

Source mechanism of the 2000 November 15 Lake Van earthquake ($M_w = 5.6$) in eastern Turkey and its seismotectonic implications

A. Pınar,^{1,2} Y. Honkura,² K. Kuge,³ M. Matsushima,² N. Sezgin,¹ M. Yılmaz⁴ and Z. Ögütçü⁴

¹Department of Geophysics, Istanbul University, 34850 Avcılar, Istanbul, Turkey. E-mail alipinar@istanbul.edu.tr

²Department of Earth and Planetary Sciences, Tokyo Institute of Technology, Tokyo 152-8551, Japan

³Department of Geophysics, Kyoto University, Kyoto 606-8502, Japan

⁴Boğaziçi University, Kandilli Observatory and Earthquake Research Institute, 81220 Çengelköy, İstanbul, Turkey

Accepted 2007 March 18. Received 2006 December 26; in original form 2005 December 7

SUMMARY

Detailed source process of the 2000 November 15 Lake Van (eastern Turkey) earthquake ($M_w = 5.6$) was retrieved using the method of Kikuchi & Kanamori for source inversion of complex body waveforms. The event has been reported by USGS as a deep (67 km) subcrustal earthquake resulting from a rupture on a normal fault beneath the Bitlis Suture zone where continental collision is in action between the Arabian and Eurasian plates. However, our source model based on the analysis of complex body-waveforms suggests that the earthquake is a shallow crustal event comprised of two subevents on a predominantly reverse fault at depths of 12.5 and 15 km with a time interval of about 18 s. The seismic moment of the second subevent ($M_o = 1.6 \times 10^{17}$ Nm; strike = 76° , dip = 55° and rake = 120°) is larger than the first subevent ($M_o = 1.0 \times 10^{17}$ Nm; strike = 87° , dip = 62° and rake = 110°). Visual inspection of the strong ground motion records and waveform inversion analysis of the near-field records from four broad-band stations confirm the occurrence of the two subevents at shallow depths. We examined whether or not the observed complex waveforms of the main shock can be modelled in terms of a single point source embedded at an intermediate depth, and the result turned out to be further evidence for the multiple rupture. The plausibility of the focal depths and mechanisms of the two subevents was also examined by retrieving the source parameters of 14 aftershocks from near-field waveform data. Most of the aftershocks were relocated at depths around 15 km, which agrees with the shallow main shock. In addition, the analysis of near-field waveform data indicates subcrustal earthquake activity neither in the source region of the Lake Van earthquake nor in the Turkey–Iran border region. In the Turkey–Iran border region, an event on 1999 February 19 was reported to have taken place at a depth of 66 km (USGS) and 77 km (ISC), which also conflicts with our focal depth (18 km) determined through CMT inversion analysis of the broad-band records at the station GNI.

Key words: aftershocks, complex earthquakes, Lake Van, near-field waveform analysis.

1 INTRODUCTION

The Lake Van region was hit by a moderate size earthquake at 15h 05m GMT on 2000 November 15 (38.397°N and 42.922°E , $h = 67$ km, $M_w = 5.6$, $mb = 5.2$, United States Geological Survey (USGS); 38.41°N and 42.95°E , $h = 48.4$ km, $M_s = 5.4$, $mb = 5.2$, International Seismological Center (ISC 2001); 38.14°N and 42.86°E , $h = 32$ km, $M_w = 5.6$, Harvard seismology group). The hypocentre is located to the north of Bitlis Suture Zone, a convergence zone between the Arabian and Eurasian plates in Eastern Anatolia (Fig. 1).

The 18 mm yr^{-1} N–NW motion of the Arabian plate results in westward escape of the Anatolian plate along the two prominently large transform faults, the North Anatolian fault (NAF) and the East Anatolian fault (EAF). GPS vectors in Fig. 1 show that the north-eastern part of the Eastern Anatolia moves towards N–NE along the Northeast Anatolian fault (NEAF). The intersection of these transform faults forms a triple junction (McClusky *et al.* 2000; Bozkurt 2001). The area to the east of the triple junction is squeezed in a nearly N–S direction, which causes NW- and SE-trending dextral and sinistral faults, thrust-to-reverse faults, E–W trending folds, strike-slip basins, Plio-Quaternary volcanic activity and elevation

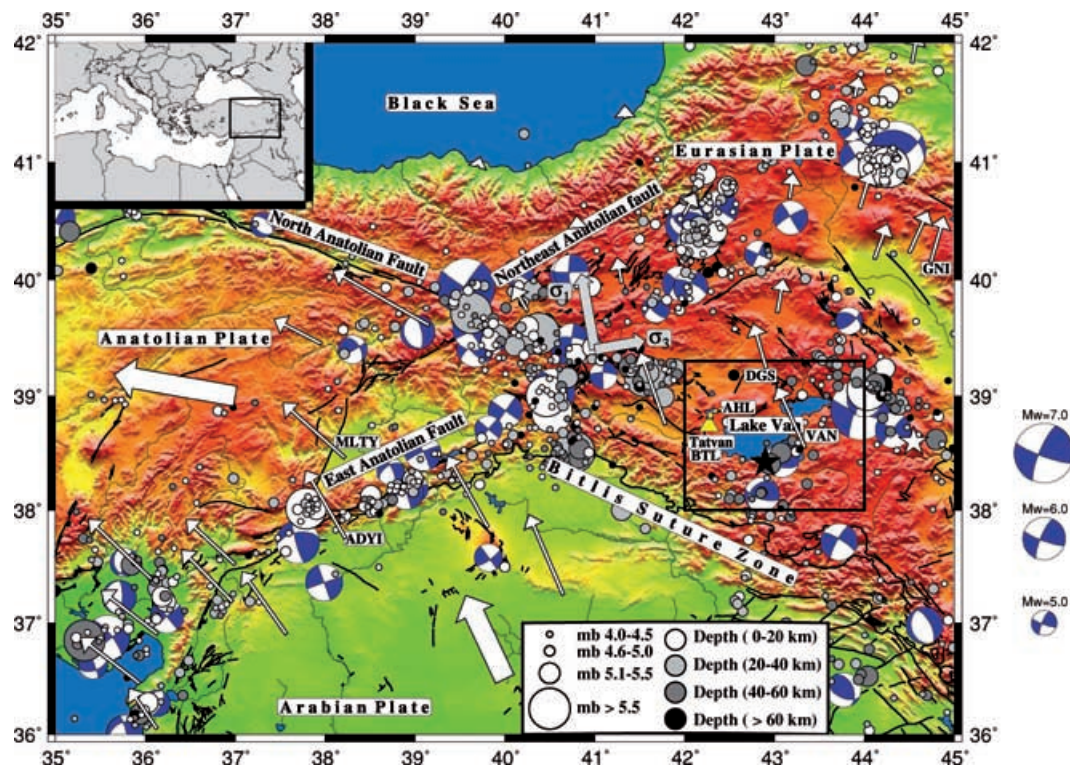


Figure 1. Seismotectonic map of Eastern Turkey. The thin white arrows are the GPS vectors showing the azimuth and slip rate of ground motion, and the thick white arrows show the plate motions given by McClusky *et al.* (2000). The best-double-couple focal mechanisms of moderate to large size earthquakes in the Harvard CMT catalogue during the period from 1976 to 2006 are illustrated on the lower-hemispheres with the equal-area projection. The size of the beach balls is proportional to the size of the events, as shown in the right-hand side of the figure. The circles with varying size and colour, as illuminated with a legend, are the ISC epicentres of earthquakes ($M > 4$) since 1964. The grey arrows in the mid of the figure show the direction of the maximum (σ_1) and minimum (σ_3) principal stress axes determined by Sezgin & Pinar (2002) using the focal mechanisms illustrated in this figure along with additionally published data. The solid rectangle shows the study area, which is enlarged in Fig. 7. Within the rectangle are shown the locations and abbreviations of the four broad-band stations used to retrieve the moment tensors of the aftershocks. The black star is the USGS epicentre of the 2000 November 15 Lake Van earthquake. The white star is the epicentre of the 1999 February 19 earthquake ($mb = 4.5$) analysed using the broad-band records at the station GNI. The solid lines indicate the active faults compiled by Şaroğlu *et al.* (1987).

of topography with the average height of 2 km (Şengör & Kidd 1979; Şengör & Yılmaz 1981; Dewey *et al.* 1986; Keskin *et al.* 1998; Bozkurt 2001; Koçyiğit *et al.* 2001; Keskin 2003; Şengör *et al.* 2003). The sense of motion on the faults, which is illuminated by the Harvard focal mechanism solutions of moderate to large earthquakes since 1976, is mostly strike-slip (Fig. 1). Stress tensor inversion of these focal mechanisms shows that the azimuth of the maximum compressive axis, (σ_1), is N350°E and the azimuth of the minimum compressive axis, (σ_3), is N80°E, that is, the compression is in the N–NW direction and the extension is in the E–NE direction (Sezgin & Pinar 2002). The focal mechanism solutions also clearly show that the present-day Arabia-Eurasia plate convergence results in continental deformation mainly taken up by the left-lateral strike-slip faults in the west of the region and the right-lateral strike-slip faults in the east. Thus, the prominent feature of strike-slip faulting suggests that the only driving force of tectonics in the region is the motion of the Arabian plate that causes escape tectonics (McKenzie 1972; Koçyiğit *et al.* 2001).

Along with the dominant strike-slip motion, ongoing compression is obvious from E to W trending compressional ramp basins. The 2000 November 15 earthquake occurred beneath one of these ramp basins, namely the Lake Van basin. Lake Van, which is located 1648 m above the sea level (Fig. 1), is the largest soda lake in the

world (Utkucu 2006). An active volcano called Nemrut is located on the western shore of the lake, which last erupted in 1441 or in 1597 (Aydar *et al.* 2003).

Several tectonic models have been proposed to explain the style of deformation in Eastern Turkey, as summarized in Keskin (2003). Some of them are as follows. (1) Tectonic escape of the microplates to the east and the west along the NAF, EAF and NEAF transform faults has been proposed by McKenzie (1972), Şengör & Kidd (1979) and Jackson & McKenzie (1988). Most of the focal mechanisms shown in Fig. 1 support this model because they are dominantly of a strike-slip type, including a little components of reverse and normal faulting. (2) The second model based on subcrustal earthquake activity is subduction of the Arabian plate beneath Eurasia (Rotstein & Kafka 1982). The circles in Fig. 1 show earthquakes ($M > 4$) compiled from the ISC catalogue since 1964. Focal depths of several events were reported to be larger than 40 km. (3) Another model that requires intermediate-depth earthquake activity stems from the study of Dewey *et al.* (1986), suggesting that the Arabia-Eurasian collision results in thickening of the lithosphere in Eastern Anatolia. In Fig. 1, the epicentre of the 2000 November 15 Lake Van earthquake is indicated by the black star near the southern shore of Lake Van. If the hypocentre is deep, the occurrence of the Lake Van earthquake will support the models (2) and (3). Therefore, the depth

of the Lake Van earthquake is important to evaluate these tectonic models in eastern Turkey.

