Altiner, Y., Söhne, W., Güney, C., Perlt, J., Wang, R., Muzli, M. (2013): A Geodetic Study of the 23 October 2011 Van, Turkey earthquake. - Tectonophysics, 588, 118-134

DOI: 10.1016/j.tecto.2012.12.005

A Geodetic Study of the 23 October 2011 Van, Turkey earthquake

Yüksel Altiner[1], Wolfgang Söhne[1], Caner Güney[2], James Perlt[1], Rongjiang Wang[3], Muzli Muzli[3]

[1)] Federal Agency for Cartography and Geodesy, Frankfurt am Main, Germany
[2)] Technical University of Istanbul, Turkey
[3)] GFZ German Research Centre for Geosciences, Potsdam, Germany

E-mail: yueksel.altiner@bkg.bund.de, Phone: +49 69 6333276, Fax: +49 69 6333425
E-mail: wolfgang.soehne@bkg.bund.de, Phone: +49 69 6333263, Fax: +49 69 6333425
E-mail: guneycan@itu.edu.tr, Phone: +90 212 2853825, Fax: +90 212 2856587
E-mail: james.perlt@bkg.bund.de, Phone : +49 69 6333496, Fax : +49 69 6333425
E-mail: wang@gfz-potsdam.de, phone: +49 331 2881209, Fax: +49 331 2881204
Email: muzli@gfz-potsdam.de, Phone: +49 331 288-28637, Fax: +49 331 288-1277

ABSTRACT
The Van (Eastern Anatolia, Turkey) earthquake occurred on Sunday, October 23, 2011 with a moment magnitude of 7.2. The tectonics of this region is characterized by strike-slip faulting on the Bitlis Suture Zone, and thrusting in the Zagros fold and thrust belt. Using high-rate (1 second) GPS data from permanent GNSS stations from the CORS-TR network, co-seismic displacements of eleven stations were determined using precise point positioning during this earthquake. We used the time series of coordinate changes for fourteen CORS-TR stations, and calculated the crust movements before and after the earthquake.

According to the PPP solutions computed using high frequency GPS data to determine the co-seismic motions of stations, we conclude for the Van earthquake an occurrence time of 10:41:22 (UTC). No pre-seismic horizontal movement of stations at the level more than 5 mm before the earthquake could be observed. That means that no kinematic warning or prediction before the earthquake exists. Along an east-west horizontal line north of the Van Sea with a length of about 100 km, the northern part of this line experienced extension of 0.2-1 ppm in a NW-SE direction. The southern part experienced N-S shortening of 0.5-1.5 ppm. The N-S shortening we estimated geodetically matches well with the N-S shortening and thrust focal mechanism derived independently using seismic data by the USGS.

Co-seismic surface displacements derived from the GPS data are consistent with the teleseismic source model given by the USGS. The geodetic source model derived from the GPS data reproduces the same moment magnitude and centroid as the teleseismic model, but shows a higher spatial resolution of the slip distribution. We also analyzed the post-seismic surface displacements derived from the GPS data within the first two weeks after the mainshock. No reasonable slip distribution on the co-seismic fault plane could be found, indicating that the sources for the early post-seismic deformation might come from the widely scattered aftershocks.

Key words: Earthquake interaction, forecasting, and prediction; Precise Point Positioning; Analytical surface deformation theory; Internal and external crust deformations; Geodetic source model; Slip distribution

1. Introduction
The region where the 2011 Van earthquake occurred is nearly 20 km north of Van City center, west of Erçek Lake, and near village of the Kasımoğlu on the East Anatolian plateau.

This M 7.2 earthquake occurred on October 23, 2011 at 10:41 UTC (DoY296; local time 13:41) (http://www.koeri.boun.edu.tr; http://www.eerc.metu.edu.tr). The East Anatolian plateau is supported by thick crust (Gök et al., 2007). Tectonics in this region are characterized by convergence of the Arabian plate with the Eurasian plate at a rate of about 26 mm/yr (NUVEL-1A) directed toward NW. This motion is taken up by strike-slip faulting along the Bitlis Suture Zone and thrusting along Zagros fold and thrust belt (Şengör et al., 2003; Sandvol et al., 2003; Bird, 2003; Talebian and Jackson, 2004; Angus et al., 2006; Tan and Taymaz, 2006; Dilek, 2010). According to the Disasters and Emergency Situations Directorate of Turkey (AFAD) the coordinates of the Van earthquake epicenter was 38.68°N, 43.47°E (Koçyiğit et al., 2012). The magnitude of the earthquake estimated by different agencies varies from M_w 7.1 to M_w 7.3 with a reported depth range between 5 to 20 km. According to AFAD, this earthquake led to the death of 604 people and to the collapse of a significant number of buildings. As of Oct. 31, 2011 more than 1700 aftershocks were recorded that had a M_w of greater than 2 (http://www.cedim.de). The epicentres of the Van earthquake and aftershocks show that the seismic activity is mostly concentrated beneath Van Lake (Fig. 1). The earthquakes during this sequence that occurred along the Bitlis-Zagros Suture Zone were probably caused through loading by the Van earthquake and aftershocks. Graphs illustrated within this study were created using the software GMT (Wessel and Smith, 1991) and GNUPLOT (http://www.gnuplot.info/).

2. Geodetic study of the earthquake

A Continuously Operating Reference Station network, called CORS-TR, was established by Istanbul Kultur University in corporation with General Directorate of Land Registry and Cadastre (GDLRC) and General Command of Mapping (GCM) in Turkey and northern Cyprus between May 2006 and May 2009 (Eren et al., 2009). The CORS-TR network consists of 147 GNSS reference stations and was mainly designed to provide RTK applications and to monitor crustal movements. 15 reference stations of the CORS-TR network are located near the Van earthquake epicenter. The reference station VAN, the closest station to the earthquake epicenter, was however not active during the earthquake. Data for the other fourteen nearby stations were available from the Strong Motion Data Base of Turkey (http://kyh.deprem.gov.tr/ftpt.htm). Most of these fourteen stations have distances of greater than 100 km to the Van earthquake epicenter.

In this study, we apply several different methods of data processing (see below) to evaluate GPS data from the fourteen permanent stations (*AGRD, BASK, HAKK, HINI, HORS, IGIR, MALZ, MURA, MUUS, OZAL, SEMD, SIRN, SIRT,* and *TVAN*) within the CORS-TR network (see Fig. 12 for location of stations). The station MURA is closest to the Van earthquake epicenter, with a distance of between 39 and 43 km. This range was determined using the two different epicentral location solutions available from the GEOFON (GFZ Helmholtz Centre in Potsdam, Germany) and KOERI (Kandilli Observatory and Earthquake Research Institute, in Istanbul, Turkey) seismic stations, respectively.

