Originally published as:

Liu, C., Zheng, Y., Xiong, X., Wang, R. (2015): Rupture Process of the 23 October 2011 Mw7.1 Van Earthquake in Eastern Turkey by Joint Inversion of Teleseismic, GPS and Strong-Motion Data. - *Pure and Applied Geophysics*, 172, 6, p. 1383-1396.

DOI: <http://doi.org/10.1007/s00024-014-0994-5>

Rupture Process of the October 23, 2011 Mw7.1 Van Earthquake in Eastern Turkey by Joint Inversion of Teleseismic, GPS and Strong-Motion Data

Chengli Liu¹, Yong Zheng^{1,*}, Xiong Xiong¹, Rongjiang Wang²

1. State Key Laboratory of Geodesy and Earth's Dynamics, Institute of Geodesy and Geophysics, CAS, Wuhan, China, 430077
2. GFZ German Research Centre for Geosciences, Telegrafenberg, 14473 Potsdam, Germany

Abstract On October 23, 2011, a strong earthquake ($M_w=7.1$) struck Van (Eastern Turkey) region and its surrounding areas at 10:41:22 UTC (hereafter referred to as Van earthquake), causing severe damage to the source region. Several studies have been carried out on the focal mechanism, seismic moment, focal depth, and rupture model of this earthquake. However, there are still significant differences in the moment release, focal depth and slip asperities among these models. Since most of these models only used a single dataset to investigate the source parameters, the reliability of these results is still a concern. In order to make clear the uncertainties and provide a reliable detailed rupture model for the main shock, we investigate the rupture process of the Van earthquake by a joint inversion of teleseismic broadband seismograms, near field static GPS displacement records and strong motion data. The inversion results indicate that the main shock is dominated by a thrust slip with a small part of left-lateral strike-slip component below the hypocenter. The rupture initiated at a focal depth of 16 km and propagated to the surface with a relatively low average rupture velocity of ~ 1.8 km/s, suggesting that the major energy of the earthquake could be released in long period band, which is the main reason why the seismic moments inverted by long period or static signals are higher than those obtained by short period data. Most of slip occurred around the hypocenter with the maximum slip of more than 3.3 m, and the associated

* Yong Zheng, corresponding author, Email: zhengyong@whigg.ac.cn

static stress drop is ~ 3 MPa. The total seismic moment of the whole fault is 5.76×10^{19} N·m, and most energy was released in the first 20 s, which is in the variation range of the released rupture models. The major slip is concentrated in the deeper depth, and extends to the depth around 25 km; meanwhile, the rupture in the ground surface is quite small, which explains why only weak ruptures observed at the surface caused by such a strong earthquake.

Key words: Van earthquake; Rupture process; Stress drop; joint inversion.

1. Introduction

Turkey is located in a tectonic active region with a high rate of seismicity and has experienced several earthquakes with intensities varying from V to X in Van and surrounding areas (Ergin et al. 1967; Tan et al. 2008; Selçuk et al. 2010) since the commence of historical seismicity records (Ergin et al. 1967; Alsan et al. 1975; Soysal et al. 1981; Ambraseys 1988; Eyidoğan et al. 1991; Turkelli et al. 2003; Fig. 1). The Van Lake Basin is located in the East Anatolian Plateau, a collision zone between the Eurasian Plate and the Arabian Plate starting from Late Miocene (Şengör and Kidd 1979; Şengör and Yılmaz 1983). Thus, many strong earthquakes have occurred in this region and caused serious damages in the source region. On October 23, 2011, a strong earthquake ($M_w=7.1$) struck the Van (Eastern Turkey) region and its surroundings at 13:41:22 (local time), and caused severe damages to the cities of Van, Erciş and nearby villages. According to the information provided by Prime Ministry Disaster and Emergency Management Presidency (AFAD, <http://www.afad.gov.tr>), this earthquake killed 604 people and injured 2,608. The main shock is located at 43.4657°N , 38.6890°E coordinates determined by the stations of the National Strong Motion Network of Turkey (TR-KYH).

Several studies have been done on the source parameters of this earthquake. According to the results from different international seismological institutions, the focal depth of the Van earthquake ranges between 10 and 19 km. The focal mechanism solution indicates that the earthquake is dominated by a major thrust fault striking 241°

and dipping 51° northward with small sinistral offsets (USGS 2011a: <http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc>; Erdik et al. 2012). However, there are big differences can be found in the seismic moment among the released models(Irmak et al. 2012; Fielding et al. 2013; Görgün 2013), moreover, significant differences can also be found in the rupture process of this earthquake, which causes some difficulties for understanding the seismogenic zone and evaluating earthquake damages.

After the occurrence of the Van earthquake, its surface deformation has been mapped by several studies by field work (Akyüz et al. 2011; Çağlan et al. 2011; Doğan et al. 2011; Ersoy et al. 2011; Emre et al. 2011; Özacar et al. 2011). These preliminary results show that the length of the major surface rupture is approximately 8 km and locates between Bardakci village and southeast of Asit village (Fig 1). In addition, the northern block (hanging wall) overlies the southern block (foot wall) along the main surface rupture. The maximum thrust offset and the maximum left lateral offset are declared as 0.15 m and 0.09 m respectively by the field observations (Doğan and Karakaş 2013). Thus, Doğan and Karakaş (2013) proposed that the fault might be interpreted as a blind thrust fault. Also, by inverting teleseismic waveforms, Irmak et al. (2012) and Fielding et al. (2013) demonstrated that the major rupture area is located around the initial failure zone with the maximum slip reaching about 3.6 m.

