**Improved source parameter constraints for five undersea earthquakes from north
component of GRACE gravity and gravity gradient change measurements**

Chunli Dai^{1*}, C.K. Shum^{1,4}, Junyi Guo¹,
Kun Shang¹, Byron Tapley², Rongjiang Wang³

¹ Division of Geodetic Science, School of Earth Sciences, The Ohio State University,
Columbus, Ohio, USA

² Center for Space Research, University of Texas at Austin, Texas, USA

³ Physics of Earthquakes and Volcanoes, GFZ German Research Centre for Geosciences,
Potsdam, Germany

⁴ State Key Laboratory of Geodesy and Earth's Dynamics, Institute of Geodesy and
Geophysics, Chinese Academy of Sciences, Wuhan 430077, China

* Corresponding author, Chunli Dai, dai.56@buckeyemail.osu.edu

25 **Abstract:**

26 The innovative processing of Gravity Recovery And Climate Experiment (GRACE) data
27 using only the north component of gravity change and its corresponding gravity gradient
28 changes allows the enhancement of the spatial resolution for coseismic deformation
29 signals. Here, we report the study of five undersea earthquakes using this technique: the
30 2004 Sumatra-Andaman earthquake, the 2007 Bengkulu earthquake, the 2010 Maule,
31 Chile earthquake, the 2011 Tohoku earthquake, and the 2012 Indian Ocean earthquakes.
32 By using the high spherical harmonic degree (up to degree 96) data products and the
33 associated GRACE data processing techniques, the retrieved north component of gravity
34 change is up to -34 ± 1.4 μGal for the 2004 Sumatra-Andaman earthquake, which
35 illustrates by far the highest amplitude of the coseismic signal retrieved from satellite
36 gravimetry among previous studies. We creatively apply the localized spectral analysis as
37 an efficient method to empirically determine the practical spherical harmonic truncation
38 degree. By combining least squares adjustment with the simulated annealing algorithm,
39 point source parameters are estimated, which demonstrates the unique constraint on
40 source model from GRACE data compared to other data sources. For the 2004 Sumatra-
41 Andaman earthquake, GRACE data produce a shallower centroid depth (9.1 km), as
42 compared to the depth (28.3 km) from GPS data. For the 2011 Tohoku earthquake, the
43 GRACE-estimated centroid location is southwest of the GPS/seismic solutions, and the
44 slip orientation is about 10° clockwise from the published GPS/seismic slip models. We
45 concluded that these differences demonstrate the additional and critical offshore
46 constraint by GRACE on source parameters, as compared to GPS/seismic data.

47

48

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71 **1. Introduction**

72 The quantification of large undersea earthquakes, including their sizes, location,
73 geometry and orientation of the faults, is critical for improving our understanding of fault
74 mechanisms, and for applications in tsunami warning. As pointed out by *Geist et al.*
75 [2007], the centroid location and seismic moment are the most essential parameters
76 [*Meng et al.*, 2012; *Tanioka and Satake*, 1996] for tsunami forecast and hazard
77 assessment models; while other first-order source parameters such as dip, fault length and
78 width also affect the tsunami wave field [*Geist et al.*, 2007]. Besides seismic data and
79 GPS data, which are traditionally used for studying focal mechanisms, other data sets
80 including the Interferometric Synthetic Aperture Radar (InSAR) data, tsunami data, and
81 repeated airborne LiDAR data, have also been used. However, there are limitations for
82 each type of data set on estimating the source parameters of undersea earthquakes.
83
84 Seismological methods have difficulty in estimating source parameters in some instances,
85 e.g., seismic moment for ruptures with long duration due to the overlap of interfering
86 arrivals [*Chlieh et al.*, 2007; *Lay et al.*, 2005], and seismic wave data are also inadequate
87 for detecting slow or aseismic slip and postseismic slip [*Chlieh et al.*, 2007; *Han et al.*,
88 2013]. One example for a long duration, slow rupture is the 2004 Sumatra-Andaman
89 earthquake [*Banerjee et al.*, 2005; *Lay et al.*, 2005; *Park et al.*, 2005], which also had
90 large magnitude of afterslip [*Chlieh et al.*, 2007]. *Lay et al.* [2010] indicate that seismic
91 inversions are sensitive to waveform types and the frequency band, which is shown by
92 the fact that different slip models are obtained from seismic waves with different

93 frequency bands. In addition, since seismic inversions are highly dependent on the
94 velocity structure, they have instabilities for shallow ruptures [*Lay et al.*, 2011].
95
96 Although geodetic measurements, such as GPS data, have the potential to overcome the
97 inadequacy of seismological data in detecting slow/aseismic slip, they are limited by the
98 spatial distribution of their ground-based sites. Particularly, for undersea earthquakes,
99 although more ocean-bottom measurements are available, GPS stations are mostly
100 located at one side of the fault area, thus providing limited constraints on source
101 parameters [*Chlieh et al.*, 2007; *Lay et al.*, 2011; *Wei et al.*, 2012]. For example, *Chlieh et*
102 *al.* [2007] and *Wei et al.* [2012] showed that inland static GPS data are quite insensitive
103 to shallow ruptures and resolutions of inverted slip models decrease rapidly away from
104 the coast.
105
106 The twin-satellite Gravity Recovery And Climate Experiment (GRACE) mission [*Tapley*
107 *et al.*, 2004] has been producing temporal global gravity field observations with monthly
108 sampling rate and a spatial resolution up to spherical harmonic degree 60. The GRACE
109 data have revolutionized our understanding of Earth's mass redistribution, including
110 terrestrial hydrologic water balance, ocean mass variations and sea level rise, and ice-
111 sheet and glacier ablation, and their possible links with anthropogenic climate change. As
112 one of such mass redistribution processes, earthquakes trigger crustal/mantle dilation or
113 compression and surface uplift or subsidence, causing permanent change in Earth's
114 gravity field. By surveying right above the rupture region over the ocean, although with a
115 coarse spatial and temporal resolution, GRACE data have been demonstrated to have the

116 feasibility to complement other data for detecting and constraining focal mechanisms of
117 large undersea earthquakes, since GRACE data have a better spatial coverage as
118 compared to GPS data, and have better capability to detect aseismic slip as compared to
119 seismic data. Several large earthquakes have been detected in GRACE data and analyzed
120 in contemporary studies, including the 2004 Sumatra-Andaman earthquake [e.g., *Han et*
121 *al.*, 2006; *Wang et al.*, 2012c], 2010 Maule, Chile earthquake [e.g., *Han et al.*, 2010, *Heki*
122 *and Matsuo*, 2010, *Wang et al.*, 2012a], and the 2011 Tohoku earthquake [e.g., *Cambiotti*
123 *and Sabadini*, 2012; *Dai et al.*, 2014; *Han et al.*, 2011, 2013; *Li and Shen*, 2015; *Matsuo*
124 *and Heki*, 2011; *Wang et al.*, 2012b]. Moreover, GRACE has also shown its unique
125 contribution to the detection of postseismic gravity signals [*Han et al.*, 2014; *Panet et al.*,
126 2010; *Tanaka and Heki*, 2014].