3. Geodetic determination of co-seismic motion

Except at the stations AGRD, BASK, and MUUS (only data with a 30 seconds interval were available for these three stations), high-rate (1Hz / 1 second) data were available at other CORS-TR stations from the Strong Motion Data Base of Turkey with a data span of one hour from 10:00 to 11:00 UTC on Oct. 23, 2011 (DoY296). Processing of the 1Hz-data was conducted using RTNet (Real Time NETwork processing engine) software and applying the method of precise point positioning (PPP). The RTNet software was developed by GPS

Solution and was designed primarily for real-time applications, with possible use for post-processing applications (http://www.gps-solutions.com/rtnet.html). PPP is a method that allows precise point positioning with a single GNSS receiver using precise satellite orbits and clock corrections. One reaches the highest possible PPP accuracy using dual-frequency GNSS receivers, together with precise orbits and satellite clock corrections. Precise orbit and clock files are routinely available through international service providers with a latency of about two weeks. In this study, we use GPS data acquired using dual frequency receivers, and satellite orbit and clock files from the International GNSS Service (IGS). Accuracy of the IGS final product is estimated to be about ~ 5cm for orbits and ~ 0.1 ns (3 cm) for clocks, respectively (http://igscb.jpl.nasa.gov/). Phase center variations of satellite and ground antennas were also used in our data processing.

As an example of how we study the co-seismic motions for the CORS-TR stations, the difference of single epoch coordinate solutions for the stations MURA and TVAN, and MALZ and SIRN are plotted in Figs. 2 and 3 for a time period spanning the earthquake. Each point of a displacement-component plotted expresses a one-second-epoch coordinate difference relative to the reference coordinates of stations. The stations SIRT, MUUS, HINI, and HORS, which are located in northwest and southwest of the epicenter of the Van-earthquake showed no significant horizontal and vertical dislocations at a confidence level of 95% (Fig. 8). The station TVAN, situated west of the Van Lake, belongs to the group of stations that showed moderate horizontal ground motions, such as the stations HAKK, IGIR, and MALZ. The horizontal co-seismic movement of the station TVAN for the mainshock was estimated to be 3.8 ±2.5 mm toward the W (Fig. 8). The station SEMD, located south-east of the mainshock's epicenter, also showed a small horizontal co-seismic movement, with a magnitude of 3.6 ±3.1 mm in a N direction (Fig. 8). Considering all of the 1Hz single epoch solutions derived by PPP for the eleven CORS-TR stations, the Van earthquake caused the highest dislocation on station MURA, located only about 43 km northeast from the epicenter (Fig. 2). Energy, dispersed in wave form from the mainshock epicenter reached station MURA at 10:41:46 (GPS time; At the occurrence date of the Van earthquake GPS time was ahead of UTC by 15 seconds, so-called leap seconds), caused a horizontal ground movement toward the SW and SE. The finite (total) horizontal ground motion was SW with a magnitude of 38 ±2.5 mm. Shaking of this station lost its acceleration at 10:42:51 (GPS time). By comparing Figs. 2 and 3 one can see that ground movement began at station MURA earlier (10:41:46 at GPS time) than at the station TVAN (10:42:00 at GPS time). On the next day, Oct. 24, 2011 (DoY297), the station MURA continued its SW movement (21.2 ±2.2 mm) due to aftershocks and afterslip (Fig. 9). The N-S shortening we derive geodetically for these near-field stations (Figs. 8, 9, 12, and 14) matches well with the N-S shortening and thrust focal mechanism derived independently using seismic data (see the web site of the U. S. Geological Survey (USGS) at: http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc/neic_b0006bqc_cmt.php). Seismic travel times of P_g for CORS-TR stations given above were calculated using the earthquake travel time calculator by the USGS to compare our observed time difference for the beginning of co-seismic ground motion at stations MALZ, MURA, SIRN, and TVAN (Fig. 4). The calculated difference of seismic travel times is in the range of 11 seconds between the stations MURA and TVAN (Fig. 4). The observed time difference for the beginning of co-seismic motions between these stations amounts to 14 seconds (compare the Figs. 2 and 4). As opposed to that, the differences of calculated seismic travel times between the stations MALZ, SIRN, and MURA (10 seconds between MURA and MALZ, and 20 seconds between MURA and SIRN) match well to the observed time difference for the beginning of co-seismic motions between these three stations (compare Figs. 3 and 4).

According to these results of comparison between the observed time of beginning of co-seismic motions at stations and the calculated seismic travel times of P_g at stations, we conclude for the Van earthquake an occurrence time of 10:41:37 at GPS time (compare Figs. 2, 3, and 4). If we subtract the leap seconds from the GPS time (The difference between the GPS and UTC amounted to 15 seconds from 01.01.2009 to 30.06.2012), we will obtain the occurrence time of the Van earthquake at UTC (10:41:22), which matches well with the time (10:41:21.73) derived seismologically by the USGS, European-Mediterranean Seismological Centre (10:41:22), and GFZ Helmholtz Centre in Potsdam, Germany (10:41:22); for more details please see the report (Report_Van_EQ_2011.pdf) from the web page http://www.eerc.metu.edu.tr.

4. Geodetic evaluation of datum-effect

Far from plate boundaries, horizontal and vertical movement of GNSS stations due to plate loading generally amounts to less than a few mm per year. Satellite geodesy (e.g., GNSS) typically estimates such small movements over time periods of at least one year. Observation periods of several years are especially good to emerge out uncertainties induced from external influences (Altiner and Seeger, 1993; Kahle et al., 1995; McClusky et al., 2000; Altiner, 2001a; Ayhan et al., 2002; Oldow, 2002; Grafarend and Voosoghi, 2003; Kreemer and Chamot-Rooke, 2004; Babbucci et al., 2004; Battaglia et al., 2004; Altiner et al., 2006a, 2006b; Hollenstein et al., 2006; Cai & Grafarend, 2007; Hefty, 2007; Caporali et al., 2009; Weber et al., 2010; Kutoğlu et al., 2011; Özyaşar and Özlüdemir, 2011).

From a geodetic point of view, crustal deformations must be derived using free-network solutions. Free-network solutions are those that are free of datum effects. This means that the internal geometry of the points, that is the relative location of the points, is well defined and is invariant relative to rigid body motions of the networks, such as shifting and rotation. In such a case, i.e. using a free-network solution, the movement of stations can be estimated relative to a single datum station. Further, residual effects, which might caused by rotation of the network, i.e. a rigid body motion, can be eliminated by applying analytical surface deformation theory (Altiner, 1999, 2001b). To derive the absolute station displacements for a 3-D network, as opposed to simply relative movements, the datum-defect due to the rank deficiency (singularity) of the normal equations needs to be removed through predefining at least seven coordinate components of a set of datum stations. In our case, we have rather weighted all nine coordinate components of three selected datum stations with an a priori uncertainty of 0.1 mm to define the net-datum we use for estimating station dislocations.

