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The 2011 Tohoku-Oki earthquake GCMT solution from the GOCE model of the Earth's crust

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ABSTRACT Space gravity missions allow us to make a step ahead in the physics of large earthquakes, M_w higher than 8.5, thanks to the gravity signal from mass rearrangement within the crust and lithospheric mantle and from the ocean water washed away from the epicentral region by co-seismic displacement of the ocean bottom. Although designed to detect the time dependent and static components of the gravity field, the Gravity Recovery And Climate Experiment (GRACE) and the Gravity and steady state Ocean Circulation Explorer (GOCE) space missions play a complementary role in retrieving this new physics by sampling the co-seismic gravity signal and the thickness of the crust, the latter of importance to determine the synthetic expression of the former within realistic, dislocation Earth's models. We present a novel procedure for estimating the global Centroid Moment Tensor (CMT) solution, which provides the principal seismic source parameters (hypocentre and moment tensor) of the 2011 Tohoku earthquake that relies solely on space gravity data from GRACE and GOCE. Increasing the GOCE crustal thickness from 13.0, to 16.5 and 20.0 km, corresponding to the error bounds, the former and the latter values, of the regional value of 16.5 km, the epicentre for the best model moves by about 20 km roughly in the SE direction and the magnitude M_w decreases from 9.19 ± 0.11 for the thinner crust to 9.07 ± 0.11, the latter concordant with the CMT solution from teleseismic waves.

Key words: compressible Earth's models, gravity data, large earthquakes.

1. Introduction

Gravity from space, thanks to Gravity Recovery And Climate Experiment (GRACE) time series, gave us the unprecedented opportunity to retrieve the Gravitational Centroid Moment Tensor (GCMT) solution for the 2011 Tohoku-Oki (M_w = 9.1) earthquake as detailed in Cambiotti and Sabadini (2013), independently from the United States Geological Survey (USGS) and the global Centroid Moment Tensor (CMT) project solutions based on teleseismic wave inversion (Gilbert and Dziewonski, 1975; Dziewonski *et al.*, 1981). GCMT yields a seismic source model closely resembling the solution based on classical seismology, although the moment magnitude is slightly higher (9.13 ± 0.11 compared to 9.08) and the hypocentre is further off-shore by about 40 km, within the oceanic plate, but still within the one-sigma error from the global CMT project solution (Cambiotti and Sabadini, 2013). The concept underlying this first GCMT solution stands on the long-wavelength gravity signature, nowadays detectable by GRACE (Gross and Chao, 2001), associated with the mass redistribution induced by great earthquakes, of magnitude M_w higher than 8.5. GRACE data have been used to study the three major seismic events in the past decade, the 2004 Sumatran (Han *et al.*, 2006; Panet *et al.*, 2007; de Linage *et al.*, 2009; Broerse *et al.*, 2011; Cambiotti *et al.*, 2011), the 2010 Maule (Han *et al.*, 2010; Heki and Matsuo, 2010) and the 2011 Tohoku-Oki (Han *et al.*, 2011; Matsuo and Heki, 2011; Cambiotti and Sabadini, 2012, 2013; Zhou *et al.*, 2012) earthquakes. These megathrust earthquakes occur at fast converging oceanic and continental plates, causing rock volume changes in the region surrounding the fault, as well as deformation of the Earth surface and internal boundaries with density contrasts and uplift of the ocean floor: the latter displaces the ocean water away from the near field, the gravitational effect of which is comparable with that from mass rearrangement within the solid Earth (de Linage *et al.*, 2009; Broerse *et al.*, 2011; Cambiotti *et al.*, 2011).

The GOCE gravity mission, designed to detect the short wavelength static components cannot apparently be used to detect the gravity signal from that large Tohoku-Oki earthquake, although after the quake the static gravity field has in fact been affected by the permanent signature left by the co-seismic mass redistribution. We may hope to see at the GOCE shorter wavelengths with respect to GRACE, the earthquake fingerprint in the gravity field at least in future data processing. GOCE data can become interesting anyway in these studies related to the new physics of large earthquake mass redistribution even at the present day state of analysis in the indirect way of providing for the first time a worldwide self-consistent crustal thickness with its own error bounds. The present analysis is based on the exploitation of the impact on the GCMT solution of the new crust model provided by GOCE (Reguzzoni and Sampietro, 2012a, 2012b), in terms of its effects on the single seismic moment tensor components, hypocentre and on the synthetic gravity pattern at the GRACE wavelengths, to be compared with other crustal models routinely used in geophysics.