The hypocentres of the Lake Van earthquake determined by USGS, ISC and Harvard University are located subcrustal, supporting the geodynamics model that requires relatively deeper seismicity. However, the results of recent studies indicate that the seismogenic thickness in Eastern Turkey is shallower than 30 km (Sandvol *et al.* 2003; Turkelli *et al.* 2003; Zor *et al.* 2003). We, therefore, investigate in detail the source parameters of the 2000 November 15 Van earthquake and its aftershocks with a special emphasis on constraining their focal depths. For this purpose, we utilize records at local and teleseismic distances. We invert the teleseismic data to analyse the main shock and the local records to analyse its aftershocks. Furthermore, we invert the local broad-band records of the 1999 February 19 earthquake ($m_b = 4.5$, $h = 66$ km (USGS), $h = 77$ km; ISC 2001) which was located in a cluster of subcrustal events to the northeast of Lake Van.

2 NEAR-FIELD RECORDS OF THE MAIN SHOCK

One of the stations that recorded the main shock of the Lake Van earthquake is the broad-band station, VAN (Fig. 1), operated by Kandilli Observatory and Earthquake Research Institute (KOERI) and equipped with a Guralp CMG-3T broad-band sensor having a flat velocity response up to 300 s. The station is located about 45 km to the NE of the epicentre. High-pass filtering with 5 Hz cut-off frequency was applied to the original records to clearly show the three distinct arrivals within 1 min after the P onset (Fig. 2a). These arrivals can be also identified on the acceleration waveforms recorded at Tatvan which is located about 50 km to the west of the epicentre (Fig. 2b). The broadband and strong motion data show that the amplitude of the third arrivals is smaller than those of the first two arrivals. The amplitude of the second arrival in the broad-band record is larger than the first one, whereas the acceleration amplitude of the second event is weaker. This could be due to eastward rupture propagation, variations in focal mechanisms and/or locations of subevents if the arrivals are manifestations of subevents during the earthquake rupture. In the next section, we show that the second arrival is attributed to the second subevent and that the second subevent has a larger size and a slightly different mechanism than the first one.

3 SOURCE INVERSION OF THE TELESEISMIC BODY WAVES

Of the three distinct arrivals observed at local distances, the first two arrivals can be depicted in the teleseismic records. The smaller amplitude and the low S/N ratio of the records could make the third arrival invisible. The second phases observed at the initial part of the teleseismic seismograms could be either surface reflection or the second subevent associated with the source process of the earthquake. In the case of surface reflection, the first phase is the P phase and the second one is the pP depth phase (Fig. 3), and the time difference between the P and pP phases provides the most reliable information on the earthquake depth. In the case of the 2000 November 15 Van earthquake, the time difference is about 18 s, corresponding to a depth of about 65 km ($pP_DEPTH = 65.2 + 0.91$; ISC). On the other hand, if the earthquake is shallow, we predict no phases for about 18 s after the P onset at teleseismic distances,

which can lead to a notion that the second phase was radiated from the second rupture. We, therefore, analyse the teleseismic records to distinguish between the two possibilities.

There are several waveform inversion methods to determine the source parameters of moderate to large earthquakes recorded at teleseismic distances. The method of Kikuchi & Kanamori (1991) is unique in that it not only determines the source parameters of the subevents but also their spatial and temporal distribution by inverting the body waves into their sources. Thus, the temporal and spatial distribution of the subevents yields a rupture model for the multiple events. In the method, the observed complex waveforms are modelled with point sources. First, a grid scheme of equally spaced point sources is constructed along the strike and dip directions of the fault plane. If the strike of the fault plane is not known, the azimuth of the grid scheme is gradually varied from 0° to 360° to search for the best azimuth that maximizes correlation between the observed and synthetic waveforms. Theoretical seismograms are first calculated for different depths at the epicentre, and then by shifting the theoretical seismograms in time, Green's functions are prepared for the other point sources located along the strike direction. Centroid Moment Tensor (CMT) parameters of an earthquake are determined by computing correlation between the observed and theoretical seismograms and looking for the maximum in the grid scheme. The node with the maximum correlation is retrieved for the centroid location of the analysed event. If that point source does not explain all the phases of the observed waveforms and we have evidence for the second rupture, as in the present case, the inversion procedure is carried out for two subevents and CMT parameters of the two subevents are retrieved through fitting calculated seismograms into observed ones. Usually, in standard inversion routines, the parameters of the second subevent are inferred in a trial-and-error manner. Fortunately, the method of Kikuchi & Kanamori (1991) can retrieve the source parameters of the whole subevents through inversion of the observed waveforms. Subevents are added until a satisfactory waveform fitting is achieved between the observed and synthetic seismograms.

Fig. 4(a) shows the waveform inversion result obtained by modelling the teleseismic waveforms with two subevents as suggested on the strong motion and near-field broad-band waveforms. We utilized 13 P and 10 SH phases recorded at different azimuths. Although the faulting mechanisms of the two subevents show predominantly reverse faulting, a considerable strike-slip component emerges in the second subevent. The following two centroid locations are determined by waveform inversion process in a grid scheme: (1) the first subevent is located at a depth of 15 km below the epicentre of USGS, (2) the second subevent is located 7.5 km to the north of the first subevent, at 12.5 km depth. The seismic moment of the second subevent ($M_0 = 1.6 \times 10^{17}$ Nm; strike 76° , dip 55° , rake 120°) is considerably larger than the first subevent ($M_0 = 1.0 \times 10^{17}$ Nm; strike 87° , dip 62° , rake 110°), which is in good agreement with our observation on the filtered waveform at VAN. The total CMT parameters obtained by summing up the seismic moment tensor components of the two subevents are as follows: seismic moment $M_0 = 2.5 \times 10^{17}$ Nm, strike $= 81^\circ$, dip $= 58^\circ$ and rake $= 117^\circ$ (Fig. 4a). The total seismic moment yields a moment magnitude $M_w = 5.6$.

We also implemented a single subevent inversion using the P waves only, in order to see whether or not the second phase can be modelled as the pP depth phase. Since the time difference between the P and assumed pP phases is about 18 s, we placed the grid scheme deeper and found out that a satisfactorily good fitting between the observed and synthetic seismograms is achieved for a depth of

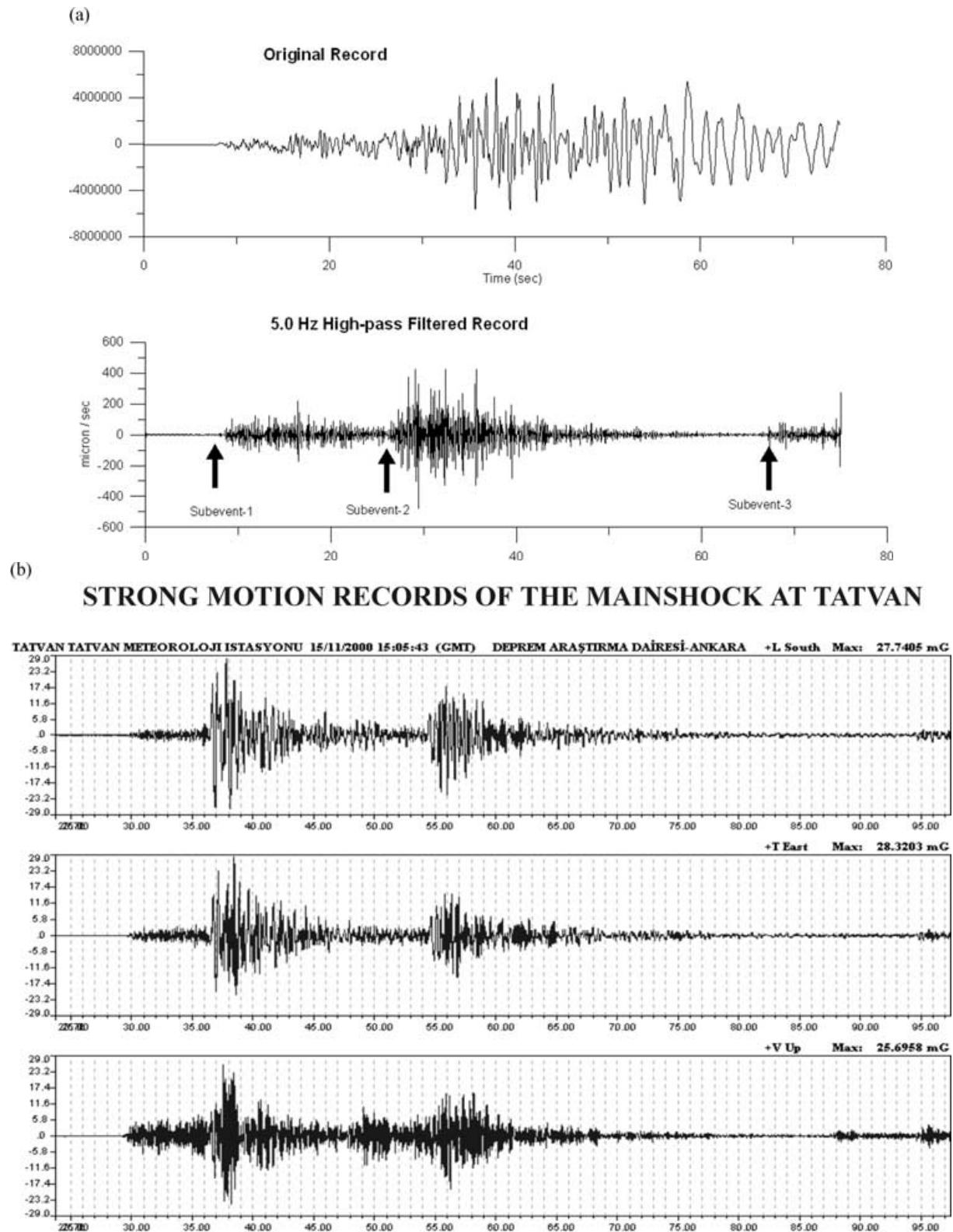


Figure 2. (a) Vertical-component broad-band seismogram of the 2000 November 15 Lake Van earthquake main shock from the station VAN. The upper trace is the original seismogram and the lower one is the 5.0 Hz cut-off high-pass filtered seismogram. The units of the original and filtered waveforms are different and denoted in the vertical axis. Note the three phases emerging after the filtering process, which correspond to three different ruptures in the source region. (b) The strong motion records of the main shock at Tatvan, also indicating the multiple ruptures associated with the main shock.

