



# Joint inversion of GPS and high-resolution GRACE gravity data for the 2012 Wharton basin earthquakes

Michel Diamant, Valentin Mikhailov, Elena Timoshkina

## ► To cite this version:

Michel Diamant, Valentin Mikhailov, Elena Timoshkina. Joint inversion of GPS and high-resolution GRACE gravity data for the 2012 Wharton basin earthquakes. *Journal of Geodynamics*, 2020, 136, pp.101722 -. 10.1016/j.jog.2020.101722 . hal-03489750

**HAL Id: hal-03489750**

**<https://hal.science/hal-03489750>**

Submitted on 22 Aug 2022

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



Distributed under a Creative Commons Attribution - NonCommercial 4.0 International License

# Joint inversion of GPS and high-resolution GRACE gravity data for the 2012 Wharton basin earthquakes

Michel Diament<sup>1</sup>, Valentin Mikhailov<sup>2</sup>, Elena Timoshkina<sup>2</sup>

<sup>1</sup> Université de Paris, Institut de physique du globe de Paris, CNRS, IGN, F-75005 Paris, France

<sup>2</sup> Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, 10 B. Gruzinskaya, Moscow, 123995 Russia

## Abstract

The Wharton basin is situated in the north-eastern part of the Indian Ocean. In 2012 it hosted the largest intraplate strike-slip earthquakes ever recorded by geophysical networks. The  $M_w$  8.6 earthquake of April 11, 2012, was preceded by a major foreshock ( $M_w$  7.2) on January 10 and was followed two hours afterward by a  $M_w$  8.2 event. These three large events occurred at the diffuse boundary between the Indian and Australian plates and were almost pure strike-slips on sub-vertical rupture surfaces. Using GRACE data, we first extracted the coseismic and postseismic gravity signals caused by these earthquakes. Then we fitted both GPS and the highest available spatial resolution of GRACE data using the geometry of the fault system suggested by Hill et al. (2015). We propose a regularization, which allows to solve for a linear problem in order to invert GPS and GRACE data under constraints on the rake angle. Our inversion yields a uniform displacement field on all elements of a given fault plane. Our solution shows that even the main displacement occurred on WNW trending faults, comparable displacement also occurred on a rupture striking NNE. Hence, we show that the deformation in this diffuse plate boundary region in 2012 was accommodated by displacements along both fault-systems.

A viscoelastic relaxation of the asthenosphere with a Maxwell viscosity  $10^{19}$  Pa·s successfully explains the postseismic displacements at GPS sites and postseismic gravity signals. The limited postseismic aftershock activity suggests small postseismic slip in the area of the 2012 Wharton earthquakes contrary to what is often observed after large subduction event. Because a part of the observed signal could be related to afterslip, our obtained Maxwell viscosity value should be considered as a lower limit of the asthenospheric viscosity below the Wharton basin.

**Keywords:** Wharton basin, Indian Ocean, 2012 earthquake, GRACE gravity, GPS, joint inversion, asthenospheric viscosity

## Introduction

Located in the north-eastern part of the Indian Ocean, to the East of the Ninety East Ridge (NER), the Wharton basin in 2012 hosted the largest intraplate strike-slip earthquakes ever recorded by geophysical networks. The  $M_w$  8.6 earthquake of April 11, 2012 with a hypocentre at a depth  $d=20$  km (NEIC catalogue of the USGS), was preceded by a major foreshock ( $M_w$  7.2,  $d=19$  km) on January 10 and was followed two hours afterward by a  $M_w$  8.2 event,  $d=25$  km (Fig.1). These three large events were almost pure strike-slips on sub-vertical rupture surfaces (e.g., Duputel et al., 2012; Satriano et al., 2012; Meng et al., 2012; Yue et al., 2012; Ishii et al., 2013; Wei et al., 2013; Hill et al., 2015).

These seismic events occurred southward of the area struck by the megathrust Sumatra-Andaman (26 December 2004,  $M_w$  9.2) and Nias (28 March 2005,  $M_w$  8.6) events in a region classified as a diffuse boundary between the Indian and Australian plates, that is supposed to accommodate about 11 mm/year of relative motion (e.g., DeMets et al., 2010).

The origin of this diffuse boundary is still debated. Space geodetic data from sites within Australia and Indian plates support the existence of distinct Indian, Capricorn, and Australian plates separated by diffuse oceanic plate boundaries (Gordon et al., 2008 and references herein). Reanalysing satellite-derived gravity anomalies and marine magnetic anomalies Jacob et al. (2014) proposed a new tectonic scenario for the study area. The “hard collision” between Indian and Eurasian plates  $\sim 40$  Ma led to a  $\sim 50^\circ$  clockwise change of the spreading direction between India and Antarctica (Patriat and Achache, 1984). Seafloor spreading along the Wharton ridge ceased consequently, as shown by the presence of a fossil spreading centre in the Wharton Basin (Liu et al., 1983). Since that time, the Indian and Australian plates have been moving in parallel, but because of the resistance due to the collision of India with Eurasia, the Indian plate is moving northward  $\sim 11$  mm/year slower than the Australian plate. The broad diffuse zone including a system of transform faults within the Wharton basin accommodates this difference. Jacob et al. (2014) model predicts a total compression of  $\sim 180$  km across this diffuse boundary.

### *Figure 1*

Earthquakes in the Wharton basin are generally occurring predominantly along preexisting and reactivated NNE trending strike-slip faults (Deplus et al., 1998; Abercrombie et al., 2003; Rajendran et al., 2011; Aderhold & Abercrombie, 2016). However, the seismic rupture of the 2012 Wharton events appeared to be more complex. The CMT solution for the main shock provided two fault planes with strike, dip, slip  $\{20^\circ, 76^\circ, 5^\circ\}$  and  $\{289^\circ, 85^\circ, 166^\circ\}$  (<http://www.globalcmt.org/CMTsearch.html>). The  $20^\circ$  plane coincides with NNE system of left-lateral reactivated faults clearly seen on seismic profiles

and bathymetry (e.g. Deplus et al., 1998). However, seismological results, including backprojection (Meng et al., 2012; Yue et al., 2012; Ishii et al., 2013), W phase inversion (Duputel et al., 2012), joint inversion of regional and teleseismic waveform data (Yue et al., 2012; Wei et al., 2013), aftershock locations, as well as finite fault models based on GPS data (Hill et al., 2015) revealed that the rupture reactivated an orthogonal system of NNE left-lateral and WNW right-lateral faults. The NNE strike-slip faults are deep structural boundaries formed at the Wharton spreading centre during the Eocene (e.g. Jacob et al., 2014). On the contrary, the WNW faults are likely to be much younger (e.g. Carton et al., 2014) and some models predict that the main shortening is mostly accommodated on the WNW trending faults (Duputel et al., 2012, Yue et al., 2012; Ishii et al., 2013).

Until recently no WNW active structures were found in the basement topography or seismic profiles (Geersen et al., 2015). But analysing new seismic profiles, Carton et al. (2014) and Qin and Singh (2015) found inclined reflectors extending into the mantle down to 35-37 km and dipping at 30-45 degree. Singh et al. (2017) reported a system of conjugate faults in the area of the  $M_w$  8.2 aftershock.

Zang et al. (2012) backprojected teleseismic P wave observed at three distant regional seismic networks. Their results indicate that the earthquake ruptured a conjugate fault system, composed of two subparallel WNW-ESE faults, and a NNE-SSW fault. They did not invoke displacement along the Ninety East Ridge as reported in other studies.

Another interesting feature of these earthquakes is that they ruptured the upper mantle. Indeed, estimated centroid depth is about 30 km (Duputel et al., 2012) and waveform inversions showed that rupture penetrated down to 50 or even 60 km deep (Yue et al., 2012, Wei et al., 2013). The results of a joint inversion of static GPS offsets and high-rate GPS data (Hill et al., 2015) support this conclusion. Even if it is well known that the mantle contributes most the strength of the oceanic lithosphere (e.g. Burov, 2011) such depth seems to contradict the results of laboratory studies (e.g. Boettcher et al., 2007), which predict that the transition from stable to unstable sliding occurs at about 600°C. According to Hill et al. (2015) this isotherm in the Wharton basin is at 30 km depth (we will discuss this point in more detail hereafter).

Because of the remote offshore location of the events, the GPS constraints on the fault geometry are limited. For example, Meng et al. (2012) concluded that only seismological data could be used to invert for the focal mechanisms. Hence, satellite GRACE gravity data might provide additional constraints on the coseismic deformation (e.g. Mikhailov et al., 2004). Studies of the Wharton events using GRACE data were performed by Han et al. (2015) and Dai et al. (2016). Analysing the low resolution GRACE models of the Center for Space Research (CSR), Houston, USA with  $N=40$  spherical harmonics, Han et al. (2015) showed that the gravity changes were predominantly produced by coseismic compression and dilatation within the oceanic crust and upper mantle and by postseismic

viscoelastic relaxation. Dai et al. (2016) analysed the temporal variations of the northern component of gravity and gravity gradient from CSR RL05 models up to degree 40 to study the coseismic signal.

Here we reinvestigate these earthquakes using more detailed gravity models from CSR and GRGS (Groupe de Recherche de Géodésie Spatiale, France) centres computed up to degree 96 and 80 spherical harmonics respectively. Below, we first discuss the extraction of coseismic and postseismic signals, then we address the comparison of real and synthetic data and present our joint GPS – GRACE modelling of the co- and postseismic processes and discuss our results. Finally, we investigate the mantle viscoelastic postseismic response and estimate the Maxwell viscosity of the asthenosphere.

## 1. Extraction of coseismic and postseismic signal from the GRACE data

Several centres deliver time-series of the GRACE gravity models. Among them are CNES/GRGS in Toulouse, France; GFZ in Potsdam, Germany; CSR in Austin, USA; JPL in Pasadena, USA; AIUB in Bern, Switzerland; TU in Graz, Austria; TONGJI in Shanghai and HUST in Wuhan, China. CNES/GRGS solutions are supposed to be ready for use and do not need any additional filtering (Lemoine et al., 2013). For all other centres, unfiltered and filtered by a dedicated DDK filter solutions are available. To regularize the normal equation, the DDK filter (Kusche, 2007, Kusche et al., 2009) uses a priori error covariance matrices derived from GRACE processing. Depending on the level of filtering, a set of GRACE CSR models is available, from more smoothed DDK1 to less smoothed DDK8 or even unfiltered. From this large set of models we analysed the most widely used not smoothed CSR solutions up to spherical harmonics (SH) degree  $N=40$ ; moderately smoothed (DDK5) CSR solutions up to the maximum available number of SH  $N=96$ ; CNES/GRGS ones with  $N=50$  and  $N=80$ . This choice covers a wide range of existing models.

To extract the coseismic and postseismic signals in the Wharton basin we fitted the gravity time-series  $V_z(\lambda, \phi, t)$  in every point  $(\lambda, \phi)$  with two linear trends in the mean squared sense:

$$V_z(\lambda, \phi, t) = \begin{cases} a_1(\lambda, \phi) + b_1(\lambda, \phi)(t - t_0) & t < t_e \\ a_2(\lambda, \phi) + b_2(\lambda, \phi)(t - t_0) & t > t_e \end{cases} \quad (1)$$

where  $\lambda$  is the latitude,  $\phi$  is the longitude,  $t_0$  is the date of the first gravity model used in our calculations,  $t_e$  is the date of the considered seismic event,  $a_1, b_1$  and  $a_2, b_2$  are parameters of the linear trends before and after the seismic event.

The coseismic jump in the gravity time series is estimated as:

$$\delta(\lambda, \phi) = a_2(\lambda, \phi) + b_2(\lambda, \phi)(t_e - t_0) - a_1(\lambda, \phi) - b_1(\lambda, \phi)(t_e - t_0), \quad (2)$$

and the coseismic jump plus the postseismic trend as:

$$a_2(\lambda, \phi) + b_2(\lambda, \phi)(t_f - t_0) - a_1(\lambda, \phi) - b_1(\lambda, \phi)(t_f - t_0), \quad (3)$$

where  $t_f$  stands for the date of the last gravity model used in calculations.