127
128 Recent studies explored different data processing methods to better retrieve co- and
129 postseismic gravity change signal from GRACE data. *Wang et al.* [2012c] showed a
130 spatial resolution enhancement by using the inferred gravity gradient changes computed
131 from GRACE temporal gravity field solutions, for the 2004 Sumatra-Andaman
132 earthquake. To optimize the spatial and temporal resolution, *Han et al.* [2011] directly
133 exploited the signal of the 2011 Tohoku earthquake in the change of inter-satellite K-
134 band range (KBR) rate. *Wang et al.* [2012a, 2012b] for the first time utilized the Slepian
135 functions [*Simons et al.*, 2006] to analyze GRACE-observed gravity changes aiming at
136 the spatial resolution enhancement of the coseismic signals. The same techniques are also
137 applied for the source parameters inversion by *Cambiotti and Sabadini* [2012].
138 Furthermore, an innovative approach of using only the north component of gravity

139 change [Dai *et al.*, 2014], the corresponding gravity gradient change, e.g., T_{xx} and T_{xz} (x ,
140 z refers to north and up directions, respectively) change by Wang *et al.* [2012c], T_{xz}
141 change by Li and Shen [2011], are found to substantially avoid the correlated errors in the
142 GRACE temporal gravity field solution, with no decorrelation nor spatial filtering
143 needed, leading to improved spatial resolution at the full wavelength corresponding to the
144 highest spherical harmonic degree of GRACE data.

145

146 The inversion for several source parameters, such as the seismic moment, dip angle and
147 rake angle, based on normal mode formulation assuming point dislocation were
148 demonstrated for the 2011 Tohoku earthquake [Han *et al.*, 2011]. Wang *et al.* [2012a,
149 2012b] adopted the Markov Chain Monte Carlo algorithm to invert for fault length and
150 width of the 2010 Chilean Maule and the 2011 Tohoku earthquakes, based on a finite
151 fault model assuming uniform slip. Cambiotti and Sabadini [2013] presented an
152 estimation of all fault parameters (centroid location and moment tensor) for a point
153 source using GRACE data. Han *et al.* [2013] solved for the seismic moment tensors of
154 multiple centroids, but with locations fixed, based on the normal mode formulation for a
155 number of large earthquakes over the last decade using GRACE data. Dai *et al.* [2014]
156 further solved for the centroid location, seismic moment, fault width, and slip rake angle
157 based on finite fault model using simulated annealing algorithm, and found that GRACE
158 data are especially effective in constraining centroid location and slip orientation.

159

160 In this paper, the new approach of GRACE data processing using only the *north*
161 component of gravity change and the corresponding gravity gradient change [Dai *et al.*,

2014] is adopted to detect the coseismic signal for five recent large undersea earthquakes using several different GRACE products, including particularly the high degree (up to degree 96) CSR RL05 data. By using the linear algorithm of gravity and gravity gradient change with respect to the double-couple moment tensor, the point source parameters are estimated through a least squares adjustment combined with the simulated annealing algorithm. From the improved GRACE data processing method, the high degree (up to degree 96) data products, and the new inversion scheme, we solve for the point source parameters for the 2004 Sumatra-Andaman (M_w 9.2) earthquake, the 2011 Tohoku (M_w 9.0) earthquake, the 2010 Maule, Chile (M_w 8.8) earthquake, the 2012 Indian Ocean (M_w 8.6 and M_w 8.2) earthquakes, and the 2007 Bengkulu (M_w 8.5) earthquake. The results show the resolving power of GRACE data on slip orientation, and centroid location and depth.

175

176 **2. Improved GRACE Data Processing**

177 This work is based on the GRACE data processing method of using only the *north*
178 component of gravity and gravity gradient changes developed in [Dai *et al.*, 2014]. To
179 enhance spatial resolution, we further advance this approach by applying the localized
180 spectral analysis as an efficient method to determine the practical spherical harmonic
181 truncation degree from high degree L2 products. As an example, this localized spectral
182 analysis suggests that only the degrees less than 70 from the high degree (up to 96) CSR
183 L2 Release 05 (RL05) monthly geopotential solutions contain reliable seismic signal for
184 the 2011 Tohoku earthquake as shown in the following paragraph.

185

186 The localized spectra analysis [Wieczorek and Simons, 2005] is used as an efficient
187 method to justify the choice of practical truncation degree. Dai *et al.* [2014] show the
188 application of this localized spectral analysis in the way of evaluating the signal and noise
189 level of each component of gravity and gravity gradient as function of spherical harmonic
190 degree. Here we compute the localized degree variance, and the results are shown in
191 Figure 1. Where, the north, east, and down components of the gravity disturbance are
192 denoted by g_N , g_E and g_D , and the north components of gravity gradient disturbances are
193 denoted by T_{xx} , T_{xy} and T_{xz} (x , y , z , is north, west, up direction). The coseismic signal is
194 estimated by fitting the GRACE-observed gravity and gravity gradient time series on a
195 regular grid with a model composed of a Heaviside step function, a linear trend and three
196 periodicities (equation A4, Figure S1). We can see that the GRACE-observed g_N change
197 (Figure 1a) agrees well with the model prediction up until around degree 70, with slightly

198 larger amplitude; while, the degree variance for GRACE-observed g_E and g_D change
 199 (Figure 1a) increases sharply starting from degrees at around 40, indicating that degrees
 200 higher than that are dominated by noise. This demonstrates that north component of
 201 gravity change contains reliable signal up to much higher degree than the east and down
 202 components of gravity change. Furthermore, as shown in Figure 1a, the degree variance
 203 for g_N change keeps in good agreement with the model prediction until around degree 70,
 204 but above degree 70, the GRACE data is much higher than the model prediction, which
 205 we interpret as noise. Based on this, we choose to discard the spherical harmonic
 206 coefficients above degree 70 and compute the gravity and gravity gradient change (Figure
 207 1b) only up to degree 70, which will lead to good signal-to-noise ratio (in spatial domain
 208 as well).

210

211 Figure 1. Comparison between GRACE-observed and model-predicted gravity and
 212 gravity gradient change in the spectral domain for the 2011 Tohoku earthquake.
 213 The localized degree variance as a function of spherical harmonic degree is shown for the
 214 observed and model-predicted g_N , g_E , g_D change (a), and T_{xx} , T_{xy} , T_{xz} change (b). The
 215 GRACE-observed gravity and gravity gradient change is computed from the CSR RL05
 216 NMAX 96 product up to maximum degree 96. The coseismic slip model by *Wei et al.*
 217 [2012] and the postseismic slip model for March 2011 by *Ozawa et al.* [2011] are used to
 218 calculate the model prediction using the forward modeling procedure in [*Dai et al.*,
 219 2014].

