The 2012 August 11 $M_W$ 6.5, 6.4 Ahar-Varzghan earthquakes, NW Iran: aftershock sequence analysis and evidence for activity migration

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SUMMARY
The Ahar-Varzghan doublet earthquakes with magnitudes $M_W$ 6.5 and 6.4 occurred on 2012 August 11 in northwest Iran and were followed by many aftershocks. In this paper, we analyse ~5 months of aftershocks of these events. The Ahar-Varzghan earthquakes occurred along complex faults and provide a new constraint on the earthquake hazard in northwest Iran. The general pattern of relocated aftershocks defines a complex seismic zone covering an area of approximately 25 × 10 km$^2$. The Ahar-Varzghan aftershock sequence shows a secondary activity which started on November 7, approximately 3 months after the main shocks, with a significant increase in activity, regarding both number of events and their magnitude. This stage was characterized by a seismic zone that propagated to the west of the main shocks. The catalogue of aftershocks for the doublet earthquake has a magnitude completeness of $M_c$ 2.0. A below-average $b$-value for the Ahar-Varzghan sequence indicates a structural heterogeneity in the fault plane and the compressive stress state of the region. Relocated aftershocks occupy a broad zone clustering east–west with near-vertical dip which we interpret as the fault plane of the first of the doublet main shocks ($M_W$ 6.5). The dominant depth range of the aftershocks is from 3 to about 20 km, and the focal depths decrease toward the western part of the fault. The aftershock activity has its highest concentration in the eastern and middle parts of the active fault, and tapers off toward the western part of the active fault segment, indicating mainly a unilateral rupture toward west.

Key words: Earthquake source observations; Seismicity and tectonics; Asia.

1 INTRODUCTION
The first of the doublet main shocks ($M_W$ 6.5) occurred at 12:23 UTC, on 2012 August 11, in northwestern Iran. No foreshock preceded this earthquake. Only 11 min later, another moderate-sized earthquake ($M_W$ 6.4) occurred at 12:34 UTC. The epicentre of the first earthquake was located approximately 17 km southeast of Varzghan and ~20 km southwest of Ahar. The epicentre of the second earthquake is located ~6 km to the west. These two main shocks were followed by three events of magnitude ≥5 and thousands of small aftershocks. For the first main shock a maximum acceleration of 478 cm s$^{-2}$ on the longitudinal component was recorded at the SAT3 station, located at the crest of the Sartarkan-Dam. For the second main shock, a maximum acceleration of 532 cm s$^{-2}$ on the longitudinal component was recorded at the VAZ station in Varzghan. The reported epicentres for these earthquakes are located between the cities of Ahar, Varzghan and Heris in East Azerbaijan province (Fig. 1). They have been variously referred to as the East Azarbaijan, Arasbaran, Ahar, Varzghan, Heris and Ahar-Varzghan events. About 46 villages were completely destroyed by the earthquakes and ~365 villages were seriously damaged. More than 330 people were killed; ~26 000 injured and more than 50 000 people had to be resettled (Miyajima et al. 2012). At the time of the Ahar-Varzghan earthquakes the majority of people were working in the field, and this reduced the number of casualties. The tectonics of the region is dominated by convergence between the Arabian and Eurasian plates. The convergence vector is trending north to north–northeast at a velocity ranging from 23 to 25 mm yr$^{-1}$ (Berberian 1976; Jackson & McKenzie 1984; Walker & Jackson 2002; McClusky et al. 2003; Vernant et al. 2004) to as much as 35 mm yr$^{-1}$ according to the NUVEL-1 model (DeMets et al. 1990). The region referred to as northwest Iran is situated between two thrust belts: the Caucasus to the north and the Zagros Mountain belt to the south (Hessami et al. 2003b). The overall shortening due to the Arabia–Eurasia collision in the region is, in general, spatially separated into right-lateral slip in the Turkish–Iranian Plateau and thrusting in the greater Caucasus (Copley & Jackson 2006). The North Tabriz Fault with northwest–southeast trend and ~60 km from the source region of Ahar-Varzghan earthquakes is the main tectonic feature in the area (Fig. 1). Deformation in northwest Iran is characterized by ~8 mm yr$^{-1}$ of right-lateral movement on the North Tabriz Fault, in agreement with a recurrence time interval of 250–300 yr suggested by historical seismicity studies (Masson et al. 2006). Based on GPS measurements, Vernant
et al. (2004) suggested \( \sim 14 \text{ mm yr}^{-1} \) of north–south shortening between the Caucasus and North Zagros Mountains. Mountain belts in this area are a result of the closure of the Neotethys Ocean, and final collision of the Arabian plate with Central Iran block. Geological evidences and fault plane solutions of earthquakes indicate the existence of both thrust and conjugate strike-slip faulting in this region (Jackson 1992; Talebian & Jackson 2002). Most of the earthquakes in this region are shallow (Siahkali Moradi et al. 2011) and commonly associated with surface faulting (Berberian & Yeats 1999; Hessami et al. 2003b; Karakhanian et al. 2004). Two prominent Neogene-Quaternary volcanoes Sahand (calc-alkaline stratovolcano) and Sabalan (andesitic volcano) are located in this region.