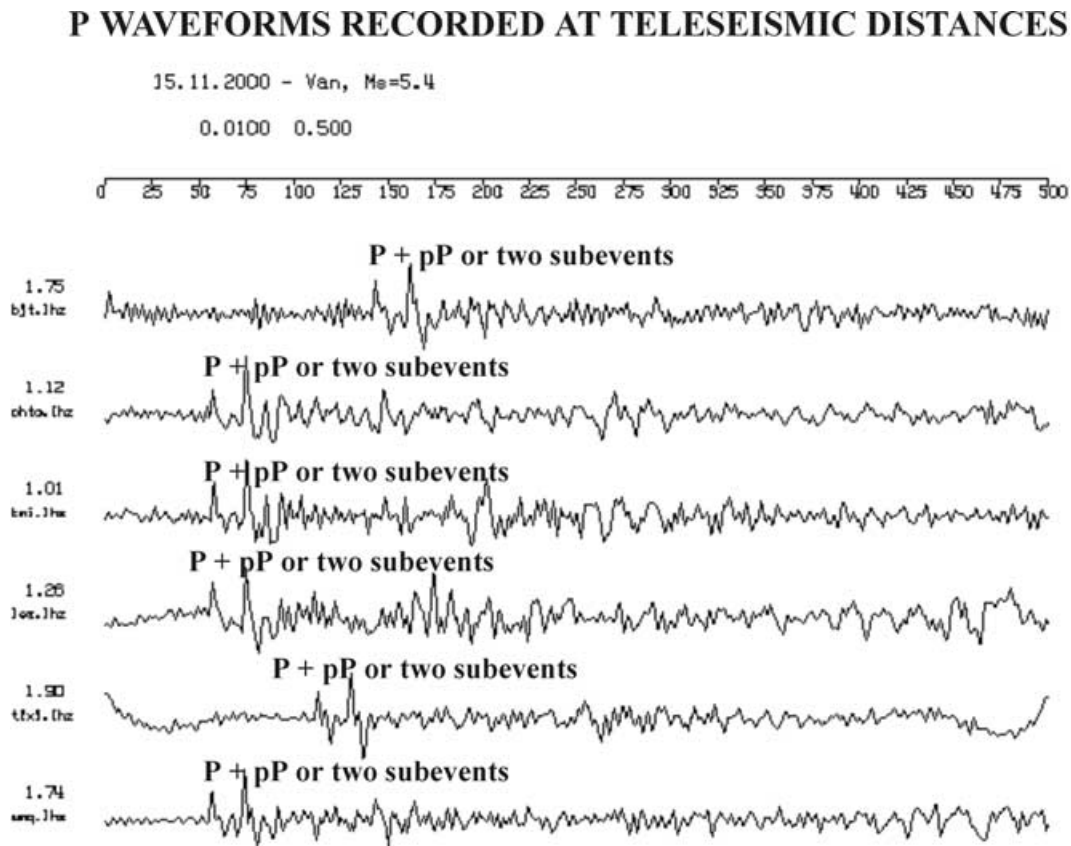


Figure 3. Telesismic waveform records of the main shock. Note the two distinct phases at the initial portion of the seismograms that could be either a combination of P and pP phases or phases radiated from two different ruptures at an interval of 18 s.

65 km. The best-fitting focal mechanism was predominantly normal faulting. Then we included the SH phases in the inversion, with small weights, and found that the solution obtained from the P waves does not fit the SH waves (Fig. 4b). Therefore, the intermediate focal depth is not plausible for the 2000 November 15 Lake Van earthquake, that is, the second phase in the observed seismograms is not the pP phase but a phase denoting a subevent. Increasing the weights of the SH phases to normal values, that is, weights according to the observed amplitude values, and including the second subevent at an intermediate depth did not provide a solution that fits the waveforms at all azimuths.

Next, we analyse the near-field broad-band records of the main shock and its aftershocks recorded at four nearby stations to verify the results from the telesismic data.

4 SOURCE PARAMETERS OF THE MAIN SHOCK AND ITS AFTERSHOCKS USING NEAR-FIELD RECORDS

The worldwide deployment of three-component broad-band stations stimulated studies on developing the near-field and regional moment tensor inversion methods, which are now widely used in recovering the source parameters of small to moderate size events (Fan & Wallace 1991; Dreger & Helberger 1993; Singh *et al.* 1997; Delouis & Legrand 1999; Kuge 1999; Legrand & Delouis 1999; Legrand *et al.* 2000; Kuge 2003; Pinar *et al.* 2003).

Fortunately, the network Eastern Turkey Seismic Experiment (ETSE) consisting of 29 broad-band stations operated in Eastern Anatolia when the 2000 November 15 Lake Van earthquake took

place. Örgülü *et al.* (2003) obtained a CMT solution for the main shock by modelling the waveforms with a single point source. Their solution is similar to the Harvard CMT solution. However, here we show that single source modelling of the near-field records does not give a satisfactory fit between the synthetic and observed waveforms bandpass filtered between 0.04 and 0.1 Hz. We used data from the three nearest ETSE stations along with the broad-band station VAN, providing a good azimuthal coverage (Fig. 1) to recover the source parameters of the main shock using the technique developed by Kuge (2003). The method is based on waveform modelling of displacement seismograms at one or more stations at local distances. A centroid location can be searched for in a 3-D-grid scheme by achieving the best fit between observed and synthetic displacement seismograms. The synthetics are calculated following Kohketsu (1985) for a horizontally layered structure given in Table 1. We fixed the crustal structure model, inspecting the velocity models of Turkelli *et al.* (2003) and Zor *et al.* (2003). By comparing our observed arrival times for P and S waves to the predicted ones, we confirmed that the velocity model based on the receiver-function analysis beneath the station AHL well explains the observed P and S traveltimes (Table 1). For most of the events, data were bandpass filtered between 0.04 and 0.1 Hz, while the waveforms of the smaller events were bandpass filtered between 0.06 and 0.15 Hz. Fan & Wallace (1991) showed that waveforms at a low frequency band are not so much sensitive to crustal models and fault parameters can be successfully determined from the waveforms even without knowledge of fine crustal structure.

During the inversion process, we give uniform weighting to all the seismograms. The goodness of fit between observed and

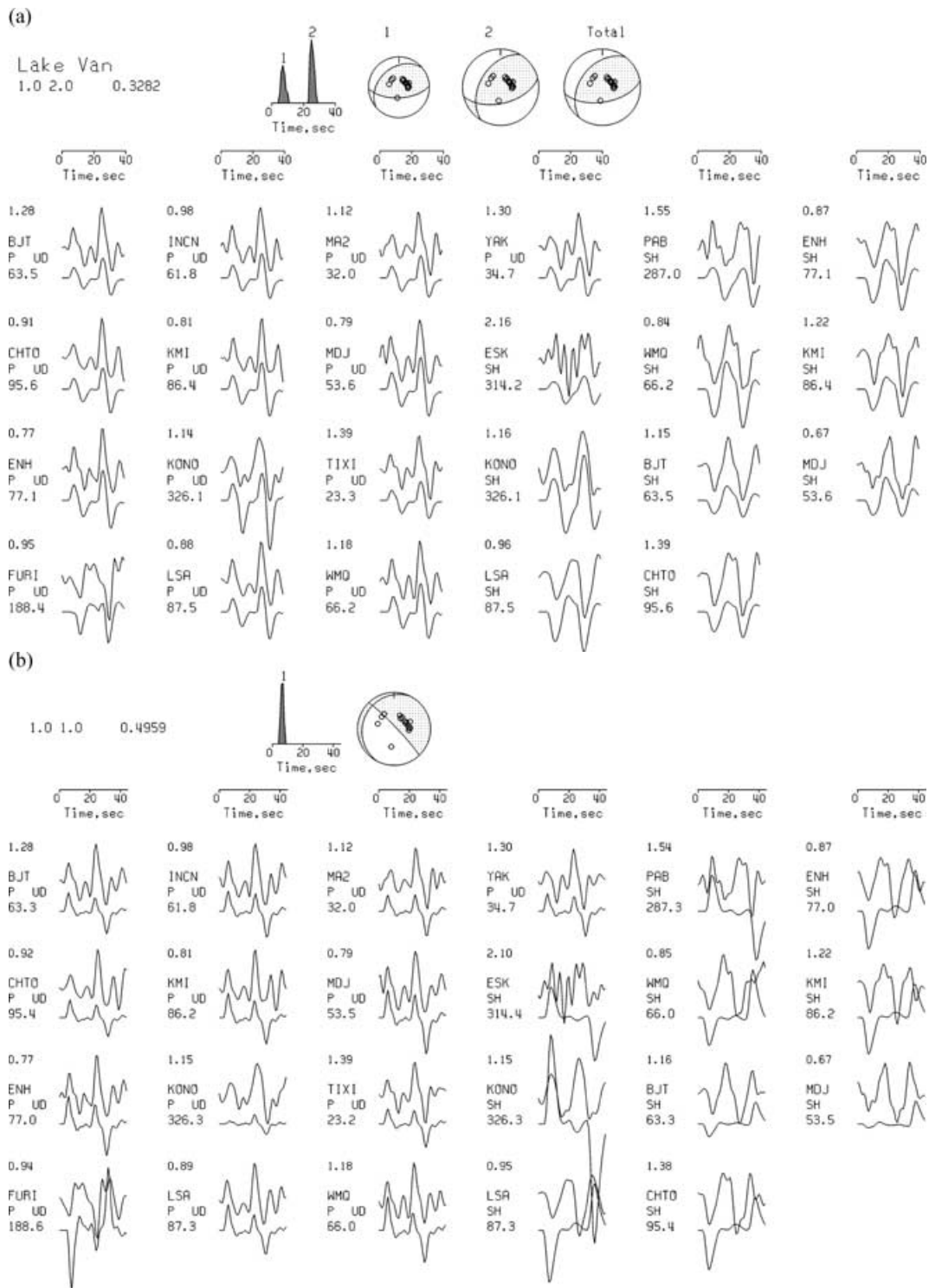


Figure 4. (a) Results of source inversion with two subevents for the main shock. The source time functions, the focal mechanisms of the first and second subevents, and the total moment tensor are shown above the seismograms. The observed (upper) and calculated (lower) waveforms are shown. The values given above the station code indicate the peak-to-peak amplitude (in microns) of the observed records, and the numbers given below the names of the component indicate the azimuths of the stations. (b) Results of source inversion with a single point source embedded at an intermediate depth. Whether or not the second subevent inferred in (a) could be explained with the pP depth phase is tested. Note a satisfactory good fit to the *P* waves but no to the SH waves.

predicted seismograms is measured by variance reduction (VR) (Kuge 2003). The larger the value of VR, the better is the fitting. The maximum of VR is 100. The variance reduction is calculated for various depths, and we select the faulting mechanism, for which

VR is the maximum, as faulting parameters of the analysed event. The result of the moment tensor inversion is illustrated in Fig. 5(a). Note that we only model the initial pulses of the main shock records (Fig. 5b) corresponding to the first subevent since the technique of

Table 1. Crustal model used to estimate the source parameters of the main shock and its aftershocks. V_p , V_s , ρ and H are P -, S -wave velocity, density and thickness of the layers, respectively.