Processing of GPS data was done on a daily basis with data interval of 30 seconds using Bernese GNSS software (BSW, v. 5.0), developed at the University of Bern (Dach et al., 2007). We also used precise ephemeris, the so-called “final orbits” from the IGS. The coordinates of stations were then estimated in the IGS08 reference frame using the standard methods of the BSW, including double differencing of phase measurements for parameter estimation, using phase center variations of satellite and ground antennas, and accounting for the effect of ocean loading. Tropospheric horizontal gradients were also considered in the data processing to increase accuracy in the estimation of the vertical components of coordinates. In addition, an ionosphere-free solution was applied to eliminate a large part of the ionospheric affects. The accuracy of a priori coordinates of CORS-TR stations were improved by including data from several IGS stations (*ANKR, DRAG, EVPA, ISTA, KTVL, NICO, RAMO, TUBI*, and *ZECK*) near the study area in our analysis (Fig. 5). To check whether the datum-effect on coordinate estimation could be caused by the coordinate accuracy or by the geometry of the selected datum stations, daily coordinates of fourteen CORS-TR stations were first estimated relative to the IGS station ZECK, located north of the network (free-network solution / relative station displacements). As a second step, coordinates of stations in the CORS-TR network were re-computed for comparison on a daily basis relative to the datum was realised through predefining all nine coordinate components of IGS stations ZECK, ANKR, and ISTA at the observed IGS08 epoch (absolute station displacements). All three datum stations within the IGS network are located north of the Van network, and were also used to define the far-field northward movement of the Arabian plate.

Because data from the station BASK were not available for some of the days considered in this study to compute the change of internal and external network geometry, the horizontal and vertical dislocations of stations were determined in IGS08 by forming coordinate differences between DoY294 (Oct. 21, 2011) and DoY307 (Nov. 03, 2011). To check the datum-effect, displacement components of the solution relative to the datum station ZECK and the additional solution relative to the datum stations ZECK, ANKR, and ISTA were then compared on daily basis with one other (Figs. 6 and 7). Coordinate differences between these two solutions are systematic (similar coordinate difference for each station of the network; net shifting) and amount to -0.5 mm for the north, -1.2 mm for the east component, and 2.3 mm for the vertical component. Considering these small differences, we will only discuss the results derived relative to the datum stations ZECK, ANKR, and ISTA (see Fig. 5). To illustrate the effects the mainshock and early aftershocks, horizontal coordinate differences between the solutions of DoY295 (Oct. 22, 2011) and DoY296 (Oct. 23, 2011), as well as between the solutions DoY296 (Oct. 23, 2011) and DoY297 (Oct. 24, 2011) are illustrated in Fig. 8 and 9, respectively. Considering our uncertainty of ±10 mm for vertical movement of stations, no significant height change of stations was observed within the network, except the station OZAL. At this station a height change of 13 mm (uncertainty ±6 mm) was computed between the solutions DoY295 and DoY296 (Fig. 18).

5. Time series determination of pre-seismic and post-seismic motions

To consider the question of whether any pre-seismic warning could be derived from the observed ground motions before the Van earthquake, the time series of coordinate differences of the fourteen stations within the CORS-TR network were determined for a time period from DoY289 (Oct. 16, 2011) to DoY307 (Nov. 03, 2011). Because of their large uncertainties, the estimated coordinates of the stations BASK, HAKK, and SEMD were removed from the daily solution for DoY292. Additionally, data for station BASK were not available on DoY295 and on DoY296. The time series of horizontal coordinate differences for the eight remaining CORS-TR stations from DoY289 to DoY307 are illustrated in Fig. 10. Fig. 10 shows no horizontal movement of stations at the level more than 5 mm before the earthquake between DoY289 and DoY295. That means that no kinematic warning or prediction exists. Eastward movements of up to 5 mm were detected on DoY291 for some stations in the south and southeast of the network, e. g. SIRT, HAKK, and OZAL, but 5 mm is our the approximate bound of accuracy of coordinate estimation from the GPS data within this study. According to the time series of coordinate differences, after the earthquake on DoY296 the stations located in the north of the Van Sea moved in a S, SW or SE direction, whereas those situated south of the Van Sea moved toward the N, NW or NE. If we assume that the Van earthquake or other external effects caused no tectonic movements for the far-field IGS stations included into our data processing from DoY289 to DoY307, then the time series of coordinate differences for IGS stations (*DRAG, EVPA, KTVL, NICO, RAMO, and TUBI*) suggest an accuracy of about 4 mm in the horizontal and 10 mm in the vertical coordinate estimate (Fig. 5). This inference is consistent with the 95% error ellipses and error bars we determined for horizontal and vertical movements of the stations shown in Fig. 11, and supports the idea that a large part of the apparent vertical motion of the stations observed from DoY300 to DoY302 was really a residual tropospheric effect.

6. Determination of change of internal and external network geometry

To determine the change of internal and external network geometry, ground deformation within the study area was derived from the horizontal and vertical displacements of stations as determined by coordinate difference between the solutions for DoY294 and DoY307 relative to the datum stations ZECK, ANKR, and ISTA. On DoY294 and DoY307 estimated coordinates for all 14 CORS-TR stations within the study area were available. Uncertainties in coordinate differences were determined using standard error propagation and scaling with a factor of 4. The observed horizontal and vertical displacements and 95% uncertainties are illustrated in Figs. 12 and 13. Stations (horizontal movement of stations and their uncertainties are given in the brackets) IGIR (13.9 ±2.1 mm), MALZ (17.0 ±2.2), AGRD (28.2 ±2.3 mm), and MURA (62.6 ±2.5 mm), located north and north-east of the Van Sea, were the stations most affected and moved to S, SE and SW, respectively (Fig. 12). HORS (8.2 ±2.2 mm) and HINI (3.8 ±3.2 mm), located north-west of the Van Sea, also showed similar motions to the SE (Fig. 12). TVAN (6.9 ±2.5 mm) moved to the W. The stations situated south and south-east of the Van Sea, OZAL (12.2 ±2.5 mm), BASK (28.7 ±3.2 mm), SIRN (11.6 ±3.9 mm), HAKK (16.8 ±3.2), and SEMD (4.6 ±3.2 mm) moved in a N to NW direction. The stations MUUS (1.4 ±3.0 mm) and SIRT (3.0 ±7.7 mm) experienced no significant horizontal and vertical movement (Figs. 12 and 13).