In the presence of a high level of noise, the estimation of the coseismic and postseismic signals by equations (2-3) appears to be an unstable problem. Therefore, to confirm the estimated coseismic signal given by equation 2 we also used the difference of average values before and after the earthquake as suggested by Han et al. (2015):

$$\sum_{i=1}^{k-1} V_z(\lambda, \phi, t_i) - \sum_{i=k+1}^N V_z(\lambda, \phi, t_i), \quad (4)$$

Here, the date of model  $k$  coincides with the month of the earthquake. Since we averaged over several months after the event, the extracted jump also includes some part of the postseismic gravity signal.

Figure 2 shows the coseismic jumps we obtained applying our different approaches to solutions from different processing centres with different spatial resolution starting from January 2008 up to April 2012 and adding annual and semi-annual components. The coseismic jump on plots A-C is calculated applying equation 4 to unfiltered CSR model truncated at  $N=40$  (A), DDK-5 filtered CSR model up to  $N=96$  (B), and GRGS  $N=80$  models (C). Plot D shows the coseismic jump estimated as the difference of two linear trends (equation. 2) using DDK-5 filtered CSR  $N=96$  model. Plot E shows the coseismic plus postseismic signals calculated using equation 3 and the same CSR  $N=96$  model as on plot (D).

Comparison of plots A to E shows that the location of all extrema is almost identical. Indeed, the half-wavelength of the shortest harmonic corresponding to  $N=40$  is  $l = \pi R_{\text{earth}}/N \approx 500 \text{ km} \approx 5^\circ$  whereas for  $N=96$   $l$  is about 209 km. Hence, for  $N=40$  the uncertainty in the position of a maximum or minimum value is about 250 km and for  $N=96$  it is about 105 km. It is worth noting that the shift of the extrema for the high resolution signals ( $N=96$  on plots B and  $N=80$  on plot C) is relatively small when comparing with the low resolution model (plot A). This results from the filtering of the higher harmonics in CSR and GRGS solutions, which significantly reduces the upper part of their spectra. One should take this into account when comparing observed and computed signals. The morphology of the anomalies shown on plots B to D on Figure 2 is very close. The amplitude of signals on plots B and C is higher than on plot D, because using equation 4, we add a part of the postseismic signal to the coseismic one.

*Figure 2*

Figure 3 shows the gravity time-series in the centres of the negative lobes for CSR and GRGS models with different number of SH. Not-smoothed CSR  $N=40$  models are rather noisy (Fig 3A) so only the more stable jump from average values (equation 4) can be extracted. Considering extracted trends and jumps as a signal and the difference between the real series and the signal as a noise, we found that the noise level is everywhere approximately the same, but of cause the signal-to-noise ratio is much higher in the points of extrema. Estimates of the co- and postseismic signals appear more reliable for higher resolution models. Therefore, we decided to use CSR  $N=96$  models. We estimated

the coseismic gravity jump (figure 2 D) as a difference of two trends (equation 2), and used equation 3 to extract the total coseismic plus postseismic signal (Figure 2E) removing annual and semi-annual components.

*Figure 3*

## 2. Comparison of synthetic and real data

When comparing a synthetic signal to the observed one, it is desirable to filter the synthetic signal with the same filter used for processing the real one. For the GRACE data this becomes critical when  $N > 50$ . Filtering applied by different agencies is adaptive, e.g. DDK filter uses an a priori error covariance matrices derived from GRACE raw data processing. Because it is impossible to reproduce the thorough processing developed by the different groups producing GRACE models, we suggest to apply a filtering calibrated upon a well-known signal from another large earthquake. We used a roll-off filter proposed by F. Pollitz in his Static1D package (<https://earthquake.usgs.gov/research/software/#STATIC1D>) applied to the spherical harmonics higher than  $l > \frac{L_{max}}{2}$ ,

$$F(l) = 1 - \cos\left(\pi\left(1 - \frac{l}{L_{max}}\right)\right), \quad (5)$$

where  $L_{max}$  is the number of SH in the synthetic model.

We calibrated the filter (5) using the rupture model of the 2004 Sumatra earthquake determined with a high accuracy from seismology, geodesy and tide gauge data by Lorito et al. (2010). We used their set of rectangular elements and GPS data and added the coseismic GRACE gravity signal. We determined the coseismic along-strike and along-dip displacements by minimizing a functional composed of: (1) the RMS residual of the calculated and measured northwards and eastwards displacements at GPS sites, normalized to the square of the maximum displacement and, (2) the RMS of the real minus synthetic gravity signal, normalized to the maximum value of the gravity signal and taken with a weight factor  $< 1$ . To calculate the displacements and the associated gravity signal, we used the Static 1D code developed by F. Pollitz, which solves the problem of a rectangular dislocation in a spherically stratified self-gravitating planet (Pollitz, 1996).

The best fit of synthetic to observed GPS displacements and GRGS gravity signals is obtained when calculating synthetic models up to  $L_{max} = 100$  (Mikhailov et al., 2018). Hence, to mimic the processing used in the GRGS solution, the synthetic field should be calculated up to a larger number of SH than there is in the GRACE gravity models (in our case, up to  $N=100$  for the GRGS models containing 80 SH). For the more noisy CSR data containing 96 SH, the best result was obtained applying a simple truncation of SH at  $N=80$ .

### 3. Comparison with fault plane models based on GPS displacements and seismic waveforms.

Different fault-plane models were proposed for the 2012 Wharton seismic events. They mostly differ by the length and strike of the faults (see Singh et al., 2017). To jointly invert GPS and GRACE data, we used the fault-plane model proposed by Hill et al. (2015), based on a joint inversion of high-rate GPS data, teleseismic observations, source time functions from broadband surface waves, and far-field static GPS displacements. The GPS sites in the nearest zone belong to the SuGAR network, which consists of GPS stations situated along the Sumatra subduction zone (Figure 4). The SuGAR network is operated and maintained by the Indonesian Institute of Sciences (LIPI) and the Earth Observatory of Singapore (EOS). In this study we used Static GPS offsets from 43 daily GPS solutions of the SuGAR network listed in the dataset S1 of Supplementary to Hill et al. (2015) adding station CBAY on the Nicobar Islands and stations HBAY, PORT, HAVE, MBDR on the Andaman Islands (Table 1 in Yadav et al., 2013.). Static GPS offsets were estimated using 10 days of data before and after the 11 April 2012 earthquakes, hence the offsets include effects of both the coseismic and early postseismic deformations. The real northing and easting GPS time series were fitted by two trends and a jump to get the offsets. We used the mean square error of the fits as a weight of every GPS station.

Our approach consists in minimizing a functional composed of the misfits of the weighted GPS northing and easting displacements, the weighted misfit of the gravity coseismic jumps and the weighted regularization condition, aimed to keep the rake close to an assigned direction. The minimization of the misfits in GPS and gravity in a mean-squared sense is a linear problem. However, adding the condition that the rake is close to a given value makes the functional nonlinear because the tangent of the rake angle is the ratio of the dip-slip  $D$  to the strike-slip  $S$  displacements on every element of a considered fault plane. To maintain linearity, we used the following approach. Let us consider a rupture model containing  $N$  planes, each plane being subdivided into  $K_i$  elements and with a rake angle at plane  $i$  being  $r_i$ ,  $i=1,2..N$ . We use the following regularization condition:

$$\alpha \sum_{i=1}^N \left\{ \sum_{j=1}^{K_i} \left[ \left( D_j \cos(r_i) - S_j \sin(r_i) \right)^2 + (D_j - \bar{D}_i)^2 + (S_j - \bar{S}_i)^2 \right] \right\} \quad (6)$$

The first term in squared brackets requires the motions at plane  $i$  to occur along a direction with an azimuth  $r_i$ . Unfortunately, two vectors satisfy this condition:  $\{D_j, S_j\}$  with the azimuth  $r_i$  and  $\{-D_j, -S_j\}$  with the azimuth  $r_i + 180^\circ$ . The last direction is geodynamically meaningless and to exclude it we introduce two supplementary conditions in (6): the dip-slip ( $D_j$ ) and strike-slip ( $S_j$ ) displacements at every element of plane  $i$  are close to their average value over all plane  $i$ .



242 Our fault plane model includes four planes - marked I – IV on figure 4A - that we subdivide into  
243 different number of elements (see below). Numerical tests proved that the condition (6) effectively  
244 keeps displacements close to  $r_i$  and eliminates displacements in the opposite direction. The goodness  
245 of fit between the observed and synthetic GPS data depends on  $\alpha$ , the relative weight of the condition  
246 (6). For the examples shown on figure 4, we iteratively found  $\alpha$  assuming almost pure strike-slip at all  
247 planes keeping the maximum misfit at GPS sites smaller than 8%.

248 We first only inverted the GPS data using the Hill et al. (2015) fault-planes geometry and found  
249 that a good fit of the 48 GPS sites was achieved with almost pure strike-slip displacements on the four  
250 considered planes (Fig. 4). Figure 4B shows a solution when planes I-IV were subdivided along strike  
251 in 3, 3, 2 and 2 elements respectively, thus totally 10 elements and therefore 20 unknown values of  $D$   
252 and  $S$ . Fig. 4A shows a solution for a more complex model in which the planes were subdivided along  
253 strike and dip into 3x3, 3x3, 2x2, and 2x2 elements respectively (26 elements and 52 unknowns).

254 For both cases, the computed displacements at GPS sites are very close to the observed ones (on  
255 Fig 4a, the synthetic arrows are above the real ones, on Fig.4b observed arrows are above). The  
256 displacement field on every plane is almost homogeneous. In the simplest model (10 elements), the  
257 displacement is on average 7.3 m on plane I, 4.8 m on plane II, 17.4 m on plane III and 3.6 m on plane  
258 IV. The difference between displacements on elements within each plane is in centimetres (Table 1).  
259 For the more detailed model, the displacements vary between 7.24 and 7.28 m, 4.77 and 4.81 m, 17.34  
260 and 17.41 m, and 3.56 and 3.57 m for planes I-IV respectively. Thus, as already noted by Hill et al.  
261 (2015) the main displacement occurred on plane III striking WNW. A significant displacement also  
262 occurred on plane I striking NNE. Figure 4b shows that the low-resolution ( $N=40$ ) gravity signal  
263 corresponding to the rupture model based on GPS data only, also fits well the observed one. This  
264 confirms Han et al. (2015) results. However, the observed and synthetic gravity signals significantly  
265 differ when taking into account a higher number of spherical harmonics (Fig. 4a). The synthetic  
266 gravity signal has larger amplitude than the observed ones. Indeed, the difference between the  
267 maximum values of the SE and NW observed and synthetic signals is 2.1 and 1  $\mu\text{Gal}$ , and reaches 3.7  
268  $\mu\text{Gal}$  for the SW one. To improve the fit to the observed gravity data, we then performed a joint  
269 inversion for GPS and high resolution CSR satellite gravity data.

270

271 *Figure 4*

272

273 Figure5 shows the solution which fits both the GPS and GRACE CSR  $N=96$  data under  
274 condition (6) on the rake angle. We assigned a weight of the GPS misfit 5 times bigger than that of the  
275 GRACE misfit and 2 times bigger than the weight of the condition (6) on the rake angle. The fit is  
276 good for both data sets (Table 1). For the GPS data the maximum misfit is 7.0%. The positive

277 observed and synthetic SW and NE gravity lobes are now superimposed, and the maximum difference  
278 between the observed and synthetic signals is 1.4  $\mu\text{Gal}$ . We however note a shift of the SE and NW  
279 negative lobes of about  $1^\circ$  in latitude and longitude. This shift is most probably caused by the DDK5  
280 filtering of the highest harmonics. Figure 2 shows that depending on the applied filtering, on the  
281 number of harmonics and on the method of trend extraction, the position of extremum values is  
282 varying up to  $1-2^\circ$ .

283

284

### Figure 5

285

286 The solution we obtain is again almost pure strike-slip. The displacements on fault planes I-IV  
287 are on average 7.0 m, 7.6 m, 12.2 m, and 0.3 m respectively with a difference of several cm between  
288 elements composing each plane (Table 1). In comparison to the inversion without gravity, the  
289 displacements on planes II increased by 2.8 m, and decreased on planes I, III and IV by 0.26 m, 5.2 m  
290 and 3.3 m. Displacement on the plane IV, situated at the Ninety East Ridge, became negligible (30  
291 cm). This difference is not surprising since plane IV is the more distant from the GPS sites and  
292 therefore satellite gravity is better suited to estimate the displacement on this plane.