As Fig. 1 shows, there was no significant seismicity close to the epicentral area of the Ahar-Varzghan earthquakes. The epicentres of historical and instrumental events are mostly located to the west and east of the 2012 August 11 earthquakes. The nearest large instrumental earthquake to the Ahar-Varzghan earthquakes in the west was a magnitude \( M_b \geq 6.0 \) event, which occurred southeast of Sabalan-Mountain near Ardabil on 1997 February 28. The Fault that ruptured in the 2012 August 11 earthquakes was previously unknown, illustrating the difficulty of seismic hazard estimation based on incomplete tectonic maps. The Iranian plateau, including the study area, part of a structurally complex and inhomogeneous collision region in which seismicity is not the result of the activity of a few faults but it is due to fault activity in zone hundred kilometres wide (Hessami et al. 2003a).

Table 1 compares the different hypocentre parameters associated with the Ahar-Varzghan main shocks reported by different seismological agencies and authors. Reported locations for the Ahar-Varzghan events by different seismological agencies/authors are rather different (Fig. 2). The two earthquakes are separated by \( \sim 5 \text{ km} \) in an east–northwest direction based on the Global Centroid Moment Tensor (GCMT) solutions. The values in Table 1 are the latest updated values. Most of agencies and authors located the second main shock to the west of the first event (Fig. 2).

All focal mechanisms from different agencies/authors are very similar for both earthquakes, respectively. Focal mechanisms for the first event suggest slip on a fault plane striking roughly east–west or north–south. In Fig. 2, the solid and dot-dashed lines show the Ahar Fault trace that was extracted from two different scale maps. Their locations are different but both of them have a strike nearly in east–west direction.

For the first earthquake the GCMT solution indicates a mostly right-lateral strike-slip motion on a fault plane dipping at 84° to the south and a focal depth of 15 km. The U.S. Geological Survey fault plane solution was very similar to the CMT solution but preferred a vertical dip angle with a focal depth of 11 km. A waveform inversion study by Institute of Geophysics, University of Tehran (IGUT) determined a similar mechanism to the GCMT solution but dipping...
Table 1. Hypocentre and focal mechanism parameters of the Ahar-Varzghan earthquakes were determined by different seismological agencies and authors.

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>Agencies &amp; authors</th>
<th>Origin time: hh:mm:ss.s</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>Depth (km)</th>
<th>Magnitude</th>
<th>Strike/Dip/Rake (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>First</td>
<td>IGUT</td>
<td>12:23:15</td>
<td>38.393</td>
<td>46.806</td>
<td>9</td>
<td>$M_w$ 6.5</td>
<td>267/−175 &amp; 176/85/−9</td>
</tr>
<tr>
<td></td>
<td>IIEES</td>
<td>12:23:16.2</td>
<td>38.55</td>
<td>46.87</td>
<td>15.4</td>
<td>$M_b$ 6.1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>BHRC</td>
<td></td>
<td>38.52</td>
<td>46.86</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>USGS</td>
<td>12:23:18.2</td>
<td>38.329</td>
<td>46.826</td>
<td>11</td>
<td>$M_w$ 6.4</td>
<td>85/−178 &amp; 355/88/−1</td>
</tr>
<tr>
<td></td>
<td>GCMT</td>
<td>12:23:18.2</td>
<td>38.31</td>
<td>46.80</td>
<td>15</td>
<td>$M_w$ 6.5</td>
<td>84/170 &amp; 175/16/81</td>
</tr>
<tr>
<td></td>
<td>EMSC</td>
<td>12:23:18</td>
<td>38.41</td>
<td>46.81</td>
<td>10</td>
<td>$M_w$ 6.4</td>
<td></td>
</tr>
<tr>
<td></td>
<td>C. et al.</td>
<td>12:23</td>
<td></td>
<td></td>
<td></td>
<td>$M_w$ 6.4</td>
<td>265/0/175 &amp; 355/85/0</td>
</tr>
<tr>
<td></td>
<td>D. et al.</td>
<td>12:23:16</td>
<td>38.399</td>
<td>46.842</td>
<td>6</td>
<td>$M_w$ 6.4</td>
<td>265/166</td>
</tr>
<tr>
<td></td>
<td>M. &amp; R.</td>
<td>12:23:15</td>
<td>38.43</td>
<td>46.82</td>
<td>10.6</td>
<td>$M_w$ 6.4</td>
<td>85/165 &amp; 175/75/1</td>
</tr>
<tr>
<td></td>
<td>This study</td>
<td>12:23:14.8</td>
<td>38.436</td>
<td>46.838</td>
<td>16.4</td>
<td>$M_w$ 6.5</td>
<td></td>
</tr>
<tr>
<td>Second</td>
<td>IGUT</td>
<td>12:34:35.9</td>
<td>38.394</td>
<td>46.814</td>
<td>4</td>
<td>$M_w$ 6.3</td>
<td>7/57/21 &amp; 265/72/146</td>
</tr>
<tr>
<td></td>
<td>IIEES</td>
<td>12:34:35</td>
<td>38.58</td>
<td>46.78</td>
<td>16</td>
<td>$M_b$ 6.1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>BHRC</td>
<td>38.45</td>
<td>46.75</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>USGS</td>
<td>12:34:35.9</td>
<td>38.389</td>
<td>46.745</td>
<td>12</td>
<td>$M_w$ 6.2</td>
<td>19/47/13 &amp; 256/60/128</td>
</tr>
<tr>
<td></td>
<td>GCMT</td>
<td>12:34:35.6</td>
<td>38.35</td>
<td>46.78</td>
<td>19.2</td>
<td>$M_w$ 6.4</td>
<td>10/50/36 &amp; 255/63/134</td>
</tr>
<tr>
<td></td>
<td>EMSC</td>
<td>12:34:36</td>
<td>38.48</td>
<td>46.75</td>
<td>10</td>
<td>$M_w$ 6.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>D. et al.</td>
<td>12:34:35</td>
<td>38.425</td>
<td>46.777</td>
<td>12</td>
<td>$M_w$ 6.2</td>
<td>267/61/125</td>
</tr>
<tr>
<td></td>
<td>M. &amp; R.</td>
<td>12:34:35.4</td>
<td>38.422</td>
<td>46.735</td>
<td>13.2</td>
<td>$M_w$ 6.2</td>
<td>85/165 &amp; 175/75/1</td>
</tr>
<tr>
<td></td>
<td>This study</td>
<td>12:34:33.6</td>
<td>38.416</td>
<td>46.815</td>
<td>15.4</td>
<td>$M_w$ 6.3</td>
<td></td>
</tr>
</tbody>
</table>