7. What happened regarding internal network geometry?

As a check on our derivation and scaling of internal and external deformation measures as well as for an area-wise study of ground deformation, the horizontal and vertical velocities of the stations were next interpolated using the spline method (Bronstein et al., 1995; Dermanis, 2009). In this analysis, we used ellipsoidal coordinates for a regular area-wide grid spanning 37.3° to 40.2° latitude and 41.5° to 44.2° longitude, and a mesh spacing of 0.1°. Station positions were defined according to the coordinate system given in Heitz (1988). The internal (largest and smallest principal strain rates) and external (change of main curvatures and change of principal curvatures) deformation measures were next evaluated using analytical surface deformation theory (Altiner, 1999, 2001b) and coordinate differences between DoY294 and DoY307 reported above. Results from the analytical surface deformation analysis the internal and external deformation measures are expressed by the station displacements and uncertainties shown in Figs. 12 and 13. In this study the internal deformation measures illustrated in Fig. 14 are all statistically significant, except the in area between the stations SIRT, MUUS, and HINI in the west of the network.

Drawing an east-west horizontal line north of Van Sea with a length of about 100 km, the northern part of this line experienced extension of magnitude 0.2-1 ppm in a NW-SE direction. The southern part experienced 0.5-1.5 ppm of N-S shortening (Fig. 14). As a comparison to the geodetic results obtained and presented here, co-seismic deformation was also qualitatively mapped using the European Macroseismic Scale and is shown as an overlay in Fig. 14. The intensity VIII zone covered a distance of 20-30 km from the earthquake epicenter, in a zone where no GPS data were available. A N-S shortening of 0.5-1.5 ppm dominates in the eastern part of the study area, between the GNSS stations MURA and OZAL. The north-eastern part of the network was subject to an extension with magnitude of 0.2-1 ppm in a NW-SE direction. West and south of GNSS stations MALA, TVAN, SIRT, and SIRN small crust deformations may have occurred, but non-zero strains determined near the map boundary could also simply be artifacts of the extrapolation.

8. How did external network geometry change?

To illustrate the change of the form in the external geometry within the study area during and after the Van earthquake between DoY294 and DoY307, the changes of the main and principal curvatures in the stations were determined (Altiner, 1999, 2001b). The changes of the mean and principal curvatures, a function of the Earth's radius, describe using normal vector to the surface at station (surface normal; perpendicular to the tangent plane to that surface at point P), in three-dimensional Euclidian space, how after an event, the external geometry within the surface area-wise changed, and enable a view about the slope direction of external changes. This differs from the point-wise demonstration of vertical changes of stations (height changes) that are perpendicular to a reference surface, and that we as previously shown in Fig. 13. To calculate out the direction of slope of external geometry, the main axes of the changes of the principal curvatures are needed. These external geometry deformation measures, including the change of mean curvature and the changes of principal curvatures have parallel meanings to the dilatation and elongation used to characterise changes in internal geometry. To better visualize the observed change of the external geometry, the area was rotated 5 degrees about the X axis and 30 degrees about the Z axis according of the global Cartesian coordinate system (Fig. 15). The relative changes obtained were then converted to the metric dimension of cm. The Van earthquake and aftershocks that occurred up to DoY307, caused an external decrease of 2-10 mm in an area stretching from SIRN in the south to MALZ in the north. The slope of the external geometry of this area was mainly toward the NW. The north-eastern and eastern parts of the Van Sea were raised up from 0.5 to 5 mm. A NE slope of the external geometry around the station AGRI in the north-eastern part of the network observed, whereas the slope of the eastern part between the stations MURA and BASK was toward the SE (Fig. 16). We would also notice here that using the analytical surface deformation theory, the accuracy of these external deformation measures are expressed by the uncertainties of the vertical station displacements illustrated in Fig. 13.

9. Modeling of co- and post-seismic surface deformation

In this section, we firstly investigate the consistency of the co-seismic surface deformation that we measured at the CORS-TR stations with the earthquake source model derived independently from the seismic observations. Based on the dislocation theory, we can simulate the co-seismic surface deformation using a given finite fault model inverted from the teleseismic broadband waveform data. In the present case, we adopt the final version of the teleseismic fault slip model provided by the USGS (see: http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc/finite_fault.php, last access September 2012). For the simulation, we use the software PSGRN/PSCMP (Wang et al., 2006) that can incorporate with the same crustal structure interpolated from CRUST2.0 (Bassin et al., 2000) as used for the teleseismic modeling. The Fig. 17 shows a comparison between the observed surface displacements from DoY295 to DoY296 and the simulated co-seismic values. Though the misfits at stations OZAL, MURA and TVAN are significant, the overall magnitude and pattern of the displacement field are in agreement particularly for the horizontal components of displacements. The relatively poor agreement for the vertical component is expected because of the known larger uncertainty in the data. Considering our uncertainty of ±10 mm, we detected a statistical significant height change of 1.3 mm at station

OZAL; measured height changes for all other stations were not significant. The results of the simulated vertical displacements, shown in Fig. 17, also supported this fact.

As can be seen from Fig. 18, the GPS network provides a relatively good coverage of the earthquake near-field area, implying that the observed surface displacement data can provide useful constraints on the earthquake source. Therefore, we test to derive the geodetic fault slip model from the observed co-seismic static displacement data and compare it with the teleseismic one. For this purpose, we use the inversion code SDM written by one of the co-authors (R. Wang) based on the constrained least-squares method, which has been used in a number of recent publications for analyzing GPS, InSAR and strong-motion based co- and post-seismic deformation data (e.g., Wang et al., 2004, 2009, 2011, 2012; Motagh et al., 2008, 2010; Diao et al., 2010; Xu et al., 2010; Zhang et al., 2011). To overcome the problem of non-uniqueness and instability of the inversion result, a smoothing constraint is applied to the slip distribution. An optimal smoothing factor is determined by analyzing the trade-off curve between the data misfit and the slip roughness (Segall and Harris, 1987). Using the same fault plane location and geometry, the geodetically inverted fault slip model is shown in Fig. 19 in comparison with the teleseismic results. Both slip models yield about the same moment magnitude (Mw 7.1), but two different slip patterns. In particular, the major slip asperity in the teleseismic model is located in depth between 20 and 30 km. In the geodetic model, however, it is clearly located near the surface, implying that the fault slip reaches the surface, which is verified by the field observation by the National Seismological Observation Network operated by Prime Ministry Disaster and Emergency Management Presidency (AFAD). The correlation between the data and the geodetic model is as good as 97%.