V_p (km s ⁻¹)	V_s (km s ⁻¹)	ρ (kg m ⁻³)	H (km)
4.50	2.6	2210	2
5.54	3.2	2540	6
6.23	3.6	2760	10
6.92	4.0	2980	12
7.78	4.5	3260	∞

Kuge (2003) does not support analysis of subevents with varying focal mechanisms. The variance reductions determined for different depths suggest a focal depth of 20 km for the first subevent (Fig. 5c), which is slightly deeper than the depth determined from the inversion of the teleseismic waveforms. Nevertheless, it is obvious from the analysis of both the near-field and teleseismic waveforms that the Lake Van earthquake occurred within the crust.

We have also determined moment tensor solutions for 14 aftershocks. In Fig. 6, we show an example of the CMT inversion using

waveforms of a small aftershock ($M_w = 3.5$). The results of the aftershocks are illustrated in Fig. 7 and Table 2 gives their CMT parameters.

5 CMT PARAMETERS OF 1999 FEBRUARY 19 EVENT ($m_b = 4.5$)

Waveform data for this event were available only at the broad-band station GNI (Garni, Armenia) of IRIS-USGS. Therefore, we used three-component waveform data from the single station. There are several studies in the literature showing that three-component broad-band data from a single station can be used to retrieve a reliable moment tensor (Fan & Wallace 1991; Delouis & Legrand 1999; Dreger & Savage 1999; Kim & Kraeva 1999; Legrand & Delouis 1999; Pinar *et al.* 2003). For example, reliability of the result obtained by inversion of waveforms from a single station was examined by Pinar *et al.* (2003) using the records of the 1999 August 31 ($M_w = 5.2$) Western Turkey event from the broad-band station ISK of KOERI and comparing their result with the Harvard CMT

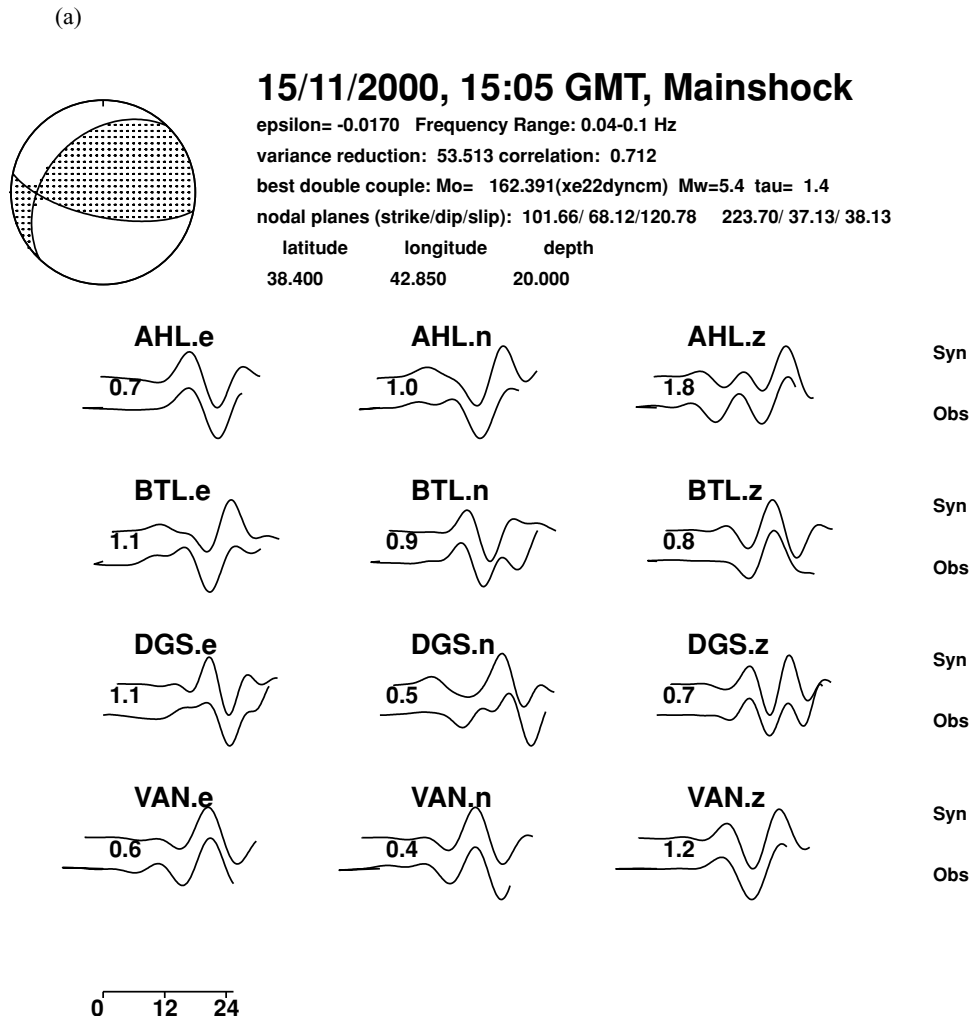


Figure 5. (a) Result of moment tensor inversion using the near-field waveform records of the main shock at four broad-band stations. Note that only the initial portions of the seismograms, which were generated by the first subevent, are used. The focal mechanism is illustrated in the upper-left part, and the source parameters are shown in the right of the focal sphere. The synthetics (upper) and observed (lower) seismograms for the East–West component (e), for the North–South component (n) and the vertical component (z) at each station are shown along with the synthetic-to-observed amplitude ratio between the two seismograms. (b) Bandpass filtered (0.03–0.1 Hz) displacement seismograms of the main shock recorded by the four broad-band stations. The boxes indicate the time windows used for CMT inversion in (a). (c) Variation of VR with respect to depth to determine the best CMT depth.

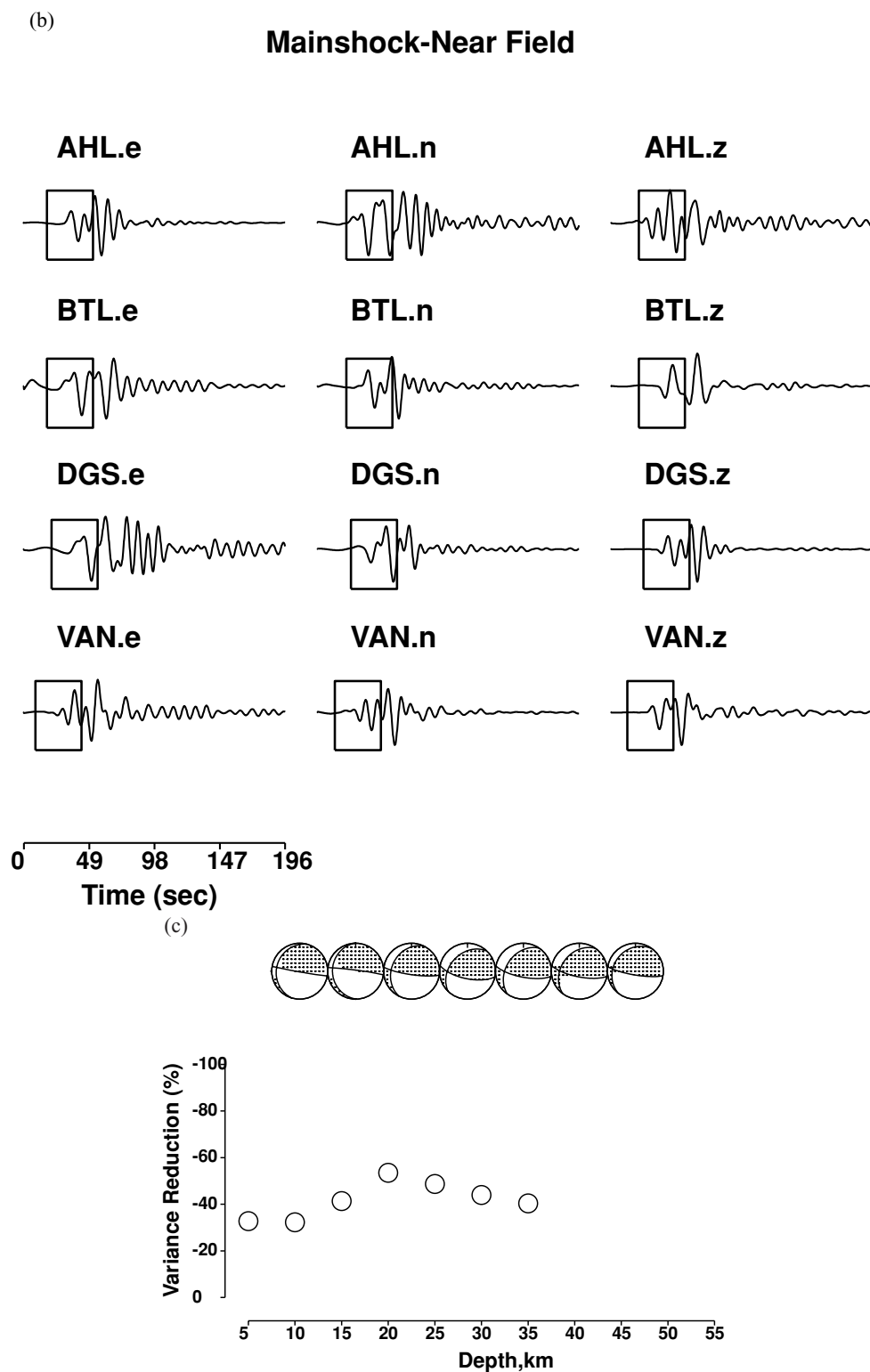


Figure 5. (Continued.)

solution. They showed that the three-component data recorded at a distance of about 100 km from the epicentre could successfully be used to retrieve a moment tensor solution of the earthquake.