*IGUT, Institute of Geophysics, University of Tehran; IIEES, International Institute of Earthquake Engineering and Seismology; BHRC, Building and Housing Research Center; USGS, U.S. Geological Survey; GCMT, Global Centroid Moment Tensor; EMSC, European-Mediterranean Seismological Center; C. et al., Copley et al. (2014); D. et al., Donner et al. (2015); M.&R., Mirdamadi & Rezapour (2015).*  

Figure 2. Epicentres determined by different agencies and authors for the first and second Ahar-Varzghan earthquakes are shown by stars and circles, respectively. The focal mechanisms of the earthquakes are also indicated. The location of cities is shown by filled squares. The solid lines show the trace of Ahar Faults, extracted from the Ahar Quadrangle Map with scale of 1:250 000 (Amidi et al. 1978). The dot-dashed line shows the trace of the Ahar Fault which was extracted from the Seismotectonic Map of Iran with a scale of 1:2 500 000 (Berberian 1976). The dashed line presents the ~12 km surface rupture which was observed by Faridi & Sartipi (2012). The inverted triangle shows the accelerograph stations which belong to BHRC.

to the north. The focal depth of the first earthquake was estimated at 9 km by the IGUT.

Copley et al. (2014) jointly inverted $P$ and $SH$ waveforms to obtain the focal parameters of the first earthquake. Their solution is dominantly strike-slip with a centroid depth of less than 10 km. Donner et al. (2015) determined a strike slip mechanism for the first earthquake by inverting regional surface waves to obtain the moment tensor. Donner et al. (2015) obtained 6 ± 1 km for the centroid depth. Also, Mirdamadi & Rezapour (2015) applied the ISOLA, ISOLated Asperities software package (Sokos & Zahradnik 2008) to local and regional waveforms to perform moment tensor inversion. They determined a pure strike-slip mechanism with a focal depth of 13.2 km for the first earthquake.

For the second earthquake, different agencies and authors determined an oblique-thrusting mechanism oriented on northeast–southwest or northwest–southeast nodal planes (Fig. 2 and Table 1).

Two moderate size earthquakes separated by ~11 min in time occurred on an unmapped fault and thousands aftershocks followed. Precise identification of the causative fault(s) of the Ahar-Varzghan
earthquakes can be facilitated by extra information such as aftershocks distribution.

Our prime interest in this study is to unravel the details of the main shock–aftershock sequence, aftershock migration and identify the causative fault(s). Our approach to this problem is to relocate as many events as possible in the sequence by gathering available phase arrival data from seismic stations in Iran and neighbour countries.

2 CATALOGUE DATA (2012 AUGUST 11 TO DECEMBER 31)

In the IGUT bulletin more than 3741 events have been reported in 143 d after the main shock. After a short introduction to the seismic network in Iran, the temporal distribution and magnitude completeness in Ahar-Varzghan sequence is analysed.

2.1 Seismic network and data acquisition

Modern digital seismic networks in Iran began in 1996 and consist of a series of local networks which were deployed by IGUT. At the present time IGUT is running 19 local networks with about 111 stations in the country. Four types of seismometers were used in this network: Kinematics SS-1 sensor at 46 stations, Nanometrics Trillium-40s at 38 stations, 25 stations with Guralp CMG-3ESP-120s sensors and two stations with Guralp CMG-3T-360s seismometers. Data acquisition is performed by Data Analysis Program (DAN) which has been produced by Nanometrics. A seismic network of about two dozen broad-band stations (Guralp 3ESP-120s and CMG-3ESP-100s sensors) operated by the International Institute of Earthquake Engineering and Seismology (IIEES) which were joined to the national seismic network in the early 2000s. Other organizations have also added seismic stations for specialized monitoring projects.

During ~5 months after the first main shock more than 3741 events have been recorded by IGUT. The Tabriz seismic network was especially important in recording the Ahar-Varzghan sequence.

