

1 **Source process of the September 12, 2007  $M_w$  8.4 Southern**  
2 **Sumatra earthquake from tsunami tide gauge record inversion**

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7

8 **Abstract**

9 We infer the slip distribution and average rupture velocity of the magnitude  $M_w$  8.4 September 12,  
10 2007, Southern Sumatra earthquake from available tide-gauge records of the ensuing tsunami. We  
11 select 9 waveforms recorded along the west coast of Sumatra and in the Indian Ocean. We assume  
12 the fault plane and the slip direction to be consistent with both the geometry of the subducting plate  
13 and the early focal mechanism solutions. Slip distribution and rupture velocity are determined  
14 simultaneously by means of a non linear inversion method. We find high slip values ( $\sim 10$  m) into a  
15 patch 100 km long and 50 km large, between 20 and 30 km of depth, about 100 km north-west from  
16 the epicenter. Our estimate of rupture velocity is of  $2.1 \pm 0.4$  km/sec. The relatively large depth of  
17 the main slip patch is the likely explanation for the small observed tsunami.

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## 26 **1. Introduction**

27 On September 12, 2007 at 11:10:26 UTC an earthquake of magnitude  $M_W$  8.4 occurred off the  
28 west coast of Sumatra about 130 km SW of Bengkulu. The epicenter was localized at  $4.517^\circ\text{S}$  and  
29  $101.382^\circ\text{E}$ , between the Sunda trench and Bengkulu (Figure 1). The induced sea-floor displacement  
30 generated a moderate tsunami that nevertheless propagated through the Indian Ocean and was  
31 recorded for example also by the tide-gauge at Salalah, Oman. Apart from minor aftershocks, it was  
32 followed, 12 hours later, by another earthquake of magnitude  $M_W$  7.9 near the Sumatra coast, 185  
33 km SSE of Padang (Figure 1). Both earthquakes caused in total 25 fatalities and 161 people were  
34 injured [ref. USGS]. The seismic sequence continued on the next two days, with the biggest event  
35 of magnitude  $M_W$  7.1 happening on September 13. This sequence took place in the same zone  
36 where the historical earthquakes of 1797 and 1833 also generated significant tsunamis [*Nalbant et*  
37 *al.*, 2005].

38 This is the fourth very large earthquake to occur on the Sunda megathrust and generating a  
39 significant tsunami in less than three years (Figure 1). We recall the huge Sumatra-Andaman  
40 earthquake and tsunami of December 26, 2004 with a magnitude  $M_W$  9.2 and a rupture extent of  
41 almost 1,300 km, that caused widespread victims and destruction on the Indian Ocean coasts. The  
42 Sumatra-Andaman 2004 event was followed later on by the  $M_W$  8.7 earthquake and tsunami of  
43 Simeulue-Nias (March 28, 2005), generated along the Sunda trench on the stretch just contiguous to  
44 the Southern end of the Sumatra-Andaman 2004 rupture zone. The imminence of the Nias  
45 earthquake was predicted as a consequence of the coseismic stress induced by the slip distribution  
46 of the Sumatra-Andaman earthquake [*McCloskey et al.*, 2005]. The rupture zone of the Simeulue-  
47 Nias earthquake has also a significant overlap with that of the historical 1861 earthquake and  
48 tsunami [*Nalbant et al.*, 2005]. The next destructive tsunami earthquake was the  $M_W$  7.7 Java event  
49 (July 17, 2006), with a rupture zone displaced southward with respect to the one of the Simeulue-  
50 Nias event. It generated an abnormally large tsunami with respect to what expected by its

51 seismically radiated energy and it has been then catalogued as a “tsunami earthquake” [*Ammon et*  
52 *al.*, 2006; *Fuji and Satake*, 2006].

53 The inversion of tsunami waveforms is an important tool for characterizing the seismic source of  
54 off-shore earthquakes. Recently, such inversions have been performed by *Piatanesi and Lorito*  
55 (2007) and by *Fuji and Satake* (2007) to characterize the kinematic rupture of the 2004 Sumatra-  
56 Andaman earthquake.