220

221 Based on the localized spectral analysis, GRACE-observed and model-predicted north
222 component of gravity and gravity gradient changes are compared for degrees up to 70,
223 which is by now the highest degree of coseismic signal achievable by GRACE. As shown
224 in Figure S2, the spatial pattern of GRACE-derived gravity and gravity gradient change is
225 evident and agrees well with the model prediction. For example, the GRACE-observed g_N
226 change reaches up to $-24.0 \pm 3.9 \mu\text{Gal}$ at 140.2°E 36.85°N (Figure S2a, Figure S1b), with
227 the magnitude slightly greater than the model prediction, $-18.5 \mu\text{Gal}$. And the T_{xx} change
228 (Figure S2b) also has good consistency with the model prediction (Figure S2f).

229

230 **3. Source Parameters Inversion Method**

231 For seismic source model inversion, instead of solving for the finite fault model as shown
232 in previous studies [*Cambiotti and Sabadini*, 2012; *Wang et al.*, 2012a, 2012b; *Dai et al.*,
233 2014], here we choose to directly solve for the moment tensor of a point source as
234 presented in [*Han et al.*, 2011, 2013; *Cambiotti and Sabadini*, 2013]. This simple point-
235 source moment tensor is adopted to represent the source model mainly because of the
236 linear relationship between the coseismic gravity and gravity gradient change and the
237 double-couple moment tensor elements, as well as the lower sensitivity of GRACE to the
238 extension of finite fault area [*Cambiotti and Sabadini*, 2012; *Dai et al.*, 2014]. The
239 seismic moment tensor can be straightforwardly resolved using least-squares adjustment
240 (see details in supplementary materials). The simulated annealing algorithm [*Kirkpatrick*
241 *et al.*, 1983] is further used to solve for the centroid depth and location, because of the
242 non-linear relationship between centroid location and coseismic gravity change.

243

244

245 **4. Inversion for Source Parameters**

246 GRACE data have already been shown to provide good constraint on seismic moment,
247 centroid depth [*Han et al.*, 2011, 2013], centroid location [*Cambiotti and Sabadini*, 2013;
248 *Dai et al.*, 2014], and slip orientation [*Dai et al.*, 2014]. The sensitivity study by
249 comparing the residuals for the inverted source parameters at various depths (Figure S3)
250 further shows that GRACE data are able to detect the source depth. Here we present the
251 inversion for all source parameters of a point source including the centroid location and

252 depth together with its seismic moment tensor, using the improved-processed high-degree
253 GRACE data products.

254 **4.1 The 2004 Sumatra-Andaman and 2005 Nias earthquakes**

255 The 26 December 2004 Sumatra-Andaman (M_w 9.2) megathrust earthquake was the
256 largest earthquake in the last 50 years, and it ruptured about 1500 km of the subduction
257 zone along the India-Burma plate boundary. Three months later, the 28 March 2005 Nias
258 (M_w 8.6) earthquake ruptured about 300 km at the adjacent Australia-Sunda plate
259 boundary. In this study, we evaluate the cumulative coseismic slip deformation for the
260 two events from GRACE data, considering the short time difference between these two
261 earthquakes. We first compare GRACE measurements with that predicted from the
262 cumulative coseismic slip models [*Chlieh et al.*, 2007; *Konca et al.*, 2007], then we solve
263 for the fault parameters using GRACE-derived gravity and gravity gradient change.

264
265 For these two megathrust earthquakes, the gravity and gravity gradient change are
266 estimated from the high degree monthly solutions (up to degree 90) of GFZ RL05a
267 NMAX 90 product. CSR RL05 NMAX 96 product are also used to study the coseismic
268 gravity change, but shows a worse model-data misfit compared to GFZ RL05a product.
269 Two more large earthquakes occurred in Sumatra and its nearby region in the last decade,
270 including the 2012 Indian Ocean earthquakes (M_w 8.6 and M_w 8.2), and the 2007
271 Bengkulu earthquake (M_w 8.5). Considering the overlap of the fault area, three coseismic
272 jumps at 26 December 2004, 11 April 2012, and 12 September 2007 are estimated jointly
273 using three Heaviside step functions corresponding to these three earthquakes and a linear
274 trend and three periodicities to fit the GRACE-observed gravity and gravity gradient time

series (Figure S1a) at each grid point. For the 2004 Sumatra-Andaman earthquake, although the coseismic signal is estimated at the epoch 26 December 2004, we consider the jump as the cumulative signal for the two earthquakes three months apart, since the number of data after the two events greatly outweighs the three months of data from January to March 2005.

The localized spectra analysis is carried out for GRACE-observed and model-predicted gravity and gravity gradient change (Figure 2). We can see that the GRACE-observed g_N change agrees well with the model prediction up until around degree 70, although the amplitude is slightly larger than the model prediction. Similarly, for the gravity gradient change, GRACE-observed value is larger than the model prediction but has good agreement until around degree 65. Above degree 65, the GRACE data is much higher than the model prediction, which we interpret as noise. Using the new application of this localized spectra analysis to justify the choice of practical truncation degree, and based on additional comparison in spatial domain, we choose to discard the spherical harmonic coefficients above degree 65.

Hence, GFZ RL05a solutions are truncated to degree 65 to retrieve g_N , T_{xx} , T_{xy} and T_{xz} time series. As shown in Figure S1a, the earthquakes cause the g_N significantly decreasing by about $34 \pm 1.4 \mu\text{Gal}$, which is by far the highest coseismic signal achievable by innovative GRACE data processing. The overall positive-negative-positive pattern for the GRACE-derived g_N change (Figure 3a) over the Aceh region and surrounding ocean is consistent with the slip model prediction (Figure 3e). The maximum g_N change (–

298 $34 \pm 1.4 \mu\text{Gal}$) is almost twice of the model prediction at the same point, $-19.4 \mu\text{Gal}$.
 299 GRACE-derived g_N change is larger than the model prediction mainly due to the small
 300 shear modulus value (30 GPa) applied in our forward model prediction, which
 301 underestimates the moment by about half. Coinciding with g_N change, the GRACE-
 302 derived gravity gradient change (Figure 3b~d), T_{xx} , T_{xy} , T_{xz} also matches well with the
 303 model prediction.
 304

305
 306 Figure 2. Comparison in the spectral domain up to degree 90 from GFZ RL05a product.
 307 The coseismic slip distribution model ([http://www.tectonics.caltech.edu/slip_history/](http://www.tectonics.caltech.edu/slip_history/2004_sumatra/update1/slipAceh_BSSA2007)
 308 [2004_sumatra/update1/slipAceh_BSSA2007](http://www.tectonics.caltech.edu/slip_history/2004_sumatra/update1/slipAceh_BSSA2007)) given by *Chlieh et al.* [2007] for the 2004
 309 Sumatra-Andaman earthquake, and the coseismic slip distribution model
 310 (http://www.tectonics.caltech.edu/slip_history/2005_sumatra/update2/static_out) by
 311 *Konca et al.* [2007] for the 2005 Nias earthquake, are used for the model prediction.
 312 Other descriptions are the same as in Figure 1.
 313
 314

Figure 3. Comparison of GRACE-produced and model-predicted gravity and gravity gradient change up to degree 65 from GFZ RL05a product for the 2004 Sumatra-Andaman earthquake. (a) ~ (d): GRACE-observed g_N (a), T_{xx} (b), T_{xy} (c), T_{xz} (d) change, respectively. (e) ~ (h): model-predicted g_N (e), T_{xx} (f), T_{xy} (g), T_{xz} (h) change, respectively. The black star is the GCMT (Global Centroid Moment Tensor Project) centroid (94.26°E, 3.09°N).