2.2 Aftershock decay

The relationship between main shocks and their aftershocks has been a topic of study for decades (Utsu 1961; Scholz 1968; Helmstetter & Shaw 2006; etc.). Fig. 3(a) shows that in the Ahar-Varzghan aftershock sequence, the seismic activity could be divided into two periods. In the first period (August 11 to November 7) more than 3308 aftershocks were located by the IGUT. The greatest activity was during the early days. The second stage started on November 7 after an \( M_N > 5 \) event approximately 3 month after the main shock, with a significant increase in activity, regarding both events and their magnitude.

Aftershocks occur with a pattern that follows Omori’s law. The modified Omori’s law is an empirical relation for the temporal decay of aftershock rates; its most popular form is

\[
N(t) = B + K/(t+c)^p + K_2/(t-T_2+c_2)^p \quad (K_2=0 \text{ for } t<T_2)
\]

where \( N(t) \) is the rate of earthquakes measured in a certain time \( t \) after the main shock, \( K \) is a magnitude-dependent productivity constant and \( c \) reflects earthquake detection limits and also prevents the Modified Omori Law from ‘blowing up’ at short times after an event, \( p \) is the log-log slope of the power law decay and typically falls in the range 0.7–1.5 (Utsu 1961). This empirical law may be used directly to invert \( c \) and \( p \) values (Kisslinger & Jones 1991) or recursively applied to all events in an epidemic type aftershock sequence (Ogata 1988, 1999). In order to model the aftershock activity of Ahar-Varzghan aftershocks, we use a modified Omori model with a second aftershock sequence starting at \( T_2 \) (Utsu et al. 1995):

\[
N(t) = B + K/(t+c)^p + K_2/(t-T_2+c_2)^p \quad (K_2=0 \text{ for } t<T_2)
\]

(1)
This model was fitted to the data by maximum likelihood estimation (Akaike, 1974); it is plotted as solid curve in Fig. 3(b). For the main sequence the fitted parameter estimates are $K = 1162.00547$ events d$^{-1}$, $c = 3.33845$ d$^{-1}$ and $p = 1.09013$. For the second sequence starting at $t = 87.75227$, the estimated values are $K_2 = 892.47771$, $c_2 = 1.92022$ and $p_2 = 2.06779$, $B$ indicates the level of background activity, which was constrained to 0.

2.3 Magnitude determination and its completeness

Among the several magnitude scales in current use, the local magnitude $M_L$ has the most direct relevance to engineering applications because $M_L$ is determined within the period range of greatest engineering interest. The IEES determines $M_L$ values for recorded events in Iran by using the empirical distance-correction term $-\log A_0$ of Hutton & Boore (1987), which was calibrated for southern California region. Rezapour & Rezaei (2011) introduced an $M_c$ scale to determine magnitude of events in NW Iran and showed that the attenuation in NW Iran is more rapid than Hutton & Boore (1987) curve.

To determine the magnitude of recorded events, the IGUT uses the modified mbLG scale (Rezapour 2005). The mbLG scale is the lowest magnitude value where the N denotes Nuttli. The International Seismological Center (ISC) and National Earthquake Information Center (NEIC) catalogues are the most common earthquake catalogues used in Iran for seismic hazard analysis. To keep uniformity of magnitude scale when the ISC/NEIC and IGUT catalogues are jointly used for seismic hazard analysis, Rezapour (2005) calibrated the $M_c$ formulae with mb(ISC/NEIC) values.

The completeness magnitude $M_c$ is defined as the lowest magnitude at which 100 per cent of the earthquakes in a space-time volume are detected (Rydelek & Sacks 1989). $M_c$ is often estimated by fitting a Gutenberg–Richter (G-R) model to the observed Frequency-Magnitude Distribution (FMD). The magnitude at which the lower end of the FMD departs from the linear trend in the log-line plot (the G-R law) is taken as an estimate of $M_c$ (Zuniga & Wyss 1995). Within the first hours to days of an aftershock sequence, the frequent larger aftershocks eclipse smaller events and this causes a large estimated value for $M_c$. The improvement in seismic stations by installing temporary seismic network causes a decrease in the $M_c$ value. Normally $M_c$ for aftershock sequences has temporal variability and a systematic tendency for $M_c$ to decrease is observed. The Ahar-Varzghan database shows this tendency. In the Ahar-Varzghan sequence during ~5 months about 3741 events were located by IGUT. To determine the $M_c$ value in the Ahar-Varzghan sequence, in addition to considering the variation of $\Delta \log (N)/\Delta (M_c)$ with time (Fig. 4a), we used the procedure derived by Shi & Bolt (1982) as implemented by Weimer (2001) in the software Zmap under Matlab. We determined a $M_c$ value of 2.0 with 95 per cent probability for Ahar-Varzghan sequence.

The $b$-value is estimated by the maximum likelihood method (Aki 1965). For the data set used here, a $b$-value of $0.737 \pm 0.02$ was found which, is less than the normal value of 1.0 (Fig. 4b). The $b$-value, however, varies from 0.5 to 1.5 depending on tectonic setting, tectonic stress, magnitude range, etc. (Scholz 1990; Wiemer & Wyss 1997). Many factors can cause perturbations of the normal $b$-value. The $b$-value for a region reflects the relative proportion of the number of large and small earthquakes, also it relates to the stress condition over the region. A lower $b$-value and the higher stress after a main shock are proposed out by many researchers (e.g. Scholz 1990; Frohlich & Davis 1993; Wiemer & Wyss 1997). A lower $b$-value in Ahar-Varzghan sequence indicates a structural heterogeneity in the fault plane with possible variations in frictional conditions along the fault.