293

294

## 4 Discussion

295

296 The joint inversion of different geophysical data based on geodynamic models is now widely  
297 used (e.g. Tiberi et al., 2003; Basuyau et al., 2013). The advantage is that various independent data  
298 (GPS, seismic waveforms, CMT solutions, gravity anomalies etc.) can be inverted simultaneously or  
299 sequentially to get a set of model parameters (in our case, parameters of the fault planes, module and  
300 rake of displacement vectors) fitting all data. However, any geodynamic model is an approximation of  
301 the natural process under study. For instance, in our case, the numerical solution for a rupture within a  
302 spherically layered planet - which is commonly used for fault-plane inversion - does not account for  
303 the lateral variations of the lithospheric thickness and its composition, or of the presence of a  
304 subducting slab and numerous faults. Therefore, it is necessary to keep a balance between the precision  
305 and resolution of data, the accuracy of the natural process description through the numerical models  
306 used in an inversion on the one hand and the number of details and parameters, which one desires to  
307 explore through the inversion on the other hand. Our preferred model contains 26 elements, but with  
308 10 elements it is possible to fit both the gravity and GPS observations with a reasonable accuracy.

309

310 In addition to a description of the faults geometry, the two main findings of the thorough study  
311 of Hill et al. (2015) were: (1) the main moment release was on young WNW trending right-lateral  
faults, contrary to previous assumptions that the reactivated NNE trending fracture zone played a

primary role in the rupture process; (2) these faults ruptured deep down in the upper mantle with high stress drops ( $>20$  MPa).

Our rupture solution agrees in general with the one of Hill et al. (2015). As should be expected, the seismic moment magnitude release on the fault planes is nearly identical to the Hill et al. (2015) results when considering GPS only solution. For plane I their estimate is  $M_w 8.2$ , our solution for GPS data only, give 8.28, plane II –  $M_w 8.1$  (8.15), plane III – 8.5 (8.44) and for plane IV 8.2 (8.0). The values we obtain when solving for both GPS and GRACE are also close except for the plane IV, where the seismic moment magnitude decreases down to 7.34 compared to 8.2. According to the NEIC seismicity catalogue, the maximum magnitude of seismic events in the area of planes I-III in the period 01.01.2012 – 30.06.2012 was 8.6, whereas in the vicinity of plane IV it was only 6.2 (shown by small red circle on figures 4 and 5). Hence our result seems to better agree with seismological data. We also recall that the rupture model of Zang et al. (2012) does not contain this fault plane. The small estimated displacements on plane IV (30 cm) simply results from the fact that the fault-plane in our model is very large. Indeed, following (Hill et al., 2015) we assigned an along strike size of plane IV of 232 km, and along dip size as 44 km. However, it is well known that the size of a rupture surface for an earthquake of magnitude 6.2 is much smaller (e.g. Wells and Coppersmith, 1994).

In our joint GPS and GRACE solution, the seismic moment release on plane I is  $2.44 \cdot 10^{21}$  N·m,  $2.65 \cdot 10^{21}$  N·m on plane II and  $3.22 \cdot 10^{21}$  N·m on plane III. Therefore, as in Hill et al. (2015) we conclude that the main seismic moment release occurred on the WNW trending plane III. In our solution the seismic moment release on the NNE trending plane I is 2.4 times smaller than the total release at planes II and III. In Hill et al. (2015) solution, this ratio is larger: 3.7.

Different data including seismic profiles, magnetic anomalies and satellite gravity show that the NNE trending system plays important role in the regional geodynamics (Jacob et al., 2014 and references herein). Numerous structures seen at the sea-bottom and on seismic profiles have been presumably formed in result of strike-slip faulting. They extend to thousands of kilometres approximately in NS direction across the Wharton basin and further to the north in the Bengal Fan (Deplus et al., 1998; Franke et al., 2008, Matthews et al., 2011; Carton et al., 2014). Therefore, the higher energy release on plane I in our solution better suits the regional geodynamics. In our model strike of planes I and IV is  $12^\circ$  when strike of planes II and III is  $108^\circ$ . To produce right lateral strike-slip at WNW faults and left lateral strike-slip at NNE faults simultaneously, the regional compressional stress should be directed NNW, somewhere in the middle between azimuth  $12^\circ$  and  $288^\circ$ , i.e. close to  $330^\circ$ . According to Singh et al. (2017) the principal compressional stress in the Wharton basin is oriented at  $335^\circ$ .

In their study, Hill et al. (2015) mention that their inversion is not able to resolve the depth to which coseismic rupture penetrated. Unfortunately, inclusion of satellite gravity data does not really

347 help. McKenzie et al. (2005) and later Jackson et al. (2008) and Géli and Sclater (2008) suggested that  
 348 the depth of the well constrained earthquake centroids is limited by the 600°C isotherm. Depth to the  
 349 600°C isotherm depends on the age of the lithosphere and thickness of sediments if no thermal  
 350 rejuvenation occurred. Hill et al. (2015) estimated the depth to the 600°C isotherm in the epicentral  
 351 area to be at about 30 km based on a diffusion model (Stein and Stein, 1992) and plate ages from  
 352 Müller et al. (1997). Both half-space cooling model and plate-cooling model predict the depth of the  
 353 600° isotherm for a 50-60 my age lithosphere to be in the interval 33-38 km (e.g. Aderhold and  
 354 Abercrombie, 2016, Fig.2). The thickness of sediments in the study area is 3-4 km (Carton et al.,  
 355 2014). Sedimentary wedge slows down the cooling and uplifts the 600° isotherm but it also causes  
 356 additional isostatic subsidence moving this isotherm down. Mikhailov & Timoshkina (1993) and  
 357 Mikhailov et al. (2007) investigated the half-space cooling model taking into account sedimentation  
 358 and latent heat of basalt crystallization at the lithosphere – asthenosphere interface. Using this solution  
 359 and assuming a rate of sedimentation of 3-4 km per 50-60 my, the depth of the 600°C isotherm should  
 360 be shifted down by ~ 4 km to a depth of about 40 km. Considering the results of waveform inversion  
 361 and the estimation of centroid depths ranging from 30 to 45 km (Duputel et al., 2012), one may  
 362 therefore conclude that the rupture penetrated slightly below the 600°C isotherm and probably some  
 363 alternative mechanism to the frictional slip such as the thermal runaway mechanism may have to be  
 364 considered (e.g., McGuire and Beroza, 2012).

## 365 366 **5 Postseismic viscoelastic relaxation**

367  
368 Finally, we investigate the postseismic relaxation within the three years following the  
 369 earthquakes. The GPS sites of the Sumatran GPS Array (SuGAR) registered coseismic and postseismic  
 370 displacements. Postseismic ones appear to be as large as one fourth of the coseismic ones. GRACE  
 371 time series also show some postseismic signal. Large-scale postseismic processes consist in creep  
 372 (afterslip) and/or viscoelastic relaxation. As shown by Panet et al. (2010), the observed postseismic  
 373 crustal displacements and gravity signals after the December 2004 great Andaman Sumatra earthquake  
 374 are well explained by a viscoelastic relaxation model to which some afterslip on the downdip  
 375 continuation of the ruptured surface is added. In this model, the asthenosphere (between 60 and 220  
 376 km depth) has a Burgers body rheology with transient and steady state viscosities equal to  $4 \cdot 10^{17} \text{ Pa} \cdot \text{s}$   
 377 and  $8 \cdot 10^{18} \text{ Pa} \cdot \text{s}$ , respectively, and the mantle below depth 220 km has a Maxwell rheology with  
 378 viscosity  $8 \cdot 10^{18} \text{ Pa} \cdot \text{s}$  for the upper mantle and  $8 \cdot 10^{20} \text{ Pa} \cdot \text{s}$  for the lower mantle. In addition,  
 379 Mikhailov et al. (2013) showed how post-seismic stress initiated a gradual (~1 m) slip localized at  
 380 ~100-km downdip extension of the coseismic rupture by modelling the seismic cycle in the area of  
 381 Sumatra-2004 using a damage rheology.

382 Some authors explained the recorded postseismic surface displacements and satellite gravity  
383 signal after large events assuming only a viscoelastic relaxation with a low viscosity in the  
384 asthenosphere. For example, to fit GPS data in the area of the Simushir 2006-2007 earthquakes on the  
385 Kuril subduction zone, a Maxwell rheology with a viscosity as low as  $2 \cdot 10^{17} \text{ Pa} \cdot \text{s}$  was suggested by  
386 Kogan et al. (2013). To fit CSR  $N=40$  postseismic gravity signal in the same area Han et al. (2016)  
387 used a Maxwell rheology with a viscosity of  $10^{18} \text{ Pa} \cdot \text{s}$  for the asthenosphere. Using high-resolution  
388 GRACE models and GPS data we previously showed (Mikhailov et al., 2018) that at high resolution  
389 the synthetic gravity signal calculated assuming a low-viscosity Maxwell rheology does not fit the  
390 signal extracted from CSR  $N=96$  GRACE models. On the other hand, the postseismic gravity signal  
391 could be well explained by postseismic creep in a wide zone around the coseismic rupture area,  
392 including a deeper continuation.

393 In the Wharton basin, postseismic processes seem to be much weaker. We estimated the  
394 postseismic displacement field and gravity signal for this area using our preferred fault plane model  
395 (26 elements) obtained by the joint inversion of GPS and gravity data (Fig. 5) and the VISCO1D  
396 software of F. Pollitz. The spherically symmetric layered Earth model used in this calculation is based  
397 on PREM (Dziewonski and Anderson, 1981) with a 60 km thick elastic lithosphere and a 160 km thick  
398 asthenosphere for which the viscosity is considered as an adjustable parameter. The postseismic  
399 viscoelastic relaxation produces displacements at GPS sites in the same direction as the coseismic  
400 ones. The computed postseismic gravity signal has the same pattern as the coseismic one. The  
401 amplitude of the postseismic gravity signal as well as the amplitude of the postseismic displacements  
402 strongly depends on the viscosity of the asthenosphere. Figure 6 shows that we fit the observed data  
403 assuming a Maxwell rheology with a viscosity of  $10^{19} \text{ Pa} \cdot \text{s}$ . For a lower viscosity of  $10^{18} \text{ Pa} \cdot \text{s}$  the  
404 amplitudes of synthetic signals are one order of magnitude larger than the observed ones, whereas for a  
405 viscosity of  $10^{20} \text{ Pa} \cdot \text{s}$ , the synthetic displacements at GPS sites are close to zero. Hence we conclude  
406 that in the region of the Wharton diffuse plate boundary, the GPS and GRACE gravity post seismic  
407 signals are well explained assuming the commonly used value for the asthenospheric viscosity of  $10^{19}$   
408  $\text{Pa} \cdot \text{s}$ . Taking into account that a part of the observed data could be also due to some postseismic  
409 creep, this viscosity value is thus a lower estimate.

410

411 *Figure 5*

412

## 413 **6. Summary**

414

415 The main results of our study are as follows:

• We fitted both GPS and high resolution GRACE gravity data using the geometry of the faults system suggested by Hill et al., (2015) for the 2012 Wharton earthquakes. Our solution shows that even if the main displacements occurred on WNW trending faults, comparable displacements occurred also on a NNE trending one. Hence deformation in this diffuse plate boundary area appears to be accommodated along both orthogonal fault-systems. Our model also shows that the displacements on the faults close to the Ninety East Ridge were small.

• We suggest to use the most detailed gravity models available for the analysis of earthquake induced gravity signals.

• To invert with constraints on the rake angle, we suggest a new regularization which allows keeping the problem linear. This method yields a rather uniform displacement field on the fault planes, without asperities. This does not mean that asperities do not exist, but that GRACE gravity models with  $N=96$  as well as displacements at GPS sites situated far from the epicentre are not sensitive enough to resolve local variations of displacement fields on the fault planes.