For our source parameters inversion, we adopt the same earth structure model [Chlieh *et al.*, 2007] as the one used for the forward model prediction (Figure 3) to reduce the bias caused by structural differences. The GRACE-estimated location (red star in Figure 4) is remarkably close to the GCMT solution (black star, <http://www.globalcmt.org/>). It is west of the USGS CMT (http://comcat.cr.usgs.gov/earthquakes/eventpage/official20041226005853450_30#scientific_summary, green beach ball), and south of the cumulative coseismic slip models (blue beach ball) [Chlieh *et al.*, 2007; Konca *et al.*, 2007]. Poisson *et al.* [2011] demonstrated that the best model that reproduces the tsunami data among five published slip models is the slip model of Rhie *et al.* [2007], which has the peak slip at south of Nicobar Island, close to our centroid location.

334
335 Our estimated centroid depth (9.1 km) is much shallower than the depth (28.6 km)
336 resolved using seismic data (GCMT and USGS CMT) and the depth (28.3 km) from
337 geodetic data [Chlieh *et al.*, 2007; Konca *et al.*, 2007]. The underestimation of slip at the
338 updip direction and overestimation of slip at downdip direction (closer to measurements)
339 from the checkerboard test [Chlieh *et al.*, 2007], might explain the deeper centroid by
340 geodetic data compared to our solution by GRACE data. In addition, the shallow
341 sediments near trench are thought to accommodate relative plate motions aseismically
342 [Lay *et al.*, 2012]. Hence, the afterslip, which releases about 35% of the coseismic
343 moment [Chlieh *et al.*, 2007], might occur at the shallow region, contributing toward a
344 shallower centroid depth. Finally, the GRACE-solved location is closer to the trench
345 (Figure 4 *bottom*), which indicates a shallow source.

346
347 Our estimated centroid moment tensor, with a total moment of 6.09×10^{22} Nm, is the total
348 of the coseismic slip of the 2004 Sumatra-Andaman earthquake, the afterslip and
349 postseismic slip over the following three months, and coseismic slip of the 2005 Nias
350 earthquake. The estimated moment, M_0 , is smaller than the published total moment
351 (10.37×10^{22} Nm), which is the sum of the coseismic moment 6.93×10^{22} Nm, one-month
352 postseismic moment 2.44×10^{22} Nm by [Chlieh *et al.*, 2007], and the coseismic moment
353 1.0×10^{22} Nm [Konca *et al.*, 2007] for the 2005 Nias earthquake. Dip angle is another
354 important source parameter since the estimation of moment has a strong dependence on
355 dip [Banerjee *et al.*, 2005; Rhie *et al.*, 2007; Han *et al.*, 2011, 2013] for shallow thrust
356 earthquakes. The sensitivity test by Rhie *et al.* [2007] shows the dip angle is not well

357 resolved using the long-period teleseismic data and near-field GPS displacements. Our
 358 estimated dip angle, δ , is 32° (Table S1), larger than the mean value (14°) of cumulative
 359 coseismic slip models by *Chlieh et al.* [2007] and *Konca et al.* [2007]. Based on the plate
 360 boundary geometry (Figure 4 profile AA'), it seems that the GRACE-estimated dip angle
 361 is also larger than the slope angle of the subducting slab. One possible cause for our
 362 smaller seismic moment and larger dip angle might be the trade-off between these two
 363 parameters for shallow thrust earthquakes [*Han et al.*, 2013], where the trade-off is
 364 revealed by their large correlation of 0.4 (Table S6, equation A2). This trade-off implies
 365 that the $M_0 \sin(2\delta)$ is better constrained than each of M_0 and dip individually [*Han et al.*,
 366 2011]. As shown in the moment-dip comparison (Figure S7), the GRACE-estimated
 367 solution is consistent with the cumulative slip model in terms of $M_0 \sin(2\delta)$. Our slip
 368 azimuth [*Dai et al.*, 2014] is about 8° clockwise of the GCMT and USGS CMT solution,
 369 and 8° anticlockwise of the cumulative slip distribution models by *Chlieh et al.* [2007]
 370 and *Konca et al.* [2007]. The correlation (Table S6) between strike and rake angle is
 371 strong, up to about one, which is consistent with the trade-off between these two angles
 372 as discussed in *Han et al.* [2011, 2013] and *Dai et al.* [2014], indicating that the slip
 373 azimuth is better constrained from GRACE data compared to the strike and rake angle
 374 individually for the small-dip thrust earthquakes. For these mega thrust earthquakes,
 375 although a single point source is only a simple approximation for the long rupture, it
 376 provides the critical prior information (the location, dip angle, strike angle) for the slip
 377 distribution model inversion when using near-field GPS displacements or seismic data. In
 378 future studies, we will investigate the potential modeling error caused by using one single
 379 point source for the case of a long rupture.

381
 382 Figure 4. Comparison of centroid moment tensor solutions. In *top* panel, the *blue* slip
 383 contours are for the cumulative slip distribution models (*red* 2D image) [Chlieh *et al.*,
 384 2007; Konca *et al.*, 2007] on the fault plane. The *red* image along the India-Burma plate
 385 boundary is for the slip model by Chlieh *et al.* [2007] for the 2004 Sumatra-Andaman
 386 earthquake, and the *red* 2D map along the Australia-Sunda plate boundary is for the slip
 387 model [Konca *et al.*, 2007] of the 2005 Nias earthquake. The *blue* beach ball represents
 388 the centroid location and moment tensor estimated from the cumulative slip distribution
 389 models, where the centroid moment tensor is computed using the summation of every
 390 moment tensor at each patch on the fault plane, and the centroid location and depth is
 391 computed using the weighted average of every patch's location, with the moment on each
 392 fault patch as the weight. The GCMT solution is denoted as the *black* beach ball, with

location at the **black** star. **Green** beach ball represents USGS CMT solution and location. **Red** beach ball is CMT solution from GRACE GFZ RL05a data, with the location at the **red** star (94.20°E, 3.00°N, 9.1 km). The thick black line is the plate boundary [Bird, 2003], with black triangular representing the subduction zone. The **green** lines indicate the depth contours of the subducting slab at 20-km intervals from Slab1.0 model [Hayes *et al.*, 2012] (<http://earthquake.usgs.gov/data/slab/>). The **bottom two** panels give the vertical profiles of the subducting slab and the crustal structure for profiles *AA'* and *BB'* as marked in the *top* panel. The crustal structure is interpolated from the CRUST 1.0 model (<http://igppweb.ucsd.edu/~gabi/rem.html>). The **green** curve denotes the subducting slab. Black lines are the interfaces between different layers, where the red layer with P-wave velocity of around 8 km/s denotes the mantle. The **red**, **blue**, and **black** arrows denote the centroids locations and dip angles for these **red**, **blue**, and **black** beach balls in the *left* panel, respectively, i.e., the **red** arrow is for GRACE solution (9 km depth, 32° dip), the **blue** arrow is for the cumulative slip distribution models (28 km depth, 14° dip), and the **black** arrow is for the GCMT solution (29 km depth, 8° dip). The GRACE-estimated depth is consistent with the shallow subducting slab near the Trench.