3 RELOCATED DATA

The IGUT located the Ahar-Varzghan aftershocks using the data recorded by the national seismic network. Because of an inadequate regional 1-D velocity structure and very poor coverage by permanent seismic stations in the study area, the uncertainty of the aftershock locations in catalogue data could be up to 20 km.
Seismic station coverage for the Ahar-Varzghan region is poor. Most of aftershocks were located with a large azimuthal gap. To identify the responsible fault for the Ahar-Varzghan earthquakes we need well located aftershocks. First a velocity model for the region is obtained, and then the aftershocks are relocated by applying single-event and multi-event approach using all available data.

3.1 Crustal velocity

The velocity model employed in the routine location procedure for events in Iran by the IGUT contains two layers: a 36-km thick layer \( (V_p = 6.2 \text{ km s}^{-1}; V_S = 3.57 \text{ km s}^{-1}) \) over a half-space \( (V_p = 8.2 \text{ km s}^{-1}; V_S = 4.7 \text{ km s}^{-1}) \) (Mottaghi et al. 2010). It is well known that an inadequate velocity structure creates large uncertainty in locations. Bayramnajad et al. (2008) used the program VELEST (Kissling 1988) to invert arrival times of events with magnitude greater than 4.0 recorded by Tabriz seismic network and suggested a 1-D velocity structure for northwest Iran. Siahkali Moradi (2008) used recorded data by a dense temporary seismographic network around Tabriz and determined a three layer crustal structure for the region. Also, Aghaei (2013) inverted arrival times of the Ahar-Varzghan aftershocks recorded by Tabriz seismic network for upper crustal structure in the region.

We adopted the Aghaei (2013) model as an initial model. To substantiate the inferred crustal-velocity structure, we also used VELEST, a 1-D inversion of the arrival times for the 238 aftershocks with \( M_S \geq 2.0 \) that were recorded by at least six stations with a root mean square (rms) of arrival-time residuals \( \leq 0.4 \text{ s} \), and an azimuthal gap of \( \leq 170^\circ \). Several tests were performed to ensure that the 1-D inversion was correctly converged. Considering these multilayered models, velocity discontinuities were explored approximately at 7, 14 and 20 km depths. The deeper velocity discontinuities such as Moho were not able to be determined with the data set which was used here. The resulting model consists of a 7-km-thick layer with \( V_p = 5.6 \text{ km s}^{-1} \), overlain by a 7-km-thick layer with \( V_p = 5.9 \text{ km s}^{-1} \), a 6-km-thick layer with \( V_p = 6.2 \text{ km s}^{-1} \) as a third layer, a lower layer with \( V_p = 6.4 \text{ km s}^{-1} \). We used a \( V_p/V_S \) ratio of 1.76, calculated from database. In the 1-D velocity inversion and relocation procedures, a mantle-layer/half-space (Moho depth) of 46 km with \( V_p = 8.1 \text{ km s}^{-1} \) (Siahkali Moradi 2008) was used.

3.2 Single-event relocation

Considering the magnitude completeness value of \( (M_L) M_N = 2.0 \), 1624 events were selected for relocation. We relocated the selected events with different models using the locating program HYPOINVERSE (Klein 1985) and Seisan software (Havskov & Ottemoller 2005). The average rms and uncertainties in both epicentral (ERH) and depth (ERZ) determinations are significantly reduced when the velocity model obtained here is used. It should be mentioned that here in addition to the recorded data by Tabriz seismic network we used all available data recorded by three seismic stations in the Republic of Azerbaijan and station GRMI belonging to IIEES. However phase reading data from these stations were available only for the larger-events. The seismic stations that were used for relocation of Ahar-Varzghan sequence are marked by filled-triangles in Fig. 5. The distribution of 1624 relocated aftershocks with magnitude \( M_N \geq 2.0 \) is shown in this figure. Fig. 5 shows that the aftershock distribution is covering an area of approximately 25 \times 10 \text{ km}^2\) trending e–w. This orientation is compatible with the orientation of Ahar Fault and \( \sim 12 \text{ km} \) of observed surface rupture (Faridi & Sartipi 2012). The compatibility of the aftershock distribution with the observed surface rupture justifies the choice of the east–west nodal plane as the rupture plane for the first main shock (see Fig. 2). Fig. 5 shows that HRS and LRK stations are the nearest and farthest station to the activated area, respectively. The insert histogram in the lower-left corner of this figure presents the contribution of each seismic station in locating and shows that the azimuthal coverage of available phases is not optimal.

The histograms of seismic stations versus epicentral distance and aftershocks as a function of focal depth were plotted in Fig. 6. Fig. 6(b) shows tow peaks at depths of 13 and 18 km. However, these observed peaks in the histogram don’t correspond to the velocity contrasts, but could be affected by unmodelled velocity structures. The velocity bounders of the crustal model used here are 7, 14 and 20 km.

