that the earthquake was most likely a mega-thrust event. We use an inverse modeling approach to numerically reconstruct the tsunami, which constrains the likely location and magnitude of the 1629 earthquake. Only, linear numerical models are applied due to the low resolution of bathymetry in the Banda Islands and Ambon. Therefore, we apply various wave amplification factors (1.5–4) derived from simulations of recent, well-constrained tsunami to bracket the upper and lower limits of earthquake moment magnitudes for the event. The closest major earthquake sources to the Banda Islands are the Tanimbar and Seram Troughs of the Banda subduction/collision zone. Other source regions are too far away for such a short arrival time of the tsunami after shaking. Moment magnitudes predicted by the models in order to produce a 15-m tsunami are M_w of 9.8–9.2 on the Tanimbar Trough and M_w 8.8–8.2 on the Seram Trough. The arrival times of these waves are 58 min for Tanimbar Trough and 30 min for Seram Trough. The model also predicts 5-m run-up for Ambon from a Tanimbar Trough source, which is inconsistent with the historical records. Ambon is mostly shielded from a wave generated by a Seram Trough source. We conclude that the most likely source of the 1629 mega-thrust earthquake is the Seram Trough. Only

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one earthquake >M_w 8.0 is recorded instrumentally from the eastern Indonesia region although high rates of strain (50–80 mm/a) are measured across the Seram section of the Banda subduction zone. Enough strain has already accumulated since the last major historical event to produce an earthquake of similar size to the 1629 event. Due to the rapid population growth in coastal areas in this region, it is imperative that the most vulnerable coastal areas prepare accordingly.

Keywords Tsunami modeling · Indonesia · Banda Arc · Seram Trough · Tanimbar Trough · Banda Islands · Ambon · Mega-thrust earthquakes

1 Introduction

During the twentieth century, Indonesia had around two hundred major earthquakes (M_s 7.5 or greater), more than all of North America or South America during the same time interval (Harris et al. 1997). At least 110 of these earthquakes were destructive; the majority jolting densely populated western Indonesia (Fig. 1).

These high rates of seismic activity in the past century are consistent with the recorded history of Indonesia, as documented in a compilation of geophysical events from the seventeenth to nineteenth centuries by Wichmann (1918). The reliability of the earthquake and tsunami catalog is demonstrated by the recent recurrence of several earthquakes that ruptured similar fault segments and are of similar magnitudes to those inferred from earlier accounts. For example, the 2005 northern Sumatra earthquake near Nias Island ruptured

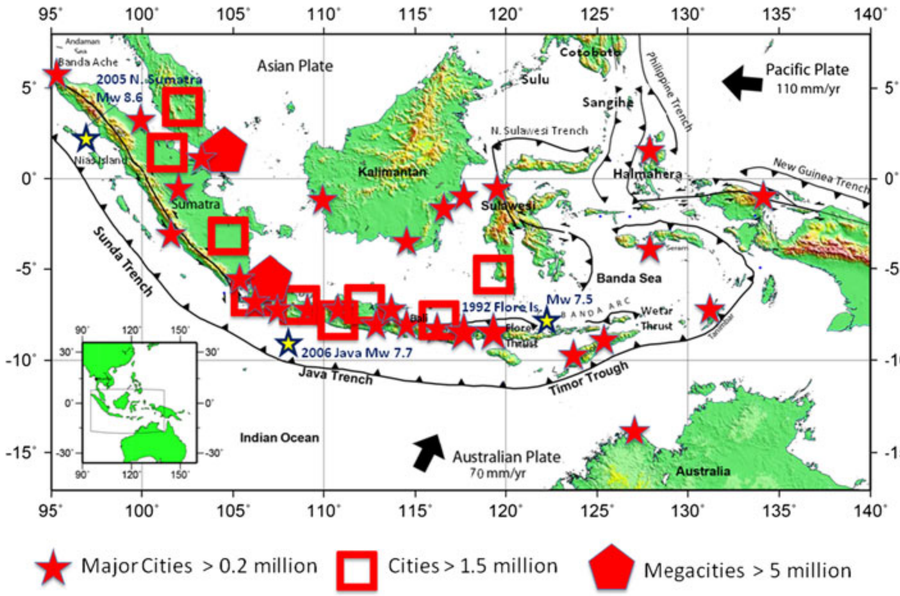

Fig. 1 Map of major cities and proximity to major plate boundary segments in the Indonesia region. The Indo-Australian Plate moves NNW at a rate of around 70 mm/year. The Pacific Plate moves WNW and sideswipes the northern part of the Banda Arc at a rate of 110 mm/year. *Blue stars*—epicentral locations and magnitudes of earthquakes mentioned in the paper

nearly the same area estimated by Newcomb and McCann (1987) from accounts in the Wichmann catalog of an earthquake and tsunami in 1861.

The M_w 9.2 Banda Ache mega-thrust earthquake, and the spatial and temporal clustering of events that followed, has stimulated new interest in the earthquake history of the western Sunda Arc. All eyes are presently on the segment of the Sumatran subduction zone to the south of Nias, which last ruptured in 1833. From reports of this event and the tsunami that followed that are recorded in the Wichmann catalog Newcomb and McCann (1987) estimate, it could have been up to M_w 9.0.

The Sunda arc-trench system continues to the east of Sumatra adjacent to the densely populated islands of Java and Bali, then becomes the Banda Arc in the Timor region (Fig. 1). Although no earthquakes larger than M_w 8.0 are documented in the Java–Bali region, there are several earthquakes in the Banda Arc region documented before 1900 by Wichmann with similar characteristics to those in Sumatra. However, the source parameters of the large events are unknown. No large earthquakes in the Banda Arc and eastern Indonesia region are included in the recent summary of mega-thrust earthquake events worldwide by Heuret et al. (2012), which is based only on the last 100 years of instrumental records. When these largely unknown events reoccur in eastern Indonesia, it will affect an order of magnitude more people and urban centers than before (Fig. 1), which raises the stakes on determining the most likely large earthquake and tsunami sources in the region. This paper investigates the most likely source of one of the largest tsunami events documented in eastern Indonesian region, which occurred in 1629.