Based on these rupture models, the detailed rupture process of the Van earthquake remains unclear, mainly because of the data set used in these studies. As it is well known, without near-field records, especially near-field high-frequency records, it will be hard to determine the detailed information of the rupture model; on the other hand, if the teleseismic data are not used in inversion, the overall rupture pattern cannot be constrained well. Thus, in this work we investigate the detailed rupture process of the Van earthquake by jointly inverting teleseismic broadband seismograms, static GPS displacements and strong motion data in the near field, and provide a reference kinematic slip model for further studies on the Coulomb failure stress distribution on the surrounding faults and assess future seismic activity under the impact of the Van earthquake.

2. Methodology

To determine the rupture process of the Van earthquake by jointly inverting near field GPS data, teleseismic broadband data and strong motion data, we utilized the finite inversion method developed by Hartzell and Heaton (1983), and employed a simulated annealing algorithm to find a best slip model with good fit to the geodetic data and the seismograms in the wavelet domain (Ji et al. 2002).

$$u(t) = \sum_{j=1}^n \sum_{k=1}^m D_{jk} \left[\cos(\lambda_{jk}) Y_{jk}^1(V_{jk}, t) + \sin(\lambda_{jk}) Y_{jk}^2(V_{jk}, t) \right] \times \hat{S}_{jk}(t) \quad (1)$$

Here, $u(t)$ is the displacement at an arbitrary station, j is the j th subfault along the strike, and k is the k th subfault down dip. D_{jk} is the average slip amplitude of each subfault; λ_{jk} is the rake angle; $\hat{S}_{jk}(t)$ is the rise time function; V_{jk} is the average rupture velocity between the hypocenter and subfault; and $Y_{jk}^1(V_{jk}, t)$, $Y_{jk}^2(V_{jk}, t)$ are subfault Green's functions, which are generated by a combination of the reflection and refraction coefficient method (source region) and the ray theory (mantle region). For the case of the static response of a finite fault, the static Green's functions reduce to Y_{jk}^i ($i=1, 2$). The near field Green's Functions were computed using a frequency-wavenumber integration method (Zhu and Rivera 2002).

As for near-field GPS co-seismic displacements, sum-squared residuals (χ^2) have been used as the criteria for measuring the difference between synthetic and observed static displacements (Hudnut et al. 1996). It can be represented as:

$$\text{err}_{\text{st}} = \sum_i^N \left(\frac{S_o^i - S_s^i}{\sigma_i} \right)^2 \quad (2)$$

Here, S_o^i and S_s^i are the i th observed and synthetic static displacements, respectively, and σ is the observed standard error.

The simulated annealing method (Sen and Stoffa 1991) is applied to find a global minimum of the error function, which can be represented as:

$$\text{err}_{\text{wf}} + W_{\text{st}} \cdot \text{err}_{\text{st}} + W_{\text{c}} = \text{minimum} \quad (3)$$

Here, err_{wf} and err_{st} are objective functions of the seismic waveforms and static displacements, respectively. Two types of constraints are chosen: the one that minimizes the difference in slip on adjacent subfaults and the other that minimizes the total seismic moment (Hartzell et al., 1996). The W_{st} and W_{c} are the weights for the static criteria and the constraints which are applied to make the error function more stable. Based on this method, we can do joint inversion for the distribution of slip, rake angels, rupture velocities and rise time simultaneously.

3. Data and Fault Geometry

We collected the GPS displacements dataset provided by Altiner et al. (2013, pers. comm.). It consists of co-seismic displacements determined from the Turkish CORS-TR GPS network at 11 sites (Fig. 2, Table 1). Since the vertical components of the 11 GPS sites contain large uncertainties (Altiner et al. 2013) and are not ideal to be used to do the inversion for the rupture model, and the precision of GPS solutions in horizontal directions is usually more accurate than that in vertical direction, we collected 11 horizontal static offsets and 2 near fault vertical displacement records in this study. Although the overall distribution of GPS stations is relatively sparse, fortunately, the network still provides good azimuthal coverage near the source region.

In order to improve the dataset coverage in the near field, we also collected data from four strong-motion stations surrounding the fault from Center for Engineering Strong Motion Data (CESMD). Among the chosen strong-motion stations, one station is about 42 km to the northeast of the fault plane, and other three stations are less than 180 km away from the fault (Fig. 2). These four stations belong to the National Strong Motion Network of Turkey (TR-KYH). The raw acceleration records were integrated once to obtain velocity with a sampling rate of 0.2 s. Records from station 6503, were band-pass filtered in the frequency band of 0.01~0.4 Hz by a zero phase-shift Butterworth band-pass filter to remove long-period drift and high-frequency noise. The records from the other three distant stations (4901, 1302, 5601) were band-pass filtered

at 0.01~0.2 Hz.