57 A better understanding of the rupture process of tsunamigenic earthquakes originating on the  
58 Sunda megathrust is in turn a key issue for evaluating the possible consequences of future events  
59 both for the risk mitigation and the warning strategies design [*Borrero et al.*, 2006; *Geist et al.*,  
60 2007; *McCloskey et al.*, 2007a; *McCloskey et al.*, 2007b].

61 Here we perform the inversion of the waveforms recorded by 9 tide-gauge stations distributed in the  
62 Indian Ocean both in the near and in the far field with respect to the source zone. Our goal is to  
63 retrieve the slip distribution and the average rupture velocity of the September 12, 2007 earthquake.

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## 65 **2. Sea-level Data**

66 The tsunami waves generated by the September 12, 2007 earthquake were recorded by tide-gauge  
67 stations distributed in the Indian Ocean in the shallow waters of harbors and coastal sites. The data  
68 were available for download in near real time at the website of the University of Hawaii Sea Level  
69 Center (UHSLC; <http://ilikai.soest.hawaii.edu/RSL/>). Most of these stations are operated by the  
70 Global Sea Level Observing System (GLOSS; <http://www.gloss-sealevel.org/>).

71 The tsunami was also recorded by the DART buoy 23401 in the deep ocean northwest of Sumatra.  
72 This buoy is owned and operated by the Thailand Meteorological Department (TMD) in  
73 conjunction with National Disaster Warning Center of the Kingdom of Thailand. Data were  
74 available in real-time at the National Data Buoy Center (NDBC) website  
75 (<http://www.ndbc.noaa.gov/dart.shtml>), maintained by the U.S. National Oceanic and Atmospheric  
76 Administration (NOAA).

77 Both the UHSLC and the DART 23401 records of this tsunami have a sampling rate of one minute.  
78 The positions of the stations we choose for the inversion are plotted on the map of Figure 1. Before  
79 using these data in the inversion, we remove tidal components to extract the tsunami signal and we  
80 select only the first cycles of the waveforms that are less sensitive to local bathymetry than later  
81 arrivals. We use an even narrower time window for Padang and Sibolga stations since they clearly  
82 show a resonant character in the later phases.

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### 84 **3. Seismic Source Parameters**

85 To define the dimension of the causative fault, we use the spatial distribution of the aftershocks  
86 occurred during 12 hours after the mainshock, that is available at the National Earthquake  
87 Information Center website of the U.S. Geological Survey (USGS;  
88 <http://neic.usgs.gov/neis/epic/epic.html>). We choose a region large enough to contain all the  
89 aftershocks, and then we end up with a length of 350 km and a width of 200 km (Figure 2).

90 The strike of the source is  $323^\circ$  roughly parallel to the Sunda trench [*Bird, 2003*]. The dip is fixed at  
91  $12^\circ$ , using the quick moment tensor solution of the Global CMT project. We adopt a slip direction  
92 (rake) of  $105^\circ$ , basing both on the focal mechanism and on the finite fault model calculated by C. Ji  
93 according to the algorithm described in *Ji et al. [2002]*, and available at  
94 [http://www.geol.ucsb.edu/faculty/ji/big\\_earthquakes/2007/09/sumatra\\_seismic.html/](http://www.geol.ucsb.edu/faculty/ji/big_earthquakes/2007/09/sumatra_seismic.html/).

95 We split the source region into 28 square subfaults of dimension 50x50 km (see Table 1 of the  
96 Online Supplementary Material). The slip amount on each subfault is to be determined as a result of  
97 the inversion, along with the average rupture velocity.

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### 99 **4. Tsunami Modeling and Bathymetric Dataset**

100 For each of the subfaults we calculate the corresponding marigrams at the coordinates of the tide-  
101 gauges, which are the Green's functions we will use for the inversion.