3.3 Aftershocks migration

After the doublet earthquakes, three relatively large aftershocks of \( M_S 5.2, 5.0 \) and 5.4 occurred on 14 August, 15 August and 7 November, respectively. To investigate the whole aftershock-migration pattern in detail, we separated all aftershocks into three time ranges: 2012 August 11–14, 2012 August 14 to November 7 and 2012 November 7 to December 31. These sequences are plotted in Fig. 7 in different colours. The epicentral distribution of relocated aftershocks and their depths versus distance along east–west strike are plotted in Figs 7(a) and (b), respectively. It seems that there are three clusters on the seismicity map. At depth, the seismicity is mainly concentrated between \( ~10 \text{ and } \sim 20 \text{ km} \), and in the western part they become shallower. We observe a migration of the seismicity with aftershocks mainly concentrated west of the main shocks, including the three large aftershocks of August 14, 15 and November 7 (Fig. 7b).

Aftershock locations in Fig. 7(b) seem to show apparently horizontal banding. The horizontal banding in Fig. 7(b), or observed peaks in the focal depth histogram in Fig. 6(b) are not artificial due to unmodelled velocity structures, because the aftershocks with focal depth of \( \sim 13 \text{ and } \sim 18 \text{ km} \) did not homogenously distribute all over active area. Most of the aftershocks with focal depth of \( \sim 13 \text{ km} \) occurred in the western part and the majority of the aftershocks with focal depth of \( \sim 18 \text{ km} \) occurred in the eastern part of the active zone (Fig. 7b). However unmodelled velocity structures may cause a systematic error in hypocentre parameters, but in our case the main source of error in hypocentre locations is due to non-optimal azimuthal coverage of available phases.

Considering the location of aftershocks with passing time shows that in the early days, aftershocks occurred in the eastern part of region. Then the activation gradually propagates to the western part of the region (Fig. 7c).

To obtain more detail the separated aftershock sequences are plotted in Fig. 8. Fig. 8(a) shows that the aftershocks in the first three days delineated an east–west trend and even demonstrate the clusters where a concentration of aftershock activity occurred. 480 aftershocks with \( M_N \geq 2.0 \) which occurred in this period (\( \sim 74 \text{ hr} \)) are distributed in two clusters. One of these clusters locates around the doublet Ahar-Varzghan main shocks. A strong aftershock (\( M_N 5.2 \)) which occurred on the fourth day is located in this cluster. The other cluster locates about 15 km in the west of main shock. Both of these clusters were simultaneously activated. After four days the activity concentrated in the west of main shocks between 38.4028–38.4513°N and 46.7885–46.666°E, where a large
Figure 5. Distribution 1924 events relocated using the hypoinverse program. The circles show the epicentres of aftershocks, scaled according their magnitude. The star symbol and filled circle represent the first and second Ahar-Varzghan earthquakes, respectively. The triangles indicate the seismic stations used for relocating events. The stations of ORD, NAX and LRK belong to the Republic of Azerbaijan; GRMI belongs to the IIEES; the other eight seismic stations which form Tabriz seismic network belong to the IGUT. The solid lines show traces of major active faults in the region which were mapped by Hessami et al. (2003a), except the Ahar Fault which was extracted from a geological map with scale of 1:250 000 (Amidi et al. 1978). The location of important cities is shown by squares. The dashed line over the aftershock cloud shows the \( \sim 12 \text{ km} \) surface rupture which was observed by Faridi & Sartipi (2012). The insert histogram in the lower-left corner shows the number of events which each station contributed in their relocations.

Figure 6. Histogram of epicentral distances and aftershock depth. (a) Number of seismic stations versus epicentral distance. (b) Number of aftershocks versus focal depth.

Aftershock (\( M_N 5.0 \)) occurred. During 201208141402 to 20121070628 (\( \sim 85 \text{ days} \)) 888 aftershocks with \( M_N \geq 2.0 \) occurred which have been distributed in the mentioned clusters (Fig. 8b). The aftershocks in this period mostly concentrate in the western part, and relative to Fig. 8(a), the activity has moved to the west. 88 days after the main shocks the activity moved westwards again where another strong aftershock (\( M_N 5.4 \)) occurred at 38.444° N and 46.5842° E. The events that occurred after the last strong aftershock, with magnitude 5.4 on 2012 November 7, are concentrated in a cluster oriented north–south (Fig. 8c). This stage was characterized by a seismic
Figure 7. (a) Distribution of 1624 Ahar-Varzghan events with magnitude $M_N \geq 2.0$ which occurred during about 5 months (from 2012/08/11 12:23 to 2012/12/31 23:29). (b) Distribution of aftershock depths versus distance along east–west strike. (c) The occurrence times of aftershocks since the 2012 Ahar-Varzghan main shock versus the distance along east–west strike. Note that the horizontal scale of plots (b) and (c) is different from (a). The aftershocks are separated into three time ranges: 2012 August 11–14, 2012 August 14 to November 7 and 2012 November to December 31, and plotted with open-square, open-circle and open-diamond symbols, respectively. In each case, the star symbol and filled circle represent the first and second Ahar-Varzghan main shocks, respectively. Symbols are scaled according to their magnitude. In panel (b), the dashed line marks the approximate slope of aftershock migration from east to west.