2 Tectonic setting

The Banda Arc region occupies a convergent triple junction of three of Earth's largest plates. Relative to the Sundaland block, a tectonic domain that is extruding eastward away from Eurasian Plate (Rangin et al. 1999), the Indo-Australian Plate converges NNE at a rate of around 70 mm/a (Nugroho et al. 2009). The Indo-Australian Plate subducts beneath the Banda Arc from three sides along the Timor, Tanimbar and Seram Troughs. These troughs are underthrust mostly by Australian continental margin lithosphere that is attached to old Indian Ocean lithosphere at deeper levels of the subduction zone (Hamilton 1979). The resulting arc-continent collision bends 180° around the western edge of NW Australia (Fig. 2). Another major tectonic player in the region is the Pacific Plate. It moves WNW at a rate of around 110 mm/a relative to the Sundaland Plate (Rangin et al. 1999) and sideswipes the northern part of the Banda Arc along an array of plate boundary segments linked by the left-lateral Sorong fault (Figs. 1, 2).

The NW part of the Australian continental margin first began to subduct beneath the Banda Arc at around 8 Ma in the Timor and Seram regions (Berry and McDougall 1986; Linthout et al. 1996; Harris 2011). The arc-continent collision has now propagated around the embayment south of the Bird's Head and involves the entire Banda Arc region (Hall 2002). Underthrusting of the Australian continental margin increases coupling along the subduction interface as indicated by partitioning of strain away from the deformation front into the forearc (Reed et al. 1986; Harris et al. 2009) and back arc regions (Silver et al. 1983). In the backarc, the volcanic arc is thrust over the southern edge of the Banda Sea Basin (Fig. 2) along the Flores and Wetar Thrusts (Breen et al. 1989). These south-dipping backarc thrust systems (Figs. 1, 2) currently take up a significant amount of the convergence between the Australia and Asia plates (Mccaffrey and Nabelek 1984; Genrich et al. 1996). The amount of movement along these thrust systems decreases to the west where

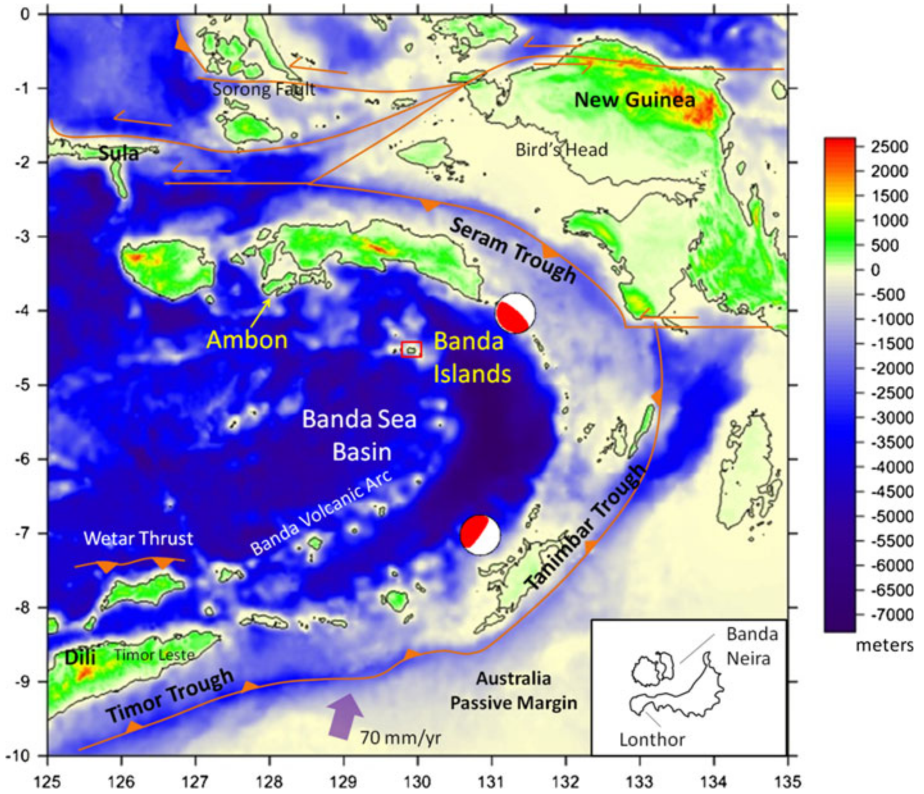

Fig. 2 Relief and bathymetry model of the Banda Sea and surrounding islands of the region. Active faults are shown in *orange*. The bathymetry is based on ETOPO-1 data with a grid size of 1 min. At this resolution, the Banda Islands of Banda Neira and Lonthor (*inset*) are identified, as is the shallow embayment protecting the city of Ambon. Two fault plane solutions we used for the modeling are marked on this map. See Sects. 4.1 and 4.2

the oblique collision is less developed (Silver et al. 1983). Even in the most advanced part of the collision in the Timor region, there is still up to 21 mm/a of convergence measured across the Timor Trough (Nugroho et al. 2009), which means that a significant amount of elastic strain energy is accumulating along this plate boundary (Harris 2011).

The low rate of seismicity along the Timor, Tanimbar and Seram Troughs is commonly cited as evidence that they are no longer active (Mccaffrey and Nabelek 1984). However, seismic reflection profiles across the deformation front at the Seram, Tanimbar and Timor Troughs show thrust faults breaking all of the way to the surface (Schluter and Fritsch 1985; Karig et al. 1987; Pairault et al. 2003). We interpret the low seismic slip rates of the Banda collision zone as evidence of a locked plate boundary interface, which potentially could produce mega-thrust earthquakes. There are several small thrust-mechanism earthquakes along the low-angle plate interface of both the Timor and Seram Troughs. Although active thrusting is observed in earthquakes along the plate boundary in the Tanimbar Trough region, there are many extensional and strike-slip events as well (Mccaffrey and Nabelek 1984).

The backarc thrust systems may also produce large earthquakes. The mapped length of the Wetar Thrust is around 350 km, which may be too short for a mega-thrust earthquake event. However, the Flores Thrust is at least 500 km in length, which, if the whole thrust system ruptured, may produce a mega-thrust earthquake.

The working definition here for a mega-thrust event is an interplate earthquake that occurs along a subduction interface, that is, $M_w \geq 8.5$ and is associated with uncommonly large rupture lengths (>400 km). The fault plane commonly has a shallow dip (<10°) and the earthquake, a shallow hypocenter (<30 km), which at a subduction zone can produce large tsunami.