Teleseismic P waveforms from 22 broadband seismic stations and SH waveforms from 14 stations are used in this study. These waveforms are downloaded from IRIS data center, with good signal-to-noise ratio and good azimuthal distribution (Fig. 2). The original records are deconvolved with instrument responses to obtain ground velocity data and are used to improve the spatiotemporal resolution of the inversion (Wald et al. 1996). Then the seismograms are band-pass filtered with frequency band of 0.002~1 Hz.

Based on the information of previous studies (Irmak et al. 2012; Doğan and Karakaş 2013), the Van earthquake is dominated by vertical slip but contains some minor strike slip components. The strike and dip angles of the rupture are in the range of N55°E to N70°E and 40° to 55°, respectively. Based on the tectonic settings and aftershock distribution, we adopted the focal mechanism provided by USGS (http://earthquake.usgs.gov/earthquakes/eqinthenews/2011/usb0006bqc/finite_fault.php) with the strike and dip angles of 241° and 51°, respectively. The strike and dip angles are quite close to those obtained by InSAR measurement (Elliott et al. 2013). Then, we designed a fault model which consists of a single plane with a spatial range of 72 km along strike and 40 km down dip. The depth of the hypocenter in the model is set to 16 km, and the fault plane is discretized into a total of 180 subfaults with the spatial dimension of 4.0 km × 4.0 km in length and width.

During the inversion, the slip amplitude for each subfault can vary from 0 to 6 m, the rake angle is searched from 21° to 81° with a search step of 2°, and the average rupture velocity is set to range from 1.6 to 3.0 km/s with an interval of 0.1 km/s. Finally, the rise time is allowed to change from 0.6 to 6.0 s with a time step of 0.6 s.

4. Result and Discussion

Source parameters from previous studies are obtained using a single dataset, e.g. teleseismic data, InSAR data, or satellite images. However, a single dataset sometimes

is hard to determine the rupture process well because of the blurring effect in teleseismic data caused by long epicentral distance. Although near-field GPS data can constrain the slip better than far field observations, since the static displacements decrease rapidly with the distance from fault trace, the precision of distant data is relatively lower. Furthermore, if the azimuthal coverage of near field data is not good enough, the overall moment of the earthquake may not be determined well. Considering these reasons, we combined all of the available data and carried out a joint inversion to constrain the rupture process. In this work, a 1D layered velocity model proposed by Novotny et al. (2001) (Table 2) was adopted to approximate the structure in the source region. And then, based on the source parameters described in section 3, the rupture process of the Van earthquake is determined from joint inversion of teleseismic, static GPS displacements and strong-motion data.

4.1 Checkerboard test

In order to evaluate the temporal and spatial solutions and reliability of our inverted model, we performed a checkerboard test to identify the rupture recovery using each of the single dataset and the joint datasets. We designed a checkerboard-like slip model (Fig. 3A). The results of the checkerboard test (Fig. 3) demonstrate that the teleseismic data can resolve the slip in the middle of the fault plane well, but relatively poor at the upper and lower depths (Fig. 3B). Compared with the teleseismic data, the strong motion data can better resolve the two asperities at both the middle and the shallow depths; while the asperity in the lower depth cannot be resolved well, and the spatial scale of the middle asperity is much smaller than its real shape (Fig. 3C). Compared with the strong motion data and the teleseismic data, the near field static GPS displacements can resolve none of the three asperities (Fig. 3D), which may be caused by two possible reasons: first, the GPS stations are relatively far away from the fault plane, and only two sites are close to the fault, which will smear the slip asperities; second, the static GPS data can only provide the permanent static deformation after the occurrence of the earthquake at the ground surface, which makes it impossible to resolve the dynamic process of the slip and hard to solve the slip in deep depths. On the

other hand, the GPS data can constrain the overall spatial scale of the slip. Thus, it is still quite useful for improving the reliability of a joint inversion.

Compared with the inversion using a single dataset, the joint inversion can resolve the rupture process much better. As shown in Fig. 3E, all of the three rupture patches are recovered very well. Both of the spatial scale and the magnitude of the slip are well constrained. Only in the deepest depth, the shape of the asperity is slightly smeared. From the middle to shallow depths, the slips are recovered very well with the difference relative to the testing model smaller than 3 %. Although there are some small artificial slips in the joint inversion model, their small magnitudes will not cause significant disturbance to the rupture model. Thus, the joint inversion model can provide much better rupture model than the inversion using a single dataset alone, and our joint inversion result can provide higher resolution and more reliable rupture process for the Van earthquake. Also, our result should be able to distinguish the differences among the previous models and the field observations.