2. The co-seismic gravity signature

The co-seismic gravity signature of the 2011 Tohoku earthquake is estimated from the GRACE time series of the Release-04 Level-2 provided by the GeoForschungsZentrum, on the basis of the scheme described in detail in Cambiotti and Sabadini (2013) where oceanic and atmospheric effects have been removed. It is worthwhile to note that the peculiar noise of GRACE data, the so-called stripes, has been reduced by making use of the anisotropic DDK3-filter (Kusche, 2007; Kusche *et al.*, 2009). This filter reduces the co-seismic perturbations of harmonic degree higher than 60, similarly to an isotropic Gaussian filter with averaging radius of 330 km, keeping perturbations of lower harmonic degrees more efficiently than the latter. In order to spatially localize in the surrounding region of the earthquake the GRACE data, consisting of Stokes coefficients which describe the time dependent gravity field over the whole Earth surface, we made use of Slepian functions (Simons *et al.*, 2006; Cambiotti and Sabadini, 2012) bandlimited to harmonic degree 60, consistently with the spatial resolution of DDK3 filtered GRACE data. The Slepian functions are optimally concentrated within the 8 degree radius circular cup centred at the 38.3°N, 142.4°E USGS mainshock. Larger circular cups

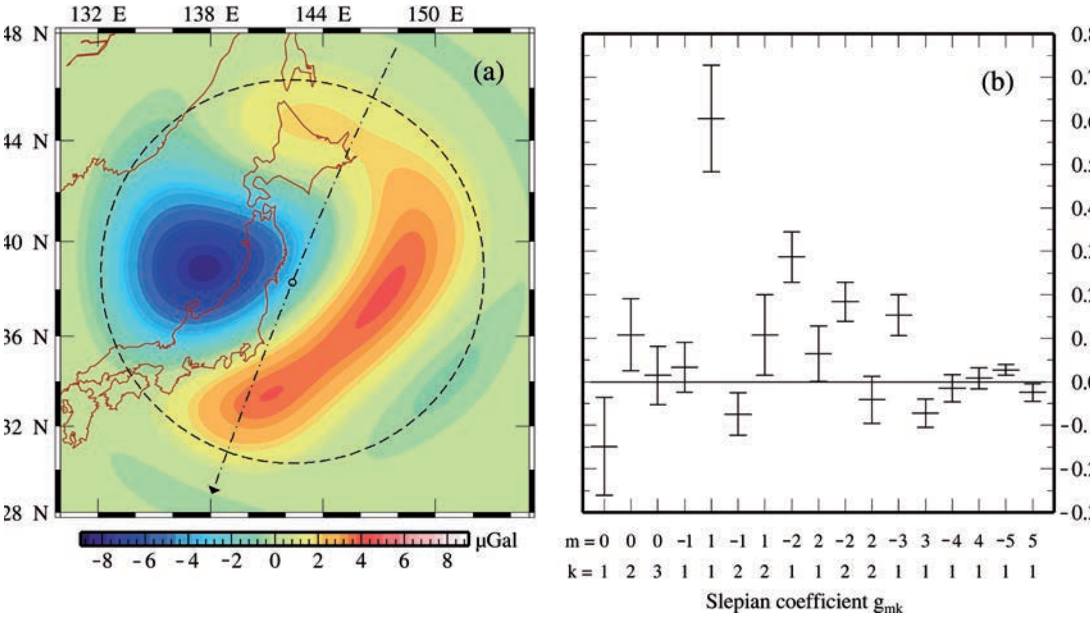

Fig. 1 - (a) Co-seismic gravity anomaly estimated from GRACE, after DDK3 filtering and spatial localization within the circular cup (dashed circle) of half-width 8 degrees and centred at the USGS mainshock; (b) Slepian coefficients *gmk* of the co-seismic gravity anomaly estimated from GRACE (horizontal segments, error bars show one-sigma errors inferred a posteriori from GRACE data analysis).

would include regions where the gravity signature of the earthquake is too small to be detected (Cambiotti and Sabadini, 2013). GRACE time series, in terms of Slepian coefficients rather than Stokes coefficients, are then interpolated looking for the step-like discontinuity at the earthquake time, representing the co-seismic gravity signature. We thus retrive from GRACE data time series 17 Slepian cofficients describing co-seismic gravity anomalies optimally concentrated within the circular cup chosen to spatially localize the signature. Fig. 1a shows the resulting co-seismic gravity change in the spatial domain, while Fig. 1b shows the estimated Slepian coefficients and their *a posteriori* one-sigma errors. The pattern is bipolar: the negative pole in the hanging-wall side, with minimum gravity anomaly of –8.6 ± 1.6 μGal at point 39.0°N, 137.3°E, and the positive pole in the foot-wall side that is characterized by two maxima of + 3.6 ± 1.5 and + 3.4 ± 1.1 μGal at points 38.3°N, 147.9°E and 33.3°N, 141.0°E.