Most mega-thrust earthquakes of the magnitude needed to produce the tsunami observed in the Banda Islands have rupture lengths of >500 km, such as 2010 M_w 8.8 Maule "Chile" earthquake (Wang et al. 2012) and 2011 M_w 9.0 Tohoku earthquake (Suzuki et al.2011). However, the most significant problem with either of the backarc thrust systems sourcing the 1629 tsunami is their distance (>450 km) from the Banda Islands where the tsunami arrived only 30 min after intense shaking (see Sect. 2.1 below).

It is also possible that unknown active faults exist close enough to the Banda Islands to generate a tsunami that arrives within 30 min. For example, active strike-slip faults are inferred to the west of Banda Neira (Mccaffrey and Nabelek 1984) that, with a significant component of oblique slip, may produce a large tsunami if the rupture zone parameters are large enough. It is also possible that an earthquake-induced landslide generated the tsunami (Brune et al. 2010). Nevertheless, the extent to which the earthquake was felt, and the prolonged seismicity following the event, indicates that the earthquake was larger than any event occurring over the past 100 years of instrumental records.

2.1 Testing for mega-thrust earthquakes

Records of geophysical events throughout the Indonesian region date back to 1600 and are mainly compiled in "The Earthquakes of the Indian Archipelago" by Authur Wichmann (1918). This volume documents several large earthquakes and tsunami throughout the Banda Arc region that are characteristic of mega-thrust earthquake events.

It is possible that one of these events happened on August 1, 1629, in the eastern Banda Sea region:

> A half hour after the termination of a violent seismic shock there formed in the sound … a high mountain of water. The tidal wave rolled westward straight against fort Nassau on Banda Neira, as well as the village on the beach …where it achieved a height of 9 fathoms [15.3 m] above the springtide stand. The mole built of stone before the fort was beaten away and the water penetrated into the fort with such force, that a 3,500 pound heavy mass of iron was displaced by 36 feet [11.3 m].

The records indicate that the tsunami arrived at the Banda Islands 30 min after the earthquake occurred. Based on the tsunami propagation equation:

$$V = \sqrt{gD} = d/t$$

where V is velocity of the tsunami wave, g is gravity, D is the depth of the bathymetry, d is the distance between rupture and the Banda Islands, t is the tsunami arrival time, and the reasonable distance between the source of the tsunami and the Banda Islands is about 200–300 km.

In the Wichmann records, the earthquake was also felt in Ambon, which is 210 km to the NW, but there is no record of a tsunami there. An unusually high rate of seismicity in the region followed the 1629 event and increasingly diminished after 9 years. Fifteen years after the 1629 earthquake, there were several other large earthquakes in and around Ambon, some of which caused tsunami. We interpret this temporal and spatial clustering of seismicity as stress contagion from the 1629 earthquake, which caused a series of aftershocks. The record of the possible aftershocks and the distance over which the earthquake was felt argues for a large, plate boundary earthquake as the most likely source of the tsunami. The extent over which the earthquake was felt may have been much larger than noted due to the lack of European outposts keeping records in the region during the seventeenth century.

The only plate boundary source region of a possible mega-thrust earthquake, that is, within 200–300 km from Banda Neira is the easternmost Banda arc-continent collision zone between Seram and Tanimbar (Fig. 2). This section of the collision zone is segmented around the Aru Trough region (Schluter and Fritsch 1985). We use this segment boundary to constrain the southern limit of Seram Trough rupture and the northern limit of Tanimbar Trough rupture (Fig. 2). Other plate boundary segments in the region, such as the Timor Trough, the Wetar or Flores backarc thrust, or the Sulawesi, Sangihe, Halmahera, Cotabato, New Guinea and Philippines Trenches, are too far away to generate a 15-m tsunami that arrives 30 min after violent shaking in the Banda Islands (Fig. 1). Seismological evidence is lacking for how the Seram, Tanimbar or Timor Troughs are segmented.

3 Method

The numerical solutions used for the model simulations are those of Satake (2002); Ma and Lee (1997) and Ma et al. (1991). The bathymetry for the model is based on ETOPO-1 data (Amante and Eakins 2009) with a grid size of 1 min (about 1.6 km) (Fig. 2). At this resolution, the Banda Islands of Banda Neira and Lonthor are identified, as is the shallow embayment protecting the city of Ambon (Fig. 2).

3.1 Vertical seafloor deformation

Vertical seafloor deformation of the simulated earthquakes is modeled using the method by Okada (1985), which computes ground deformation caused by faulting in a homogeneous half-space. Since the fault parameters of the 1629 earthquake are unknown, we use those from focal mechanisms of instrumented, but smaller events along what is likely the collisional interface (Global CMT database, http://www.globalcmt.org/CMTsearch.html). Fault width and length are calculated from empirical equations (Table 1) derived from earthquake observations (Wells and Coppersmith 1994; Ma and Lee 1997).

3.2 Tsunami propagation simulation

From the vertical surface deformation and bathymetry models, tsunami propagation is computed using a finite difference method. Here, we use the linear relationship between tsunami amplitude and fault slip (Ma and Lee 1997; Satake 1995). We assume that the water surface is uplifted instantaneously exactly in the same way as the bottom

Table 1 Empirical equation used for all numerical simulations, based on Wells and Coppersmith (1994)

Empirical equation of thrust fault
$M_w = (5.00 \pm 0.22) + (1.22 \pm 0.16)\log(L)$
$M_w = (4.37 \pm 0.16) + (1.95 \pm 0.15)\log(W)$
$M_w = (6.52 \pm 0.11) + (0.44 \pm 0.26)\log(D)$

deformation. Computations were made for a total duration of 8 h at time increments of 5 s. A linear model is used to compute wave amplitudes along the coastal regions of Banda Island and Ambon that are based on the ratio of computed tsunami run-up heights to observed heights, which is known as amplification factor.

3.3 Tsunami run-up amplification factor

We calibrated the amplification factor by modeling the nearby 1992 tsunami of Flores Island (Hidayat et al. 1995) and 2006 slow earthquake tsunami in Java (Koshimura 2006) (Fig. 1). These two very different tsunamigenic earthquakes in the region provide the widest possible range of direct measurements of wave and run-up heights.