4.2 Rupture process of the Van earthquake

Assuming the north-dipping plane is the fault plane, we obtained the slip model shown in Fig. 4 and the ground surface projection of the slip distribution shown in Fig. 5. The main shock is dominated by thrust slip with a significant left-lateral strike-slip component below the hypocenter. Although the majority slip of the Van earthquake occurred in a relatively compact area, the distribution of co-seismic slip displays three distinct asperities with relatively large slip. One asperity with large slip is located in the hypocenter area, a circular region with a radius of 10 km, and gradually expanded to the surroundings of the hypocenter with the maximum slip more than 3.3 m. The second asperity with a belt like distribution is smaller and shallower than the major asperity, and its center is located southwest to the hypocenter with a depth of about 5 km. The third asperity was located northeast to the hypocenter at a greater depth of about 25 km. The aftershocks are mostly concentrated on the overlying northern block and at/around the hypocenter. After the rupture of the mainshock, most of the stress and strain have been released in the hypocenter areas of the mainshock, while in the surrounded areas,

the stresses were not released completely with significant stress still left in these areas (Das and Henry, 2003). Thus, the stress concentration area was shifted to the west part of the fault. After the occurrence of the Van earthquake, the surface deformation caused by this earthquake was investigated by many researchers (e.g., Elliott et al. 2013; Akyüz et al. 2011; Çağlan et al. 2011; Doğan et al. 2011; Ersoy et al. 2011; Emre et al. 2011; Özacar et al. 2011). These surveys show that the length of main surface rupture was only approximately 8 km. Whereas, based on the slip distribution of our best model (Fig. 4a and Fig. 5), the smaller slip asperity extends to the Van Lake, which is hard to be observed from the surface, and moreover, most of the slips are concentrated in the great depth of the fault plane. Thus, our rupture model provides a reasonable explanation for why the observed surface rupture in the field survey cannot compensate for the magnitude of the earthquake.

4.3 Evolution of rupture process

For better understanding the rupture process of the Van earthquake, we present the model in a snapshot form. Figure 6 exhibits the evolution of the rupture process with time interval of 4 s. In the first 4 s, the rupture initiated from the epicenter, and propagated to the left side to a distance of ~8 km; then, in the following 4 s, the rupture propagated right and downward, together with some smaller slips propagated upward. In 8-12 s, the rupture got smaller and mainly propagated right and upward. During the first 10-12 s, most ruptures took place, but some weak ruptures still propagated farther away, even between 16 and 28 s, there were some small ruptures propagating to the southwest end of the fault. It is also noteworthy that at the first 4 s the northeastern rupture front already exceeded the solid contour, corresponding to a constant rupture velocity of 1.8 km/s, and then the rupture velocity slowed down after 8 s. Overall, the average rupture velocity of the Van earthquake is about 1.8 km/s, which is relatively small compared with normal intraplate earthquakes. This may be due to that the earthquake mainly initiated in the deep depth, and the slip was mainly composed by thrust component. Based on the dynamic analysis of earthquake rupture process (Chen et al. 2006), the rupture velocity of the strike-slip earthquake with apparent surface

rupture is usually faster than the thrust earthquake without obvious surface rupture.

4.4 Comparison between the synthetic and observed seismograms

The slip model determined by our joint inversion can explain most of the observed co-seismic displacements (Fig. 7), especially for the vertical components. However, two stations (AGRD, MALZ) located northwest to the fault plane have relatively larger fitting errors. Figure 8 shows the comparison between the observed teleseismic records and synthetic seismograms predicted by the slip model. The predicted waveforms are quite similar to the data in the direct P wave segments, but the mismatches at the tail of the waveforms cannot be ignored, especially for the SH waves. As for the strong-motion records (Fig. 9), the synthetic waveforms fit the data quite well, especially the velocity waveforms of the UD component and the NS component, but the slip model cannot predict the EW component of the station 4901 after 45 s.

Several factors may be responsible for the large residuals, such as the path effects caused by complicate structures and the heterogeneities in the earth. In our study, a 1D crustal model is adopted in the joint inversion, which is suitable for the inversion with long-period data, especially for data with periods longer than 10 s. However, in our inversion, the frequency range of the band-pass filter is 0.002~1 Hz. Because the source region is in a complex tectonic block, the 1D model may be a little oversimplified. Perhaps, a fine 3D model of the crustal structure beneath the Van region and its surrounding areas could help us to better model the rupture process. Besides the effect of crustal structure, the fault geometry may be another factor responsible for the misfits between the observed and predicted data. Based on the investigation of Elliott et al. (2013), the rupture may occur on two diverging fault planes with different dip and strike angles, and the geometries of the fault planes change gradually from surface to deep depth. The geometry of the fault plane may influence the distributions of slips. But overall, the observed data, including the static GPS displacements, local strong motion seismograms, and the teleseismic data are all well explained by our joint inversion. The rupture model shows good consistency with the distribution of the aftershocks.

4.5 Comparison with other models

A number of slip models have been derived for the Van earthquake. Elliott et al. (2013) obtained the static slip model for this earthquake using InSAR measurements, and the moment tensor using body wave inversion; Fielding et al (2013) inverted body wave data and geodetic data separately to obtain two kinds of slip models with different datasets. In order to analyze the differences between our model and the other models and evaluate the reliability of our model, we compared our results with the models proposed by Elliott et al. (2013) and Fielding et al. (2013).