102 The initial seawater elevation is assumed to be equal to the coseismic vertical displacement of the  
103 sea bottom corresponding to an assumed “unitary” slip of 3 meters [Piatanesi and Lorito, 2007],  
104 and computed through the Okada’s analytical formulas [Okada, 1992], while the initial velocity  
105 field is assumed to be identically zero. The boundary conditions are pure wave reflection at the  
106 solid boundary (coastlines) and full wave transmission at the open boundary (open sea). The  
107 tsunami wave propagation from the initial field to the tide-gauges is calculated by solving the non  
108 linear shallow water equations, including Coriolis force and bottom friction, with a finite  
109 differences scheme on a staggered grid. More detailed description of the equations as well as of the  
110 numerical method can be found in Lorito *et al.* [2008], Piatanesi and Lorito [2007] and Mader  
111 [2001].

112 The computational domain is shown in Figure 1. We choose a grid resolution for tsunami  
113 propagation of 1 arc-minute. The bathymetry grid is obtained by merging different bathymetric  
114 datasets [e.g. Geist *et al.*, 2007; Fujii and Satake, 2007]. We use as a background the 1 arc-min  
115 GEBCO bathymetry [British Oceanographic Data Center, 2003], version 1.02, available at the  
116 British Oceanographic Data Center website ([http://www.bodc.ac.uk/data/online\\_delivery/gebco/](http://www.bodc.ac.uk/data/online_delivery/gebco/)).  
117 To improve the bathymetry in shallow waters, particularly along the coast of Sumatra and around  
118 some of the tide-gauges, we digitize, where available to us, 9 nautical charts (*United Kingdom*  
119 *Hydrographic Office*, 2005; see Figure 1). We then remove the shallow water points from GEBCO  
120 in correspondence of the digitized dataset and interpolate both on local 0.5 arc-min subgrids, before  
121 resampling to 1 arc-min and merging to the whole domain.

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## 123 **5. Inversion**

124 We use a non linear inversion method based on the simulated annealing technique to  
125 simultaneously infer both slip distribution and average rupture velocity, with a L1-L2 hybrid norm  
126 as cost function. The details can be found in Piatanesi and Lorito [2007] and in Lorito *et al.* [2008],  
127 and in the references therein.

128 The average rupture velocity is assumed to be the uniform velocity of a circular front propagating  
129 out from the hypocenter on the fault plane. We consequently add the appropriate delay to the  
130 Green's function corresponding to each subfault, according to their distance along the rupture path.  
131 The resolving power of the inversion setup (azimuthal coverage, sampling rate, fault  
132 parametrization, etc.) is tested by means of a checkerboard experiment, with target slip values of  
133 alternatively 2 and 6 meters on adjacent subfaults. We first tune the subfaults size by means of  
134 several checkerboard experiments, starting with larger sizes and ending up with 50 by 50 km. We  
135 moreover test for different rupture velocities. The synthetic waveforms generated with the  
136 checkerboard slip distribution are corrupted with Gaussian noise, with a variance that is 20% of the  
137 clean time series variance. We let the slip range between 0 and 15 meters, at 1 meter steps, whilst  
138 the velocity ranges between 0.5 and 3.5 km/sec, at 0.5 km/sec steps.

139 The target slip distribution, i.e. the checkerboard pattern, is reproduced fairly well by the best model  
140 (Figure 2, and Table 1 of the Online Supplementary Material), with a root mean square of the  
141 differences between target and inverted values of 0.8 meters. Conversely, the target rupture velocity  
142 of 2.5 km/sec is not inferred exactly, as our best model estimation is 2 km/sec. Following *Piatanesi*  
143 *and Lorito* [2007], we estimate the uncertainty of each inverted parameters through an appraisal of  
144 the ensemble of the models explored during the search stage (see Table 1 of the Online  
145 Supplementary Material). In particular, the marginal distribution of the explored velocity values  
146 shows quite a broad peak ranging between 2 and 2.5 km/sec (see inset of Fig. 2). The average  
147 model estimation of the rupture velocity is  $2.2 \pm 0.3$  km/sec, that is consistent with the target velocity  
148 of 2.5 km/sec. Later, in the real case inversion, we will then assume the average model value, in  
149 place of the best model, as our estimate of the rupture velocity.