3.3.1 December 12, 1992, Flore Island, Indonesia, earthquake

On December 12, 1992, an M_w 7.7 earthquake occurred near the north shore of Flores Island along the Flores backarc thrust system (Fig. 1). The earthquake and the ensuing tsunami killed around 4,000 people. Most tsunami run-up heights along the northern shore of Flores Island were 2–5 m. Here, we reproduce the same fault model used by Hidayat et al. (1995) who derive fault parameters from inverting teleseismic, broadband P and SH waves, as well as PP waves for the seismic moment rate tensor. The model uses a two-fault source to generate vertical seafloor deformation. The fault parameters and vertical deformation can be found in Hidayat et al. (1995). The only difference between earlier models and ours is the higher bathymetric resolution we use. Hidayat et al. (1995) use the ETOPO-5 data with grid size of 5 min versus the much higher-resolution ETOPO-1 model we use with a 1-min grid size. Previous studies (Satake 1995) show that observed and theoretical tsunami heights are closer with a smaller grid size.

Tsunami run-up heights were measured along the northeast coast of Flores Island by multiple international survey teams (Table 2). However, only one tide gauge, Palpo, which is 650 km north of Flore Island, recorded open ocean wave heights (Fig. 3). Comparing wave heights computed by our model waveforms with those observed by the tide gauge yields an amplification factor of 1.5 for this tsunami (Table 2). These results improved upon those of Hidayat et al. (1995) by better fitting the observed run-up heights. Locally, tsunami run-up heights were amplified by a submarine landslide caused by the earthquake.

3.3.2 July 17, 2006, south of Java Island, Indonesia, earthquake

A large earthquake (M_w 7.7) occurred along the subduction interface south of Java Island on July 17, 2006, which generated a tsunami causing around 800 casualties. Maximum run-up heights along the southern shore of Java are from 1 to 4.6 m (Hariri and Bilek 2011). Here, we reproduce the fault model in Koshimura (2006) based on Harvard CMT fault parameters, but with the higher-resolution bathymetry (ETOPO-1 versus ETOPO-2). Run-up heights (Table 3) are taken from those measured along the southern Java coast by Kongko et al. (2006). Comparison of our modeled tsunami heights with the observed data indicates an amplification factor of around 4 in this case.

However, the Java earthquake is a unique case of slow rupture that was hardly felt by those on shore (Koshimura 2006). Kato et al. (2007) made GPS observations and tsunami height measurements during the event period, but no co-seismic displacement was detected. In their paper, the observed tsunami heights are systematically higher than those predicted from numerical simulations based on seismic wave analysis, which indicates that

Table 2 1992 Flores earthquake observed and calculated tsunami run-up heights (in m) based on Hidayat et al. (1995)

Station	Name	Lat	Lon	Observed (m)	Two-fault model by Hidayat et al. (1995)	Model predicted (m) with amplification factor 1.5
1	Mage, Palu Is.	−8.3	121.75	2.8	1.79	1.24
2	Mausanbi	−8.5	121.78	3.4	3.15	3.27
3	Awora	−8.48	121.85	2.9	2.90	4.71
4	Deteh	−8.53	122.03	2.3	5.88	3.62
5	Patisomba	−8.55	122.15	4	4.11	4.97
6	Nangahureh	−8.55	122.17	1.9	3.44	3.23
7	Wailiti	−8.57	122.18	2.1	3.43	3.42
8	Wuring	−8.6	122.2	3.2	3.67	3.29
9	Maumere	−8.62	122.23	3	2.95	2.82
10	Waioti	−8.63	122.27	2.5	3.46	3.47
11	Geliting	−8.63	122.28	3.3	3.63	3.84
12	Egon	−8.6	122.42	1.8	4.05	1.95
13	Wodung	−8.58	122.48	2.3	4.69	4.11
14	Nangahale	−8.55	122.5	1.5	3.58	3.83
15	Talobura	−8.52	122.52	2.4	5.52	2.94
16	Ngolo, Pomana Is.	−8.35	122.32	3.2	1.97	3.68
17	Buton, Pomana Is.	−8.33	122.33	1.5	2.23	3.41
18	Taot, Desar Is.	−8.87	122.35	2.8	1.33	0.01
19	Kusung, Besar Is.	−8.87	122.42	4.1	2.62	0.01
20	Permahan Is.	−8.45	122.45	3.4	3.21	3.02
21	Babi Is. N	−8.4	122.52	4	2.12	3.08
22	Babi Is. W	−8.42	122.5	7.1	2.40	2.72
23	Babi Is. S	−8.43	122.52	4	3.76	3.39
24	Babi Is. E	−8.42	122.53	5.6	2.69	2.40
25	Nebe	−8.45	122.53	4.6	5.11	3.39
26	Wailamung	−8.42	122.58	5.5	6.92	3.11
27	Larentuka	−8.37	122.98	1.8	1.97	1.17
28	Pantai Lato	−8.37	122.77	3.8	3.59	4.62

The tsunami run-up heights in last column of this table are predicted by our model

fault offset may be larger than estimates using seismic analysis and the rupture was very slow. Thus, in our model, we set the maximum limit of amplification as a factor of 4.

In summary, the tsunami run-up amplification factors of 1.5 and 4 that we determined from modeling two different earthquakes in the eastern Indonesia region provide upper and lower limits on the moment magnitudes for earthquakes we model on the Tanimbar and Seram Troughs.

4 Results

The most likely active faults that could produce the violent shaking (felt over 300 km radius), a tsunami with run-up heights at least up to 15 m, 9 years of possible aftershocks and perhaps stress contagion to nearby faults are the Tanimbar (South Source) and Seram

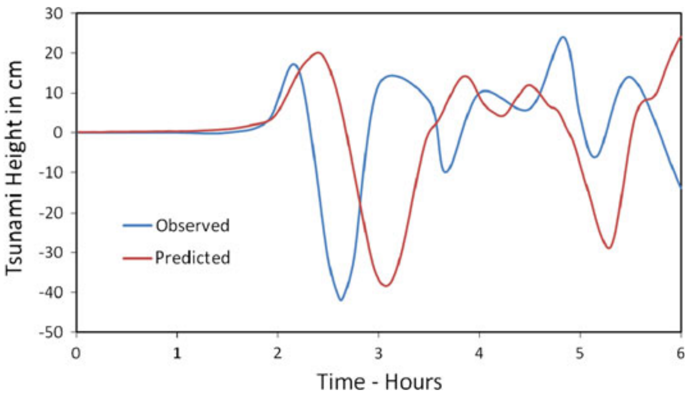

Fig. 3 Comparison of observed tsunami waveform (*blue*) and computed waveform (*red*) in the tide gauge, Palpo, which is 650 km north of the December 12, 1992, Flore Island, Indonesia, earthquake event