• We successfully explain the postseismic displacements at GPS sites and the postseismic gravity signal by viscoelastic relaxation with a commonly used asthenospheric Maxwell viscosity of  $10^{19} \text{ Pa} \cdot \text{s}$ . Taking into account the limited postseismic aftershock activity we conclude that the postseismic slip in the area of the 2012 Wharton earthquakes was limited, contrary to what is often observed after large subduction event. Nevertheless, because a part of the observed signal could be attributed to afterslip, the obtained viscosity value should be considered as a lower limit of the asthenospheric viscosity below the Wharton basin.

## 7. Acknowledgements

We much appreciate the help of F. Pollitz (USGS) who shared his knowledge and software for coseismic and postseismic modelling. We thank K. Chanard for her comments on an early version of our manuscript. We thank SuGAR, CSR and GRGS teams who generously shared their data. We appreciate comments of three anonymous reviewers which allowed to largely improve the manuscript.

VM thanks Université Paris-Diderot for financial support during his stay in Paris. Development of methods of GRACE time-series analysis and regularization was supported by the Megagrant from the Ministry of Science and Education of the Russian Federation under project no. 14.W03.31.0033 “Geophysical studies, monitoring, and forecasting the development of catastrophic geodynamical processes in the Far East of the Russian Federation”. Our study was also supported by CNES through Tosca committee. Work on data analysis of VM and ET was supported by State task of IPE RAS. This study is a result of the cooperation between IPE Moscow and IPGP and is IPGP contribution #

451  
452  
453  
454  
455  
456  
457  
458  
459  
460  
461  
462  
463  
464  
465  
466  
467  
468  
469  
470  
471  
472  
473  
474  
475  
476  
477  
478  
479  
480  
481  
482  
483  
484  
485

## References

- Abercrombie, R. E., Antolik, M. & Ekström G. (2003). The June 2000 Mw 7.9 earthquakes south of Sumatra: Deformation in the India–Australia Plate. *Journal of Geophysical Research*, 108 (B1). doi:10.1029/2001JB000674
- Aderhold, K., & Abercrombie, R. E. (2016). Seismotectonics of a diffuse plate boundary: Observations off the Sumatra-Andaman trench. *Journal of Geophysical Research: Solid Earth*, 121, 3462–3478. doi:10.1002/2015JB012721
- Basuyau, C., M. Diament, C. Tiberi, G. Hetényi, J. Vergne, and A. Peyrefitte (2013), Joint inversion of teleseismic and GOCE gravity data: Application to the Himalayas *Geophys. J. Int.*, 193(1 ), 149 - 163. doi: 10.1093/ggi/ggs110
- Boettcher, M. S., Hirth, G., & Evan B. (2007). Olivine friction at the base of oceanic seismogenic zones. *Journal of Geophysical Research*, 112, B01205. doi:10.1029/2006JB004301
- Bull, J. M., & Scrutton, R. A. (1990). Fault reactivation in the central Indian Ocean and the rheology of oceanic lithosphere. *Nature*, 344 (# 6269), 855.
- Burov E. (2011). Rheology and strength of the lithosphere. *Marin Petroleum Geology*, 28, 1402-1443. doi:10.1016/j.marpetgeo.2011.05.008
- Carton, H., Singh, S. C., Hananto, N. D., Martin, J., Djajadihardja, Y. S., Udrek, Franke, D., & Gaedicke C. (2014). Deep seismic reflection images of the Wharton Basin oceanic crust and uppermost mantle offshore Northern Sumatra: Relation with active and past deformation. *Journal of Geophysical Research: Solid Earth*, 119, 32-51. doi:10.1002/2013JB010291
- Dai, Ch., Shum, C.K., Guo, J., Shang, K., Tapley, B., Wang, R. (2016). Improved source parameter constraints for five undersea earthquakes from north component of GRACE gravity and gravity gradient change measurements. *Earth and Planetary Science Letters*, 443, 118–128. doi.org/10.1016/j.epsl.2016.03.025
- DeMets, C., Gordon, R. G., & Argus, D. F. (2010). Geologically current plate motions. *Geophysical Journal International*, 181(1), 1-80. doi: 10.1111/j.1365-246X.2009.04491.x
- Deplus, C., Diament, M., Hébert, H., Bertrand, G., Dominguez, S., Dubois, J., Malod, J., Patriat, P., Pontoise, B., & Sibilla, J.-J. (1998). Direct evidence of active deformation in the eastern Indian oceanic plate. *Geology*, 26, 131–134.
- Duputel, Z., Kanamori, H., Tsai, V. C., Rivera, L., Meng, L., Ampuero, J.-P., & Stock J. M. (2012). The 2012 Sumatra great earthquake sequence. *Earth and Planetary Science Letters*, 351-352, 247–257.
- Dziewonski, A.M., & Anderson, D.L. (1981). Preliminary Reference Earth Model (PREM). *Physics of the Earth and Planetary Interiors*, 25, 297-356.

486 Geersen J., Bull J. M., McNeill, L. C., Henstock, T. J., Gaedicke, C., Chamot-Rooke, N., Delescluse  
487 M. (2015). Pervasive deformation of an oceanic plate and relationship to large> Mw 8 intraplate  
488 earthquakes: The northern Wharton Basin, Indian Ocean. *Geology*, 43(4), 359-362.  
489 doi.org/10.1130/G36446.1

490 Géli L., & Sclater J. (2008). On the depth of oceanic earthquakes: Brief comments on “The thermal  
491 structure of oceanic and continental lithosphere”, by McKenzie, D., Jackson, J. and Priestley, K.,  
492 Earth and Planetary Science Letters, [2005], 233, p. 337–349. *Earth and Planetary Science*  
493 *Letters*, 3 (265), 766-772. doi:10.1016/j.epsl.2007.08.029

494 Gordon, R.G., Argus, D.F. & Royer, J.-Y. (2008). Space geodetic test of kinematic models for the  
495 Indo-Australian composite plate, *Geology*, 36, 827–830, doi:10.1130/G25089A.1.

496 Han, S.-C., Sauber, J., & Pollitz F. (2015). Coseismic compression/dilatation and viscoelastic  
497 uplift/subsidence following the 2012 Indian Ocean earthquakes quantified from satellite gravity  
498 observations. *Geophysical Research Letters*, 42, 3764–3722. doi:10.1002/2015GL063819.

499 Han S., Sauber J., & Pollitz F. (2016). Postseismic gravity change after the 2006-2007 great  
500 earthquake doublet and constraints on the asthenosphere structure in the central Kuril Island.  
501 *Geophysical Research Letters*, 43(), 3169-3177. doi: 10.1002/2016GL068167

502 Hill E. M., Yue, H., Barbot, S., Lay, T., Tapponnier, P., Hermawan, I., Hubbard, J., Banerjee, P., Feng,  
503 L., Natawidjaja, D., Sieh K. (2015). The 2012 Mw8.6 Wharton Basin sequence: A cascade of  
504 great earthquakes generated by near-orthogonal, young, oceanic mantle faults, *Journal of*  
505 *Geophysical Research. Solid Earth*, 120, 3723–3747. doi:10.1002/2014JB011703

506 Ishii, M., Kiser, E., & Geist, E. L. (2013). Mw 8.6 Sumatran earthquake of 11 April 2012: Rare  
507 seaward expression of oblique subduction, *Geology*, 41(3), 319–322. doi:10.1130/G33783.1.

508 Jackson, J., McKenzie D., Priestley K., Emmerson B. (2008). New views on the structure and rheology  
509 of the lithosphere. *Journal of Geological Society*, 165, 453–465.

510 Jacob J., Dymant J., Yatheesh V. (2014). Revisiting the structure, age, and evolution of the Wharton  
511 Basin to better understand subduction under Indonesia. *Journal of Geophysical Research: Solid*  
512 *Earth*, 119 (1), 169-190. doi:10.1002/2013JB010285

513 Kogan M.G., Vasilenko, N.F., Frolov, D.I., Frymueller J.T. (2013). Rapid postseismic relaxation after  
514 the great 2006–2007 Kuril earthquakes from GPS observations in 2007–2011. *Journal of*  
515 *Geophysical Research: Solid Earth*, 118, 3691-3706. DOI:10.1002/jgrb.50245.

516 Kusche J. (2007), Approximate decorrelation and non-isotropic smoothing of time-variable GRACE-  
517 type gravity field models. *Journal of Geodesy*, 81(11), p. 733-749.

518 Kusche, J., Schmidt, R., Petrovic, S., Rietbroek, R. (2009). Decorrelated GRACE time-variable  
519 gravity solutions by GFZ, and their validation using a hydrological model. *Journal of geodesy*,  
520 83(10), 903–913.



521 Lemoine, J. M., Bruinsma, S., Gégout, P., Biancale, R., Bourgogne, S. (2013). Release 3 of the  
522 GRACE gravity solutions from CNES/GRGS. In: *EGU General Assembly Conference Abstracts*,  
523 15.

524 Liu, C. S., J. R. Curray, and J. M. McDonald (1983), New constraints on the tectonic evolution of  
525 eastern Indian Ocean, *Earth Planet. Sci. Lett.*, 65, 331–342.

526 Lorito, S., Piatanesi, A., Cannelli, V., Romano, F., Melini, D. (2010). Kinematics and source zone  
527 properties of the 2004 Sumatra-Andaman earthquake and tsunami: Nonlinear joint inversion of  
528 tide gauge, satellite altimetry, and GPS data. *Journal of Geophysical Research: Solid Earth*, 115,  
529 B02304. doi:10.1029/2008JB005974

530 McGuire, J. J., & Beroza G. C. (2012). A rogue earthquake off Sumatra. *Science*, 336, 1118–1119.

531 McKenzie, D., Jackson, J., & Priestley K. (2005). Thermal structure of oceanic and continental  
532 lithosphere, *Earth and Planetary Science Letters*, 233, 337–349.

533 Meng, L., Ampuero, J. P., Stock, J., Duputel, Z., Luo, Y., & Tsai, V. C. (2012). Earthquake in a maze:  
534 Compressional rupture branching during the 2012 Mw 8.6 Sumatra earthquake. *Science*,  
535 337(6095), 724-726. DOI: 10.1126/science.1224030

536 Mikhailov, V.O., Timoshkina, E.P. (1993). Analysis of data on the Nansen cordillera, assuming a  
537 thermal model of an oceanic lithosphere. *Proceedings (Doklady) of Russian Academy of Sciences*,  
538 331, 497-499.

539 Mikhailov, V., Tikhotsky, S., Diament, M., Panet, I., & Ballu, V. (2004). Can tectonic processes be  
540 recovered from new gravity satellite data? *Earth and Planetary Science Letters*, 228(3-4), 281-  
541 297. doi:10.1016/j.epsl.2004.09.035

542 Mikhailov V., Lyakhovsky V., Panet, I., van Dinther, Y., Diament, M., Gerya, T., deViron, O.,  
543 Timoshkina, E. (2013). Numerical modelling of post-seismic rupture propagation after the  
544 Sumatra 26.12.2004 earthquake constrained by GRACE gravity data. *Geophysical Journal*  
545 *International*, 94 (2), 640-650. doi: 10.1093/gji/ggt

546 Mikhailov V., Parsons, T., Simpson, R. W., Timoshkina, E., Williams, C. (2007). An explanation for  
547 deep earthquakes under the Sacramento Delta, California, in terms of deep structure and thermal  
548 history. (*Izvestiya*) *Physics of the Solid Earth*, 43 (1), 75-90.

549 Mikhailov V. O., Diament, M., Timoshkina, E. P., Khairtdinov, S. A. (2018). Assessment of the  
550 Relative Roles of Viscoelastic Relaxation and Postseismic Creep in the Area of the Simushir  
551 Earthquake of November 15, 2006, Using Space Geodesy and Gravimetry. *Moscow University*  
552 *Physics Bulletin*, 73 (5), 551–557. DOI: 10.3103/S0027134918050120

553 Mikhailov V. O., Timoshkina E. P. (2019). Geodynamic Modeling of the Process of the Formation and  
554 Evolution of Lithospheric Structures: the Experience of Schmidt Institute of Physics of the Earth,

555 RAS. *Izvestiya, Physics of the Solid Earth*, 55(1), pp. 102–110. DOI:  
 556 10.1134/S1069351319010063

557 Müller, R. D., Roest, W. R., Royer, J.-Y., Gahagan, L. M., & Sclater, J. G. (1997). Digital isochrons of  
 558 the world's ocean floor, *Journal of Geophysical Research: Solid Earth*, 102(B2), 3211–3214.