150 The checkerboard experiment results indicate that the dataset has the capability of resolving the  
151 main features of the rupture kinematics, with a good resolution for the slip and a relatively worst  
152 resolution for the velocity.

153 To retrieve the rupture features of the September 12, 2007 earthquake, we adopt the same subfaults  
154 geometry and parameter ranges as in the checkerboard experiments. Moreover we adjust of a few  
155 minutes the arrival times of the Green's functions at the stations of Cocos, Sibolga, Colombo,  
156 Diego Garcia and Pointe La Rue, at which we observed systematic differences between recorded  
157 and simulated phases. *Arcas and Titov* [2006] discussed the difficulty in reproducing coastal in  
158 comparison to offshore propagation, due to the poor bathymetry knowledge along with the  
159 contamination from unmodeled coastal processes. We actually use the GEBCO bathymetry only at  
160 some of the stations and did not simulate inundation, likely leading to travel times inaccuracy. An  
161 adjustment of travel times has been performed also by *Fuji and Satake* [2006], who observed  
162 discrepancies at the Cocos station while studying the 2006 Java tsunami. In the present case,  
163 however, we verified that the inverted rupture velocity is only slightly dependent on such arrival  
164 time adjustments.

165 The slip distribution of the best model we find is shown in Figure 3. According to our results, the  
166 rupture propagated with a velocity of  $2.1 \pm 0.4$  km/sec, with a relatively low slip (2 to 3 meters)  
167 around the hypocentral zone. The rupture then propagated to the North-West featuring the highest  
168 slip concentration (up to 12 meters) at about 3-3.5°S and 101°E, at a depth ranging between 20 to  
169 30 km. We notice that this main patch is surrounded by most of the aftershocks. We also plot the  
170 resulting coseismic vertical displacement, to be compared with the geodetic data when available.  
171 The whole set of results, best model and average model with associated errors is reported in Table 1  
172 in the online supplementary material. The best model slip values always fall well inside the error  
173 bars. Conversely, the best model velocity lies at the upper edge of the error bar, confirming the best  
174 model is an extreme one in this context and as velocity is concerned.

175 The match between the recorded and inverted waveforms is generally good (Figure 4). In some of  
176 the cases, as those of Padang (the nearest station to the source) and the DART buoy (the most  
177 reliable data) is excellent. The worst case is Diego Garcia, with at least the period well reproduced,

178 while missing its amplitude. Overall, we satisfactorily reproduce the main features of the September  
179 12, 2007 tsunami wave amplitudes and periods, and, in some cases, the arrival times.

180

## 181 **6. Conclusions**

182 We find that the source of the September 12, 2007 Southern Sumatra earthquake has been  
183 characterized by its highest slip values (~10 m) concentrated into a patch 100 km long and 50 km  
184 large, located between 20 and 30 km of depth, about 100 km north-west from the epicenter. The  
185 occurrence of such a slip amount has been suggested by *Nalbant et al.* [2005], who indicated that  
186 the greatest current seismic threat from the Sunda megathrust may come from a section overlapped  
187 with the source of the September 12, 2007 earthquake, with a slip as great as in 1833: that is, up to  
188 10 meters.

189 The slip release has been very low or even absent at depths lower than 10-15 km. A shallower slip  
190 patch of about 2-3 meters occurred only around the epicentral latitudes.

191 These results are in fair agreement, at least as regard the localization of the main slip patches, with  
192 those of [http://www.geol.ucsb.edu/faculty/ji/big\\_earthquakes/2007/09/sumatra\\_seismic.html/](http://www.geol.ucsb.edu/faculty/ji/big_earthquakes/2007/09/sumatra_seismic.html/), and in a

193 lesser extent with those of Yuji Yagi ([http://www.geo.tsukuba.ac.jp/press\\_HP/yagi/EQ/20070912/](http://www.geo.tsukuba.ac.jp/press_HP/yagi/EQ/20070912/)).