Table 3 2006 Java earthquake observed and calculated tsunami run-up height (in m)

Station	Name	Lat	Lon	Observed (m)	Model predicted (m) with amplification factor 4
1	Keburuhan	−7.88	109.90	2.5	1.3
2	Suwuk	−7.75	109.50	1.1	1.6
3	Ayah	−7.70	109.38	1.5	2.1
4	Widarapayung	−7.65	109.25	4.6	1.9
5	Bunton	−7.62	109.13	1.5	1.7
6	Cikembulan 1	−7.72	108.55	3.1	3.2
7	Cikembulan 2	−7.65	108.60	3.6	2.8
8	Pengandaran 1	−7.62	108.65	2.9	2.9
9	Pengandaran 2	−7.62	108.70	2.7	1.5

(East source) sections of the Banda arc-continent collision zone. The Seram Trough is from 220 to 300 km to the east of Banda Neira, while the Tanimbar Trough is at least 400–440 km away, respectively (Fig. 2). We primarily use the tsunami arrival times and wave heights predicted by earthquakes of various magnitudes to constrain which of the source regions are most likely. The direction of the wave and the fact that it was not observed in Ambon help us further constrain the most likely source region. We follow our modeling procedure to identify which fault parameters and range of earthquake magnitudes for these two potential sources best coincide with the historic records.

4.1 Tanimbar Trough (south source)

For fault parameters of the Tanimbar Trough, we extrapolate focal mechanism data associated with the 2004/10/22 earthquake event recorded in the Global CMT database (Tables 1, 4), which is consistent with seismic reflection profiles of the plate boundary interface (Schluter and Fritsch 1985). The epicenter of the earthquake is at 7.27°S and 130.5°E. The hypocenter is on the plate boundary interface shown by Welc and Lay (1987). A simple two-fault-segment model is used to match the curved shape of Tanimbar Trough.

Table 4 Fault parameters used for numerical simulations

Length (km)	Width (km)	Strike 1 (°)	Strike 2 (°)	Dip (°)	Depth (km)	Max slip (m)	Min slip (m)
South source–Tanimbar Trough (maximum M_w: 9.8; minimum M_w: 9.2)							
450	112	220	260	10	20	40.9	14.96
East source–Seram Trough (maximum M_w: 8.8; minimum M_w: 8.2)							
500	97	118	160	10	15	11.6	6.2

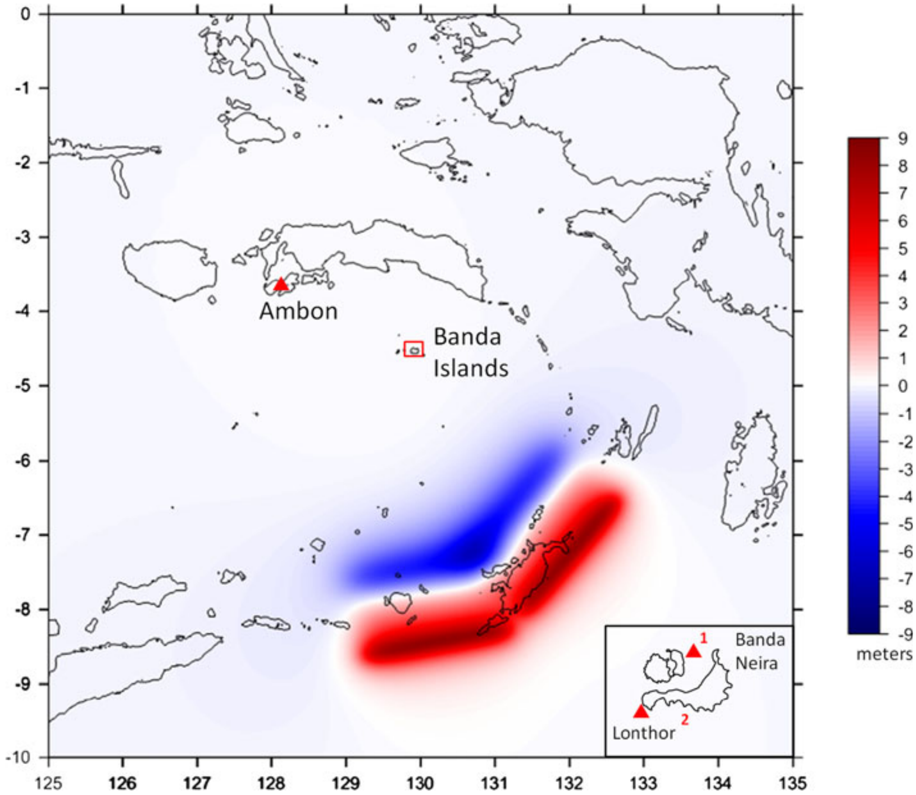

Fig. 4 The vertical component of surface deformation estimated from the two-fault-segment model of the Tanimbar forearc. The fault plane parameters can be found in Table 4. Uplift is shown as *red* and subsidence in *blue*. The city of Ambon and the simulated gauge station are indicated as a *red triangle*. *Triangles 1* and *2* represent the simulated gauge stations for Banda Neira and Lonthor islands

A vertical crustal deformation of 6 m is required to produce a tsunami as large as that observed in the Banda Islands with the maximum amplification factor of 4 (Fig. 4). According to the scaling relations, this deformation would require an earthquake of M_w 9.2 as a minimum magnitude. However, the maximum earthquake magnitude with amplification factor of only 1.5 is 9.8 for a tsunami run-up of 15 m in Banda Islands. The finite-element-based tsunami simulation predicts a tsunami arrival time at the Banda Islands of about 58 min after the initiation of the earthquake (Fig. 5). Although the numerical model predicts tsunami heights of 15 m, which are close to those observed at the Banda Islands, it also predicts first arrival tsunami heights >5 m for Ambon, which was not observed.