3.4 Multi-event relocation

The accuracy of hypocentre locations is controlled by several factors including velocity structure, the network geometry, arrival time reading accuracy and available phases. The uncertainty due to crustal structure is reduced by applying multi-event location approaches. Normally, in order to minimize the location errors the multi-event location approaches such as Joint Hypocenter Determination (JHD) and Hypocenter Double-Difference (hypoDD) methods are used. In the JHD technique, originally proposed by Douglas (1967), the location of events and station corrections are simultaneously determined. The JHD method accounts for lateral velocity variations, which are neglected in the 1-D model used to locate events, through the station corrections. Waldhauser & Ellsworth (2000, 2001) proposed the hypoDD method to minimize errors due to poorly determined velocity structure. In the double-difference method the residuals between observed and predicted phase travel-time difference (double-difference) are minimized for pairs of earthquakes at a common station. The double-difference algorithm has been extensively used for relocation aftershock sequence around the world, where it has demonstrated an ability to improve the location of clustered-events.

We applied the hypoDD method to relocate the Ahar-Varzghan sequence. From 1624 events that were located by single-event method, we selected 686 events for multi-event relocation by hypoDD technique. The sensitivity of linking parameters such as maximum separation and minimum observation etc. were tested for data set. The parameters we felt would best optimize our data set are: minimum
Ahar-Varzghan doublet earthquakes

The distribution of aftershocks into three time ranges: 2012 August 2000 to 2012 August = 10; maximum number ∼ 10 M ∼ 10^9 M 200 to 1982 0; maximum distance in km = 7; maximum number = 1978 5.2 on 14 August and = 1976 5.4 on 7 November). The epicentres of these events = 2012 August 14. The epicentres and filled circle represent the first and second Ahar-Varzghan earthquakes, respectively. The size of symbols is proportional to the magnitude value. The square and triangle symbols show the city and the nearest seismic station, respectively. The solid lines show the trace of faults which were extracted from Ahar Quadrangle Map with scale 1:250 000 (Amidi et al. 1978). The dot-dashed line shows the trace of Ahar Fault which extracted from the seismotectonic map of Iran with a scale of 1:2 500 000 (Berberian 1976). The dashed line over the aftershock cloud shows the ~12 km surface rupture observed by Faridi & Sartipi (2012). The beach balls represent the focal mechanism of events reported by the global CMT project. The GCMT has not reported the focal mechanism of event M_w 5.2, which occurred on 2012 August 14. The epicentres of these events are those determined by the hypoinverse program.

The selected data include event pairs with at least seven phases observed at common stations. The uncertainty of S-picks is between cluster centroid and station [DIST = 2 km for hypoDD epicentres and about 4 km for their depths, in this study is proposed. However in this study by applying hypoDD method, the location errors reduced, but due to especially non-optimal azimuthal coverage of available phases and other error sources, the uncertainties in determined hypocentres are still large. Our relative relocation results, while an improvement over catalogue locations, could be further improved using waveform cross-correlation data and temporary network data.

### 4 CAUSATIVE FAULT(S) GEOMETRY

The estimated faulting geometry from distribution of aftershocks clearly depends on the precision of hypocentre locations. To obtain a better image of the causative fault, the multi-event relocation approach of hypoDD was applied to the database to relocate 668 events. The distribution of these events provides a precise picture of the active zone (Fig. 10a). This figure shows that the aftershock sequence depicts an e–w trending area, about 25 km long and 10 km wide (Fig. 10a). This general trend is in agreement with the azimuth of surface rupture reported by Faridi & Sartipi (2012). They mapped a right-lateral surface rupture of about 12 km in e–w direction with maximum horizontal displacement of 70 cm and maximum vertical component 25 cm. Faridi & Sartipi (2012) reported azimuthal values of 85° to 100° along the rupture. The epicentral distribution of aftershocks apparently displays three or even four clusters of epicentre locations as input for hypoDD, the results of this test and seismic station coverage, an average relative uncertainty of ~2 km for hypoDD epicentres and about 4 km for their depths, in this study is proposed. However in this study by applying hypoDD method, the location errors reduced, but due to especially non-optimal azimuthal coverage of available phases and other error sources, the uncertainties in determined hypocentres are still large. Our relative relocation results, while an improvement over catalogue locations, could be further improved using waveform cross-correlation data and temporary network data.
aftershocks, the first at the eastern end of the active zone around the main shock and the second in the middle (46.62°–46.73° E). The latter cluster is located at the western end of the active fault and it delineates a north–south trend.

In order to understand the fault dip and focal-depth range of the aftershocks, a section approximately parallel to the fault strike and four normal sections are shown in Fig. 10. These cross-sections project data from a 5-km-wide swath parallel to the section line. Section A–A′ which is inferred to be close to the direction parallel to the fault strike shows that the focal depths of the aftershocks decreased toward the western part of the fault (Fig. 10b). In the cross-sections B–B′ and C–C′ which locate in the eastern part of active zone, the dominant depth range of aftershocks is from about 8 to 20 km, which we interpret to be the depth range of the faulting in this part. But, in the section D–D′ the depth range reduces to about 5 to 18 km. Therefore, in the eastern part the aftershocks reach the surface, which is consistent with the observations that there is surface faulting for the earthquake. In the section E–E′ the dominant depth range is about 4 to 15 km.