4.2 The 2011 Tohoku earthquake

For the 11 March 2011 Tohoku earthquake (M_w 9.0), the GRACE-derived north component of gravity and gravity gradient change up to degree 70 from CSR RL05 NMAX 96 product as presented in Section 2 are used to solve for the centroid moment tensor of the earthquake source. The estimated centroid location and source parameters from this high degree L2 product are basically consistent with the finite fault parameters inverted using GRACE CSR RL05 NMAX 60 product [Dai *et al.*, 2014]. The location is at 142.2°E, 37.65°N (**red beach ball** in Figure 5), which is only 13 km away from the centroid location determined by Dai *et al.* [2014]. The consistency of the two solutions demonstrates the coherence of the two GRACE products, as well as the coherence of the finite fault modeling and the centroid moment tensor modeling. As discussed by Dai *et al.* [2014], the GRACE-estimated centroid location is almost parallel in latitude with the GCMT centroid, and it is close to the centroid latitude from GRACE data by Cambiotti and Sabadini [2013]. As regards to its distance to the coast, GRACE-estimated centroid

425 is between the location estimated by GPS data only [Ozawa *et al.*, 2011] and the location
426 (blue beach ball) estimated by the combined seismic waves data and on-land/offshore
427 GPS data [Wei *et al.*, 2012]. The GRACE-estimated depth (16 km) indicates a shallow
428 rupture, and agrees well with the centroid depth (18 km) by combined seismic data and
429 GPS data [Wei *et al.*, 2012], as well as the centroid depth (17 km) constrained by
430 GRACE KBR data [Han *et al.*, 2011] and that (16 km) by differently processed GRACE
431 L2 data [Cambiotti and Sabadini, 2013].

432

433 The GRACE-estimated seismic moment, $(4.03 \pm 0.01) \times 10^{22}$ Nm, is slightly smaller than
434 the published results [Han *et al.*, 2011; Ozawa *et al.*, 2011; Wei *et al.*, 2012]. The
435 GRACE-estimated dip angle ($12 \pm 0.1^\circ$) agrees well with the published results, with
436 difference less than 2 degrees, especially agrees well with the dip angle of the subducting
437 slab (Figure 5 *right*). Although the strike ($236 \pm 0.5^\circ$) and rake angles ($113 \pm 0.5^\circ$) are about
438 30 degrees larger than those inverted from the finite fault model [Dai *et al.*, 2014] that
439 fixed the strike angle at 201° , considering the strong correlation between the strike and
440 rake angle (Table 1 *left*), our slip azimuth (123°) is almost the same with the slip azimuth
441 (124°) in Dai *et al.* [2014]. This slip azimuth is consistent with the earlier conclusion that
442 the GRACE-inverted slip direction is about 10° clockwise from that inverted by GPS and
443 seismic waves data [Wei *et al.*, 2012], but it is only about 3° clockwise from the USGS
444 CMT slip direction. Furthermore, this clockwise rotation of the slip direction is validated
445 by another satellite gravity gradiometry measurements - Gravity field and steady-state
446 Ocean Circulation Explorer (GOCE) [Fuchs *et al.*, 2013]. One explanation for the
447 clockwise 10° difference in the slip azimuth is that the observed onshore deformation by

the GPS network is more sensitive to the deep fault slip close to coast but less sensitive to the shallow slip near the trench, while the GRACE-observed gravity data are most sensitive to the largest offshore shallow slip near the trench. Based on finite-fault inversions [Wang *et al.*, 2013], there is an indication to a variable rake angle increasing with depth. Therefore the average slip azimuth angle estimated by GPS can be smaller than that by GRACE.

	Strike	Dip	Rake	M_0	Strike	Dip	Rake	M_0	Strike	Dip	Rake	M_0
Strike	1	0.4	1	0	1	0	0.9	-0.2	1	0.2	0.2	0
Dip	0.4	1	0.2	0.1	0	1	-0.1	0.6	0.2	1	0.4	0
Rake	1	0.2	1	0.1	0.9	-0.1	1	-0.2	0.2	0.4	1	0
M_0	0	0.1	0.1	1	-0.2	0.6	-0.2	1	0	0	0	1

Table 1. The correlation between parameters.

Left panel for the 2011 Tohoku earthquake, the **middle** panel for the 2010 Maule, Chile earthquake, and the **right** panel is for the 2012 Indian Ocean earthquakes. The correlation coefficient, which is calculated from the covariance and standard deviations of the estimated parameters (equation A2 in the supplementary material), evaluates the degree of linear relationship between variables.

465

466 Figure 5. Comparison of centroid moment tensor solutions for the 2011 Tohoku
 467 earthquake. In the *left* panel, the **red** image overlapped on the topography/bathymetry
 468 along the trench is for the cumulative coseismic and postseismic (March 2011) slip model
 469 [Wei *et al.*, 2012; Ozawa *et al.*, 2011], with contours represented by the **blue** lines and
 470 CMT denoted by the **blue** beach ball. **Red** beach ball is CMT solution from GRACE data
 471 (CSR RL05 NMAX 96 product), located at 142.2°E, 37.65°N, with depth as 16 km. The
 472 *right* three panels give the vertical profiles of the subducting slab and the crustal structure
 473 for the profiles *MM'*, *BB'*, and *AA'*, as shown in the *left* panel. The **green** curves in *BB'*
 474 and *AA'* denote the subducting slab. The *MM'* profile denotes the velocity structure at the
 475 Japan Trench deduced from a seismic experiment [Miura *et al.*, 2005], showing details of
 476 the subducting slab, demonstrating consistent characteristics with the slab along the
 477 profile *BB'* from Slab1.0. Black lines are the interfaces between different layers, where
 478 the red layer with P-wave velocity of around 8 km/s denotes the mantle. The **red**, **blue**,
 479 and **black** arrows denote the centroid location and dip angles given by GRACE data (16
 480 km depth, 12° dip), by the slip distribution model (18km, 10° dip), and that by the GCMT
 481 solution (20 km depth, 10° dip). Other markings are the same as in Figure 4. The
 482 GRACE-estimated dip angle (12°) (**red** arrow in *AA'*) agrees well with the dip angle of
 483 the subducting slab.