For the moment rate function of the rupture model, our rupture model shows that the major rupture occurred in the first 10 s, and most seismic moment is released in the first 20 s, which is in good agreement with the model of Fielding et al. (2013). In their work, the major rupture occurred in the first 10 s, followed by a secondary asperity, which occurred between 10 and 16 s. Although the moment rate functions between these two models are quite similar, there are still some minor differences. Compared with the model of Fielding et al. (2013), the biggest difference is that the rupture time of our model is a little bit longer. In our model, although most moment is released at the first 20 s after the onsets of the rupture, there is another small peak between 20 and 30 s. The appearance of this peak is due to the asperity extending to the Van Lake (Fig. 4b). In the model of Fielding et al. (2013), the rupture terminated before 20 s, and there is no significant rupture occurring in later time. Thus, we examined the original teleseismic waveforms (Fig. 8) and observed small energy peaks after 20 s in waveforms from all of the stations, especially in the P waveforms. Two possible reasons may generate such kind of energy peaks in the teleseismogram: the first reason is that the multiple reflected P phases from the source region, such as pP or sP phases. However, if the later signal is generated by the reflect phase from the ground surface, the focal depth must be deeper than 40 km given the phase is 20 s later than that of the main phase. A deeper than 40 km focal depth is highly unlikely because this depth is even greater than the Moho depth in this area. The second possible reason is that it may be caused by a small asperity, considering the good fit between the synthetic waveforms and the observed seismograms. The later small asperity may not be caused by the

artificial effect caused by the model parameters, but is a real rupture asperity on the fault. It is hard to determine whether the later asperity is a later asperity of the mainshock or just is an earlier aftershock close to the Van Lake, More work is needed to clarify these issues in the future.

For the spatial distribution of the rupture model, all of the four models obtained by using teleseismic data (Fielding et al. 2013), geodetic data (Fielding et al. 2013), InSAR observation (Elliott et al. 2013) or joint inversion of multiple datasets (our model) show that the Van earthquake is dominated with thrust slip component, especially in the deep depth. The other common characteristic is that the slip in the shallow depth is small. This is the main reason why the length of the surface rupture observed in the field is so small. Our rupture model is close to the two models proposed by Fielding et al. (2013), and all of the three models show that there are some sinistral strike slip components. The percentage of the strike slip component in our model is in the range between the two models obtained by Fielding et al. (2013). The slip model of Elliott et al. (2013), on the other hand, is quite different. Since the rupture mainly gathers in the deep depth, the surface slip is quite weak and it is hard to constrain the rupture model in deep depth using InSAR measurements. So, although Elliott et al.(2013) study is based on InSAR measurements, more work is still needed to determine whether the rupture occurred on two or a single fault plane.

4. 6 Seismic moment released by the Van earthquake

The total seismic moment of the whole fault is 5.76×10^{19} N·m, and the major rupture zone is located at and around the hypocenter based on our joint inversion model. This result is quite close to the seismic moment reported by Fielding et al. (2013). Their results show that the total moments constrained by inverting different data have small differences: the moment constrained by body wave data is about 5.0×10^{19} N·m, and for geodetic data which is about 5.84×10^{19} N·m for shear modulus 30 GPa. The scalar seismic moment of both our joint inversion model and the geodetic model (Fielding et al. 2013) are slightly larger than that constrained by body wave data (Irmak et al. 2012).

This may imply that some of the energy was released in low frequency band, which is underestimated by the teleseismic body waves. However, the moment tensor of both of our results and Fielding et al. (2013) results are slightly smaller than the Global Centroid Moment Tensor M_0 of 6.40×10^{19} N·m (from <http://globalcmt.org>) and that determined by Görgün (2013) with long period (50~100 s) local broadband seismic data. Since both the global GMT solution and Görgün (2013) study depend on the long period signals, the differences between their work and our results may reflect the difference of the energy distribution in frequency band. The Van earthquake may release more energy at the long periods but less at the shorter periods. In fact, based on our joint inversion result, the average rupture velocity of the earthquake is only about 1.8 km/s, which is relatively slower than an ordinary earthquake, and more energy was released at long periods.

4.7 Stress drop on the fault caused by the Van earthquake

The joint inversion model shows that the slip mainly occurred in a relatively compact area (Fig. 5, red circle area). Assuming a circular slip asperity, the static stress drop can be estimated using the formula of Kanamori and Anderson (1975).

$$\Delta\sigma = \frac{7\pi}{16} \mu \frac{\Delta\bar{D}}{R} \quad (4)$$

Here μ is the shear modulus, $\Delta\bar{D}$ is average slip on a circular fault of radius R . Assuming $\mu=30,000$ MPa, based on our joint inversion model, we estimated the static stress drop is about 3 MPa, which is consistent with global studies showing that stress drops of intraplate earthquakes have a median of around 3.3 MPa (e.g. Allmann and Shearer 2009; Kanamori and Anderson 1975).

5. Conclusion

The Van earthquake in Turkey has attracted the attention of many researchers. Several studies have been carried out for its rupture process and source parameters. However, significant differences in the focal depth, slip asperities and moment rate

release among different models still exist. The main reason for these differences may be caused by the factor that these studies only use a single dataset in their inversions. In order to obtain a more reliable rupture process model, we analyzed the rupture process of the Van earthquake by jointly inverting teleseismic, GPS and strong motion data. The checkerboard test shows that a single type of data cannot explain the rupture process well; on the other hand, a joint inversion with teleseismic data, near field strong motion waveforms and near field GPS data can determine the rupture process very well, especially from the middle to shallow depths of the focal plane.