As shown in previously sections, the surface displacement was observed also within two weeks after the earthquake, which is particularly significant at stations AGRD, MURA and MALZ. Although there may be different source mechanisms for the post-seismic deformation including aftershocks, slow slip (also called afterslip) on the mainshock fault, poroelastic rebound, viscoelastic relaxation of the co-seismically induced stress changes, and so on. However, the most possible sources for the early post-seismic deformation within a few weeks after the earthquake should be dominated by the aftershocks and afterslip (Wang et al., 2009). Thus, we attempted to invert the afterslip sources on the mainshock fault plane from the post-seismic GPS data. No reasonable results could be obtained, i.e., no clear relationship is found between the afterslip and the co-seismic slip. Therefore, we interpret the source of the observed post-seismic deformation to be the large aftershocks scattered around the mainshock fault.

10. Conclusions

The principal geodetic results presented here, derived by PPP using GPS data from the CORS-TR network, are illustrated in Figs. 2 and 3 for a period of ~7 minutes during the Van earthquake. Energy, dispersed from the epicenter in wave form reached the station MURA (~ 43 km to the epicenter) at 10:41:31 UTC (Fig. 2). Our PPP solutions using high frequency GPS data (1 Hz) support an occurrence time of 10:41:22 UTC for the Van earthquake (compare Figs, 2, 3, and 4).

Within an uncertainty of ±5 mm, no pre-seismic horizontal movement of stations, which could potentially serve as a warning or prediction, was observed (Fig. 10). During the mainshock, stations located in north of the Van Sea moved systematically to the S, SW, and SE, and stations south of the Van Sea moved systematically N, NW, and NE (Fig. 8). Due to large aftershocks on DoY296 and on DoY297, stations AGRD and MURA, located north and south of the epicenter, continued moving to the SE and SW with the magnitudes of 11.4 ±2.3 mm and 21.2 ±2.2 mm relative to the datum stations ZECK, ANKR, and ISTA (Fig. 9).

The change of the internal and external network geometry determined between DoY294 and DoY307, indicate that the northeastern part of the network experienced an extension of 0.2-1 ppm, directed to the NW-SE (Fig. 14). Similarly, the southern part of the network experienced shortening of 0.5-1.5 ppm mainly in a N-S direction. The Van earthquake and aftershocks that occurred through to November 03, 2011, also caused a decrease in the external geometry of 2-10 mm in the southwestern part of the network. The slope direction of the external network geometry in this area was mainly toward the NW (Fig. 16). The eastern and northeastern parts of Van Sea in an external reference frame inclined from 0.5 to 5 mm with a slope direction toward the SE and NE, respectively.

Co-seismic surface deformation determined using the GPS data is consistent with the earthquake source model derived from teleseismic observations, though the misfits at a few near-field stations are significant. The geodetic source model is inverted from the observed co-seismic displacement data, which shows the same moment magnitude (Mw 7.1) as the teleseismic source model, but a higher spatial resolution of the fault slip distribution. The observed early post-seismic deformation is interpreted to be caused by the large aftershocks scattered around the mainshock fault.

Our study has shown that geodetic results, such as PPP, could provide an important contribution for derivation of occurrence time of an earthquake using the beginning time of co-seismic motions at network stations. Applying analytical surface deformation theory, the change of external geometry, also change of slope direction, can be determined more realistically.

Acknowledgements

The authors thank Laurent Jolivet (associate editor), John Weber from the Grand Valley State University, and an anonymous reviewer for constructive and useful comments and for revising the text. We thank Gavin Hayes from the USGS for his finite fault model of the Oct 23, 2011 Mw 7.1 Eastern Turkey Earthquake published at web page of the USGS (http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc/finite_fault.php).

References

Altiner, Y., Seeger, H., 1993. Is the motion of the Eastern Mediterranean region faster than expected? Geological Journal 28(3-4), pp. 319–325, doi: 10.1002/gj.3350280310.

Altiner, Y., 1999. Analytical surface deformation theory for detection of the Earth’s crust movements. Springer-Verlag, Berlin Heidelberg New York.

Altiner, Y., 2001a. The contribution of GPS data to the detection of the Earth’s crust deformations illustrated by GPS campaigns in the Adria region. Geophys. J. Int. 145, 550–559, doi: 10.1046/j.0956-540x.2001.01411.x.

Altiner, Y., 2001b. Analytical surface deformation theory for detection of the Earth’s crust movements (in Chinese translated by Gao rongsheng and Li Zhengyuan from the English edition). Seismological Press, Beijing.

Altiner, Y., Marjanovic, M., Rasic, L., Medved, M., 2006a. Active Deformation of the Northern Adriatic Region: Results from the CRODYN Geodynamical Experiment. In: Pinter, N., et al. (Ed.), The Adria microplate: GPS Geodesy, Tectonics and Hazards. Springer-Verlag, pp. 257–268.

Altiner, Y., Bačić, Ž., Bašić, T., Coticchia, A., Medved, M., Mulić, M., Nurçe, B., 2006b. Present-day tectonics in and around the Adria plate inferred from GPS measurements. In: Dilek, Y., Pavlides, S. (Ed.), Postcollisional tectonics and magmatism in the Mediterranean region and Asia. Geological Society of America Special Paper 409, pp. 43–55, doi: 10.1130/2006.2409(03).

Angus, D.A., Wilson, D.C., Sandvol, E., Ni, J.F., 2006. Lithospheric structure of the Arabian and Eurasian collision zone in eastern Turkey from S-wave receiver functions. Geophys. J. Int. 166, pp. 1335–1346.

Ayhan, M. E., Demir, C., Lenk, O., Kilicoglu, A., Altiner, Y., Barka, A. A., Ergintav, S., Özener, H., 2002. Interseismic Strain Accumulation in the Marmara Sea Region. Bulletin of Seismological Society of America 92, pp. 216–229.

Babbucci, D., Tamburelli, C., Viti, M., Mantovani, E., Albarello, D., D'Onza, F., Cenni, N., Mugnaioli, E., 2004. Relative motion of the Adriatic with respect to the confining plates: Seismological and geodetic constraints. Geophys. J. Int. 159, pp. 765–775, doi: 10.1111/j.1365-246X.2004.02403.x.

Bassin, C., Laske, G., Masters, G., 2000. The Current Limits of Resolution for Surface Wave Tomography in North America. EOS Trans AGU 81, F897.

Battaglia, M., Murray, M.H., Serpelloni, E., Burgmann, R., 2004. The Adriatic region: An independent microplate within the Africa-Eurasia collision zone. Geophys. Res. Lett. 31, pp. L09605, doi: 10.1029/ 2004GL019723.

Bird, P., 2003. An updated digital model of plate boundaries. Geochemistry Geophysics Geosystems by AGU and Geochemical Society 4, pp. 1–52, doi :10.1029/2001GC000252.