3. Dislocation and crustal models

Synthetic co-seismic gravity anomalies, and synthetic Slepian coefficients, are obtained from our dislocation model described in Cambiotti and Sabadini (2013), characterized by a compressible, stratified Earth based on PREM (Dziewonski and Anderson, 1981) except for the crust that, in the present case, is based on the GOCE crustal model of Reguzzoni and Sampietro (2012a), rather than on the regional average of CRUST2.0 of Bassin *et al*. (2000). The gravitational effect of ocean water redistribution is also taken into account by means of a global ocean layer (Cambiotti *et al*., 2011), without considering the realistic coastline geography which does not affect our conclusions.

Within the GCMT solution, the earthquake is described as a simple point-like source, which means that we retrieve only the principal seismic source parameters, the hypocentre r and the moment tensor m, from which we can infer the source mechanism and the fault plane geometry: the finite extension of the rupture is thus neglected. The relation between the data y, consisting of the first 17 Slepian coefficients of the co-seismic gravity anomaly, and the moment tensor m, of the point-like seismic source is linear:

$$y = G\,(r)\,m \tag{1}$$

where G is the data kernel which non-linearly depends on the hypocentre r. Dislocation theory shows that the data kernel depends on the elastic parameters at the depth of the hypocentre and, thus, can be discontinuous across the internal interfaces of the Earth model (Cambiotti *et al.*, 2011). A seismic source should be represented by a pure double couple, equivalent to the shear dislocation across the fault's plane. In this respect, we have a priori assumed that the trace of the moment tensor is zero, which is equivalent to assume that no centre of compression is triggered at the source. Removing a priori also the residual dipoles would require a non-linear constraint that increases the complexity of the inverse problem, but we will show that this is not necessary because the residual dipoles of our seismic solutions are week compared to the seismic moment of the double couple.

Due to the specific dependence of the data kernel on the elastic parameters of the layer in which the seismic source is located, the choice of the shallow stratification of the Earth model deeply characterizes modelled co-seismic gravity anomalies and, thus, the reliability of the inversion of the co-seismic gravity anomaly observed from GRACE. Particularly, a seismic source located at a given depth would produce very different gravity anomalies, both in the pattern and in the amplitude, if we use Earth models where the Moho discontinuity is below or above the seismic source, i.e., if the seismic source is located within the crust or within the lithospheric mantle (Cambiotti *et al.*, 2011). In order to define an effective crust layer to be used in our spherical Earth model, we consider the crustal model obtained from inversion of GOCE data by Reguzzoni and Sampietro (2012a). Fig. 2 portrays the depth of the Moho and its one-sigma error in a wide region around Japan and Fig. 3 shows the averages of the crustal thickness and its one-sigma error on cups centred at the USGS mainshock for different half-width of the latter. For half-widths from 3 degrees to 20 degrees, corresponding to spatial resolutions from 330 to 2200 km, the average crustal thickness ranges from 15.4 to 17.4 km, with the minimum at the half-width of 10 degrees. For half-widths larger than 20 degrees, the averaged crust thickness increases almost linearly mainly due to the inclusion of the Eurasian plate, where the crustal thickness may exceed 50 km, within the cup. The averaged one-sigma error decreases from 4.7 to 3.2 km increasing the half-width, reflecting the larger errors of the crustal model from GOCE within the continental areas and at the subduction zone (the one-sigma error is about 7 km at the USGS epicentre, Fig. 2) and the small errors over the oceans (about 2 km). In light of these remarks, a reasonable choice is that of an average crust thickness of 16.5 km, in the middle of the range from 15.4 to 17.4 km obtained for half-width smaller than 20 degrees and close to that inferred from CRUST2.0 model of 16.1 km (Cambiotti and Sabadini, 2013), with one-sigma error of about 3.5 km. This is surely a conservative choice because the heterogeneities of the crust thickness is very large, especially in proximity of subduction zones,