Comparing hypocentres located by the dense ETSE network with ones of USGS, Türkelli *et al.* (2003) showed that the USGS epicentres are on the average about 20 km away from their epicentres for

events in Eastern Turkey. Taking this fact into account, we established a 3-D grid scheme with grid points extending 20 km to the east, west, north and south directions from the USGS epicentre, at intervals of 5 km. To compute the Green's function, we used the same crustal structure of Table 1 as used in the analysis of the aftershocks.

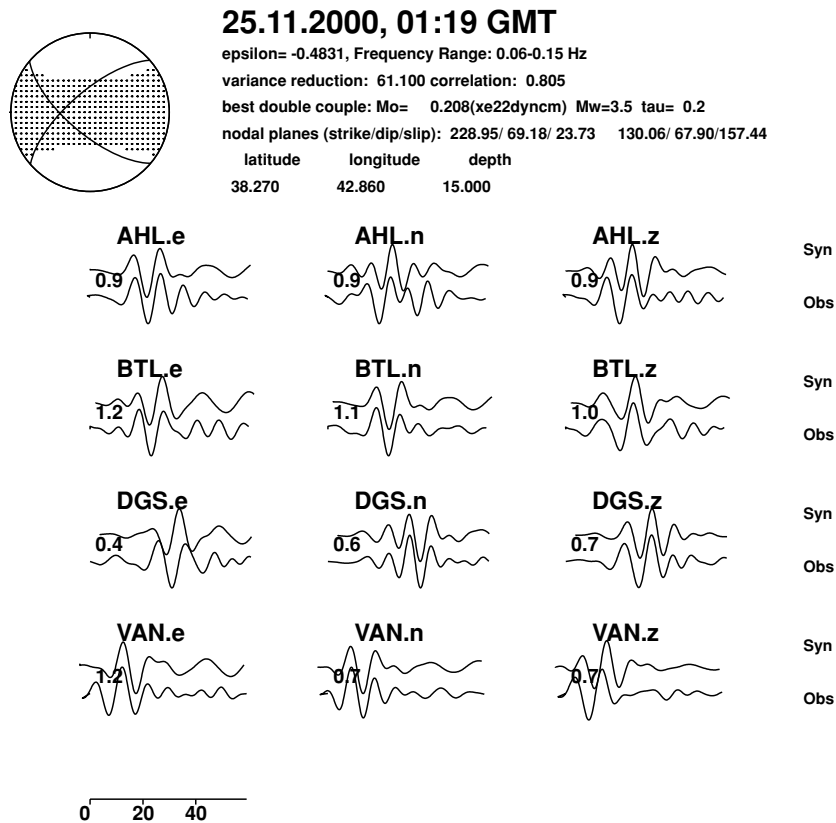


Figure 6. Results of moment tensor inversion for the 1999 November 25, 01:19 GMT, aftershock ($M_w = 3.5$) as an example for a small event. The focal mechanism is illustrated in the upper-left part, and the source parameters are denoted to the right of the focal sphere. The synthetics (upper) and observed (lower) seismograms for the East–West (e), North–South (n) and vertical (z) components at each station are shown along with the synthetic-to-observed amplitude ratio between the two seismograms.

For the sake of computing time, we first allowed the grid points up to a depth of 90 km with a grid spacing of 10 km, and good fittings were obtained for focal depths around 20 km. Then, for more precisely estimating the focal depth, we tested depths ranging from 3 to 21 km with an increment of 3 km. The final inversion results show that the 1999 February 19 event is also a shallow event located at a depth of 15–18 km. In Fig. 8, we show the focal mechanism solution and depth for which the largest variance reduction in waveform fitting is achieved at each grid node. Our preferred focal mechanism solution is predominantly normal faulting, striking in the NE–SW direction and having a slip vector parallel to the GPS vectors shown in Fig. 1. Although the focal mechanisms of the large events in the vicinity of the 1999 February 19 event are predominantly right-lateral strike-slip faulting, for example, the focal mechanism of the 1976 November 24 Çaldıran earthquake ($M_s = 7.3$), a normal faulting mechanism for the February 19 event is plausible since the large NW–SE oriented right-lateral strike-slip faults in the region are segmented and offset, giving rise to local extensional features (Toksöz *et al.* 1997; Barka & Kadinsky 1988; Koçyiğit *et al.* 2001) where NE–SW striking normal faults may occur.

6 STRESS TENSOR ANALYSIS

The focal mechanisms illustrated in Fig. 1 show mostly predominantly strike-slip mechanisms, whereas the fault-plane solutions in Fig. 7 are mostly oblique thrusting. This difference implies that a discrepancy can exist between the local stress field in the source region of the Lake Van earthquake and the regional stress field acting

throughout Eastern Turkey. We now investigate the local stress field in the source region and compare the result with the regional stress field. In our stress tensor estimations, we assume that the static stress increase associated with the Lake Van earthquake is small due to its moderate size and the influence on the regional stress field is negligible. Eventually, the focal mechanisms of the aftershocks can be used to derive the stress field in the source region of the Lake Van earthquake. Along with the aftershock data in Table 2, we use the faulting parameters of the two subevents of the main shock and the Harvard CMT solution of the 1988 June 25 earthquake in the proximity of the Lake Van earthquake.

The method we use to derive the stress tensor acting in the source region is described in Gephart & Forsyth (1984), Gephart (1985) and Gephart (1990). Data are orientations of P - and T -axes from fault plane solutions. The distribution of used P - and T -axes is shown in Fig. 9(c). The method yields three principal stress directions, namely, directions of maximum compression (σ_1), intermediate compression (σ_2) and minimum compression (σ_3) and a stress magnitude ratio defined as $R = (\sigma_2 - \sigma_1)/(\sigma_3 - \sigma_1)$.

A combination of these four parameters (σ_1 , σ_2 , σ_3 and R) is called a stress model, and the model that most closely match the whole observed data set is called the best-fitting stress model. The best-fitting model is searched for in grid points over the four model parameters, adjusting systematically one at a time through a wide range of possibilities (Gephart 1990). The measure of misfit is given by the smallest rotation angle about an axis of any orientation that brings slip directions of data into shear orientations predicted by the stress model.

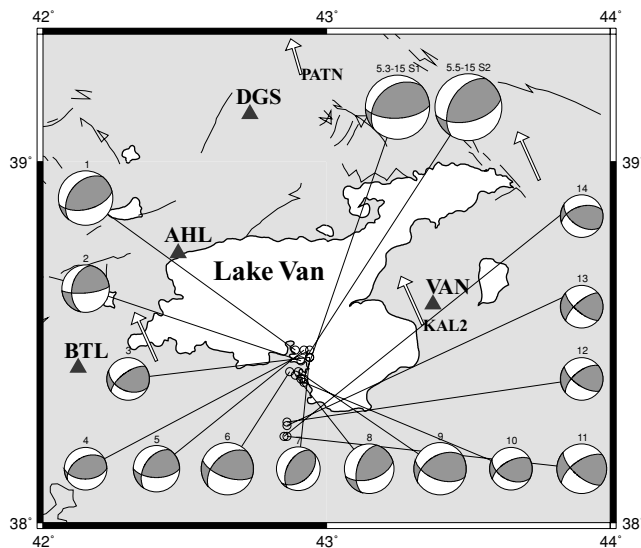


Figure 7. Locations of fourteen aftershocks and their best-double-couple focal mechanism solutions obtained from our moment tensor inversions. The locations and focal mechanisms of the two subevents during the main shock are also shown. Note the two clusters of aftershocks corresponding to the locations of the two subevents. The southernmost third cluster may correspond to the third subevent observed at the broad-band station VAN (Fig. 2a) and the strong motion station Tatvan (Fig. 2b). The source parameters are given in Table 2.

We start the inversion from the stress model of Sezgin & Pinar (2002), with azimuths of 350° and 80° for horizontal σ_1 and σ_3 compression axes, respectively. A fine grid search with an increment of 5° was performed around the σ_1 and σ_3 axes to find the best-fitting stress model. The inversion yields the following results: the azimuth and plunge pairs for the three principal stress axes, σ_1 , σ_2 and σ_3 , are 335, 16; 255, 32 and 42, 53, respectively. The best-fitting local stress model derived for the Lake Van earthquake source region is different from the regional stress model of the whole Eastern Turkey with the azimuth and plunge pairs of (350, 0) for σ_1 , (260, 81) for σ_2 and (80, 9) for σ_3 (Sezgin & Pinar 2002). In Fig. 9, we show both our result and the result of Sezgin & Pinar (2002) to compare with each other, and discuss the difference in the next section.

7 DISCUSSION

7.1 Identifying fault planes from GPS data and SLIP vectors

Incorporating studies of teleseismic, near-field strong ground motion and broad-band data, and the CMT solutions of the aftershocks has excellently revealed that the 2000 November 15 Lake Van earthquake is a multiple rupture in the upper brittle crust. The focal mechanisms of the two subevents of the main shock and most of the aftershocks have nodal planes dipping towards NNW and SSE (Fig. 7). Given the focal mechanisms, the next question for further understanding the collision process along the Bitlis Suture Zone is ‘which of the nodal planes is the fault plane?’ The aftershock distribution gives no hint on the dip directions of the fault planes because most of the aftershocks have CMT focal depths at 15 km. To answer the question, we use GPS vectors around the lake (Fig. 7) and the azimuths of the slip vectors on the two nodal planes.

First of all, we show cases when GPS vectors or differential GPS vectors can constrain slip directions on faults, that is, fault planes. Around the right lateral strike-slip NAFZ (Fig. 1), the GPS vectors are parallel to the westward motion of the Anatolia plate with respect to the Eurasia plate (McClusky *et al.* 2000). For strike-slip faulting mechanisms, a nodal plane with a slip vector parallel to the GPS vector is likely to be the fault plane. Therefore, in the vicinity of the NAFZ, comparison of the GPS vectors with two slip vectors of a fault plane solution can identify its fault plane. Alternatively, a difference between two GPS vectors at opposite sides of a fault or a fault zone can be used to compare with slip vectors on two nodal planes. For example, the relative motion between the GPS vectors at MLTY and ADYI (Table 1 in McClusky *et al.* 2000) is towards N60E or N240E, corresponding to the slip motion for the East Anatolian fault zone. This direction perfectly fits the slip vectors on the NE–SW trending nodal planes of the focal mechanisms (Fig. 1). On the contrary, for normal and reverse faulting mechanisms, comparison between GPS data and slip vectors on two nodal planes fails to determine the fault plane because the slip vectors look nearly parallel when they are projected on the horizontal plane. In cases of oblique faulting mechanisms, however, GPS vectors could be still used to determine the fault plane by examining whether a minor strike-slip motion on each nodal plane is left- or right-lateral. Our knowledge on the general pattern of deformation will then help to decide the fault plane. Based on all these techniques using the GPS data around

Table 2. Earthquake parameters and best double-couple solutions of fourteen aftershocks. *H* is depth, *Str* is strike angle and *VR* is variance reduction to represent the goodness of the CMT solutions (also see the text).