Projecting the aftershocks onto other directions within a range of about ±10° from the A–A′ profile and corresponding normal sections, does not change the distribution significantly and does not show a clear trend in dips. However, the profile C–C′ apparently shows a trend in dip to the north. Also, the offset of the aftershock cloud to the north of the observed surface rupture suggest that the causative fault dips to the north. But considering the large uncertainty of ~2 km in epicentres and about 4 km in the depths of relocated locations, a vertical slope with the potential error about 5° is suggested for the rupture zone.

5 DISCUSSION AND CONCLUSIONS

The temporal-spatial distribution of aftershocks showed that the aftershocks migrated in both along-strike and up-dip directions; the activity extended west to west–northwest (Figs 7 and 8). The pattern of relocated epicentres in Fig. 10(a) can be characterized as a cloud of aftershocks that spreads about 25 km e–w and about 7–10 km N–S. The single-event and multi-event relocation places the doublet main shocks at the eastern part of the aftershock cloud, and close to the eastern end of the observed surface rupture. Aftershocks depth in the AA′ profile which is in the direction of the aftershocks cloud strike shows that the focal depths of the aftershocks decreased toward the western part of the fault (Fig. 10b).

The profiles along and normal to the aftershock delineation show that approximately a vertical rupture initiated at the eastern part of active zone and propagated to the west. While Fig. 9 shows that the aftershocks locations in both of hypoinverse and hypoDD solutions have an offset of 2–3 km to the north of the observed surface ruptures. This offset in aftershock locations from surface rupture trace is an unusual feature for a roughly vertical fault plane. This could be due to a systematic uncertainty in the aftershocks.
location. The geometry of seismic stations contributed in locating process and unmodelled lateral variations in the velocity structure can cause a systematic uncertainty in the hypocentre locations. Also it is possible the observed offset in the aftershock locations be due to the dipping the fault plane to the north in agreement with the IGUT and Donner et al. (2015) solutions (see Table 1). This is an ambiguity and it is difficult to verify which one of the possible factors mentioned above is responsible of the offset. A 3-station temporary network was deployed in the region by Institute for Advanced Studies in Basic Sciences (IASBS). Distribution of 372 aftershocks recorded by this network during 2012 August 13–24 shows an offset to the north of the observed surface ruptures (Ghods et al. 2015).

Another ambiguity is which one of the doublet events is responsible of the surface rupture observed by Faridi & Sartipi (2012). Copley et al. (2014) believe that the ruptures are probably formed during only one of the two events, and that the other was either blind or any surface ruptures it produced were not observed. Donner et al. (2015) argue that only the first main shock produced the observed surface rupture, whereas the second main shock ruptured on a north–south-oriented nodal plane. In the Ahar-Varzghan sequence the aftershocks distributed in a broad zone rather than a linear feature, which is expected for a vertical fault, enhanced the hypothesis of two causative faults. However, the broaden aftershocks could be due to inadequate locations. Separation in epicentres of the doublet events located by each one of agencies/authors (also this study) shows that this separation is not due to the location errors (Fig. 2). Also, in addition to the ~12 km surface rupture was observed by Faridi & Sartipi (2012), two independent small segments of surface ruptures with different dips were observed by Faridi & Sartipi (2012). Therefore, we believe that the doublet Ahar-Varzghan earthquakes occurred in two separate faults.

The 2012 August 11 Ahar-Varzghan earthquake is one of the largest intraplate events ever to be instrumentally observed in the Ahar-Varzghan area. We analysed the aftershock sequence of this earthquake during about five months. The 2012 Ahar-Varzghan earthquake sequence shows a secondary activity which started on November 7, approximately 3 months after the main shock, with a
significant increase in activity, regarding both number of events and their magnitude. This stage was characterized by a seismic area that widened to the west of the main shocks. The temporal and spatial distribution of relocated aftershocks shows that they are related to a significant tectonic feature instead of being a random occurrence. We found that the aftershocks migrated in both along-strike and up-dip directions.

The observed surface faulting was dominantly right-lateral strike-slip. Therefore the Ahar-Varzghan earthquake was originated by activation of a dextral fault in the upper crust and followed by many aftershocks. The general pattern of relocated aftershocks defines a complex seismic zone covering an area of approximately $25 \times 10\, \text{km}^2$. The dominate range of the aftershocks is from 1 to about 25 km. Among the location parameters, the depth has normally a large uncertainty, due to large uncertainty in determined depths; it is difficult to conclude that this corresponds to the top and bottom of the coseismic rupture. These events cluster in a broad zone to the west with a tendency to the northwest direction rather than a linear feature that might have been expected from a nearly vertical strike-slip fault. It is possibly due to activation of two different faults with nearly the same strike. With the available information and considering the accuracy of locations we are not able to state which of the two events caused the surface ruptures.

The cross-sections through the aftershock locations do not show a clear trend in dip, but delineates the east–west trend, which is interpreted as the fault plane of the main shock and justifies the choice of e–w nodal plane as the rupture plane for the first main shock ($\text{M}_\text{w} \, 6.5$). This geometry is consistent with a highly unilateral rupture with a westward trend. We believe that the Ahar-Varzghan doublet earthquakes probably occurred on two separate faults with slightly different strikes in the east–west direction. But, due to uncertainty in our locations we could not determine the relative locations in a north–south direction, however most agencies/authors and our relocation show that the second earthquake occurred at western part of the first earthquake with $\sim 6\, \text{km}$ distance. Further studies may clarify this ambiguity.

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REFERENCES