484

485 4.3 The 2010 Maule, Chile earthquake

486 The 27 February 2010 Maule, Chile (M_w 8.8) earthquake ruptured about 500 km along
 487 Nazca-South American plate boundary, with the Nazca oceanic crust subducting under
 488 the continent of the South American plate to the east direction. For the coseismic gravity
 489 change detection, although g_N change from GRACE contains reliable signal up to much

490 higher degree than g_D change, the magnitude of coseismic g_N change can be smaller
491 compared to g_D change depending on fault parameters. For example, for this east-west
492 slip earthquake, the model-predicted g_N change is only about 40% of the g_D change up to
493 degree 60. Nevertheless, GRACE-observed g_N change still shows its advantage on
494 coseismic signal detection for this 2010 Maule, Chile earthquake as shown later.

495

496 The coseismic gravity and gravity gradient change up to degree 60 are retrieved using
497 monthly solutions from CSR RL05 product. For the forward model prediction, we start
498 from the USGS finite fault model by *Hayes* [2010], shown in red overlapped on top of
499 the local topography/bathymetry in Figure 7. The GRACE-observed g_N , T_{xx} , T_{xy} , T_{xz}
500 change agrees with the model prediction with slightly larger magnitude in both spectral
501 domain and spatial domain up to degree 60. For the GRACE-derived g_N change, the
502 quadruple spatial pattern is consistent with the model prediction. The peak g_N change is
503 $10.6 \pm 1.3 \mu\text{Gal}$ at 69.4°W 32.45°S as shown in Figure S1c and Figure 6a, slightly larger
504 than the predicted value, $7.2 \mu\text{Gal}$. Even for this adverse case of east-west slip, the
505 GRACE-derived g_N change still has slightly larger magnitude than the peak g_D change of
506 $-8.0 \mu\text{Gal}$ from GRACE data [*Wang et al.*, 2012a], due to the smoothing filter applied for
507 g_D to reduce stripes. The north component of gravity gradient change (Figure 6b~d) also
508 match well with the model prediction, with peak value of T_{xz} as $-0.75 \pm 0.11 \text{ mE}$, slightly
509 larger than its model prediction, -0.51 mE .

510

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514

515 Figure 6. Comparison of GRACE-produced and model-predicted gravity and gravity
 516 gradient change up to degree 60 from CSR RL05 product for the 2010 Maule, Chile
 517 earthquake. (a) ~ (d): GRACE-observed g_N (a), T_{xx} (b), T_{xy} (c), T_{xz} (d) change,
 518 respectively. (e) ~ (h): model-predicted g_N (e), T_{xx} (f), T_{xy} (g), T_{xz} (h) change,
 519 respectively. The model prediction is corresponding to the slip model in Hayes [2010]
 520 (<http://on.doi.gov/yVUcUQ>). The black star is the GCMT centroid (73.15°W, 35.98°S).
 521

522

523 GRACE-solved centroid location (73.4°W, 35.2°S, **red star** in Figure 7) using CSR RL05
 524 product is northwest of the centroid solved by other data sets [Hayes, 2010; Vigny *et al.*,
 525 2011]. For example, it is about 40 km north and 50 km west of the centroid location (**blue**
 526 **beach ball**) by Hayes [2010] from broadband seismic waves, and it is about 20 km west
 527 and 80 km north of the GCMT centroid (**black beach ball**). To check the uncertainty on
 528 solved source parameters caused by different GRACE L2 data processing, another two
 529 L2 products, the latest JPL RL05.1 and GFZ RL05a data products, are also used for the
 530 inversion. Both products are truncated at degree 50 to reduce the noise at high degrees.

531 We can see that the GRACE-solved CMT locations by CSR RL05 and JPL RL05.1
 532 products are almost identical to each other, which can be explained by the fact that the
 533 two data products are solved using similar background models and data processing
 534 strategies. While the GFZ RL05a product produce a location about 50 km north and 50
 535 km east of that by the CSR RL05 product. The discrepancy on the locations by three data
 536 products may, to a certain extent, represent the uncertainty of the centroid location
 537 inverted from GRACE data. Nevertheless, for this earthquake, GFZ RL05a data product
 538 shows a relatively higher noise level on the gravity and gravity gradient change (figures
 539 not shown), which is also shown by its larger RMS (Root Mean Square) data-model
 540 differences (e.g. $1.0 \mu\text{Gal}$ for g_N) as compared to the value ($0.93 \mu\text{Gal}$ for g_N) by the JPL
 541 RL05.1 product, where the model is the GRACE-inverted CMT solution using each data
 542 product.

543 GRACE-estimated depth (20.7 km) by the three L2 data products are the same (Table
 544 S2), which indicates its small uncertainty. This GRACE-estimated depth is consistent
 545 with the published depths from geodetic data [*Vigny et al.*, 2011; *Pollitz et al.*, 2011],
 546 which show major slip locates at the the shallow depth (<25 km). It is also close to the
 547 depth from seismic waves data by GCMT. The shallow slip given by GRACE and GPS
 548 data is in consistency with the generation of a strong tsunami and the numerous
 549 aftershocks near the trench.

550 GRACE-resolved seismic moment, $(2.17\sim 2.6)\times 10^{22}$ Nm, is close to the M_0 , 2.39×10^{22}
 551 Nm, by *Hayes* [2010]. But it's slightly larger than the M_0 , $(1.55\sim 1.97)\times 10^{22}$ Nm, in
 552 *Pollitz et al.* [2011], and *Vigny et al.* [2011], as well as that by GCMT and USGS CMT.
 553 Nevertheless, GRACE-resolved dip angle ($25\sim 26^\circ$) is about 8° larger than the average

554 value (18°) of dip angle by seismic waves and that by geodetic data [Pollitz *et al.*, 2011].
 555 It has been long known that there is a trade-off between seismic moment and dip angle
 556 for shallow thrust earthquakes (rake $\approx 90^\circ$) [Han *et al.*, 2013]. For this earthquake with
 557 rake angle (99°) close to 90° , this trade-off might play a role in the larger GRACE-
 558 inverted dip angle, where the coupling is also revealed by their large correlation of 0.6
 559 (Table 1 *middle*). Moreover, the GRACE-estimated moment and dip are in accordance
 560 with the solution by Hayes [2010] in terms of $M_0 \sin(2\delta)$ (Figure S7). The azimuth of slip
 561 direction from CSR RL05 and JPL RL05.1 products is close to that in Hayes [2010] and
 562 that by USGS CMT, while the slip azimuth from GFZ RL05a is about 10° different. As a
 563 trial, we also solve for the CMT parameters with location and depth fixed at the GCMT
 564 solution. By fixing the centroid location, the estimated slip azimuth is 13° anticlockwise
 565 of that in Hayes [2010], leading to a worse relative difference (equation A3), 59.5%,
 566 compared to the 53.9% from the SA inversion. This test shows that an inaccurate
 567 presumed location may affect the solution for the CMT parameters.