195 Nevertheless, the above teleseismic inversions infer a much lower slip amount than our inversion of  
196 tsunami waveforms.

197 Our best model features a seismic moment of  $4.8 \times 10^{21}$  N·m, if we use a shear modulus  $\mu=6.0 \times$   
198  $10^{10}$  N/m<sup>2</sup> [*Geist and Bilek*, 2001], corresponding to a magnitude  $M_w$  8.39 earthquake. This result is  
199 consistent with the USGS estimation of a  $M_w$  8.4 earthquake magnitude.

200 The depth of the earthquake is an important factor in controlling the tsunami amplitude especially in  
201 the near field. On one hand, slip at depth produces relatively small vertical displacement of the sea  
202 bottom. On the other hand, high slip at intermediate depths on the megathrust produces maximum  
203 sea floor displacements in shallow waters thus generating tsunamis with relatively small potential

204 energy. The absence of significant slip at shallow depths is then a likely explanation for the  
205 generation of a relatively moderate tsunami. As a comparison, the tsunami generated by the smaller  
206 and shallower Java 2006  $M_w$  7.7 earthquake was by far more destructive, and it has been then  
207 catalogued as a “tsunami earthquake” [Fuji and Satake, 2006].

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210 SISMOS laboratory, who scanned the nautical charts we used in this work.