559 Panet, I., Pollitz, F., Mikhailov, V., Diamant, M., Banerjee, P., & Grijalva, K. (2010). Upper mantle  
 560 rheology from GRACE and GPS post-seismic deformations after the 2004 Sumatra-Andaman  
 561 earthquake. *Geochemistry, Geophysics, Geosystems (G3)*, 11 (6), Q06008.  
 562 doi:10.1029/2009GC002905.

563 Patriat, P., and J. Achache (1984), India-Eurasia collision chronology has implications for crustal  
 564 shortening and driving mechanism of plates, *Nature*, 311, 615–621.

565 Petroy D.E. et Wiens D. A. (1989). Historical seismicity and implications for diffuse plate  
 566 convergence in the northeast Indian Ocean. *Journal of Geophysical Research: Solid Earth*, 94  
 567 (B9), 12301-12319. DOI:10.1029/JB094iB09p12301

568 Pollitz, F.F. (1996), Coseismic deformation from earthquake faulting on a layered spherical Earth.  
 569 *Geophysical Journal International*, 125, 1-14. doi.org/10.1111/j.1365-246X.1996.tb06530.x

570 Qin Y., & Singh S. C., (2015). Seismic evidence of a two-layer lithospheric deformation in the Indian  
 571 Ocean. *Nature communications*, 6, 8298. DOI: 10.1038/ncomms9298

572 Rajendran K., Andrade, V., Rajendran, C. P. (2011). The June 2010 Nicobar earthquake: Fault  
 573 reactivation on the subducting oceanic plate. *Bulletin of the Seismological Society of America*, 101  
 574 (5), 2568-2577. doi:10.1785/0120110002

575 Satriano, C., Kiraly, E., Bernard, P., & Vilotte, J.-P. (2012), The 2012 Mw 8.6 Sumatra earthquake:  
 576 Evidence of westward sequential seismic ruptures associated to the reactivation of a N-S ocean  
 577 fabric. *Geophysical Research Letters*, 39, L15302, doi:10.1029/2012GL052387.

578 Singh S. C., Hananto, N., Qin, Y., Leclerc, F., Avianto, P., Tapponnier, P.E., Carton, H., Wei, S.,  
 579 Nugroho, A.B., Gemilang, W.A., Sieh, K., Barbot, S. (2017). The discovery of a conjugate system  
 580 of faults in the Wharton Basin intraplate deformation zone. *Science Advances*, 3(1), e1601689.  
 581 DOI: 10.1126/sciadv.1601689

582 Stein, C. A., & Stein, S. (1992). A model for the global variation in oceanic depth and heat flow with  
 583 lithospheric age. *Nature*, 359, 123–129.

584 Tiberi, C., M. Diamant, J. Déverchère, C. Petit-Mariani, V. Mikhailov, S. Tikhotsky, and U. Achauer  
 585 (2003), Deep structure of the Baikal rift zone revealed by joint inversion of gravity and  
 586 seismology, *J. Geophys. Res.*, 108(B2), 2109, doi:10.1029/2002JB001880.

587 Yadav R.K., Kundu B., Gahalaut K., Catherine J., Gahalaut V.K., Ambikapthy A., Naidu M. S. (2013).  
 588 Coseismic offsets due to the 11 April 2012 Indian Ocean earthquakes (Mw 8.6 and 8.2) derived  
 589 from GPS measurements. *Geophysical Research Letters*, 40, 3389–3393, doi:10.1002/grl.50601.

590 Yue, H., Lay, T., & Koper, K. D. (2012). En échelon and orthogonal fault ruptures of the 11 April  
591 2012 great intraplate earthquakes. *Nature*, 490, 245–249, doi:10.1038/nature11492.

592 Wells D.L., Coppersmith K.J., (1994) New empirical relationships among magnitude, rupture length,  
593 rupture width, rupture area, and surface displacement. *Bulletin of the Seismological Society of*  
594 *America*. 84 (4). 974–1002.

595 Wei, S., Helmberger, D. , & Avouac, J.-P. (2013). Modeling the 2012 Wharton basin earthquakes off-  
596 Sumatra: Complete lithospheric failure. *Journal of Geophysical Research: Solid Earth*, 118,  
597 3592–3609, doi:10.1002/jgrb.50267.

598 Zhang, H., Chen, J., & Ge, Z. (2012). Multi-fault rupture and successive triggering during the 2012  
599 Mw 8.6 Sumatra offshore earthquake. *Geophysical Research Letters*, 39, L22305, doi:10.1029/  
600 2012GL053805.

601

602

## Figure captions

*Fig.1 Location of the  $M_w$ 8.6 strike-slip seismic event of 11 April 2012 in the Wharton Basin.*

*Green and yellow stars mark epicenters of the main aftershock  $M_w$ 8.2, which occurred 2 hours later and the main foreshock  $M_w$ 7.6 of 10 January 2012. Colored dots show position and depth of seismic events from the USGS NEIC catalogue for the period 10 January-28 December 2012. Solid black lines show the rupture model of Hill et al. (2015). The topography and bathymetry are from ETOPO1 (<https://maps.ngdc.noaa.gov/>).*

*Figure 2. Comparison of the gravity signal (in  $\mu$ Gals) for the April 2012 Wharton earthquakes using the GRACE models of CSR and GRGS for the period 01/2008 - 06/2014 with different resolution. Dashed-dot line shows Sumatra trench, solid curve marks the shoreline. A – coseismic jump calculated as the difference between average values before and after the earthquake (eq. 4) using not filtered CSR models truncated at  $N=40$ . Crosses show the locations of extrema from Han et al. (2015). B – same as A, but using CSR  $N=96$  models filtered by DDK5 and applying (eq. 4). C – same as B, but based on GRGS  $N=80$  models. D – coseismic jump estimates as the difference of two trends (2) with annual and semi-annual components using CSR  $N=96$  models. E – coseismic plus postseismic signal calculated using eq. (3) and CSR  $N=96$  models also with annual and semi-annual components.*

*Fig. 3 Time-series ( $\mu$ Gals) in the vicinity of (95E, -3S). Note the higher noise level in the un-filtered CSR  $N=40$  model.*

*Fig.4. Gravity anomaly in  $\mu$ Gals calculated using the fault planes model geometry of Hill et al. (2015) in result of inversion of GPS data only. Black lines – projection of vertical fault planes to the Earth surface. Green and black arrows show synthetic and real GPS coseismic displacements. Coloured maps on plot A and B are the observed coseismic gravity signals from Fig.2B and A. Dotted isolines show the synthetic signals: (a) truncated at  $N=80$  for the more detailed model, (b) truncated at  $N=40$  signal for the simple model. Crosses on the right plot show position of extrema in Han et al. (2015). Red circles mark the main seismic events (size proportional to magnitude). Small circle close to plane IV shows the strongest event  $M_w$ 6.2 which occurred in vicinity of this plane.*

*Fig.5. Comparison of real (Fig. 2b) and synthetic gravity signals and GPS displacements for the fault plane model based on Hill et al. (2015) geometry and joint inversion of GPS and CSR  $N=96$  data with regularization (6).*

638  
639  
640  
641  
642  
643  
644  
645  
646  
647  
648  
649  
650  
651  
652  
653

*Fig. 6*

*a - Postseismic gravity signal (in  $\mu\text{Gal}$ ,  $N=80$  SH) during two years after the April 2012 event in result of viscoelastic relaxation calculated using F. Pollitz code VISCO1D and our model based on GPS and GRACE data. For notations see Fig 4,5.*

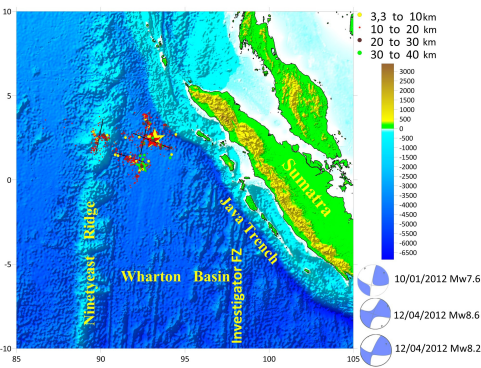
*b – Comparison of coseismic plus postseismic signals. Coloured scale shows real data (Fig.2E), isolines – synthetic coseismic (Fig.2 D) plus postseismic (Fig.6a) signals.*

*c – Solid curves show amplitude of total (black), easting (red) and northing (blue) displacements at two GPS sites of the Sumatran GPS Array (SuGAR). Symbols of the same colour show modelled displacements assuming an asthenospheric viscosity of  $10^{19} \text{ Pa} \cdot \text{s}$ . Horizontal axis shows time, vertical axis is displacements (meters).*