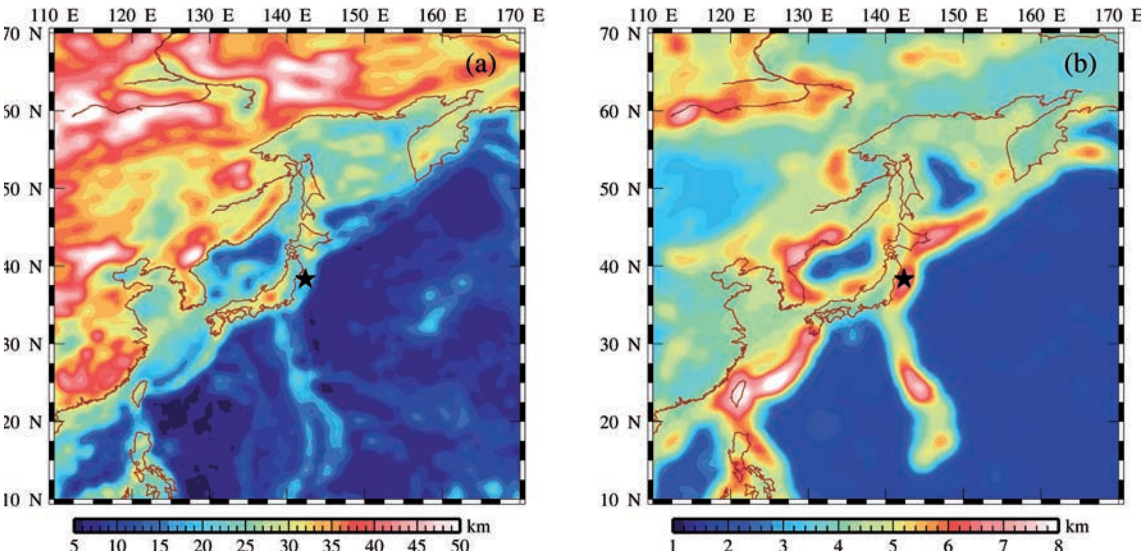

Fig. 2 - Crustal thickness (a) and its one-sigma error estimated from GOCE (b). The star indicates the USGS mainshock at 38.3°N, 142.4°E.

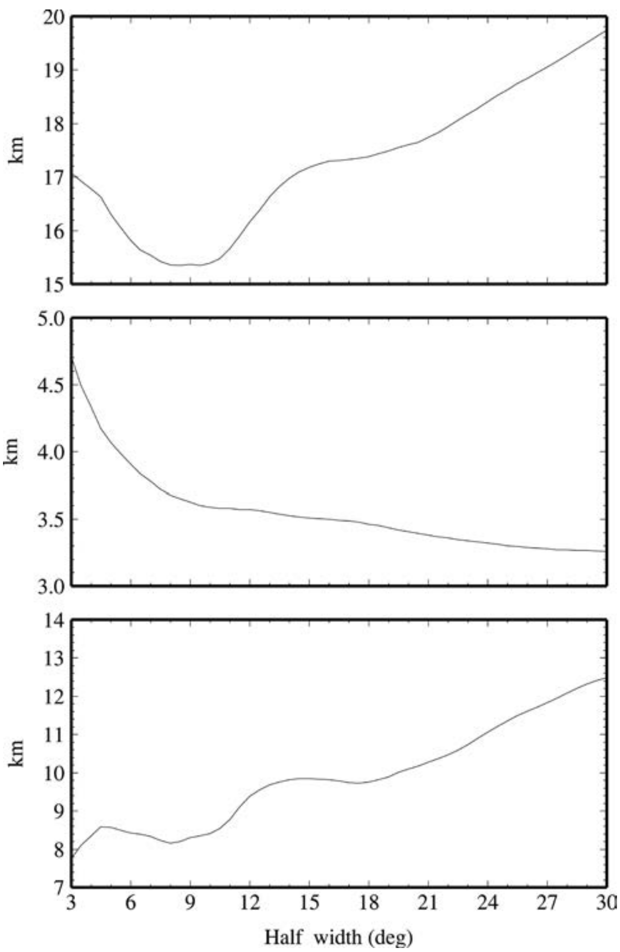

Fig. 3 - Crustal thickness (a) and its one-sigma error (b) averaged on spherical cups of half-width ranging from 3 to 30 degrees; Root mean square of the crustal thickness with respect to its average on the spherical cups (c).

as pointed out in Fig. 3c where we show the root mean square of the crust thickness with respect to its average on the spherical cups, which takes values greater than 8 km.

In order to investigate the sensitivity of our model and inversion method to the crust thickness, in the following we will thus consider three Earth models with the same stratification based on PREM but for the thicknesses of the crust that are 16.5, 13.0 and 20.0 km, with the two latter values corresponding to lower and upper bound errors in the thickness of the crust self-consistently estimated from GOCE solution of the Moho. To better focus on this issue and to not add further complexity, in the following we will not consider a finer stratification of the crust, i.e., we will not discriminate between lower, middle and upper crust.

4. Gravitational Centroid Moment Tensor (GCMT) solutions

Following the probabilistic approach to non-linear inverse problem (Mosengard and Tarantola, 2002; Tarantola, 2005), we obtain the posteriori probability distribution P for principal seismic source parameters, Eq. (B42) of Appendix B in Cambiotti and Sabadini (2013):