211 Some figures have been produced with Generic Mapping Tools software (Wessel and Smith, 1998).

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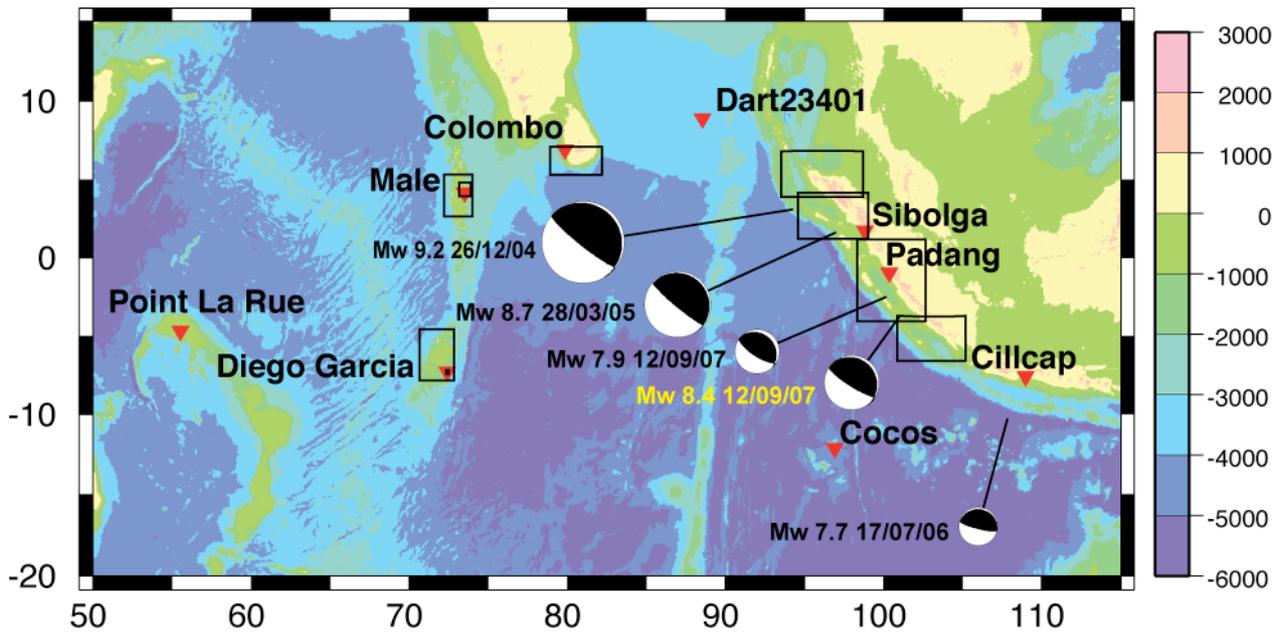
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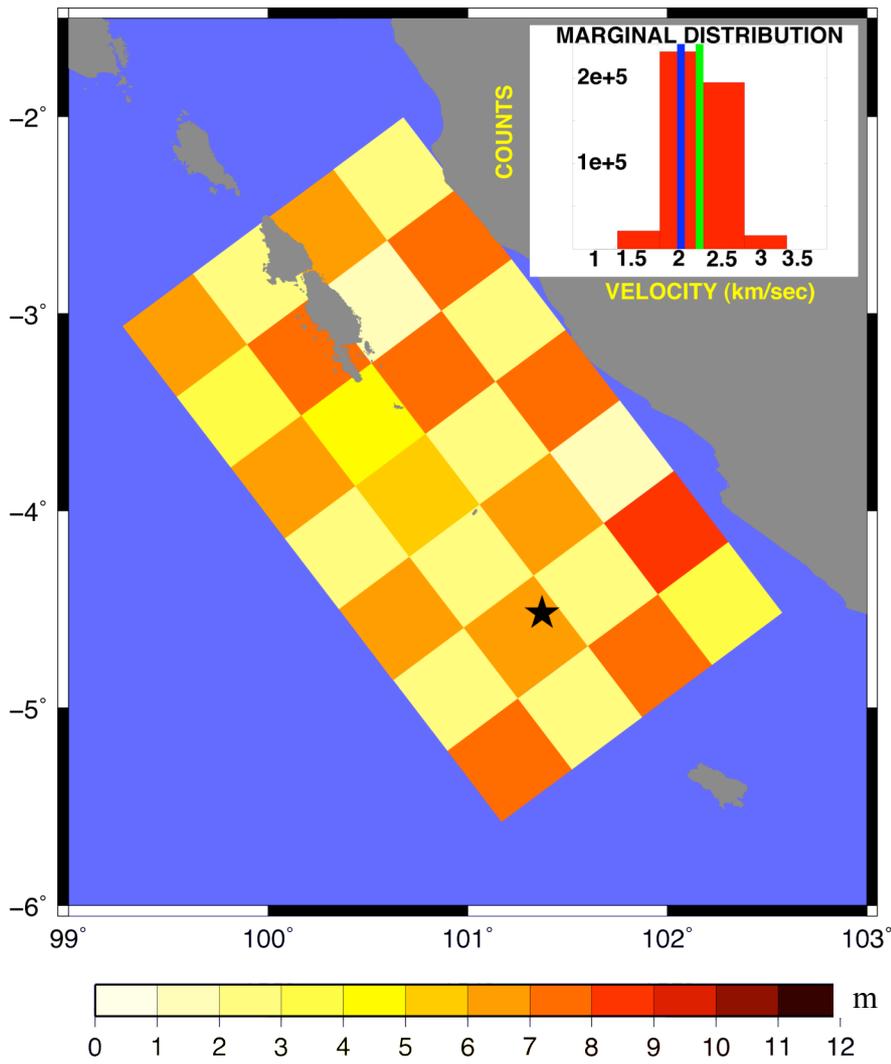
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264 **Figure 1.** Epicenter and focal mechanism of the  $M_w$  8.4 earthquake of 12 September 2007, along  
 265 with the epicentral positions and focal mechanisms of the major recent tsunamigenic earthquakes on  
 266 the Sunda trench. The triangles indicate the positions of the tide-gauge stations and of the DART  
 267 buoy 23401. The bathymetry used for the simulation is represented by the colorscale. The  
 268 rectangles are the limits of the nautical charts digitized to construct the bathymetric dataset.