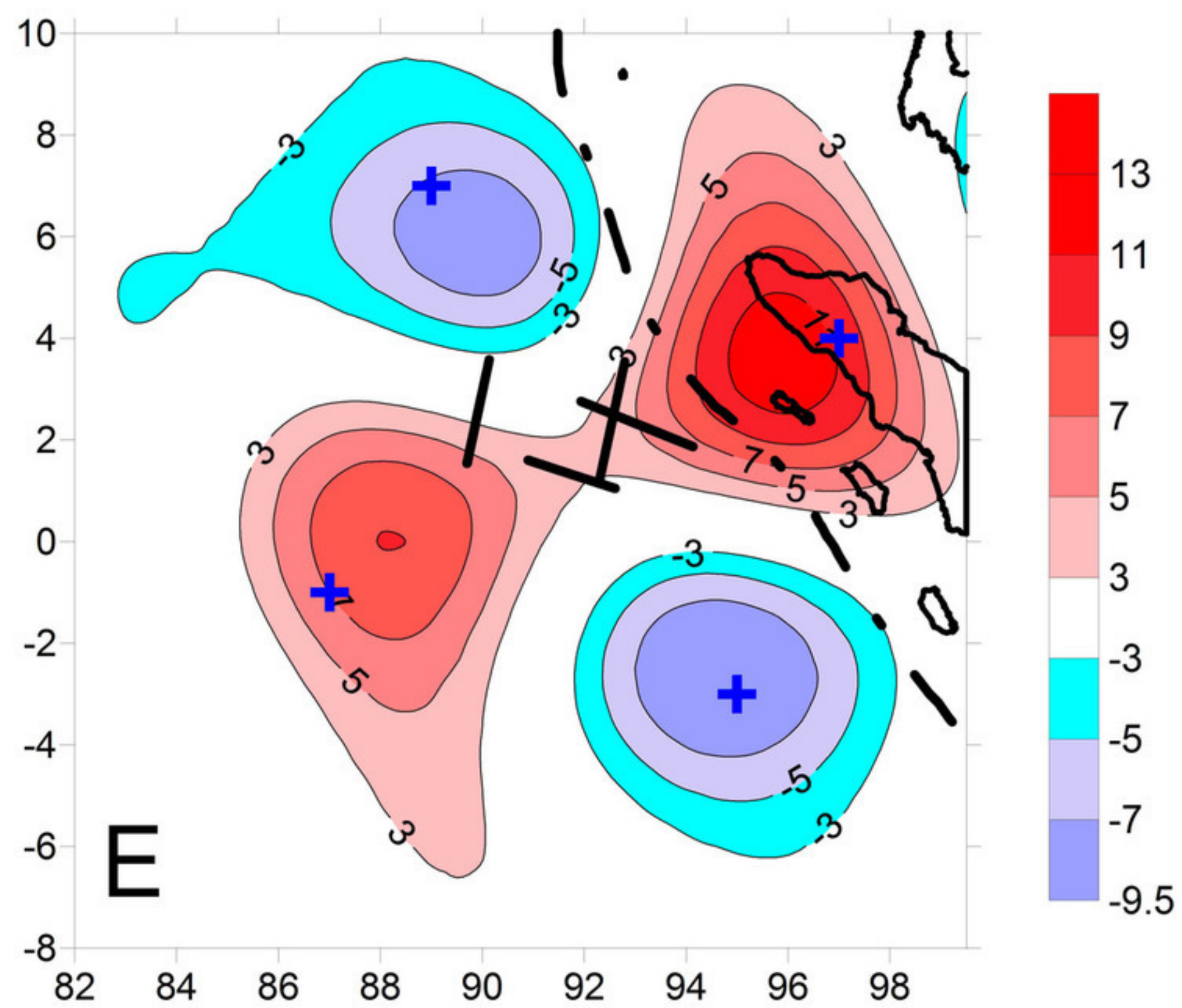
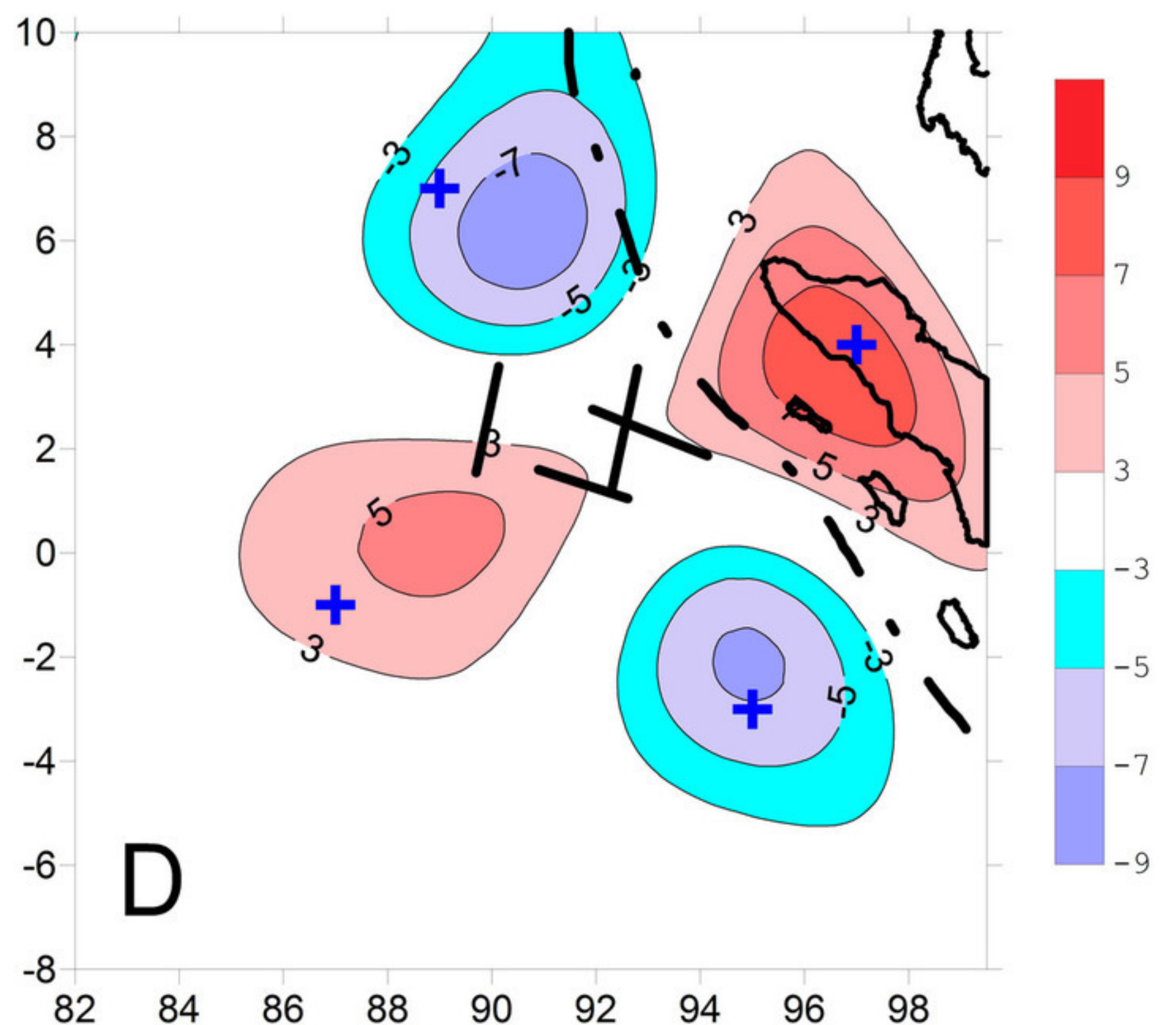
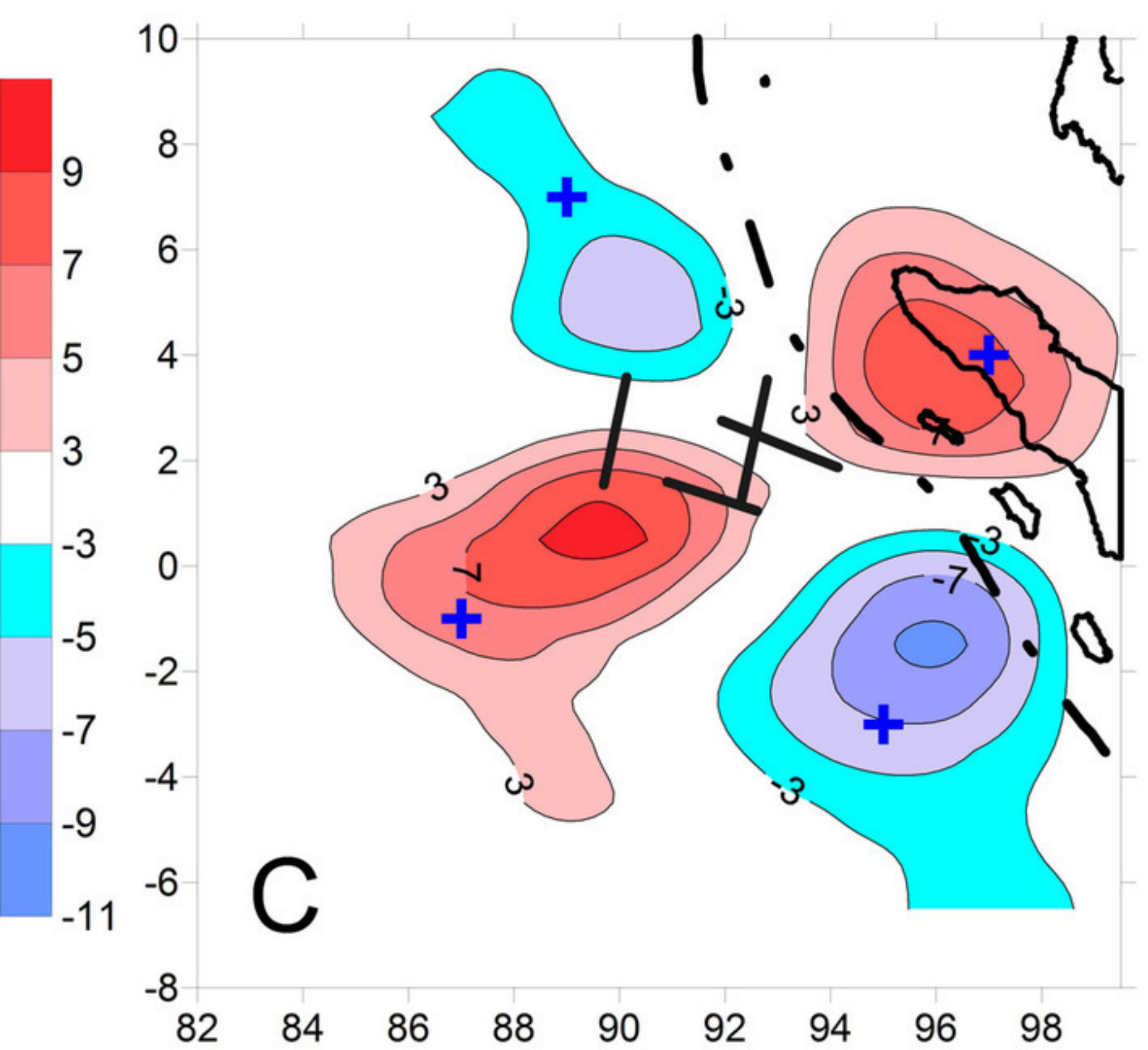
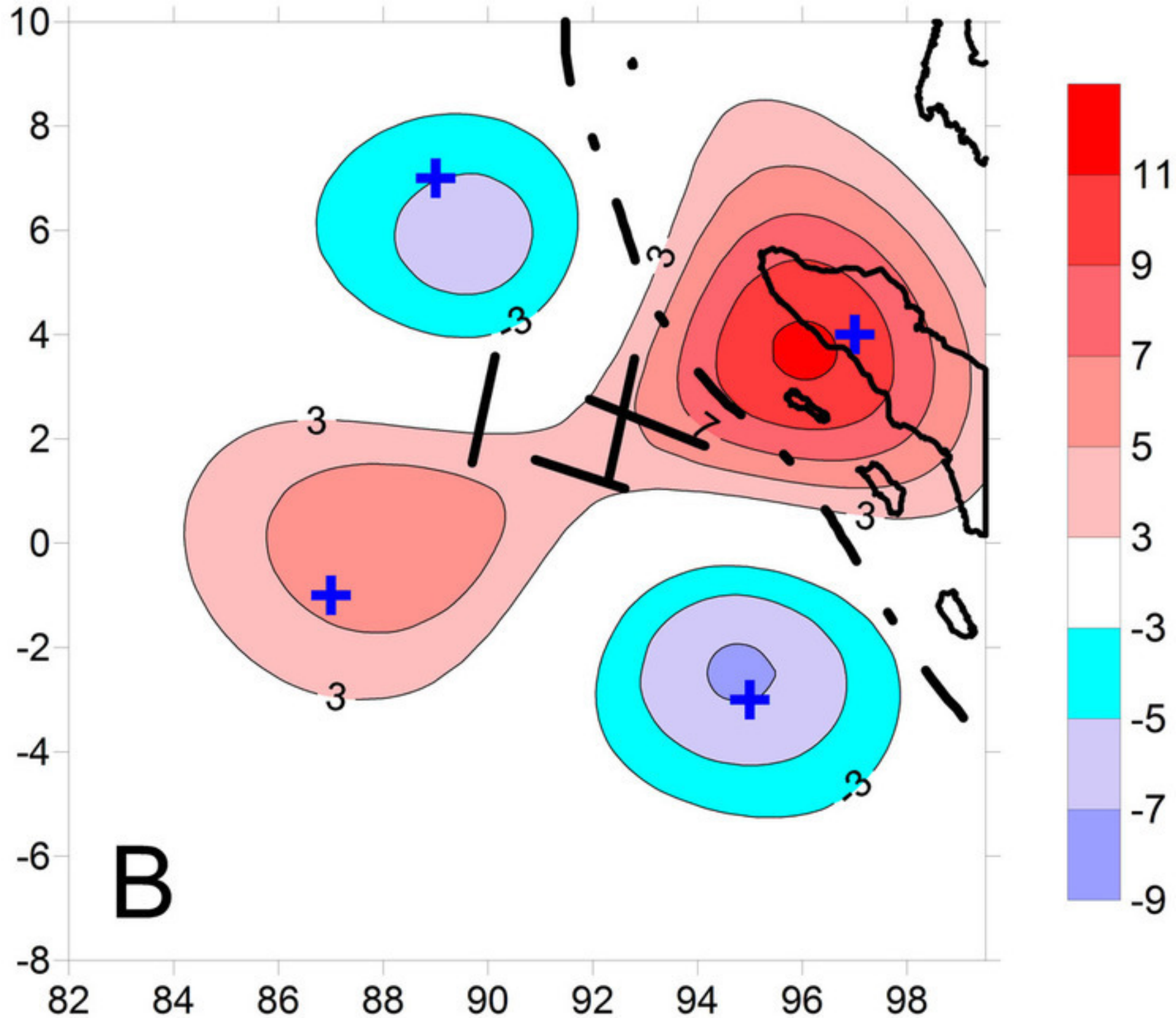
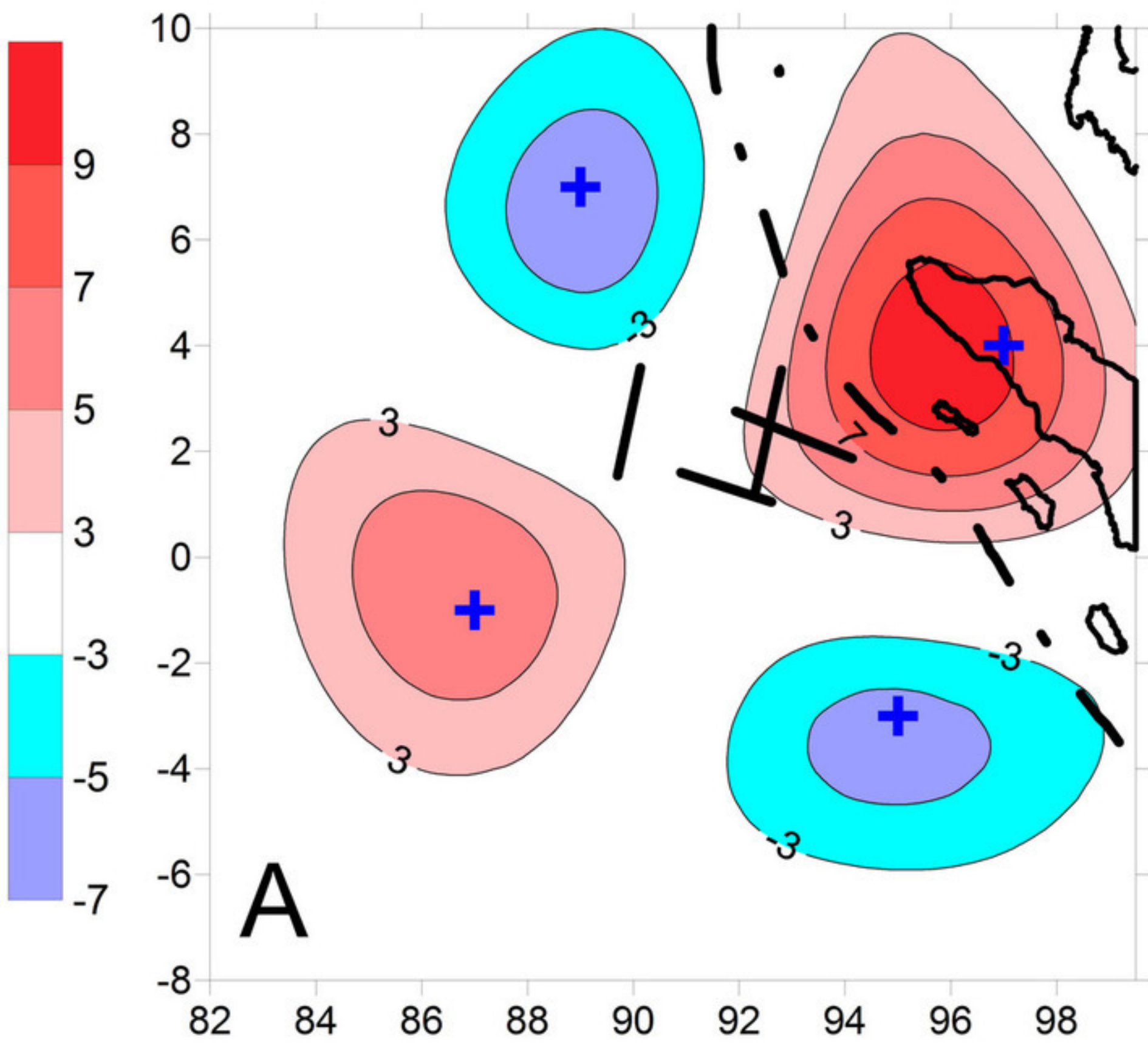
*Table 1 Solutions for different number of elements in planes I-IV using GPS only (lines 1, 2) and GPS+gravity (line 3). GPS misfit is maximum value of difference between real and synthetic absolute displacements at GPS sites in %.*