No.	Date	Time (GMT)	Lat N	Long E	H (km)	M_w	Str deg	Dip ($^\circ$)	Rake ($^\circ$)	VR
1	15.11.2000	16:06	38.48	42.89	18	4.5	76	56	113	62
2	15.11.2000	16:43	38.45	42.91	15	3.9	85	75	130	65
3	15.11.2000	17:07	38.46	42.94	12	3.5	114	43	142	65
4	15.11.2000	17:44	38.48	42.94	15	3.5	89	50	110	63
5	15.11.2000	18:16	38.48	42.92	15	3.8	90	59	127	58
6	15.11.2000	19:30	38.42	42.87	15	4.3	109	63	130	63
7	15.11.2000	21:17	38.46	42.94	15	3.6	49	42	104	68
8	16.11.2000	21:13	38.41	42.89	15	4.1	73	55	129	59
9	17.11.2000	00:27	38.42	42.90	15	4.3	109	55	124	65
10	17.11.2000	09:36	38.39	42.92	15	3.5	117	46	136	54
11	19.11.2000	00:02	38.24	42.86	15	3.7	131	65	149	56
12	25.11.2000	00:59	38.28	42.86	15	3.5	125	64	147	57
13	25.11.2000	01:19	38.27	42.86	15	3.5	130	68	157	61
14	30.11.2000	10:30	38.24	42.85	12	3.4	123	51	127	61

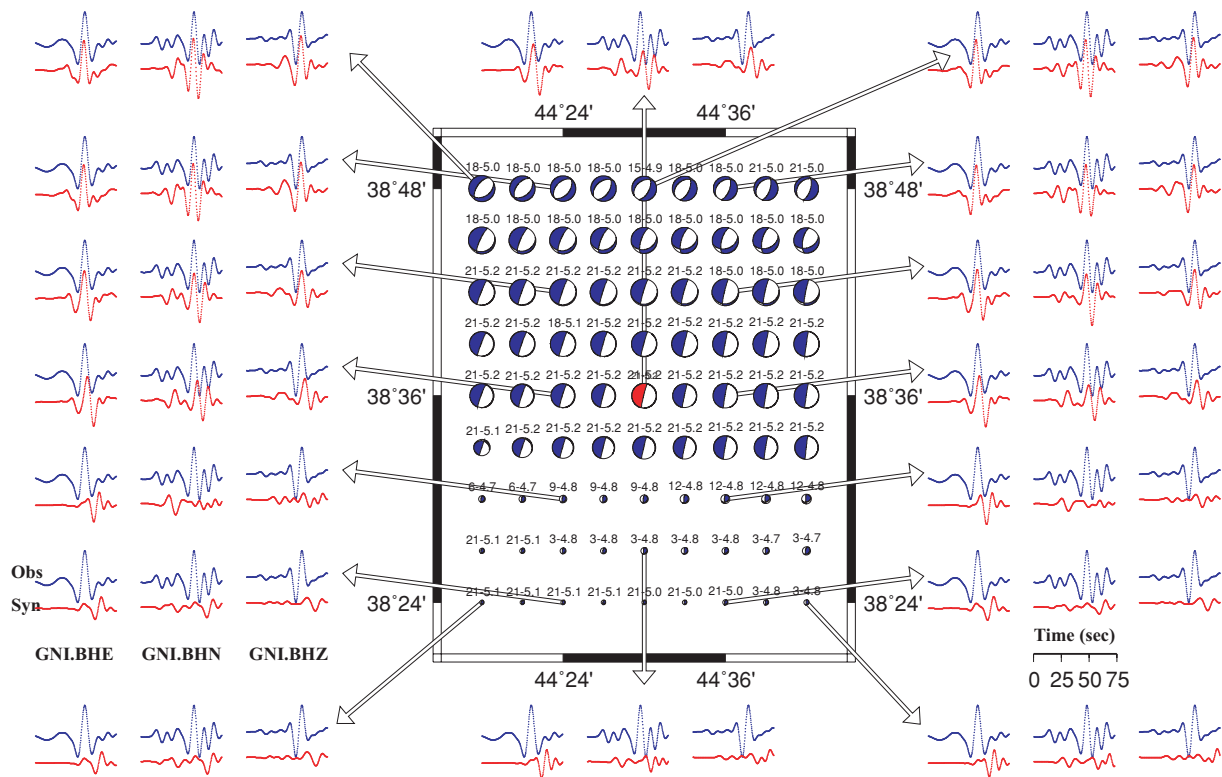


Figure 8. Result of CMT inversion for the 1999 February 19 earthquake ($m_b = 4.5$) using the broad-band records at the station GNI. The inversion was performed using a 3-D grid scheme at equally spaced nodes along x , y and z axes, where the x , y and z axes are set to longitude, latitude and depth, respectively. The focal mechanisms show the best solution at z for a grid node located at (x, y) . The values above the beach balls indicate the depth and moment magnitude obtained for the best CMT solution. The observed (upper) and calculated (lower) seismograms are shown for some grid points. The size of the beach balls is proportional to the variance reduction (VR) estimated in the inversion process. GNL.BHZ, GNL.BHN and GNL.BHE denote vertical, North–South and East–West components, respectively.

Lake Van and the slip vectors of the focal mechanisms (Fig. 7), we next try to determine the fault planes of the main shock and aftershocks.

The strike, dip and rake angles of the first subevents of the main shock are (87° , 62° , 110°) and (229° , 34° , 57°) for the first and second nodal planes, respectively. Hereafter, we call the first and second nodal planes NP1 and NP2, respectively. Following the notations of Aki & Richards (1980), the slip vectors on the two nodal planes are projected on the horizontal plane. For both NP1 and NP2, the azimuths of the projected slip vectors are nearly N320°E. However, the slip on NP1 is thrusting with a right-lateral strike-slip component, whereas the slip on NP2 is thrusting with a left-lateral strike-slip component. The tectonic deformation model proposed by Jackson (1992) suggests that the fault motion in the Lake Van region has a right-lateral strike-slip component. Therefore, the NP1 could be the fault plane.

We derive a similar conclusion from differential GPS vectors. In addition to the GPS vectors of McClusky *et al.* (2000) in Lake Van region, two more GPS vectors are available from Fig. 3 in Reilinger *et al.* (2006) (Fig. 7), so we have the three GPS vectors to the northeast, east and west of the lake. The three vectors are almost parallel, indicating uniform NW motion of the Lake Van region relative to the Eurasia. On the other hand, the relative motions between the GPS vectors at PATN and KAL2 (Table 1 in McClusky *et al.* 2000) and between the two GPS vectors parallel to KAL2 are towards N323E and N143E, respectively. Both directions are very close to the slip vector direction of NP1. Consequently, the differential GPS data point out that the NP1 is the fault plane. The

azimuth of the slip vector on NP2 is N177E, which is 34° deviated from the direction of the differential GPS vectors.

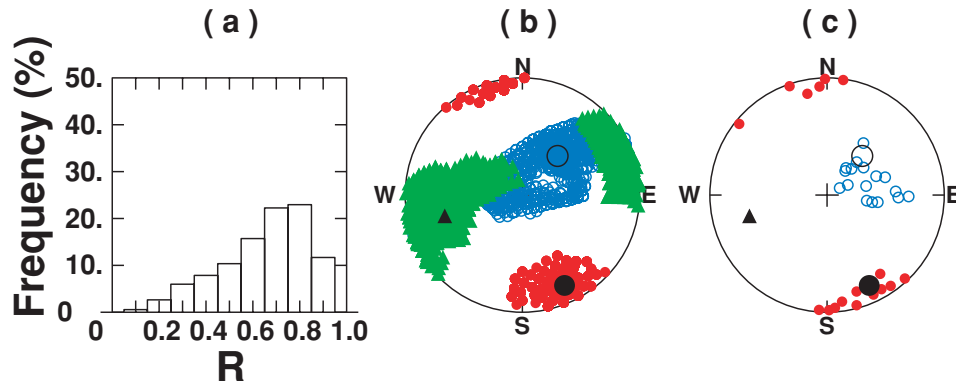
The strike, dip and rake angles of the second subevent are (76° , 55° , 120°) and (211° , 45° , 54°) for the two nodal planes, which are slightly different from those of the first subevent. The difference between the focal mechanisms of the two subevents could be partially due to errors in the waveforms because the onset of the second subevent is in the coda of the first subevent (Fig. 2a). For the second subevent, the azimuths of the slip vectors on the two nodal planes are N300°E and N166°E. Both slip vectors are similarly deviated from the differential GPS vectors, so it was impossible to identify the fault plane.

As for the aftershock source parameters, based on comparison between the differential GPS vectors and the slip vectors on the nodal planes, we found that nine aftershocks (event numbers 3, 4, 6, 7 and 9–13) are thrusting with a right-lateral strike-slip component. Discrepancies between the differential GPS vectors and the slip vectors projected on the horizontal plane are in a range from 0° to 7° (Fig. 7 and Table 2).