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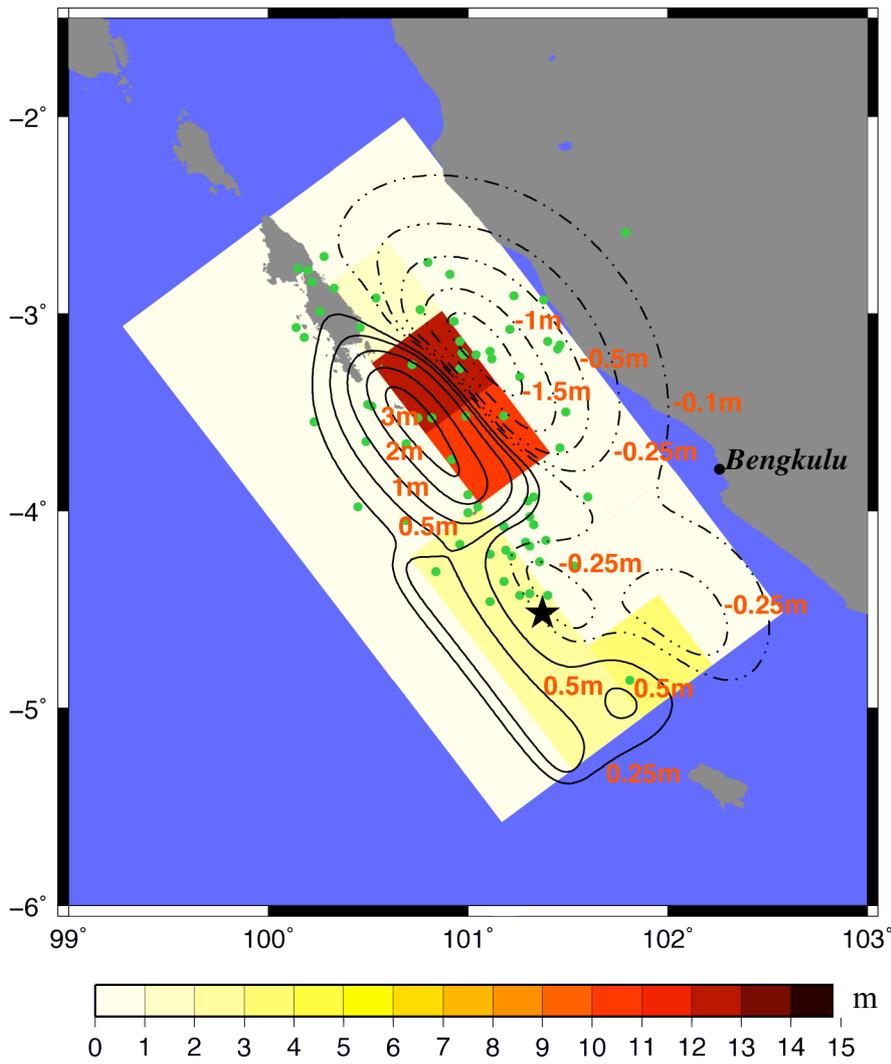
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271 **Figure 2.** Slip distribution of the best model resulting from the checkerboard test: the rectangles  
 272 represent the projection of the subfaults to the Earth's surface. The black star indicates the epicenter  
 273 position. The inset shows the marginal distribution of the rupture velocity; the blue and green  
 274 vertical solid lines represent the best and average model value respectively.