## Highlights

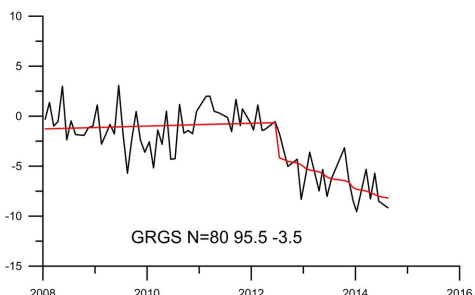
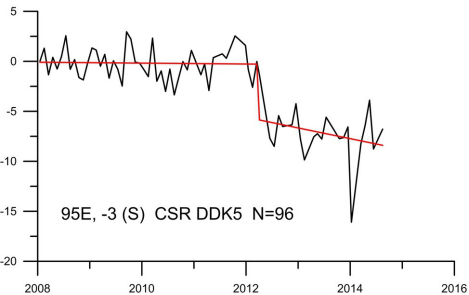
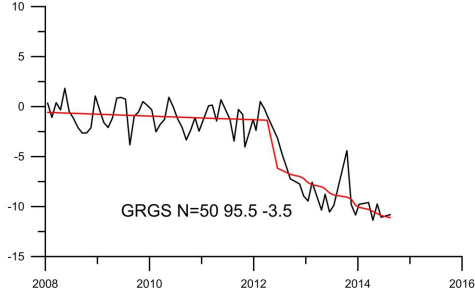
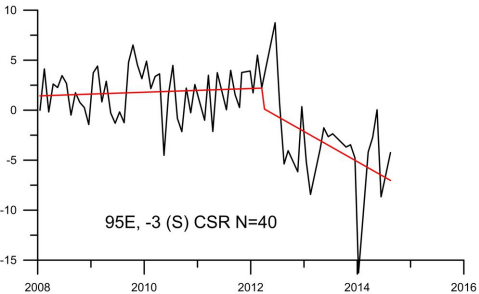
- We fitted both GPS and high resolution GRACE gravity data using the geometry of the faults system suggested by Hill et al., (2015) for the 2012 Wharton earthquakes. To invert with constraints on the rake angle, we suggest a new regularization, which allows keeping the problem linear. This method yields a rather uniform displacement field on the fault planes, without asperities.
- Our solution shows that even if the main displacements occurred on WNW trending faults, comparable displacements occurred also on a NNE trending one. Hence, deformation in this diffuse plate boundary area appears to be accommodated along both orthogonal fault-systems.
- A viscoelastic relaxation with a commonly used asthenospheric Maxwell viscosity of  $10^{19} \text{ Pa} \cdot \text{s}$  successfully explains the postseismic displacements at GPS sites and the postseismic gravity signal. Because part of the observed signal could be attributed to afterslip, the obtained viscosity value should be considered as a lower limit of the asthenospheric viscosity below the Wharton basin.