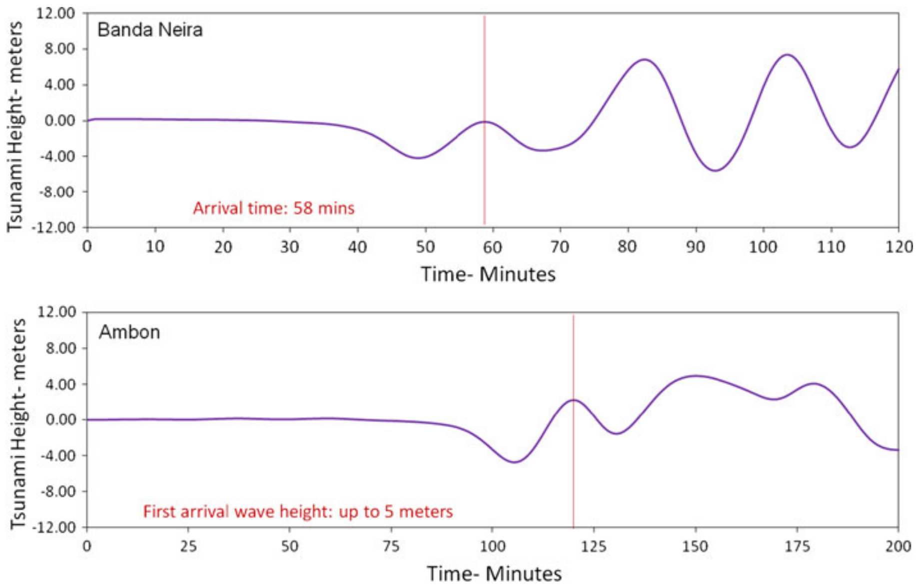

Fig. 5 Computed tide gauge records and tsunami waveform for Banda Neira and Ambon city stations from the Tanimbar Trough model earthquake. The tsunami arrival time at the Banda Islands is 58 min after the initiation of the earthquake, which is nearly twice as long as observed. The model also predicts tsunami run-up heights >5 m at Ambon city using the minimum amplification factor of 1.5, which was not observed

4.2 Seram Trough (east source)

For fault parameters of the Seram Trough, we use focal mechanism data from the 1993/12/04 earthquake (Table 4), which is along the boundary interface. A simple two-fault-segment model is employed to match the curved shape of the Seram Trough fault zone with L = 500 km, W = 97 km and D = 116 cm (Fig. 6 and Table 4). The epicenter of the model event is 4°S and 131.14°E. The maximum earthquake magnitude of M_w 8.8 is required to generate a 15-m tsunami in the Banda Islands if the minimum amplification factor of 1.5 is applied. The minimum earthquake magnitude of M_w 8.2 is estimated with a tsunami amplification factor of 4. The Seram trough source simulation yields a slip =11.6 m in order to produce a 15.3-m tsunami. This result is consistent with "Plafker's rule of thumb" (Okal and Synolakis 2004) that a seismic dislocation does not produce run-up heights much in excess of its own amplitude of slip. The minimum estimate of tsunami arrival at the Banda Islands is 30 min after the initiation of the earthquake (Figs. 7, 8). Another important result is that the best-fit model tsunami produces wave heights in Ambon of less than one meter, which is consistent with the observations of feeling the earthquake, but no tsunami noted.

4.3 Sensitivity analysis

We conducted a sensitivity analysis of our numerical methods in order to better understand how variations in hypocentral depth and fault dip from Seram Trough earthquake influence modeled tsunami heights (Fig. 9). The other fault parameters (Table 4) remained fixed for

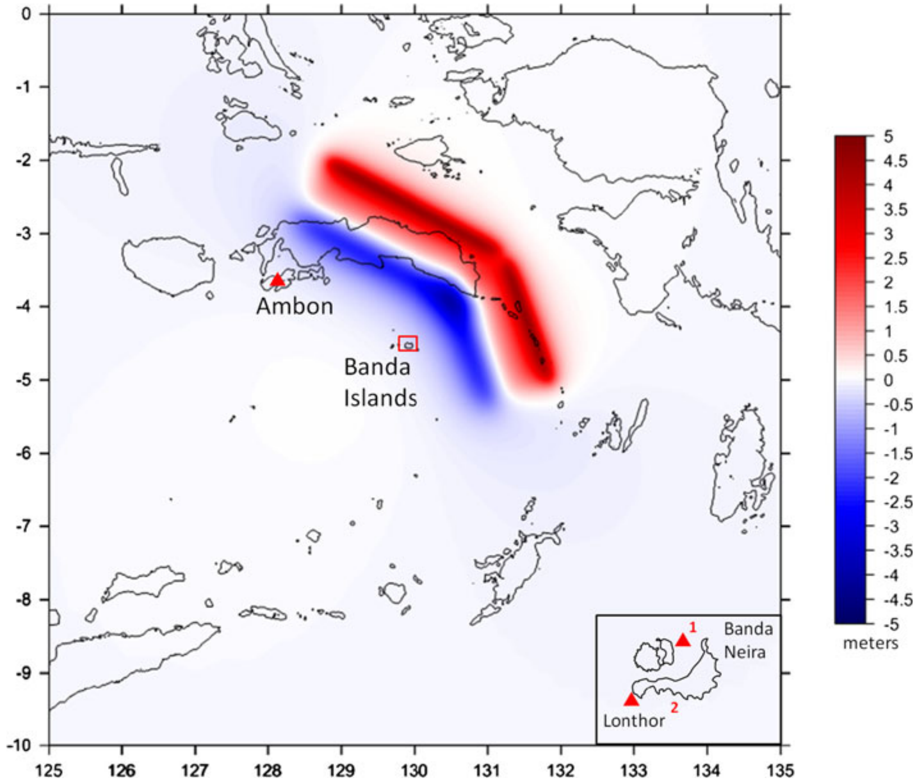

Fig. 6 The vertical component of surface deformation for the Seram forearc, estimated from the two-fault-segment model. The fault plane parameters are given in Table 4. The color scheme and stations are the same as Fig. 4

each model. For analyzing the influence of fault dip, we vary dip values by increments of 5° between dips of 10° and 45°. For analyzing the influence of hypocentral depth, we vary depths by increments of 5 km from depths of 10 to 30 km.

Changes in the dip angle of the fault influence tsunami height most, by a factor of >2 between 10° and 45° (Fig. 9a). This is not surprising due to the increase in vertical displacement caused by increasing dip angles. Variations in hypocentral depth have very little influence (<10 %) on tsunami height.