$$P\,(r, m) = M\,(r, m)\,R\,(r) \tag{2}$$

where M denotes the conditional probability distribution for the moment tensor m at fixed hypocentre r and R is the marginal probability for the hypocentre. P is characterized by step-like discontinuities at the internal interfaces of the Earth model due to discontinuities of the data kernel at these model interfaces, particularly at the Moho discontinuity, namely at 16.5, 13.0 or 20.0 km for the three Earth models considered in light of the GOCE crustal model of Reguzzoni and Sampietro (2012a). The joint probability P for the epicentre and the moment tensor is bell shaped at fixed depth which means that for these model parameters we can rely on model estimators, such as mean and best models as function of depths, and discuss separately the depth resolving power of GRACE data. In order to accomplish the latter goal, we consider the marginal probabilities for the depth of the hypocentre in Fig. 4. These marginal probabilities are not bell shaped for all three Earth models and, in particular, they are discontinuous at the Moho discontinuities which makes meaningless to establish uncertainties for the depth of the hypocentre corresponding to one-sigma error, as for symmetric, bell shaped Gaussian distributions. However, probabilities can be established within the crust and lithospheric mantle by integrating the distributions of Fig. 4 within these layers. Particularly, the largest probabilities are in the uppermost part of the lithospheric mantle, downwards with respect to the cups just below the Moho, at 16.5 or at 13.0 and at 20.0 km for the three Earth models.

There is an interesting explanation for the cusp-like shape of the marginal probabilities and for the preference for the GCMT solutions to be localized in the lithospheric mantle, just below the Moho, related to the asymmetry of the typical dipolar pattern of the gravity earthquake anomaly of large megathrust earthquakes, characterized by a large negative gravity value on the hanging wall side and a smaller positive value on the footwall, as shown in the Fig. 1. As discussed in Cambiotti *et al.* (2011), to whom we refer for a detailed discussion on the matter, the asymmetry between the maximum and minimum co-seismic gravity anomalies results to be strongly dependent on the gravity reduction due to ocean water displaced away from the near

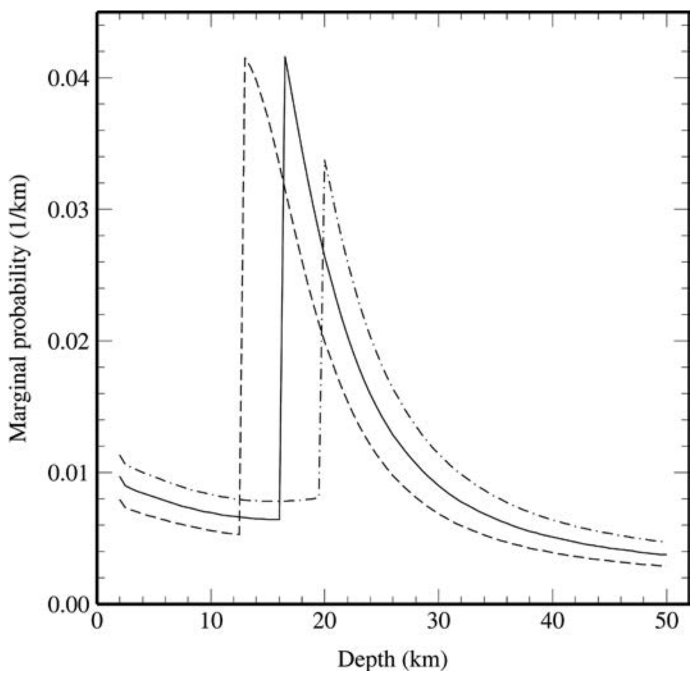

Fig. 4 - Marginal probabilities for the depth of the hypocentre obtained using averaged crustal thickness of 13.0, 16.5 and 20.0 km (dashed, solid and dash-dotted lines, respectively).

field of the earthquake. The amount of water displaced away is related to the co-seismic uplift of the ocean bottom and the latter is about inversely proportional to the elastic parameter β of the layer of the Earth model where the seismic source is located:

$$\beta = \lambda + 2\,\mu \tag{3}$$

where λ and μ are the two Lamé parameters. The β parameters equals 156 GPa in the crust and 223 GPa in the lithospheric mantle, and, thus, the asymmetry of the co-seismic gravity anomaly of the 2011 Tohoku earthquake favours the highest β value of the lithospheric mantle than the lower crustal one, that causes a smaller gravity reduction due ocean water redistribution than that corresponding to a source in the crust. This explains the discontinuity of the marginal probability at the crust-lithospheric mantle interface, where β is discontinuous, and the highest marginal probability being within the lower layer. Fig. 4 thus shows that the preferred hypocentres are close to the Moho discontinuity, but possibly within the lithospheric mantle, in such a way to prevent a too large uplift of the ocean bottom and a too large amount of ocean water displaced away from the near field of the earthquake. The shape of the marginal probability is the same for the three GOCE crustal thicknesses, but the thicker crust of 20.0 km, corresponding the crust upper bound, carries the lowest probability.