The inversion results indicate that the main shock is dominated by a thrust fault with a small left-lateral strike-slip component below the hypocenter, and a significant left lateral strike-slip component is dominant in the shallow part of the fault plane southwest to the epicenter. The rupture model is in good agreement with the historical fault movement mapped by the GPS and field observations. The rupture initiated at a focal depth of 16 km and propagated to the surface with an average rupture velocity of ~ 1.8 km/s, which is a quite slow rupture velocity. The slow rupture velocity indicates that the major energy is released at long periods, which is the reason why, in the models using long period data, the seismic moment is larger than those determined using shorter period data. The total seismic moment of the whole fault is 5.76×10^{19} N·m, and most of the moment was released at the first 20 s after the onsets of the rupture, which is within the range of previous models. The majority slip of the Van Earthquake occurred in a relatively compact area, and most of slip occurred near the hypocenter with the maximum slip of 3.3 m. Overall, most of the slips occurred in the deep depth, and only a weak slip reached the ground surface, which is the reason why the length of the surface rupture observed in the field does not compensate for the magnitude of the earthquake.

Acknowledgement

We are grateful to the IRIS for providing the teleseismic data, and to the CESMD for providing strong motion data. We thank Yüksel Altiner from Federal Agency for

Cartography and Geodesy and Frankfurt am Main from Germany for providing us the static GPS data, and also thank Prof. Michael H. Ritzwoller in the University of Colorado at Boulder for his constructive advice. The work is supported by NSFC grant No. 41174086, CEA grant No. 201308013, and Hubei Science foundation grant No. 2012FFA026.

REFERENCES

1. Akyuz, S., C. Zabcı, and T. Sancar (2011). Preliminary Report on the 23 October 2011 Van Earthquake, Tech. Rep., Istanbul Teknik Universitesi.
2. Allmann, B. P., and P. M. Shearer (2009). Global variations of stress drop for moderate to large earthquakes, *J Geophys Res-Solid*, 114, B01310.
3. Alsan, E., Tezucan, L., and Bath, M. (1976). An earthquake catalogue for Turkey for the interval 1913–1970. *Tectonophysics*, 31(1), T13-T19.
4. Altiner, Y., Söhne, W., Güney, C., Perlt, J., Wang, R., and Muzli, M. (2013). A geodetic study of the 23 October 2011 Van, Turkey earthquake. *Tectonophysics*. 588: 118-134.
5. Ambraseys, N. N. (1988). Engineering seismology, *Int. J. Earthq. Eng. Struct.Dyn.*, 17, 1–105.
6. Çağlan, D., Alan, H., Dirik, K.R., Bozkurt, E., Topal, T., Sözbilir, H., and Özkaymak, Ç. (2011), Van (Tabanlı - Edremit) Depremleri Raporu. TMMOB, Jeoloji Mühendisleri Odası Yayını, Ankara (in Turkish).
7. Chen, X.F., and Zhang, H. M. (2006). Modling rupture dynamics of a planar fault in 3-D half space by boundary integral equation method: an overview. *Pure. Appl. Geophys.*, 163:267-299.
8. Das, S., and Henry, C. (2003). Spatial relation between main earthquake slip and its aftershock distribution. *Reviews of Geophysics*, 41(3).
9. Doğan, B., Karakaş, A., and Karaağaç, S. (2011). 23.10.2011 tarihli Van (Bardakçı-Kozluca köyleri) Depremi Ön Değerlendirme Raporu. University of Kocaeli, Engineering Faculty, Department of Geological Engineering, Izmit-Kocaeli (in Turkish).
10. Doğan, B. and Karakaş, A. (2013). Geometry of co-seismic surface ruptures and tectonic meaning of the 23 October 2011 Mw 7.1 Van earthquake (East Anatolian Region, Turkey). *Journal of Structural Geology*, 46, 99-114.