568

569

Figure 7. Comparison of centroid moment tensor solutions for the 2010 Maule, Chile earthquake. In *left* panel, the **red** image overlapped on the topography/bathymetry along the Nazca/South American plate boundary is for the slip model by *Hayes* [2010], with contours represented by the **blue** lines and CMT denoted by the **blue** beach ball. **Red** beach ball is CMT solution from GRACE CSR RL05 product, located at 73.4°W, 35.2°S (**red** star), with depth as 20.7 km. **Yellow** star is the location (73.4°W, 35.1°S, 20.7km) of CMT solution (**yellow** beach ball) from GRACE JPL RL05.1 product, which is almost overlapped by the **red** star since they are only about 4 km apart. **Magenta** beach ball is the CMT solution from GRACE GFZ RL05a data product, located at 72.8°W, 34.6°S, with depth as 20.7 km. USGS CMT (**green beach ball**) is from http://earthquake.usgs.gov/earthquakes/eqinthenews/2010/us2010tfan/neic_tfan_cmt.php. The *right three* panels give the vertical profiles of the subducting slab and the crustal structure for the profiles *AA'*, *BB'*, and *CC'*, as shown in the *left* panel. The **red**, **blue**, and **black** arrows denote the centroid location and dip angles given by GRACE CSR RL05 data (21 km depth, 26° dip), by the slip distribution model (28 km, 18° dip), and that by the GCMT solution (23 km depth, 18° dip). Other markings are the same as in Figure 4.

4.4 The 2012 Indian Ocean earthquakes

The 11 April 2012 Indian Ocean (M_w 8.6 and M_w 8.2) earthquakes ruptured within the Indian Oceanic plate near the India-Australia plate boundary off the west coast of northern Sumatra. The mainshock of M_w 8.6 is the largest strike-slip earthquake on record [*Delescluse et al.*, 2012; *Meng et al.*, 2012; *Yue et al.*, 2012], and the aftershock occurred just two hours later with the magnitude of M_w 8.2. Because of its remote offshore location, the geodetic constraints on the static fault geometry are limited [*Meng et al.*, 2012]; thus only seismological data are used to invert for the focal mechanisms. Hence, GRACE data become an important source of independent constraint on the coseismic static deformation. Although the gravity change for vertical strike-slip earthquakes is expected to be small, we show that the coseismic gravity change from GRACE is still considerable, up to -5.7 ± 0.7 μGal for g_N (Figure S1d) and 0.26 ± 0.03 mE for T_{xz} .

Three products are used to estimate the coseismic gravity and gravity gradient change, including the OSU product generated by *Shang et al.* [2015] using the improved energy integral approach [*Guo et al.*, 2015; *Shang et al.*, 2015], the CSR RL05 NMAX 60 product and the GFZ RL05a NMAX 90 product. All solutions are truncated to degree 40 based on the localized spectral analysis and the signal's spatial pattern. The RMS of the difference between the GRACE-estimated g_N , T_{xx} , T_{xy} , T_{xz} and the model prediction are evaluated as shown in Table S3. It shows that GFZ RL05a gives the largest RMS difference, while the CSR RL05 and OSU product produce comparable and small RMS difference. Here, we choose to show the results from the CSR RL05 product.

Guided by the localized spectral comparison, we use only the north component of gravity and gravity gradient change up to degree 40 to study the coseismic signal. As shown in Figure S4a, the GRACE-derived g_N change and gravity gradient change agree well with the cumulative slip model predictions [*Yue et al.*, 2012] in spatial pattern and signal magnitude. The peak g_N change is at 88.4°E, 3.45°N about -5.7 ± 0.7 μGal (Figure S4a), slightly larger to the model prediction at the same location, -3.7 μGal .

As discussed above, three products are all carried out to solve for the source parameters (Table S4) with centroid location fixed at GCMT solution. The CMT parameters agree amazingly well within these three products, especially for strike angle. The rake angle also has the small variation of only 1-2°, while a slightly larger divergence for moment and dip angle are shown with the smaller value given by OSU product. The solved M_0 (10.13×10^{21} Nm) from CSR RL05 product, the cumulative moment for both the

625 mainshock and the aftershock, is close to the other published values. The M_0 is slightly
 626 smaller than the cumulative M_0 by GCMT solution (12.03×10^{21} Nm), USGS CMT
 627 solution (10.7×10^{21} Nm, http://earthquake.usgs.gov/earthquakes/eqinthenews/2012/usc000905e/neic_c000905e_cmt.php), and the published moment of the cumulative slip
 628 model (16.79×10^{21} Nm) by *Yue et al.* [2012]. Our dip angle is also consistent with the dip
 629 angle given by other solutions, just about 4-5° smaller. As shown in Table 1 *right*, the
 630 correlation between dip and moment is about zero for this vertical strike-slip source,
 631 indicating the independence of the estimated dip angle and moment.
 632
 633
 634 The rupture process of these intraplate earthquakes over the diffuse deformation zone is
 635 very complicated, and it raises speculations on whether it is dominated by east-west right-
 636 lateral slips [*Meng et al.*, 2012; *Yadav et al.*, 2013; *Yue et al.*, 2012] or meridian-aligned
 637 left-lateral slips [*Delescluse et al.*, 2012; *Shao et al.*, 2012]. Numerous preliminary results
 638 characterize the fault as a meridian-aligned left-lateral strike-slip [*Shao et al.*, 2012].
 639 Nevertheless, the back projection source imaging from seismic data shows that the
 640 rupture of faults is dominated by east-west right-lateral slip [*Meng et al.*, 2012; *Yue et al.*,
 641 2012]. The east-west right-lateral slip model is further validated by the consistency with
 642 the remote GPS displacements [*Yadav et al.*, 2013]. Coherent with the fault plane by *Yue*
 643 *et al.* [2012], the GRACE-solved CMT also coincides with an east-west right-lateral
 644 strike slip with strike angle as 292°. The reason why the strike angle by GRACE and
 645 GCMT solution is about 180° larger than that by *Yue et al.* [2012] and USGS CMT
 646 solution is that, for this nearly vertical strike-slip source, the dip direction is almost
 647 vertical and it is slightly tilted toward north for the former solutions and toward south for

648 the latter as shown in Figure 8. Taking into consideration of this 180° difference,
649 GRACE-solved slip direction is amazingly consistent with the slip azimuth by other data,
650 only 1-2° different from the GCMT and USGS CMT solutions and 5° different from the
651 cumulative model. For this vertical strike-slip earthquakes, the correlation (Table 1 *right*)
652 between strike and rake angle is significantly reduced.
653

654
655 Figure 8. Comparison of centroid moment tensor solutions for the 2012 Indian Ocean
656 earthquakes. The white/red dot is the fault patches projected to the Earth's surface for the
657 cumulative slip model by Yue *et al.* [2012] for mainshock and GCMT solution for
658 aftershock, with its CMT located at the blue star and represented by the blue beach ball.
659 Red beach ball is CMT solution from GRACE CSR RL05 data, fixed at the GCMT
660 location (black star). Other markings are the same as in Figure 4.
661

662 4.5 The 2007 Bengkulu earthquake

663 The 12 September 2007 Bengkulu (M_w 8.5) earthquake occurred off the west coast of
664 Bengkulu, Indonesia [Konca *et al.*, 2008; Gusman *et al.*, 2010]. It ruptured the plate
665 interface at the Sumatra subduction zone where the Australia plate subducts beneath the