In agreement with the findings of Fig. 4, we fix the hypocentre within the lithospheric mantle, just below the Moho discontinuities at 16.5, and at 13.0 and 20.0 km as suggested by the GOCE crustal thickness and its error bounds, and we determine the maximum likelihood epicentre and moment tensor components for each depth by minimizing the negative logarithm of the posteriori probability distribution *-lnP* by means of the step descent algorithm (Tarantola, 2005). The uncertainties are then established from their covariance matrix given by the inverse

Table 1 - Best principal seismic model parameters of the GCMT solutions obtained using crustal thickness of 13.0, 16.5 and 20.0 km (left, middle and right columns, respectively). Note that all hypocentres are located within the lithospheric mantle, just below the Moho, as indicated by the superscript + to the corresponding depths.

	GCMT SOLUTIONS		
CRUSTAL THICKNESS	13.0 km	16.5 km	20.0 km
HYPOCENTER			
Depth	13.0^{+}	16.5^{+}	20.0^{+}
Latitude	37.79 ± 0.45N	37.75 ± 0.46N	37.72 ± 0.47N
Longitude	143.34 ± 0.43E	143.48 ± 0.46E	143.54 ± 0.48E
SEISMIC MOMENT (10^{22} N m)			
Double couple	7.70 ± 3.66	6.18 ± 2.87	5.15 ± 2.34
Residual dipoles	0.14 ± 0.68	0.34 ± 0.54	0.56 ± 0.45
MOMENT MAGNITUDE			
Double couple	9.19 ± 1.1	9.13 ± 0.11	9.07 ± 1.1
Residual dipoles	8.04 ± 0.42	8.30 ± 0.31	8.43 ± 0.24
FAULT PLANE GEOMETRY			
Dip	10.83 ± 2.84°	12.27 ± 3.16°	12.68 ± 3.15°
Slip	90.01 ± 7.61°	89.70 ± 8.65°	89.56 ± 9.82°
Strike	202.13 ± 6.94°	201.90 ± 7.92°	201.80 ± 8.90°

of the Hessian matrix of *-lnP*. The maximum likelihood epicentres, given in Table 1, for the three hypocentres from 13.0 to 16.5 and 20.0 km, are located at points (37.79 ± 0.45°N, 143.34 ± 0.43°E), (37.75 ± 0.46°N, 143.48 ± 0.46°E) and (37.72 ± 0.47°N, 143.54 ± 0.44°E), thus moving to SE by about 20 km from the thinner to the thicker GOCE crust and they are collectively shifted NE-wards by about 40 km with respect to the global CMT project at (37.5°N,143.1°E).

It is interesting to highlight that the magnitude of the double couple of the 2011 Tohoku earthquake is sensitive to the crustal thickness, as shown in Table 1, decreasing from $M_w = 9.19 \pm 0.11$ from the thinner crust of 13.0 km to 9.13 ± 0.11 for 16.5 and to 9.07 ± 0.11 for 20.0. It is notable that these magnitudes well agree, within one-sigma error, with that from the global CMT project of $M_w = 9.08$. A similar argument is valid for the seismic moment, Table 1, where the maximum likelihood seismic moment tensor mainly describes the source mechanism of the tangential dislocation that, for the GOCE crustal thickness of 16.5 km, provides a seismic moment of $6.18 \pm 2.87\ 10^{22}$ N · m ($M_w = 9.13 \pm 0.11$), larger by one order of magnitude than the seismic moment of the residual dipoles $0.34 \pm 0.54\ 10^{22}$ N · m ($M_w = 8.30 \pm 0.31$). The one-sigma error of the seismic moment is large, almost half of the estimate, due to the uncertainties in the Slepian coefficients g_{01} and g_{11} of Fig. 1 estimated from GRACE data. As shown in Fig. 5 of Cambiotti and Sabadini (2013), these coefficients describe the gravity reduction due to ocean water removal and the bipolar pattern perpendicular to the trench, respectively, which are both typical for thrust earthquakes (Cambiotti *et al.*, 2011; Cambiotti and Sabadini, 2012). On the other hand, the fault plane geometry is better constrained by all the other Slepian coefficients due to their smaller uncertainties, Fig. 5, that contribute to define the pattern of the co-seismic gravity anomalies rather than its amplitude. Note that the geometry of the fault is also affected by the thickness of the crust and is also given in Table 1, in terms of dip, slip and strike angles that for 16.5 km are 12.27 ± 3.16°, 89.70 ± 8.65°, 201.90 ± 7.92°, respectively, to be compared with the dip, for example of the 13 km crust, of 10.83 ± 2.84°, with a difference of about 1.5°,