7.2 The local and regional stress tensors

We presented the results of the stress tensor inversion in Fig. 9. The measure of the goodness of the analysis, which is evaluated by the misfit values and the areas of 95 per cent confidence limit, is shown in Fig. 9(b). The number of the stress models within the 95 per cent confidence limit is 721, with the average misfit ranging between 3.4° and 6.0° . All can well represent the stress field in the region. The

Lake Van Region



Eastern Turkey - Sezgin and Pinar (2002)

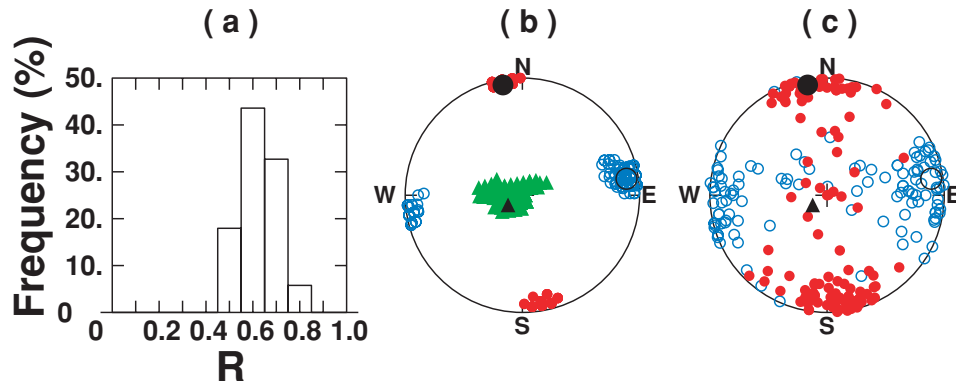


Figure 9. Results of the stress tensor analysis for the P - and T -axes of the focal mechanisms given in Table 1: (a) The histogram of R -value, (b) the distribution of the estimated principal stress axes and (c) the distribution of the observed P - and T -axes. In (b), solid circles, open circles and triangles indicate the azimuths and plunges of the maximum compression axis σ_1 , the minimum stress axis σ_3 and the intermediate stress axis σ_2 , respectively. In (c), solid and open circles indicate the P - and T -axes, respectively. Black symbols denote the principal stress axes for the best-fitting stress model. For the Lake Van region, the best fitting was attained for $R = 0.8$ and for the azimuth and plunge pair of $(335^\circ, 16^\circ)$ for σ_1 , $(255^\circ, 32^\circ)$ for σ_2 and $(42^\circ, 53^\circ)$ for σ_3 , respectively. The lower panels are the results of Sezgin & Pinar (2002) for Eastern Turkey based on focal mechanisms shown in Fig. 1 along with additionally published data.

721 theoretical stress models can be divided into two groups. One is a reverse faulting regime, and the other is a strike-slip faulting regime. The maximum compression σ_1 axis is nearly horizontal for both groups, but σ_2 and σ_3 axes are interchanged from horizontal to vertical directions. According to Anderson's theory of faulting, stress field is in a strike-slip faulting regime when the principal stress axis, σ_3 , is horizontal and σ_2 is vertical. When the principal stress axis, σ_3 , is vertical and σ_2 is horizontal, stress field is characterized by a reverse faulting type (Twiss & Moores 1992; pp. 203). Most of our stress models show horizontal σ_2 -axes and vertical σ_3 axes, which suggests that transpressive stress field dominates in the source region of the Lake Van earthquake. Only 10 out of the 721 stress models denote the strike-slip tectonic regime that the whole Eastern Turkey undergoes.

The value of R shown in Fig. 9(a) is another measure to characterize the stress field. Examples include $R \approx 0$ when $\sigma_1 \approx \sigma_2$ (biaxial deviatoric compression or state of confined extension), $R \approx 1$ when $\sigma_2 \approx \sigma_3$ (uniaxial deviatoric compression or state of confined com-

pression), and $R \approx 0.5$ when $\sigma_1 \approx \sigma_2 \approx \sigma_3$ (triaxial compression) (Christova & Tsapanos 2000). More detailed explanations on R are given by Bellier & Zoback (1995). In a strike-slip faulting stress regime where σ_1 and σ_3 are horizontal and σ_2 is vertical ($\sigma_1 = \sigma_{Hmax}$, $\sigma_3 = \sigma_{hmin}$, $\sigma_2 = \sigma_v$; σ_{Hmax} = maximum horizontal stress, σ_v = vertical stress and σ_{hmin} = minimum horizontal stress), the stress ratio $R \approx 1$ represents transition from strike-slip to reverse faulting stress regime (in which $\sigma_{hmin} \approx \sigma_v$). Accordingly, for Eastern Turkey, the stress magnitude ratio ($R = 0.6$) and the orientation of the principal stress axes suggest that the present-day active tectonics undergoes a predominantly strike-slip stress regime. For the Lake Van region, however, the large $R = 0.8$ implies transpression.

Finally, comparing the regional stress model in Eastern Turkey with the local stress model in the Lake Van region, we observe a significant discrepancy between the azimuths of the maximum compression in the two stress models (Fig. 9). The feature is also evident from the azimuths of the GPS vectors in Fig. 1, which are located along the north-south direction to the east of $41^\circ E$.

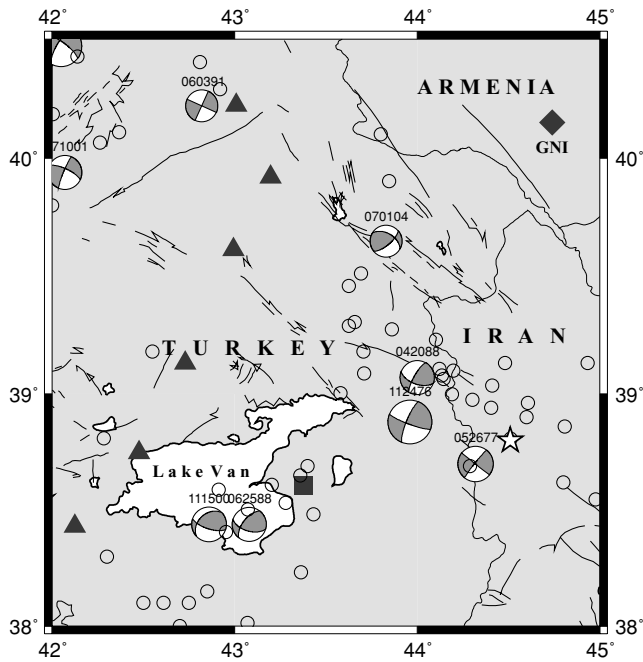


Figure 10. Subcrustal earthquake activities in the Lake Van and Turkey–Iran border regions. The white star is the epicentre of the 1999 February 19 earthquake analysed in this study. The Harvard CMT solutions of the large events are also shown. The numbers above the beach balls are the dates of the events in DDDMMYY (day, month, year) format. The solid triangles are the stations of the ETSE experiment, the diamond is the broad-band station GNI of GSN and the square is the broad-band station VAN of KOERI. The circles are the epicentres of the ISC events with hypocentral depths larger than 45 km since 1964.

7.3 Subcrustal seismic activity in Turkey–Iran border region

Some of the geodynamic models mentioned in the first section propose that the Eurasia–Arabia collision in Eastern Turkey results in subduction (Rotstein & Kafka 1982) or thickening of the lithosphere (Dewey *et al.* 1986). These models are primarily based on the subcrustal seismic activity reported by the international seismological agencies. Actually, the ISC and USGS earthquake catalogues include dozens of subcrustal events. Fig. 10 shows the ISC reviewed events ($M \geq 4$) with hypocentral depths larger than 45 km. They are considered as subcrustal events because the average thickness of the crust beneath Eastern Turkey is about 45 km and becomes thinner towards Iran (Zor *et al.* 2003). Although the KOERI seismic network covers most of the region, we do not illustrate the Turkish data because the ISC and USGS depth estimations are primarily based on pP depth phases and they include the phase arrivals from all the neighbouring countries including Turkey to constrain depth better.

The triangles in Fig. 10 show the easternmost stations of the ETSE experiment carried out in Eastern Anatolia. The main objective of the experiment was to test the several geodynamic models and determine the style of deformation in the region (Sandvol *et al.* 2003). An important result from the experiment was that no subcrustal earthquake activity occurs in Eastern Turkey.

Similarly, in this study we have revealed that the November 15, Lake Van earthquake and its aftershocks are also shallow crustal events, which conflicts with the USGS and ISC earthquake reports.

Thus, we suggest that the subcrustal activity around Lake Van should be handled with special care.

Moreover, subcrustal events were reported along a lineament extending to NW–SE in the Turkey–Iran border region by the ISC and USGS catalogues in a period from 1964 to 2000. Although broad-band stations were deployed in the ETSE experiment, unfortunately, none of the subcrustal events were reported in the region during the period from 1999 October 23 to 2001 August 9 when ETSE operated. So, we were not able to use the data to examine CMT depths for reported subcrustal events. The only permanent GSN (Global Seismic Network) broad-band seismograph that has been operated in the region since 1991 July is GNI (Garni, Armenia). For the 1999 February 19, 18:00 event ($m_b = 4.5$) with reported depth of 66 km (USGS) and 77 km (ISC 2001), we have inverted the three-component broad-band records from GNI, and the result provides an additional piece of evidence that seismic activity in the Turkey–Iran border region is shallow rather than subcrustal.

We have shown that the 2000 November 15 Lake Van earthquake was a shallow double event, and it is likely that the event was located unrealistically deep beneath the crust by ISC and USGS interpreting the phase of the second subevent as the pP depth phase. However, we have also shown that although the 1999 February 19 earthquake was a single shallow event, it was also identified as a subcrustal event by ISC and USGS. Therefore, further studies may shed light on the question why all the other ‘subcrustal’ events during the period between 1964 and 2000 were located at the depths.

8 CONCLUSIONS

(1) The teleseismic waveform modelling, the near-field data, and the aftershock distribution clearly depict that the 2000 November 15 Lake Van earthquake ($M_w = 5.6$) is not a subcrustal event as reported by ISC and USGS, but is a multiple event comprised of two subevents within the crust. Incorporation of the teleseismic, strong motion and aftershock data enabled to precisely constrain the focal depth, which is a crucial parameter for rapid earthquake impact assessment.

(2) The inversion results suggest that the first subevent is located beneath the epicentre of USGS but at the depth of 15 km, while the second one is located 7.5 km to the north of the epicentre at the depth of 12.5 km. The seismic moment of the second subevent ($M_0 = 1.6 \times 10^{17}$ Nm; strike 76° , dip 55° , rake 120°) is larger than the first subevent ($M_0 = 1.0 \times 10^{17}$ Nm; strike 87° , dip 62° , rake 110°). The total CMT parameters obtained by summing up the seismic moment tensor components of the two subevents are as follows: seismic moment $M_0 = 2.5 \times 10^{17}$ Nm ($M_w = 5.6$), strike $= 81^\circ$, dip $= 58^\circ$ and rake $= 117^\circ$.