Bronstein, I.N., Semendjajew, K.A., Musiol, G., Mühlin, H., 1995. Taschenbuch der Mathematik. Harri Deutsch Verlag, Frankfurt am Main.

Cai, J., Grafarend, E.W., 2007. Statistical analysis of geodetic deformation (strain rate) derived from the space geodetic measurements of BIFROST Project in Fennoscandia. J. of Geodyn. 43(2), pp. 214–238, doi : 10.1016/j.jog.2006.09.010.

Caporali, A., Aichhorn, C., Barlik, M., Becker, M., Fejes, I., Gerhatova, L., Ghitau, D., Grenerczy, G., Hefty, J., Krauss, S., Medak, D., Milev, G., Mojzes, M., Mulic, M., Nardo, A., Pesesc, P., Rus, T., Simek, J., Sledzinski, J., Solaric, M., Stangl., G., Stopar, B., Vespe, F., Virag, G., 2009. Surface kinematics in the Alpine-Carpathian-Dinaric and Balkan region inferred from a new multi-network GPS combination solution. Tectonophysics 474, pp. 295–312, doi: 10.1016/j.tecto.2009.04.035.

Dach, R., Hugentobler, U., Fridez, P., Meindl, M., 2007. Bernese GPS software version 5.0. Astronomical Institute of the University of Bern.

Dermanis, A., 2009. The evolution of geodetic methods for the determination of strain parameters for earth crust deformations. In: Arabelos, D., et al. (Ed.), Terrestrial and Stellar Environment: Volume in honor of Prof. G. Asteriadis, pp. 107–144, Publication of the School of Rural and Surveying Engineering, Aristotle University of Thessaloniki.

Diao, F., X. Xiong, R. Wang, Zheng, Y., Hsu, H., (2010). Slip model of the 2008 Mw 7.9 Wenchuan (China) earthquake derived from the co-seismic GPS data. Earth, Planets and Space 62, pp. 869–874.

Dilek, Y., 2010. Eastern Mediterranean geodynamics. International Geology Review 52, pp. 111–116, doi: 10.1080/00206810902951031.

Eren, K., Gülal, E., Yildirim, O., Cingöz A., 2009. Results from a Comprehensive GNSS Test in the CORS-TR Network: Case Study. Journal of Surveying Engineering (ASCE) 135 (1), pp. 10–18.

Gök, R., Pasyanos, M.E., Zor, E., 2007. Lithospheric structure of the continent-continent collision zone: eastern Turkey. Geophys. J. Int. 169, pp. 1079-1088, doi: 10.1111/j.1365-246X.2006.03288.x.

Grafarend, E. W., Voosoghi, B., 2003. Intrinsic deformation analysis of the Earth's surface based on displacement fields derived from space geodetic measurements. Case studies: present-day deformation patterns of Europe and of the Mediterranean area (ITRF data sets). Journal of Geodesy 77, pp. 303–326, doi 10.1007/s00190-003-0329-2.

Heitz, S., 1988. Coordinates in Geodesy. Springer-Verlag, Berlin Heidelberg New York.

Hefty, J., 2007. Geo-kinematics of central and south-east Europe resulting from combination of various regional GPS velocity fields. Acta Geodyn. Geomater 4(148), pp. 169–185.

Hollenstein, C., Kahle, H.-G., Geiger, A., 2006. Plate tectonic framework and GPS-derived strain-rate field within the boundary zones of the Eurasian and African plates. In: Pinter, N., et al. (Ed.), The Adria microplate: GPS Geodesy, Tectonics and Hazards. Springer-Verlag, pp. 35–50.

Kahle, H.-G., Müller, M.V., Geiger, A., Danuser, G., Mueller, S., Veis, G., Billiris, H., Paradisis, D., 1995. The strain field in northwestern Greece and the Ionian Islands: results inferred from GPS measurements. Tectonophysics 249, pp. 41–52.

Koçyiçit, A., Özer, M. F., Lenk, O., Çolakoğlu, Z., Çelebi, M., Holzer, T., Sharer, K., Havskov, J., 2012. Report_on_October_23_2011_Van_Earthquake_ Mw 7.0.pdf. http://www.deprem.gov.tr (last access September 2012).

Kreemer, C., Chamot-Rooke, N., 2004. Contemporary kinematics of the southern Aegean and the Mediterranean Ridge. Geophys. J. Int. 157 (3), pp. 1377–1392.

Kutoglu, H.S., Celik, R.N., Ozludemir, M.T., Guney, C., 2011. New findings on the effects of the ˙Izmit Mw=7.4 and Duzce Mw=7.2 earthquakes. Natural Hazards and Earth System Sciences 11, pp. 267–272.

McClusky, S., Balassanian, S., Barka, A., Demir, C., Ergintav, S., Georgiev, I., Gurkan, O., Hamburger, M., Hurst, K., Kahle, H., Kastens, K., Kekelidze, G., King, R., Kotzev, V., Lenk, O., Mahmoud, S., Mishin, A., Nadariya, M., Ouzounis, A., Paradissis, D., Peter, Y., Prilepin, M., Reilinger, R., Sanli, I., Seeger, H., Tealeb, A., Toksöz, M.N., Veis, G., 2000. Global Positioning System constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus. Geophys. Res. Lett. 105, pp. 5695–5719, doi: 10.1029/1999JB900351.

Motagh, M., Wang, R., Walter, T.R., Bürgmann, R., Fielding, E., Anderssohn, J., Zschau, J., 2008. Coseismic slip model of the August 2007 Pisco earthquake (Peru) as constrained by Wide Swath radar observations. Geophys. J. Int., doi: 10.1111/j.1365-246X.2008.03852.x.

Motagh, M., Schurr, B., Anderssohn, J., Cailleau, B., Walter, T.R., Wang, R., illotte, J.-P., 2010. Subduction earthquake deformation associated with 14 November 2007, Mw 7.8 Tocopilla earthquake in Chile; Results from InSAR and aftershocks. Tectonophysics 490, pp. 60–68, doi:10.1016/j.tecto.2010.04.033.

Oldow, J.S., Ferranti, L., Lewis, D.S., Campbell, J.K., D'Argenio, B., Catalano, R., Rappone, G., Carmignani, L., Conti, P., Aiken, C.L.V., 2002. Active fragmentation of Adria, the North African promontory, central Mediterranean orogen. Geology 30, pp. 779–782, doi: 10.1130/0091-7613.

Özyasar, M., Özlüdemir, M. T., 2011. The contribution of engineering surveys by means of GPS to the determination of crustal movements in Istanbul. Natural Hazards and Earth System Sciences 11, pp.1705–1713, doi: 10.5194/nhess-11-1705-2011.