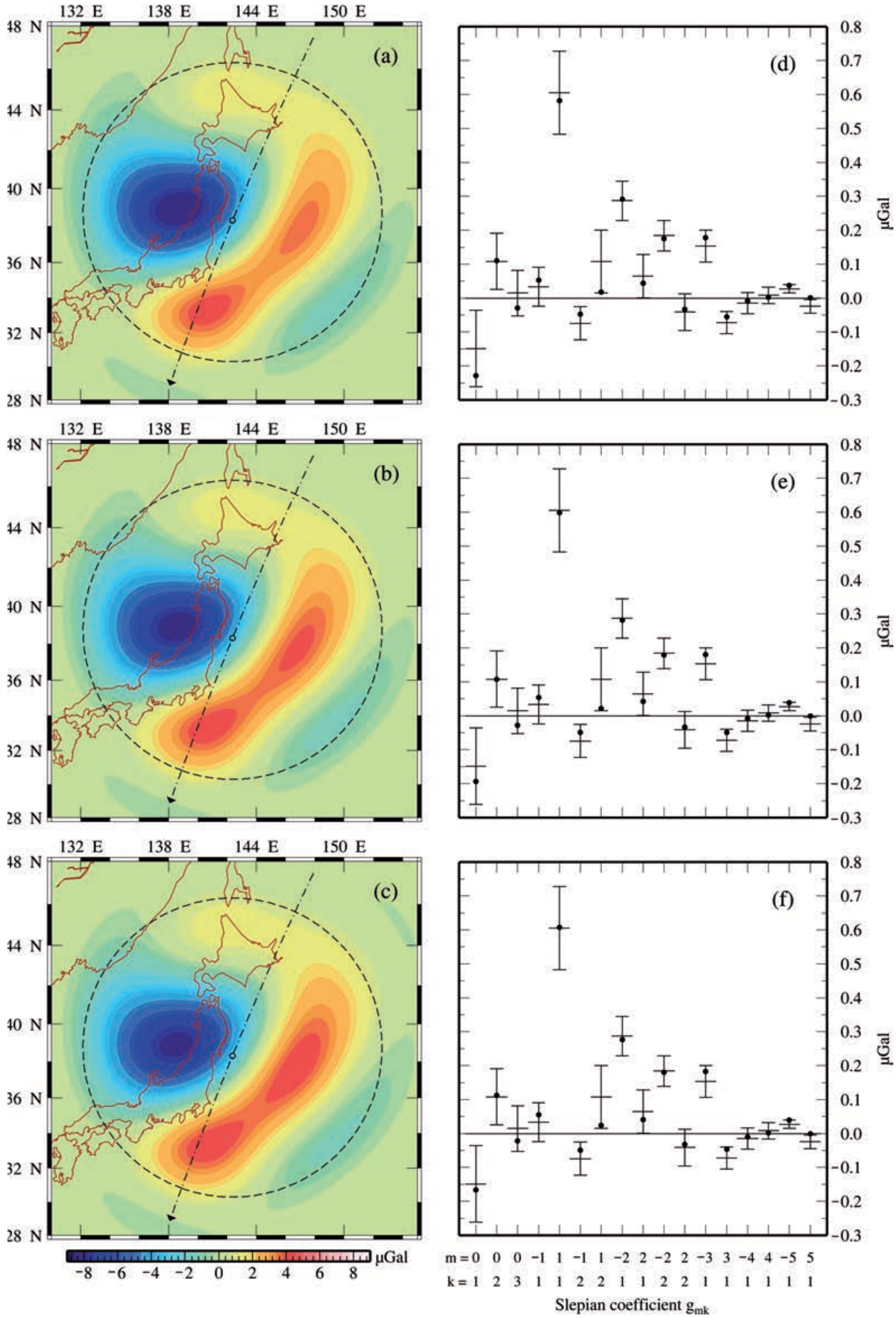

Fig. 5 - Co-seismic gravity anomalies modelled using the GCMT solutions with crustal thickness of (a) 13.0 km, (b) 16.5 km and (c) 20.0 km, after DDK3 filtering and spatial localization within the circular cup (dashed circle) of half-width 8 degrees and centred at the USGS mainshock. Slepian coefficients *gmk* of the co-seismic gravity anomaly estimated from GRACE (horizontal segments, error bars show one-sigma errors inferred a posteriori from GRACE data analysis) and modelled (dots) using the GCMT solutions with crustal thickness of (d) 13.0 km, (e) 16.5 km and (f) 20.0 km.

which is not negligible for such a shallow dipping earthquake. These findings are consistent with the geology of the subduction zone (Takahaski *et al.*, 2004) and with the expectation of a thrust earthquake, thus confirming the reliability of our method and the quality of space gravity data for the study of great earthquakes.