5 Discussion

5.1 Evaluation of the source of 1629 Banda mega-thrust earthquake and tsunami

The poorest fit between model predictions and observation is with the Tanimbar Trough source. There are four important misfits: (1) waves approach Banda Neira from the south rather than from the east as observed in the record, (2) a much larger magnitude is required along the Tanimbar Trough (M_w = 9.8–9.2) to produce the observed run-up heights in Banda Neira, (3) the wave takes twice as long to arrive as observed and (4) run-up heights

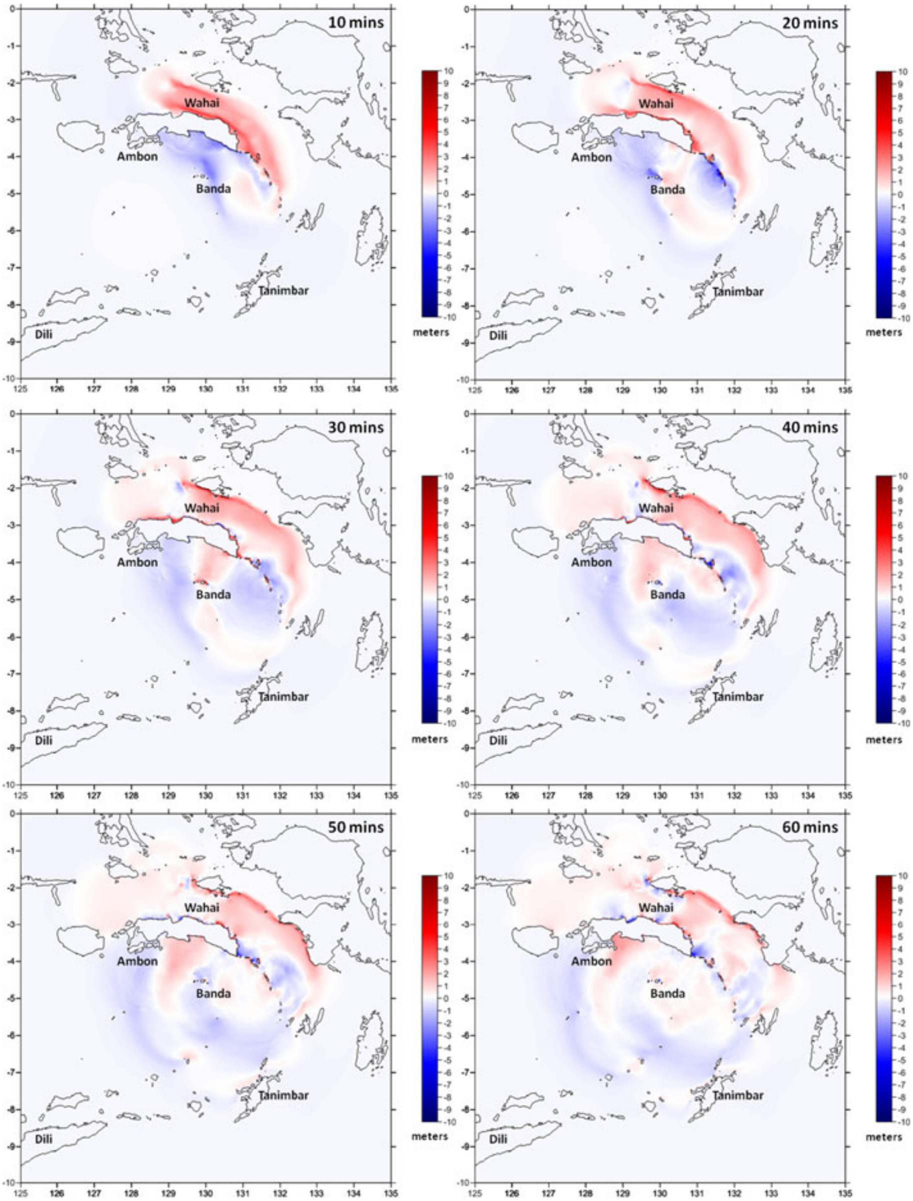

Fig. 7 Snapshots of the first 60 min of tsunami propagation for the two-fault-segment model of Seram Trough earthquake. Positive wave amplitude is shown in *red*, and negative wave amplitude is shown in *blue*. The large positive wave amplitude hits the Banda Islands after 20 min and remains large in shallow coastal areas

in Ambon are at least 5 m, which would have destroyed much of Ambon. Therefore, we conclude that the most likely source of the 1629 mega-thrust earthquake is the Seram Trough.

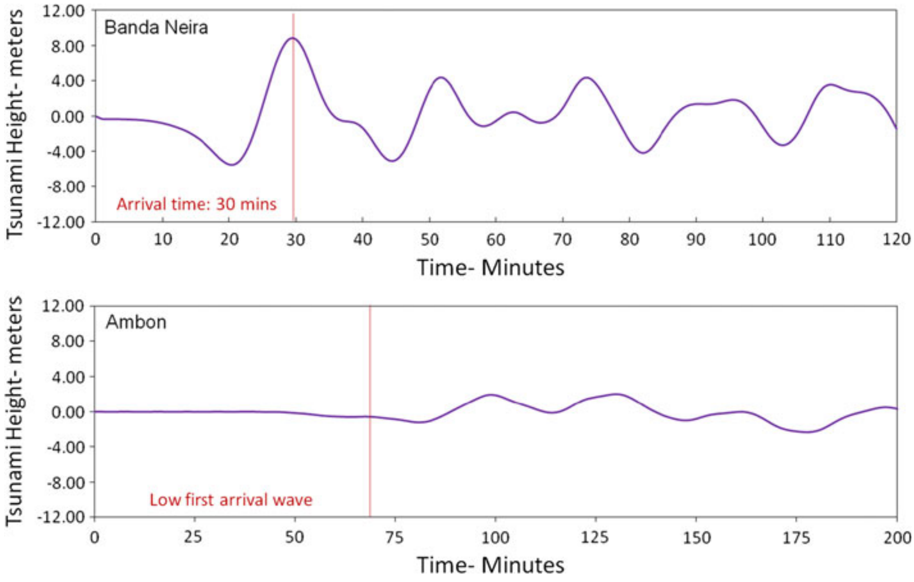

Fig. 8 Computed tide gauge records and tsunami waveform for Banda Neira and Ambon city stations from the Seram Trough model earthquake. The best-fit model tsunami produces wave heights in Ambon of <1 m, which is consistent with no mention of the tsunami there. The tsunami arrival time at the Banda Islands is 30 min after the initiation of the earthquake

We are also confident that the fault parameters we estimated for the Seram Trough are viable and not influencing the model adversely. High tsunami run-ups in Banda Neira may also result from local bathymetric effects that are below the resolution of available bathymetric data. However, this uncertainty does not change the better fit of a Seram Trough versus a Tanimbar Trough source, or the requirement for an earthquake along the Seram Trough larger than any observed since 1629.