Sandvol, E., Türkelli, N., Zor, E., Gök, R., Bekler, T., Gürbüz, C., Seber, D., Barazangi, M., 2003. Shear wave splitting in a young continent-continent collision: an example from eastern Turkey. Geophys. Res. Lett. 30(24), pp. 8041, doi: 10.1029/2003GL018912.

Segall, P., Harris, R., 1987. Earthquake deformation cycle on the San Andreas fault near Parkfield, California. J. Geophys. Res. 92, pp. 10,511-10,525.

Şengör, A. M. C., Özeren, S., Genç, T., Zor, E., 2003. East Anatolian high plateau as a mantle-supported, north-south shortened domal structure. Geophys. Res. Lett. 30(24), 8045, doi:10.1029/2003GL017858.

Talebian, M., Jackson J., 2004. A Reappraisal of Earthquake Focal Mechanisms and Active shortening in the Zagros Mountains of Iran. Geophys. J. Int. 156 (3), pp. 506–526, doi:10.1111/j.1365-246X.2004.02092.x.

Tan, O., Taymaz, T., 2006, Active tectonics of the Caucasus: Earthquake source mechanisms and rupture histories obtained from inversion of teleseismic body waveforms. In: Dilek, Y., Pavlides, S. (Ed.), Postcollisional tectonics and magmatism in the Mediterranean region and Asia. Geological Society of America Special Paper 409, pp. 531–578, doi: 10.1130/2006.2409(03).

Wang, R., Xia, Y., Grosser, H., Wetzel, H.-U., Kaufmann, H., Zschau, J., 2004. The 2003 Bam (SE Iran) earthquake: precise source parameters from satellite radar interferometry. Geophys. J. Int. 159, pp. 917–922.

Wang, R., Lorenzo-Martín, F., Roth, F., 2006. PSGRN/PSCMP–a new code for calculating co- and post-seismic deformation, geoid and gravity changes based on the viscoelastic-gravitational dislocation theory. Computer & Geosciences 32, pp. 527–541.

Wang, L., Wang, R., Roth, F., Enescu, B., Hainzl, S., Ergintav, S., 2009. Afterslip and viscoelastic relaxation following the 1999 M 7.4 Izmit earthquake from GPS measurements. Geophys. J. Int. 178(3), pp. 1220–1237.

Wang, R., Schurr, B., Milkereit, C., Shao, Zh., Jin, M., 2011. An improved automatic scheme for empirical baseline correction of digital strong-motion records. Bulletin of the Seismological Society of America 101(5), pp. 2029–2044, doi: 10.1785/0120110039.

Wang, R., Parolai, S., Ge, M., Ji, M., Walter, T. R., Zschau, J., 2012. The 2011 Mw 9.0 Tohoku-Oki Earthquake: Comparison of GPS and Strong-Motion Data. (accepted by BSSA).

Weber, J., Vrabec, M., Pavlovčić-Preseren, P., Dixon, T., Jiang, Y., Stobar, B., 2010, GPS-derived motion of the Adriatic microplate from Istria Peninsula and Po Plain sites, and geodynamic implications. Tectonophysics 483 (3-4), pp. 214–222.

Wessel, P., Smith., W.H.F., 1991. Free software helps map and display data. EOS Trans. AGU 72, pp. 441–446.

Xu, C., Liu, Y., Wen, Y., Wang, R., 2010. Coseismic Slip Distribution of the 2008 Mw 7.9 Wenchuan Earthquake from Joint Inversion of GPS and InSAR Data. Bulletin of the Seismological Society of America 100(5B), pp. 2736–2749, doi: 10.1785/0120090253.

Zhang, G. Ch. Qu, X. Shan, X. Song, G. Zhang, Ch. Wang, J.-Ch. Hu & Wang, R., 2011. Slip distribution of the 2008 Wenchuan Ms 7.9 earthquake by joint inversion from GPS and InSAR measurements: a resolution test study. Geophys. J. Int. 186, pp. 207–220, doi: 10.1111/j.1365-246X.2011.05039.x.

FIGURES:

Fig. 1: (A) Location of the epicenters of the Van earthquake and aftershocks as of Nov. 24, 2011. The circles in red show earthquakes along the Bitlis-Zagros Suture Zone. (B) Time and magnitude of earthquakes illustrated in Fig. 1A. The lines in red correspond to the earthquakes that occurred along the Bitlis-Zagros Suture Zone.

Fig. 2: Co-seismic displacements for station MURA, our closest station to the earthquake epicenter (~ 43 km), and TVAN. Each calculated displacement corresponds to a one second data interval. Energy, dispersed in wave from the epicenter, reached the station MURA at 10:41:46 (GPS time). The mainshock effect caused a SW horizontal ground motion with a total magnitude of 38±2.5 mm. The station TVAN began at 10:42:00 (GPS time) to move and reached a magnitude of 3.8±2.5 mm in W direction. B–Beginning of dislocation; M–End of maximum displacement; E–End of acceleration of the motion.

Fig. 3: Differences in one-second-epoch coordinate solutions for stations MALZ and SIRN. The differences of calculated seismic travel times between the stations MALZ, SIRN, and MURA (10 seconds between MURA and MALZ, and 20 seconds between MURA and SIRN) match well to the observed time difference for the beginning of co-seismic motions between these three stations (compare Figs. 2, 3, and 4). The co-seismic motions for the stations MALZ and SIRN amounted to 5.8±1.9 mm and 4.5±3.9 mm, respectively. B–Beginning of dislocation.

Fig. 4: Graph of calculated seismic travel times of P_g for CORS-TR stations in seconds versus distances in degrees for the Van earthquake. Values were calculated using the earthquake travel time calculator from the USGS (Coordinates of epicenter in degrees: 38.710° N, 43.446° E; Occurrence time of the Van earthquake: 10:41:21.73 (UTC); M_w=7.3; http://neic.usgs.gov/neis/travel_times/index.html).

Fig. 5: GNSS stations of the CORS-TR network in eastern Turkey and IGS stations in the surroundings included into the data processing. The violet star indicates the location of the epicenter of the Van earthquake. Boundaries of the Aegean Sea and Anatolian plates are shown in heavy colored lines following to Bird (2003) and Dilek (2010). NAF–North Anatolian fault; EAF–East Anatolian fault; DSF–Dead Sea fault; BZSZ–Bitlis-Zagros Suture Zone.

Fig. 6: Calculated horizontal coordinate differences between DoY294 (Oct. 21, 2011) and DoY307 (Nov. 03, 2011). Coordinates of daily solutions were estimated to study of datum-effect relative to the datum station ZECK (S1) and relative to the datum stations ZECK, ANKR, and ISTA (S3), respectively. The differences of coordinate components between these two solutions are systematic and amount to -0.5 mm for the north component, -1.2 mm for the east component. The violet star shows the Van earthquake epicenter.