Fig. 5 shows the modelled co-seismic gravity anomalies (panels a, b and c) and the comparison between observed and modelled Slepian coefficients (panels d, e and f) for the three GCMT solutions related to crustal thicknesses of 13.0, 16.5 and 20.0 km. The negative poles in the hanging-wall sides have the minimum gravity anomalies -8.48, -8.28 and -8.13 μGal and the positive poles in the foot-wall sides that are characterized by two maxima, the south-western ones of 4.18, 4.24 and 4.32 μGal and the north-eastern ones of 3.68, 4.04, 4.29 μGal, respectively. Note that this two-dome structure of the positive pole in the offshore region is present for the three crustal models considered and it is also present in the observations. It is due to the use of the anisotropic DDK3-filter and to the gravity reduction caused by ocean water removal from the uplifted crust, a phenomenon that leaves two small domes, remnants of the broader and higher central positive pole caused by mass rearrangement of the solid Earth (Cambiotti and Sabadini, 2012). The modelled Slepian coefficients (panels d, e and f, black dots) show that the g_{01} coefficient deviates the most, in percentage, with respect to the observed one, for the thinner crust model, panel (d), and becomes more concordant with the observations when we move to panel (f) for the thicker crust of 20.0 km.

Fig. 6 shows the squares of the difference between synthetic and observed Slepian coefficients normalized by their one-sigma errors for the three best seismic solutions of Fig. 5, panels d to f. The thinner crust carries the highest differences for the g_{01} Slepian coefficient with respect to the thicker crust, meaning a too large ocean feed back of the thinner crust, which is also discordant the most compared to the CMT solution from teleseismic waves. Similarly, the

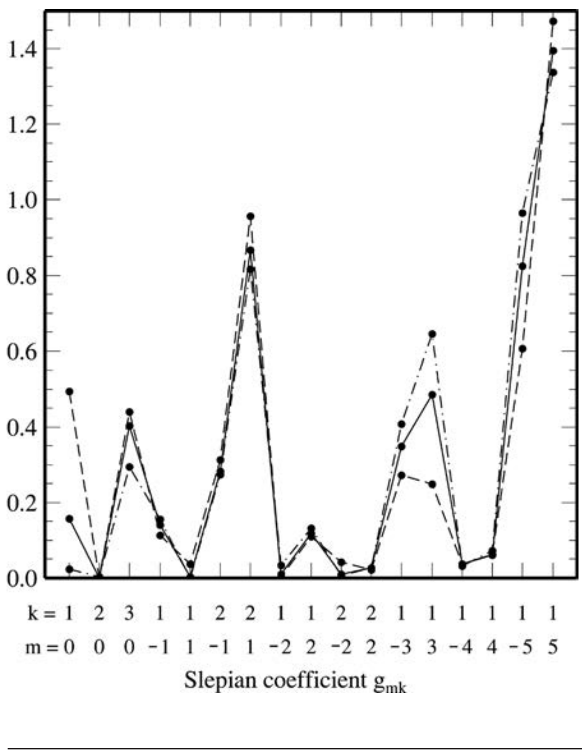

Fig. 6 - Squares of the difference between synthetic and observed Slepian coefficients normalized by their one-sigma errors for the best seismic solutions obtained using the three Earth models with crustal thickness of 13.0, 16.5 and 20.0 km (dashed, solid and dash-dotted lines, respectively).

g_{-31} and g_{31} Slepian coefficients carry the highest difference for the thinner crust, which has to do with the detailed structure of the two domes in the gravity pattern.

5. Conclusions

Our GCMT solution from GRACE and GOCE data is that of a thrust earthquake, whose geometry is consistent with the geological structure of the subduction zone. This solution closely resembles that from the global CMT project based on the inversion of teleseismic waves within the error bounds, in terms of magnitude and epicentre. Our higher estimate of the moment magnitude may, in part, reflect the influence of afterslip in GRACE measurements. This afterslip is thought to contribute to the seismic moment in the weeks following the mainshock (Ozawa *et al*., 2012).

These findings prove that principal seismic source parameters of great earthquakes can be inferred using space gravity data within a new and independent way with respect to the CMT analysis based on the inversion of teleseismic waves or with respect to the USGS solution. Our analysis is indicative of the quality of gravitational solutions for the earthquake that, although based on a different physics with respect to the classical CMT one, proves to be so robust to complement the latter and to provide extra information in terms of the new physics associated with mass readjustment in the epicentral area. On the other hand, the sensitivity of the gravitational solution for the earthquake to the crustal thickness of the spherical Earth's model is significant. Particularly, increasing the crustal thickness from 13.5 to 20.0 km, within the error bounds provided by the GOCE crustal model, the epicentre moves by about 20 km roughly in the SE direction and the magnitude decreases from 9.19 ± 0.11 to 9.07 ± 0.11. Thus, further efforts shall be made in order to account in the inversion of GRACE gravity data for the uncertainties of the crustal thickness and, within the framework of spherical Earth models, for the errors in simply assuming an average crustal thickness. In light of our preliminary results, GOCE data can become interesting in this perspective as they provide for the first time a worldwide model of crustal thickness with higher spatial resolution and consistent error bounds.

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