666 Sunda plate, with the strike angle parallel to the Sunda trench. The relatively small
667 seismic moment of this event renders it to be one of the ‘small’ earthquakes detectable by
668 GRACE data [Tanaka *et al.*, 2015]. Same as for the 2012 Indian Ocean earthquakes, to
669 crosscheck the reliability of the coseismic signal, three products for GRACE data are
670 used to estimate the gravity and gravity gradient change corresponding to this 2007
671 Bengkulu earthquake. All solutions are truncated to degree 40 based on the localized
672 spectral analysis (Figure S5), as well as additional comparison in spatial domain. The
673 RMS values of the model-data misfit for g_N , T_{xx} , T_{xy} , T_{xz} show that OSU product seems
674 to give the lowest noise level with the smallest RMS misfit, while GFZ RL05a product
675 gives the worst model-data misfit. Considering the comparison of the residual RMS as
676 well as the comparison in spatial domain, we choose to present the gravity and gravity
677 gradient change from OSU GRACE data product.

678

679 The localized degree variance (Figure S5) shows that the difference between the
680 GRACE-observed and model-predicted [Konca *et al.*, 2008] g_N change is large for both
681 the low degree part and high degree part, which indicates the GRACE noise level is
682 relatively high for this comparatively small earthquake. The GRACE-derived T_{xx} , T_{xy} ,
683 T_{xz} changes agree with the model predictions for degree below 50 but with larger
684 magnitude. With additional comparison in spatial domain by different maximum
685 truncation degree, we choose to use the spherical harmonic coefficients below 40 to solve
686 for fault mechanisms. As shown in Figure 9, the spatial pattern of the GRACE-derived
687 north component of gravity and gravity gradient change still has good consistency with
688 the model prediction. Nevertheless, the surrounding noise is relatively high for this small

689 earthquake. For example, for the g_N change (Figure 9a), there is an abnormal positive
 690 signal west coast of North Sumatra, which is absent in the model prediction (Figure 9e).
 691 Similarly, over the same region, there is uncategorized negative signal for T_{xz} (Figure 9d)
 692 change. Due to this large surrounding noise, it is not suggested to solve for the source
 693 location since the noise can be mistakenly treated as a signal and produce wrong source
 694 parameters.

695

696

697 Figure 9. Comparison of GRACE-produced and model-predicted gravity and gravity
 698 gradient change up to degree 40 from OSU product for the 2007 Bengkulu earthquake.
 699 (a) ~ (d): GRACE-observed g_N (a), T_{xx} (b), T_{xy} (c), T_{xz} (d) change, respectively. (e) ~
 700 (h): model-predicted g_N (e), T_{xx} (f), T_{xy} (g), T_{xz} (h) change, respectively. The slip
 701 distribution model by *Konca et al.* [2008] is used to model the GRACE-commensurable
 702 gravity and gravity gradient change. The black star is the GCMT centroid (100.99°E,
 703 3.78°S).

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706 All three products are used for the inversion of centroid moment tensor by fixing the
 707 centroid location at the GCMT solution. We can see that the divergence for three angles

708 from the three products is large (Table S5), e.g. the difference is up to 11° for strike
709 angle, 10° for dip angle, and 36° for rake angle. This larger variation for the solved
710 source parameters, compared to the case for the 2012 Indian Ocean earthquakes, indicates
711 again that the noise level from GRACE data is relatively high for this magnitude M_w 8.5
712 earthquake. Hence the large variation should be considered when interpret the GRACE-
713 inverted source angles. Nevertheless, the self-consistency for the seismic moment is good
714 from these three products of GRACE data.

715

716 For the seismic moment, GRACE-inverted M_0 from all three products is overall smaller
717 than that from GCMT solution (6.71×10^{21} Nm) and that (4.47×10^{21} Nm) in *Konca et al.*
718 [2008], with GFZ RL05a producing the closest M_0 (4.16×10^{21} Nm) to that in *Konca et al.*,
719 [2008]. The GRACE-solved dip angle is overall larger than the published values, and the
720 dip angle is 18° by CSR RL05 product with the least discrepancy. During the three
721 products, the OSU product produces the closest strike and rake angles with other results,
722 e.g. rake angle (108°) is in-between the value given by GCMT and *Konca et al.* [2008]
723 and it's only 2° larger than the rake angle by the joint inversion using tsunami waveforms
724 and InSAR data in *Gusman et al.* [2010]; but the strike angle is about 20° smaller causing
725 the at least 16° smaller slip azimuth. On the other hand, the slip azimuth from CSR RL05
726 and GFZ RL05a is consistent with the slip azimuth by GCMT and *Konca et al.* [2008],
727 but the individual strike and rake angle are about $20\sim 30^\circ$ smaller.

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738 **5. Conclusions**

739 Advanced GRACE data processing procedure is developed in this paper using localized
740 spectral analysis and using the north component of gravity and its corresponding gravity
741 gradient changes. By establishing the observation model for north component of gravity
742 and gravity gradient change as a linear function of the double-couple moment tensor, the
743 point source parameters are resolved through the least-squares adjustment combined with
744 the simulated annealing algorithm. The GRACE-inverted source parameters generally
745 agree well with the slip models estimated using other data sets, including seismic, GPS,
746 and their combination. While, their differences illustrate the additional and critical
747 offshore constraint from GRACE data for improved estimates of source parameters, as
748 compared to GPS/seismic data. For the 2004 Sumatra-Andaman and 2005 Nias
749 earthquakes, GRACE data infer a shallower centroid depth (9.1 km), which may be
750 explained by the closer-to-trench centroid and by the aseismic slip over the shallow
751 region. For the 2011 Tohoku earthquake, GRACE-estimated centroid location and slip
752 azimuth from two different GRACE data products and two different forward modeling
753 are consistent with each other, demonstrating the coherence of the CSR RL05 NMAX 60
754 and CSR RL05 NMAX 96 products, as well as the coherence of the finite fault modeling
755 and the point-source moment tensor modeling. The slip direction from GRACE data
756 shows a clockwise rotation compared to the slip direction by GPS and seismic waves
757 data, which is also observed by GOCE measurements [*Fuchs et al.*, 2013]. The GRACE-
758 estimated dip directions are steeper than those from GPS/seismic data, about 18° steeper

759 for the 2004 Sumatra-Andaman and 2005 Nias earthquakes, about 8° steeper for the 2010
760 Maule, Chile earthquake, and about $3\sim 13^\circ$ steeper for the 2007 Bengkulu earthquake.
761 The larger dip angles for the 2004 Sumatra-Andaman earthquake and the 2010 Maule,
762 Chile earthquake might be explained by the trade-off between the seismic moment and
763 dip angle, as also indicated by their high correlations. For future studies, the question of
764 how systematic errors from GRACE data and the errors in the forward modeling affect
765 the inverted source parameters needs to be further investigated.

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