Fig. 7: Vertical coordinate differences determined between solutions S1 and S3 between DOY294 (Oct. 21, 2011) and DoY307 (Nov. 03, 2011). Coordinate differences are systematic and amounts to 2.3 mm. The violet star shows the Van earthquake epicenter.

Fig. 8: Horizontal coordinate differences between DoY296 and DoY295 which show the co-seismic ground motions caused by the Van earthquake. Owing to the fact that GPS data for the station BASK, located in the south-east, were not available for observation days DoY295 and DoY296, ground motion of this station determined between the daily solutions DoY294 and DoY297 was added into the graph (arrow in red) to obtain an overlook about the direction and amount of movements of all stations within the network. We assume that the station BASK conducted this horizontal movement (24.2 ±3.4 mm) mainly due to the earthquake on DoY296, because the stations located in the south of the network had no significant movement on DoY297 and later. The uncertainties shown were determined using standard error propagation and scaled with a factor of 4. Error ellipses for horizontal movements are shown at 95% confidence level. The centroid moment solution determined by the USGS, is shown at the epicenter location, and indicates a visual representation of the style of faulting derived from the moment tensor. Shaded areas show quadrants of the focal sphere in which the P-wave first-motions are away from the source, and unshaded areas show quadrants in which the P-wave first-motions are toward the source. The N-S shortening we derive geodetically matches well with the N-S shortening and thrust focal mechanism derived independently using seismic data by the USGS

(http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc/neic_b0006bqc_cmt.php).

Fig. 9: To demonstrate the effects of the early aftershocks, horizontal coordinate differences between the solutions of DoY296 (Oct. 23, 2011) and DoY297 (Oct. 24, 2011) are illustrated. Due to strong aftershocks, some stations continued to move long after the mainshock, e.g., stations AGRD (11.4 ±2.3 mm) and MURA (21.2 ±2.2 mm). Uncertainties shown were determined using standard error propagation and scaled with a factor of 4. Error ellipses for horizontal movements are shown at 95% confidence level. The centroid moment solution from the USGS is also shown.

Fig. 10: Time series of daily horizontal coordinate differences of stations within the CORS-TR network (red-north, green-east). Stations located north of the blue line experienced a SW to SE motions, whereas the stations situated south of the blue line moved in a NE to NW direction (b). The vertical red line shows the approximate boundary between stations that moved SW (a1) and SE (a2) movements in the northern part of the network. Stations MURA (right top) and AGRD (center top) also include lines (blue) showing the difference in height from DoY300 to DoY302. The estimated vertical movements of the stations from DoY300 to DoY302 are interpreted as resulting from residual effect of the troposphere, rather than as being true post-seismic motions. BZSZ–Bitlis-Zagros Suture Zone.

Fig. 11: Time series of coordinate differences for far-field IGS stations DRAG, NICO, RAMO, and TUBI (red-north, green-east, and blue-height) which suggest an accuracy of about 4 mm for the north and east, and 10 mm for the height component of coordinates calculated here.

Fig. 12: Total (co-seismic and post-seismic) horizontal displacements between DoY294 (Oct. 21, 2011) and DoY307 (Nov. 03, 2011). Motions shown are relative to datum stations ZECK, ANKR, and ISTA. The uncertainties shown were determined using standard error propagation and scaled with a factor of 4. Horizontal error ellipses are shown at 95% confidence level. The centroid moment solution from the USGS indicates the Van earthquake epicenter. The N-S shortening we derive geodetically matches well with the N-S shortening and thrust focal mechanism derived independently using seismic data by the USGS illustrated in Figs. 8 and 9.

Fig. 13: Total (co-seismic and post-seismic) vertical displacements of stations between DoY294 (Oct. 21, 2011) and DoY307 (Nov. 03, 2011). Motions shown are relative to datum stations ZECK, ANKR, and ISTA. Red vertical bars correspond to surface sinking, whereas blue vertical bars mean surface uplifts. Uncertainties shown were determined using standard error propagation and scaled with a factor of 4. Error bars give a confidence of 0.95. Most calculated vertical movements of stations are not statistically significant. The violet star shows the Van earthquake epicenter.

Fig. 14: Figure shows the values of elongation determined using the analytical surface deformation theory (Altiner 1999, 2001b). Two kinds of crustal deformation dominate in investigation area: The north-eastern part of the network is characterized by extension with a magnitude of 0.2-1 ppm, mainly directed northwest-to-southeast. The southern part of the network experienced N-S shortening of 0.5-1.5 ppm. Observed co-seismic ground shaking areas were defined using the European Macroseismic Scale (EMS). The EMS is also shown as an overlay for comparison. In shaking zone VIII at a distance of 20-30 km, we have no GPS coverage. Strong N-S shortening dominates in the center of the study area. VIII–Crust

deformation of 2-3 ppm, VII–Crustal deformation of 1-2 ppm, VI–Crustal deformation of 0.5-1 ppm, V–Crustal deformation up to 0.5 ppm; BZSZ–Bitlis-Zagros Suture Zone. The centroid moment solution from the USGS indicates the Van earthquake epicenter.

Fig. 15: (A) The change of main curvature calculated illustrates how the external geometry was changed by the Van earthquake. For an optimal visualization, the figure was rotated 5 degrees about the X axis and 30 degrees about the Z axis of the global Cartesian coordinates. (B) The orientation of the network after the rotation described above.

Fig. 16: Figure illustrating amount and direction of the change in principal curvatures. Red shows the zone of decrease, whereas blue corresponds to the zone of increase. The Van earthquake and aftershocks up to November 03, 2011, caused principal curvature changes in the southwestern part of the network to decrease by 2-10 mm, whereas external geometry in the eastern and north-eastern parts of Van Sea inclined from 0.5 to 5 mm. The direction of red and blue arrows shows the slope direction of the changed external network geometry.

Fig.17. Comparison of the observed surface displacements from DoY295 to DoY296 (OBS) with the predicted co-seismic displacements based on the USGS teleseismic fault model (SYN). (a): The displacement vectors projected on the horizontal plane. (b): The vertical component of the displacements. The star is the epicenter of the earthquake and rectangle is the assumed mainshock fault plane projected on the surface, both given by the USGS.

Fig.18. Same as Fig. 17, but the predicted displacements are derived using the source model inverted from the GPS data.

Fig.19. Comparison between the teleseismic fault slip distribution given by the USGS and the geodetic fault slip distribution derived in this study. The uniform fault geometry (strike = 241° and dip = 51°) and the variable rake angle (0°-90°) are used in both models. The surface projection of the fault plane can be seen in Figs. 17 and 18.