11. Elliott, J. R., A. C. Copley, R. Holley, K. Scharer, and B. Parsons. (2013). The 2011 Mw 7.1 Van (Eastern Turkey) earthquake, *J. Geophys. Res.*, 118: 1619-1637.
12. Emre, Ö., Duman, T. Y., Özalp, S., and Elmacı, H. (2011). 23 Ekim 2011 Van Depremi Saha Gözlemleri ve Kaynak Faya İlişkin Ön Değerlendirmeler. Maden Tetkik ve Arama Genel Müdürlüğü, Jeoloji Etütleri Dairesi, Yer Dinamikleri Araştırma ve Değerlendirme Koordinatörlüğü, Aktif Tektonik Araştırmaları Birimi, Ankara (in Turkish).
13. Erdik, M., Kemer, Y., Demircioğlu, M.B., and Şeşetyan, K. (2012). 23 October 2011 Van (Turkey) earthquake. *Natural hazards*, 64(1): 651-665.
14. Ersoy, Ş., Görüm, T., Arbatlı, F., and Dölek, İ. (2011). 23 Ekim Van Depremi (Mw 7.2) Jeolojik ve Jeomorfolojik Ön Değerlendirme Raporu. Yıldız Teknik Üniversitesi, Doğa Bilimleri Araştırma Merkezi, İstanbul (in Turkish).
15. Ergin, K., U. Guolu, and Z. Uz. (1967). A catalog of earthquakes for Turkey and surrounding areas. Tech. Univ. Istanbul Mining Eng. Fac., Publ. No. 24,74 pp.
16. Eyidoğan, H., U. Güçlü, Z. Utku, and E. Değirmenci (1991). Türkiye Büyük Depremleri Makro-Sismik Rehberi (1990–1988). İTÜ Maden Fak. Jeofizik Müh. Böl., İstanbul, (in Turkish).
17. Fielding, E. J., P. R. Lundgren, T. Taymaz, S. Yolsal-çevikbilen, and S. E. Owen (2013). Fault-Slip Source Models for the 2011 M 7.1 Van Earthquake in Turkey from SAR Interferometry, Pixel Offset Tracking, GPS, and Seismic Waveform Analysis. *Seismol. Res. Lett.*, 84(4): 579-593.
18. Görgün, E. (2013). The 2011 October 23 Mw 7.2 Van-Erciş, Turkey, earthquake and its aftershocks. *Geophys. J. Int.*, 195, 1052-1067.
19. Hartzell, S., P. Liu, and C. Mendoza (1996). The 1994 Northridge, California, earthquake: Investigation of rupture velocity, risetime, and highfrequency radiation. *J. Geophys. Res.*, 101, 20,091– 20,108.
20. Hartzell, S. H., and T. H. Heaton (1983). Inversion of strong ground motion and teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley, California earthquake. *Bull. Seismol. Soc Am.* 73, 1553–1583.
21. Hudnut, K. W., Y. Bock, M. Cline, P. Fang, J. Freymueller, K. Gross, D. Jackson, S. Larson, M. Lisowski, Z. Shen, and J. Svarc (1996). Coseismic displacements of 1994 Northridge,

- California earthquake. *Geol. Soc. Am. Bull.* 86, no. 1B, S19–S36.
22. Irmak, T. S., Dogan, B., and Karakaş, A. (2012)., Source Mechanism of the 23 October 2011 Van (Turkey) Earthquake (Mw=7.1) and aftershocks with its tectonic implications. *Earth, Planets and Space*, 64, 1-13.
 23. Ji, C., D. J. Wald, and D. V. Helmberger (2002). Source description of the 1999 Hector Mine, California earthquake; part I: Wavelet domain inversion theory and resolution analysis. *Bull. Seismol. Soc. Am.*, 92, 1192-1207.
 24. Kanamori, H., and D. L. Anderson (1975). Theoretical basis of some empirical relations in seismology. *Bull. Seismol. Soc. Am.*, 65, 1073-1095.
 25. KOERI-UDIM (2012). Historical Earthquake Catalog, Boğazici University Kandilli Observatory and Earthquake Research Institute, National Earthquake Monitoring Centre, Istanbul, Turkey.
 26. Novotny, O., J. Zahradnik, and G-A. Tselentis (2001) North-western Turkey earthquakes and the crustal structure inferred from surface waves observed in the Corinth Gulf, Greece, *Bull. Seismol. Soc. Am.* 91, 875–879.
 27. Özacar, A., Kaymakçı, N., Akkar, S., Azari, A., Sandıkkaya, A., Kale, Ö., and Ay, B.Ö. (2011). 23 Ekim 2011 Mw7.2 Van Depremi Sismik ve Yapısal Hasara İlişkin Saha Gözlemleri. Rapor No: METU/EERC 2011-04. Middle East Technical University (in Turkish).
 28. Selçuk, L., Selçuk, A.S., and Beyaz, T. (2010). Probabilistic seismic hazard assessment for Lake Van basin, Turkey. *Natural Hazards*, 54, 949-965.
 29. Şengör, A. M. C., and Kidd, W. S. F. (1979). Post-collisional tectonics of the Turkish Iranian plateau and a comparison with Tibet, *Tectonophysics*, 55, 361-376.
 30. Şengör, A. M. C. and Yılmaz, Y. (1983). Türkiye’de Tetis’in Evrimi: Levha Tektoniği Açısından bir Yaklaşım, *Türkiye Jeoloji Kurumu Yerbilimleri Özel Dizisi*, no. 1, İstanbul (in Turkish).
 31. Sen, M. K., and P. L. Stoffa (1991). Nonlinear one-dimensional seismic waveform inversion using simulated annealing. *Geophysics*, 56, 1624–1638.
 32. Soysal, H., S. Sipahioğlu, D. Kolçak, and Y. Altınok (1981). Historical earthquake catalogue of Turkey and surrounding area (2100 B.C.–1900 A.D.). Technical Report, TUBITAK, No. TBAG-341.

33. Tan, O., Tapırdamaz, M.C., Yörük, A. (2008). The earthquake catalogues for Turkey. *Turkish Journal of Earth Science* 17, 405-418.
34. Turkelli, N., E. Sandvol, E. Zor, R. Gök, T. Bekler, A. Al-Lazki, H. Karabulut, S. Kuleli, T. Eken, C. Gürbüz, S. Bayraktutan, D. Seber, and M. Barazangi (2003). Seismogenic zones in Eastern Turkey, *Geophys. Res. Lett.*, 30(24).
35. Wald, D. J., T. H. Heaton, and K. W. Hudnut (1996). The slip history of the 1994 Northridge, California, earthquake determined from strong-motion, teleseismic, GPS, and leveling data. *Bull. Seismol. Soc. Am.*, 86, S49-S70.
36. Zhu, L. P., and L. A. Rivera (2002). A note on the dynamic and static displacements from a point source in multilayered media. *Geophys. J. Int.*, 148(3), 619-627.