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279 **Figure 3.** Slip distribution of the best model resulting from the September 12, 2007 Sumatra event;

280 rectangles represent the projection of the subfaults to Earth's surface. The black star indicates the

281 epicenter of the mainshock and the green circles are the epicenter of the aftershocks occurred 12

282 hours after the main event. Solid and dashed contour lines represent the positive and negative

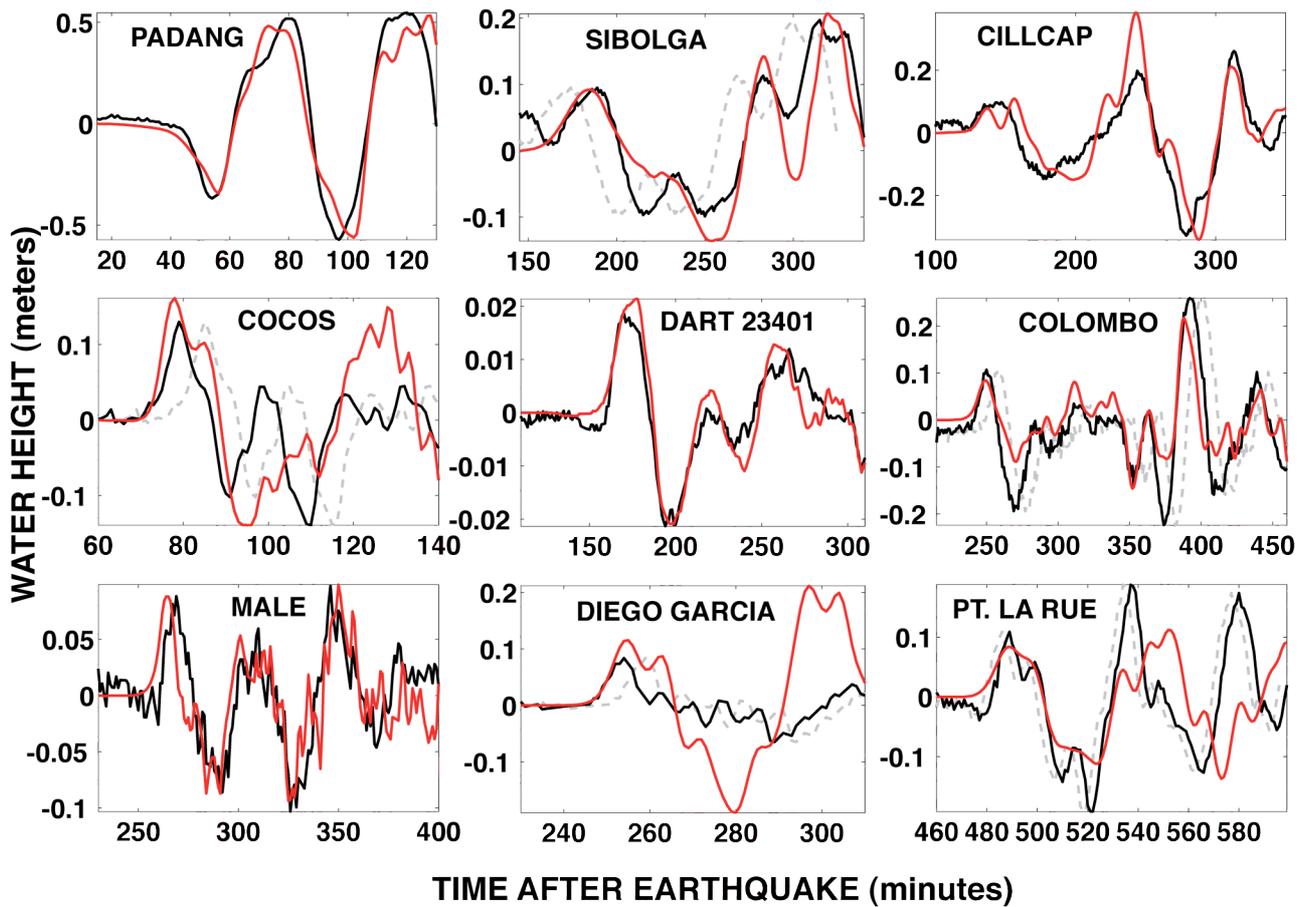
283 vertical displacement respectively.

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289 **Figure 4.** Comparison between the observed time-shifted records (black solid lines) and computed  
 290 waveforms using the inverted best model (red solid lines). Grey dashed lines are the original  
 291 observed records.

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