5.2 Implication to tsunami hazard for eastern Indonesia

Only one earthquake $>M_w$ 8.0 is recorded instrumentally from the eastern Indonesia region. It happened in 1938 in the middle of the Banda Sea north of Tanimbar. Little is known about the event (Okal and Reymond 2003), but the tsunami was small and caused no damage. In 1899, a $M_s = 7.8$ earthquake struck the Seram region and generated a 10-m tsunami, that was perhaps landslide-assisted, killing around 3,500 persons (Brune et al. 2010). With elastic strain energy accumulating across the Seram Trough at a rate of 50–80 mm/a (Bock et al. 2003), it is likely that earthquake recurrence intervals are short ($\sim$100 years).

With little to no large earthquakes along the Seram Trough since 1899 and perhaps much earlier, we are concerned about the hazard potential of this very active tectonic region. During the past century, population and urbanization in the region have increased tenfold, particularly in coastal inundation zones (Fig. 1). The tsunami potential also threatens large cities outside the region. For example, our Seram Trough tsunami model predicts a 3-m wave in Dili, which is the capital of Timor Leste and a 5.3-meter wave in the north coast of Tanimbar (Fig. 10 and Table 5).

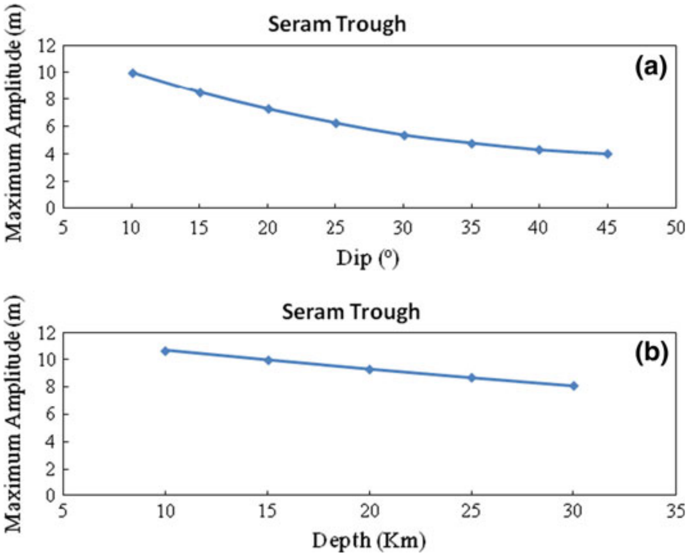

Fig. 9 Sensitivity of tsunami amplitude to fault parameters for Seram Trough. Amplitude as function of fault dip (**a**) and hypocentral depth (**b**)

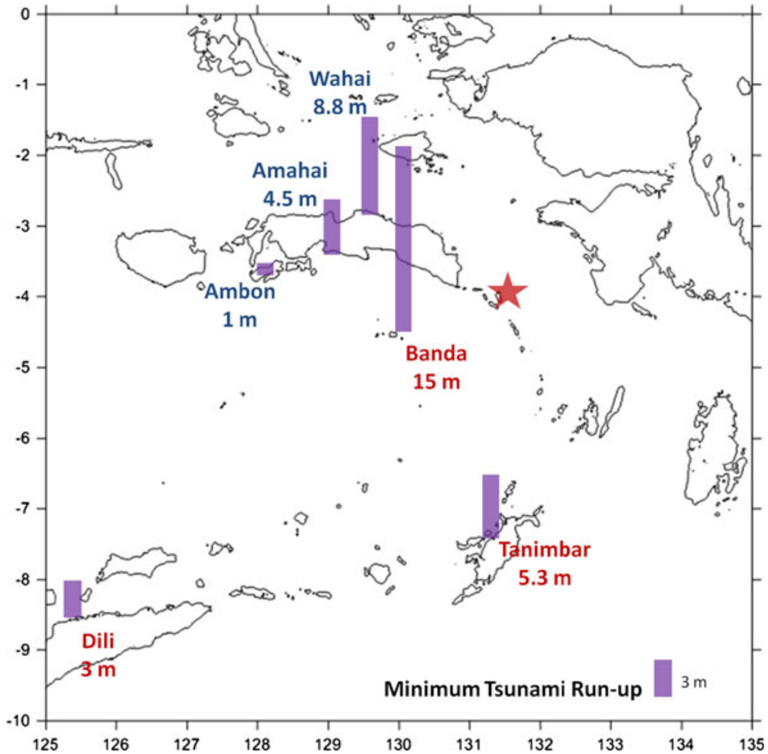

Fig. 10 Minimum tsunami run-up estimates for coastal cities from 1629 Seram Trough mega-thrust earthquake simulation. *Star* is the epicenter for the simulation

Table 5 Predicted tsunami run-up using minimum amplification factor of 1.5 from M_w 8.8 earthquake and maximum amplification factor of 4 from M_w 8.3 earthquake along Seram Trough

Major city	Banda	Ambon	Wahai Seram	Amahai	Tanimbar	Dili	Darwin Australia
Minimum tsunami height (m)	15.5	1	8.8	4.5	5.3	3	1

6 Conclusion

This study reveals for the first time the potential of the Seram Trough for generating subduction interface mega-thrust earthquakes and associated large tsunami that carries the destructive force of the earthquake to many urbanized coastal areas of the eastern Indonesian region. The Banda Sea region has the highest number of historic tsunami in Indonesia (Latief et al. 2000). Yet, it also is one of the regions least aware of tsunami hazards. We urge local officials to not underestimate the disaster potential of the active Banda arc-continent collision and imminent threat of tsunami hazards.

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