Tables

Table 1. Co-seismic displacement used in this study

Station name	Latitude (°)	Longitude (°)	Height(m)	North (cm)	East (cm)	Vertical (cm)
SEMD	37.30515	44.57389	1422.944	0.26961	-0.02350	...
IGIR	39.92920	44.01825	884.293	-0.66481	-0.13014	...
BASK	38.04345	44.01731	2312.263	2.30273	-0.76479	...
OZAL	38.65731	43.98868	2037.262	0.78939	0.11454	-1.28482
MURA	38.99010	43.76312	1736.402	-3.40219	-1.69544	-0.90505
HAKK	37.57432	43.73913	1758.027	1.00864	-0.14889	...
AGRD	39.71756	43.03288	1672.602	-1.23237	0.23214	...
MALZ	39.14316	42.53080	1547.596	-0.57525	0.10767	...
SIRN	37.52503	42.45684	1459.332	0.43195	0.13314	...
TVAN	38.52949	42.29046	1717.503	-0.01148	-0.35443	...
HINI	39.36879	41.69579	1742.639	-0.08507	0.08754	...

“...” means not used during the joint inversion.

Table 2. 1-D velocity models derived by Novotny et al. (2001)

No.	Thickness (km)	V_p (km·s ⁻¹)	V_s (km·s ⁻¹)	ρ (kg·m ⁻³)
1	1.0	2.31	1.3	2100.0
2	1.0	4.27	2.4	2750.0
3	3.0	5.52	3.1	2800.0
4	11.0	6.23	3.5	3100.0
5	17.0	6.41	3.6	3350.0
6	Half-space	8.37	4.7	3350.0

Figure Caption

Figure 1. Simplified tectonic settings of Van and its surrounding areas, and some focal mechanism solutions of both historical and the instrumental period earthquakes around the city of Van. The red star and the red beachballs indicate the epicenter and focal mechanism of the Van earthquake respectively. The black beachballs represent the earthquakes recorded by the instruments and the green stars show the historical earthquakes. The gray circles show the aftershocks of the earthquake (KOERI-UDIM, 2012). Some active faults in Van and surrounding areas are shown with black lines. The black rectangle is the assumed fault plane,

Figure 2. The source region of the Van earthquake. The epicenter of the earthquake is represented by the black star, the black rectangle outlines the fault rupture plane, the black line represents the rupture caused by this earthquake, and the dash line is the non-failure part of the Bardakçı-Saray thrust fault zone (BSTFZ) (Doğan and Karakaş 2013). The GPS displacements are shown with black arrows. The gray triangles represent the location of near field accelerometers. The upper left inset shows the location of the teleseismic stations used in this study.

Figure 3. Checkerboard test for the model inversion. (A) The checkerboard slip model used to generate synthetic data. The inversion results obtained by using teleseismic data (B), strong motion data (C), near field static GPS data (D) and joint teleseismic, strong motion and GPS data (E).

Figure 4. (a) Slip distribution calculated from the inversion mode. The direction of the fault plane is indicated by the long black arrow on the top of the figure, and the star indicates the hypocenter, the color bar scales the slip amplitude, the black arrows represents the slip directions and contours display the rupture initiation time in second. (b) The moment-rate function of this earthquake.

Figure 5. The surface projection of the inverted slip distribution, black circles show the aftershocks reported by KOERI-UDIM (2012). The red dash circles outline the major rupture zone of the Van earthquake.

Figure 6. Snapshots of the Van earthquake source rupture at every 4 s. The numbers at the upper right corner of each panel indicate the time interval of this snapshot. A black star indicates the location of the hypocenter, and the dashed contours show the positions of the rupture fronts with a speed of 1.8 km/s every 4 s, and a solid contour specifies the rupture front with the same speed at the end of each snapshot.

Figure 7. Comparison between the observed displacements and those predicted from our model. The left panel shows the horizontal displacement and the right one is the vertical displacement. The black arrows are the GPS data, the red arrows are synthetics. The rectangle is the assumed fault plane, black star shows the epicenter of the earthquake.

Figure 8. Comparison of teleseismic velocity records in black and synthetic seismograms in red predicted by the slip model, the seismograms are band-pass filtered with frequency band of 0.002~1 Hz. Both data and synthetic seismograms are aligned on the P and SH arrivals. The number at the end of each trace is the peak velocity of the data in micrometers per second, and is used to normalize both records and synthetics. The azimuth and distance in degrees are shown at the beginning of each record with the azimuth on top.

Figure 9. Comparison of the strong-motion records (black line) and synthetic seismograms (red line) derived from our model. Both data and synthetics are aligned by the first P arrivals. The number at the first of each seismogram indicates the station name, the number at the right top is the maximum velocity of the records in cm/s. For station 6503, data were band-pass filtered in the frequency band of 0.01~0.5 Hz, for the distant records (4901, 1302, 5601), the frequency band is 0.01~0.2 Hz.

Figure1

Figure2

Figure3

Figure4

Figure5

Figure6

Figure7

Figure8

Figure9