(3) Using our focal mechanisms of the aftershocks and the two subevents of the main shock as well as the Harvard CMT solution of the June 25, 1988 earthquake ($M_w = 5.5$), our analysis of the stress tensor inversion for the source region of the Lake Van earthquake yields the azimuth and plunge pairs for the three principal stress axes, σ_1 , σ_2 and σ_3 , being 335, 16; 255, 32 and 42, 53, respectively. On the other hand, the regional stress field for the whole Eastern Turkey was obtained by Sezgin & Pinar (2002) providing the strike and plunge pairs of the principal stress axes (350, 0) for σ_1 , (260, 81) for σ_2 and (80, 9) for σ_3 . The difference between the two stress models suggests that the Eastern Anatolia rotates clockwise with respect to the Lake Van region.

(4) Besides the deep seismic activity in the Lake Van region, the ISC and USGS bulletins reported subcrustal events to the northeast

of the lake in the Turkey–Iran border region. An example is an event at 18:00 on 1999 February 19 for which ISC and USGS estimated the depth to be 66 and 77 km, respectively. However, our CMT analysis of the three-component broad-band data suggests that the hypocentral depth (18 km) is much shallower and the event occurred within the crust.

ACKNOWLEDGMENTS

We greatly appreciate the contribution of all the people maintaining the *IRIS Global Seismographic Network*, the *ETSE PASCALL Network*, the *IRIS Data Management System at Seattle, Washington*, the *International Seismological Center at Thatcham*, United Kingdom and the United States Geological Survey. The first author (AP) was a Japan Society for the Promotion of Science (JSPS) fellow when this study was taken up. The comments raised by two anonymous referees improved considerably the early version of the manuscript. This work was supported by Research Fund of the University of Istanbul, project number UDP-693/08032007.

REFERENCES

- Aki, K. & Richards, P., 1980. *Quantitative Seismology*, W.H. Freeman & Co., NY.
- Aydar, E. et al., 2003. Morphological analysis of active Mount Nemrut stratovolcano, eastern Turkey: evidences and possible impact areas of future eruption, *J. Volcanol. Geotherm. Res.*, **123**(3–4), 301–312.
- Barka, A.A. & Kadinsky-Cade, K., 1988. Strike-slip fault geometry in Turkey and its influence on earthquake activity, *Tectonics*, **3**, 663–684.
- Bellier, O. & Zoback, M., 1995. Recent state of stress change in the Walker Lane zone western basin and Range Province-USA, *Tectonics*, **14**, 564–593.
- Bozkurt, E., 2001. Neotectonics of Turkey—a synthesis, *Geodinamica Acta*, **14**, 3–30.
- Christova, C. & Tsapanos, T., 2000. Depth distribution of stresses in the Hokkaido Wadati-Benioff zone as deduced by inversion of earthquake focal mechanisms, *J. Geodyn.*, **30**, 557–573.
- Delouis, B. & Legrand, D., 1999. Focal mechanism determination and identification of the fault plane of earthquakes using only one or two near-source seismic recordings, *Bull. Seism. Soc. Am.*, **89**, 1558–1574.
- Dewey, J.F., Hempton, M.R., Kidd, W.S.F., Şaroğlu, F. & Şengör, A.M.C., 1986. Shortening of continental lithosphere: the neotectonics of Eastern Anatolia—a young collision zone, in *Collision Tectonics*, 19 (R. M. Scheckleton volume), pp. 3–36, eds. M.P. Coward & A.C. Ries, Geol. Soc. London Spec. Pub.
- Dreger, D. & Helmberger, D., 1993. Determination of source parameters at regional distances with single station or sparse network data, *Bull. Seism. Soc. Am.*, **98**, 8107–8125.
- Dreger, D. & Savage, B., 1999. Aftershocks of the 1952 Kern County, California, Earthquake Sequence, *Bull. Seism. Soc. Am.*, **89**, 1094–1108.
- Fan, G. & Wallace, T., 1991. The determination of source parameters for small earthquakes from a single, very broadband seismic station, *Geophys. Res. Lett.*, **18**, 1385–1388.
- Gephart, J.W. & Forsyth, D.W., 1984. An improved method for determining the regional stress tensor using earthquake focal mechanism data: application to the San Fernando earthquake sequence, *J. Geophys. Res.*, **89**, 9305–9320.
- Gephart, J.W., 1985. Principal stress directions and the ambiguity in fault plane identification from focal mechanisms, *Bull. Seism. Soc. Am.*, **75**, 621–625.
- Gephart, J.W., 1990. FMSI: a Fortran program for inverting fault/slickenside and earthquake focal mechanism data to obtain the regional stress tensor, *Comp. Geosci.*, **16**, 953–989.
- International Seismological Centre, 2001. On-line Bulletin, <http://www.isc.ac.uk/Bull>, Internatl. Seis. Cent., Thatcham, United Kingdom.
- Jackson, J., 1992. Partitioning of strike-slip and convergent motion between Eurasia and Arabia in Eastern Turkey and the Caucasus, **97**(B9), 12 471–12 479.
- Jackson, J.A. & McKenzie, D.P., 1988. The relationship between plate motions and seismic moment tensors, and the rates of active deformation in the Mediterranean and Middle East, *Geophys. J.*, **93**, 45–73.
- Keskin, M., 2003. Magma generated by slab steepening and breakoff beneath a subduction-accretion complex: an alternative model for collision-related volcanism in Eastern Anatolia, Turkey, *Geophys. Res. Lett.*, **30**(24), 8046, doi:10.1029/2003GL018019.
- Keskin, M., Pearce, J.A. & Mitchell, J.G., 1998. Volcano-stratigraphy and geo-chemistry of collision-related volcanism on the Erzurum-Kars Plateau, North Eastern Turkey, *J. Volcanol. Geotherm. Res.*, **85**, 355–404.
- Kikuchi, M. & Kanamori, H., 1991. Inversion of complex body waves-III, *Bull. Seism. Soc. Am.*, **81**, 2335–2350.
- Kim, S.G. & Kraeva, N., 1999. Source parameter determination of local earthquakes in Korea using moment tensor inversion of single station data, *Bull. Seism. Soc. Am.*, **89**, 1077–1082.
- Koçyiğit, A., Yılmaz, A., Adamia, S. & Kuloshvili, S., 2001. Neotectonics of East Anatolian Plateau (Turkey) and Lesser Caucasus: implication for transition from thrusting to strike-slip faulting, *Geodinamica Acta*, **14**, 177–195.
- Kohketsu, K., 1985. The extended reflectivity method for synthetic near-field seismograms, *J. Phys. Earth*, **33**, 121–131.
- Kuge, K., 1999. Automated determination of earthquake source parameters using broadband strong-motion waveform data, *EOS Trans.*, **80**, F661.
- Kuge, K., 2003. Source modeling using strong-motion waveforms: toward automated determination of earthquake fault planes and moment-release distributions, *Bull. Seismo. Soc. Am.*, **93**, 639–654.
- Legrand, D. & Delouis, B., 1999. Determination of the fault plane using a single near-field seismic station with a finite-dimension source model, *Geophys. J. Int.*, **138**, 801–808.
- Legrand, D., Kaneshima, S. & Kawakatsu, H., 2000. Moment tensor analysis of near-field broadband waveforms observed at Aso Volcano, Japan, *J. Volc. Geoth. Res.*, **101**, 155–169.
- McClusky, S. et al., 2000. Global Positioning System constraints in plate kinematics and dynamics in the eastern Mediterranean and Caucasus, *J. Geophys. Res.*, **105**, 5695–5719.
- McKenzie, D.P., 1972. Active tectonics of the Mediterranean region, *Geophys. J. R. Astr. Soc.*, **30**(2), 109–185.
- Örgülü, G., Aktar, M., Türkelli, N., Sandvol, E. & Barazangi, M., 2003. Contribution to the seismotectonics of Eastern Turkey from moderate and small size events, *Geophys. Res. Lett.*, **30**(24), 8040.
- Pinar, A., Honkura, Y. & Kuge, K., 2003. Moment tensor inversion of recent small to moderate sized earthquakes: implication for seismic hazard and active tectonics beneath the Sea of Marmara, *Geophys. J. Int.*, **153**, 133–145.
- Reilinger, R. et al., 2006. GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions, *J. Geophys. Res.*, **111**, B05411, doi:10.1029/2005JB004051.
- Rotstein, Y. & Kafka, A.L., 1982. Seismotectonics of the southern boundary of Anatolia, Eastern Mediterranean region: subduction, collision, and arc jumping, *J. Geophys. Res.*, **87**, 7694–7706.
- Sandvol, E., Türkelli, N. & Barazangi, M., 2003. The Eastern Turkey Seismic Experiment: the study of a young continent-continent collision, *Geophys. Res. Lett.*, **30**(24), 8038, doi:10.1029/2003GL018912.
- Şaroğlu, F., Emre, Ö. & Boray, A., 1987. The active faults and seismicity of Turkey, Min. Res. Expl. Inst., Turkey Report no. 8174.
- Şengör, A.M.C. & Kidd, W.S.F., 1979. Post-collisional tectonics of the Turkish-Iranian Plateau and a comparison with Tibet, *Tectonophysics*, **55**, 361–376.

- Şengör, A.M.C. & Yılmaz, Y., 1981. Tethyan evolution of Turkey: a plate tectonic approach, *Tectonophysics*, **75**, 181–241.
- 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.
- Sezgin, N. & Pinar, A., 2002. Present day active tectonics of Eastern Turkey), *The Tectonics of Eastern Turkey and the Northern Arabian Plate*, p. 38, Erzurum.
- Singh, S.K., Pacheco, J., Courboux, F. & Novelo, D.A., 1997. Source parameters of the Pinotepa Nacional, Mexico, earthquake of 27 March, 1996 ($M_w = 5.4$) estimated from near-field recordings of a single station, *J. Seism.*, **1**, 39–45.
- Toksöz, M.N., Arpat, E. & Şaroğlu, F., 1997. East Anatolian earthquake of 24.11.1976, *Nature*, **270**, 423–425.
- Türkelli, N. *et al.*, 2003. Seismogenic zones in Eastern Turkey, *Geophys. Res. Lett.*, **30**(24), 8039, doi:10.1029/2003GL018023.
- Twiss, R.J. & Moores, E.M., 1992. *Structural Geology*, W.H. Freeman and Co., New York.
- Utkucu, M., 2006. Implications for the water level change triggered moderate ($M \geq 4.0$) earthquakes in Lake Van basin, Eastern Turkey, *J. Seismol.*, **10**, 105–117.
- Zor, E., Sandvol, E., Gürbüz, C., Türkelli, N., Seber, D. & Barazangi, M., 2003. The crustal structure of the East Anatolian plateau (Turkey) from receiver functions, *Geophys. Res. Lett.*, **30**(24), 8044, doi: 10.1029/2003GL018192.