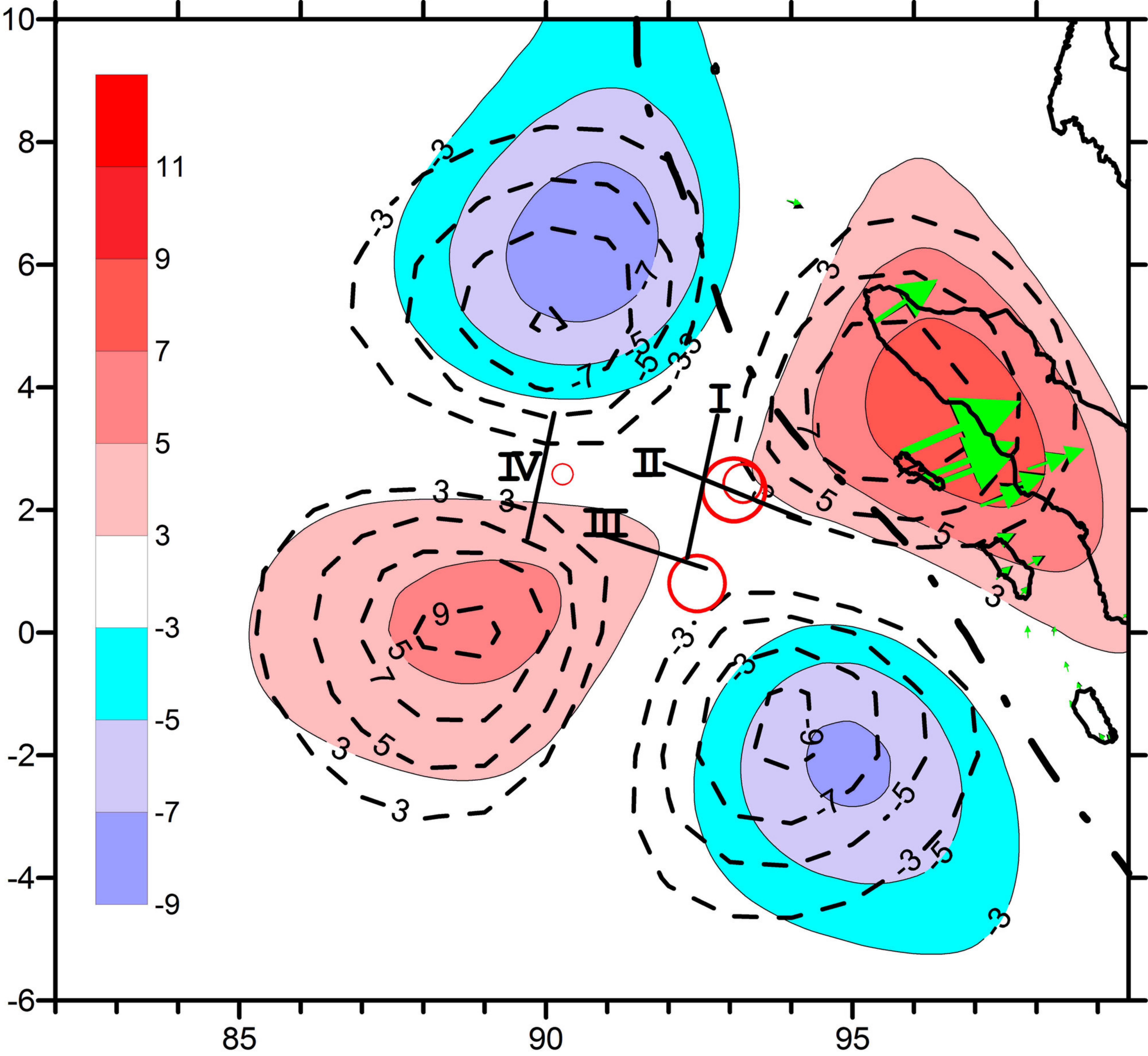




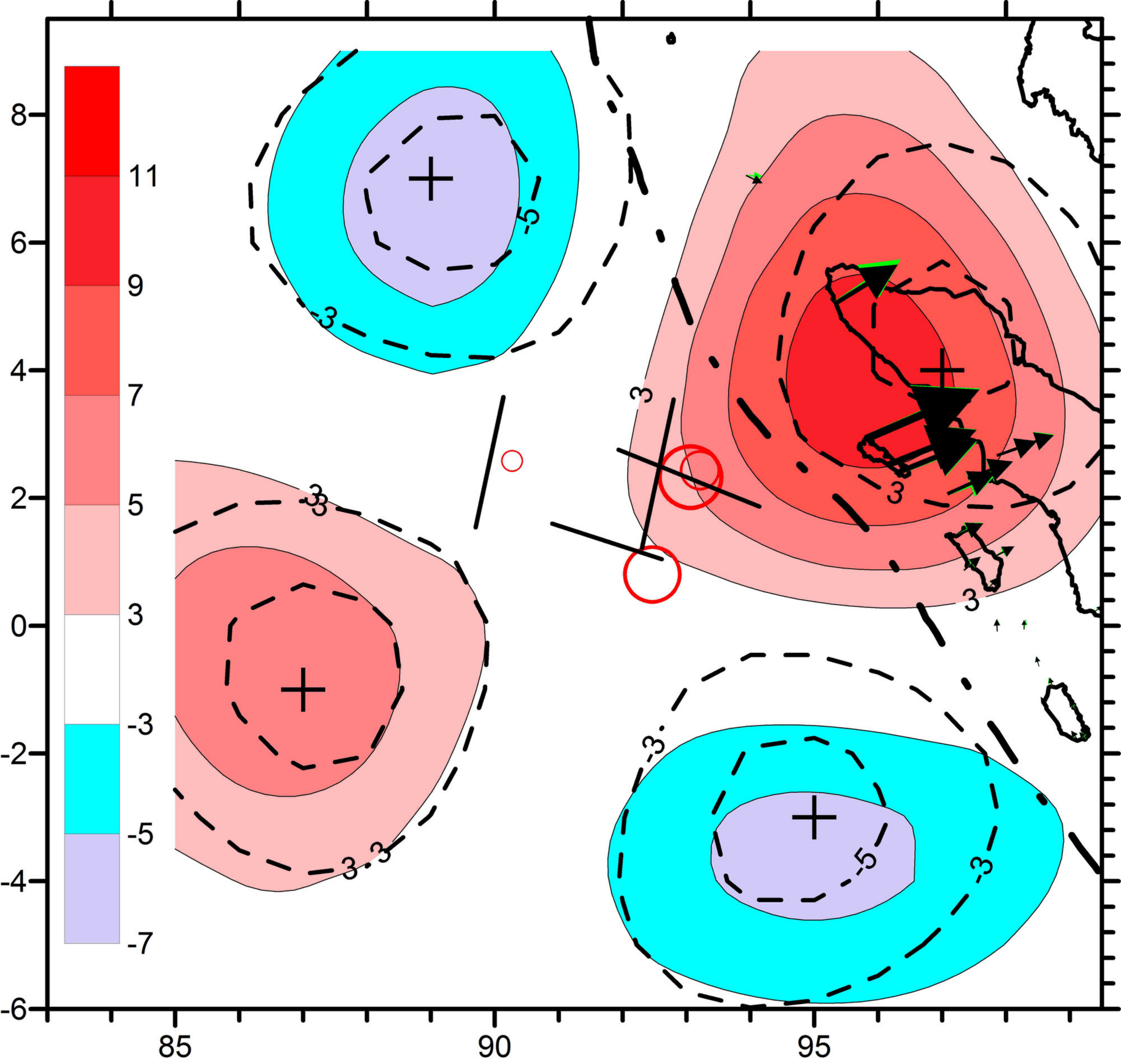




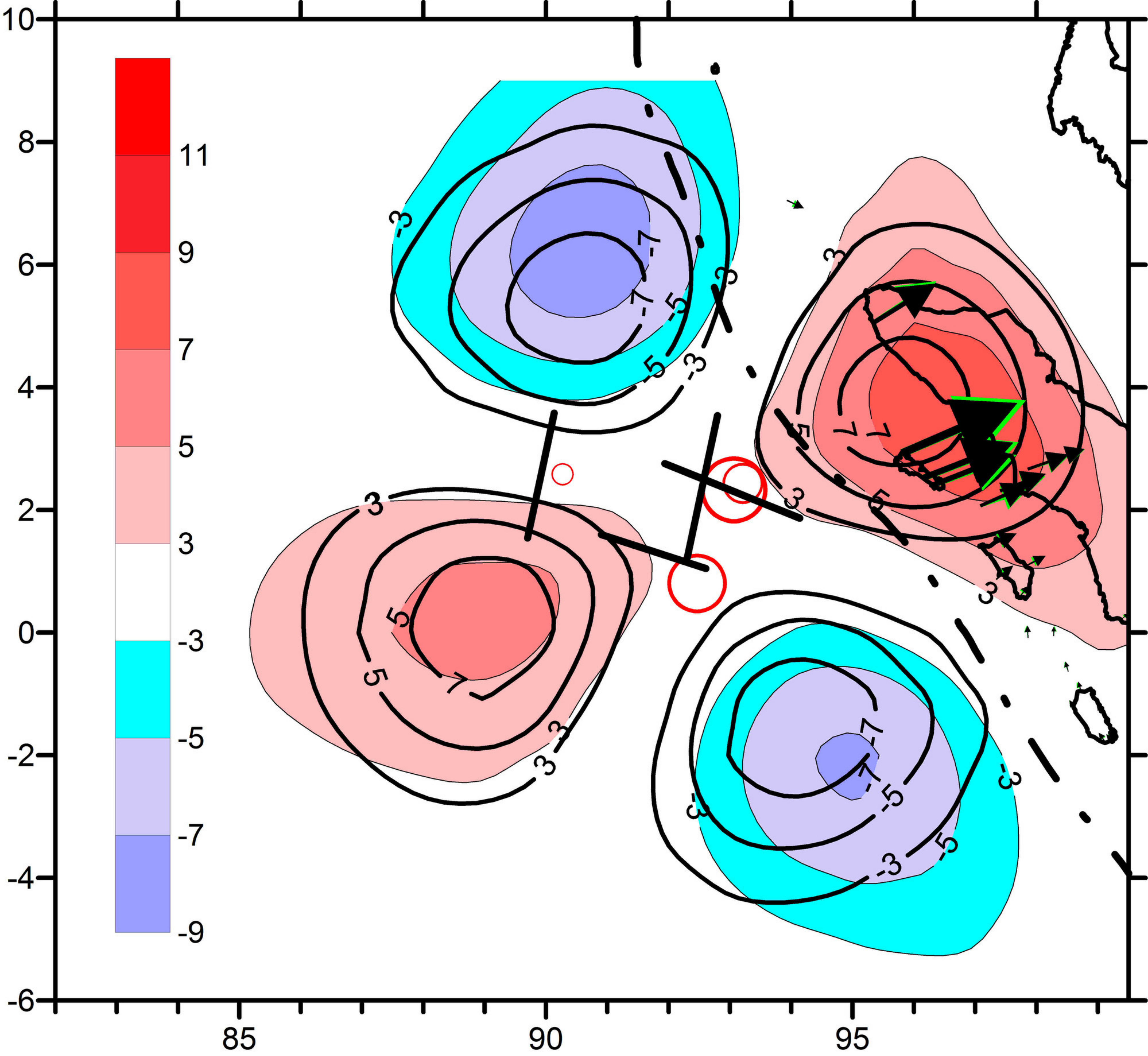




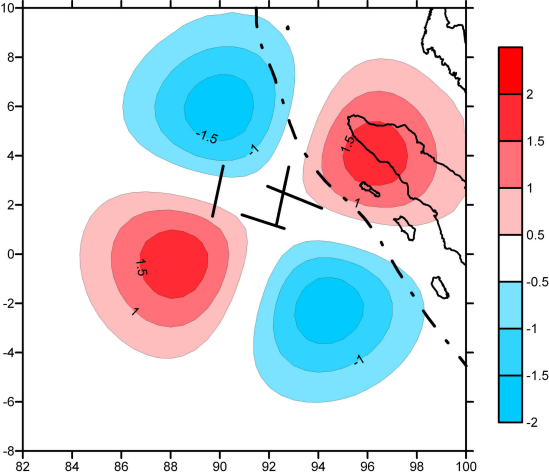




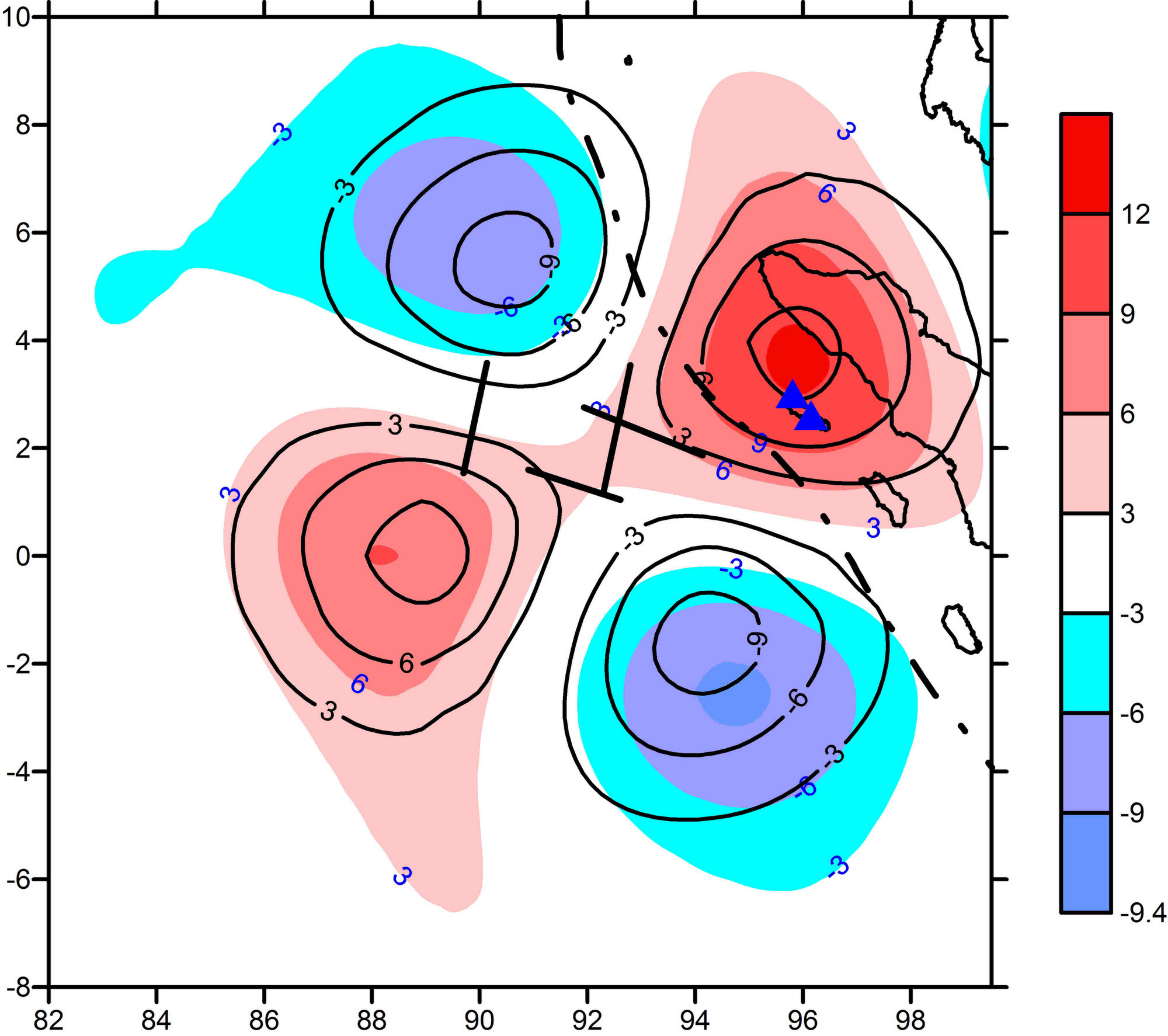














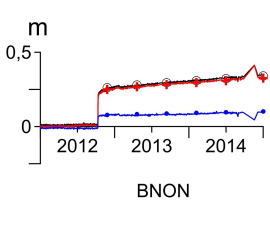
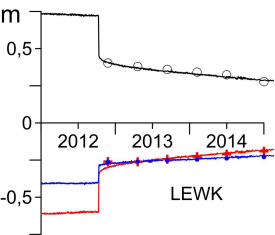


Table 1 Solutions for different number of elements in planes I-IV using GPS only (lines 1, 2) and GPS+gravity (line 3). GPS misfit is maximum value of difference between real and synthetic absolute displacements at GPS sites in %.

Number of elements	Dip slip (min, max) / Strike slip (min, max)				GPS misfit	V <sub>z</sub> misfit
	plane I	plane II	plane III	plane IV	$\Delta$ module (%)	RMS ( $\mu$ Gal)
10	(-.010;-0.10)/ (7.23-7.28)	(-0.17;-0.15)/ (4.77;4.79)	(0.04;0.04)/ (17.40-17.42)	(-0.05;-0.05)/ (3.55;3.55)	9.1	2.4
26	(-.005;-0.04)/ (7.24-7.28)	(-0.06;-0.09)/ (4.77;4.81)	(0.02;0.03)/ (17.34-17.41)	(-0.03;-0.03)/ (3.56;3.57)	9.2	1.9
26+V <sub>z</sub>	(-0.06;0.06)/ (6.93;7.04)	(-0.40;-0.23)/ (7.56;7.64)	(-0.34;-0.18)/ (12.11;12.32)	(0.27;0.37)/ (0.26;0.30)	7.0	1.5