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Analysis of the 1990 Fork (Darab), southern Iran, earthquake sequence

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Abstract

The Fork earthquake occurred in Zagros mountains at a distance about 120 km from Darab city in Fars province, producing extensive destruction but relatively low rate of human loss. Field investigation and the distribution of aftershocks suggest an east-west trend faulting with a reverse mechanism having a small strike-slip component. The locally recorded aftershock activity was extended to a length of about 40 km and a depth of about 30 km. The majority of aftershocks took place at a depth range 10-20 km and was scattered indicating a complex mode of faulting. The result of waveform inversion indicated that the mainshock had mainly reverse mechanism and the source process included two main fault slip. The total seismic moment was calculated to be $M_0 = 3.1\times10^{25}$ dyne cm. The calculated maximum dislocation was about 50 cm and the obtained moment magnitude was Mw = 6.2. The average stress drop was estimated to be 25 bar and the average dislocation was 25 cm. The Fork earthquake is one of the rare events that has occurred in Zagros suture zone with magnitude greater than 6. Therefore, the ground-motion characteristics during the mainshock should be considered for the high safety design of structures in the damaged area.

Key words: Fork (Darab) earthquake, seismicity and seismotectonics of Zagros, active faults, aftershock activity, source mechanism, source parameters

1. Introduction

On the 6th of November1990 at 22:15:53.8 GMT, 22:15:53.8 local time, a moderate but considerable destructive earthquake occurred at the southeast of Zagros suture zone in Fars province. The epicentral region was located at a distance 120 km southwest of Darab city in southern Iran. Using the onset times of 707 stations, the epicenter of mainshock was computed as 28.23N-55.47E by ISC. The magnitude of the mainshock, given by ISC, was mb=6.1, Ms=6.6 and focal depth determination indicated a shallow focal depth of 16 kilometers. The mainshock was followed by many aftershocks. The largest one occurred about 45 minutes after the mainshock at 19:30:20.0 GMT and caused more destruction to the structures that were damaged during the mainshock but were not completely destroyed. The epicenter of largest aftershock was computed as 28.20N-55.37E by ISC for the largest aftershock, using the onset times of 317 stations. Its magnitude, given by ISC, was mb=5.3, Ms=5.7 and the focal depth was 18 kilometers.The Fork earthquake is the largest instrumentally recorded shock in the vicinity of Darab and one of the rare earthquakes with magnitude greater than 6 in Zagros suture zone. The quake killed 23 people, injured 80 and left 12,105 homeless in a remote, desert area.

Shortly after the occurrence of mainshock, the Institute of Geophysics, University of Tehran deployed a temporary seismic network in the damaged area and carried on monitoring the aftershock sequence. This paper mainly presents the results of field and seismological investigations. First, the seismicity and the seismotectonics background of the affected area are presented. Then, the source parameters of mainshock are obtained by waveform inversion. Next, the locally recorded aftershock sequence is analyzed and the empirical relation for the rate of aftershock decay is determined. Finally, the macroseismic evidence of the Fork earthquake is presented and disscused.

2. Seismicity and seismotectonic background

The epicentral region of Fork earthquake, which is the subject of this study, is located along the Zagros suture zone. This zone is the most seismically active region in Iran. Deformation in this region is large and complicated partly due the fact that the continent is older and inherits the old faults and structures which become reactivated (Gheitanchi, 1987). The mechanisms by which the seismic activity is accomplished probably involve ductile creep at deeper levels and folding, faulting and fracturing at shallower levels. A simplified tectonic map based on the works done by Berberian (1976) and Jackson and McKenzie (1984) is shown in Figure 1.

Historical seismicity of Iran has been studied by Ambraseys and Melville (1982). No historical

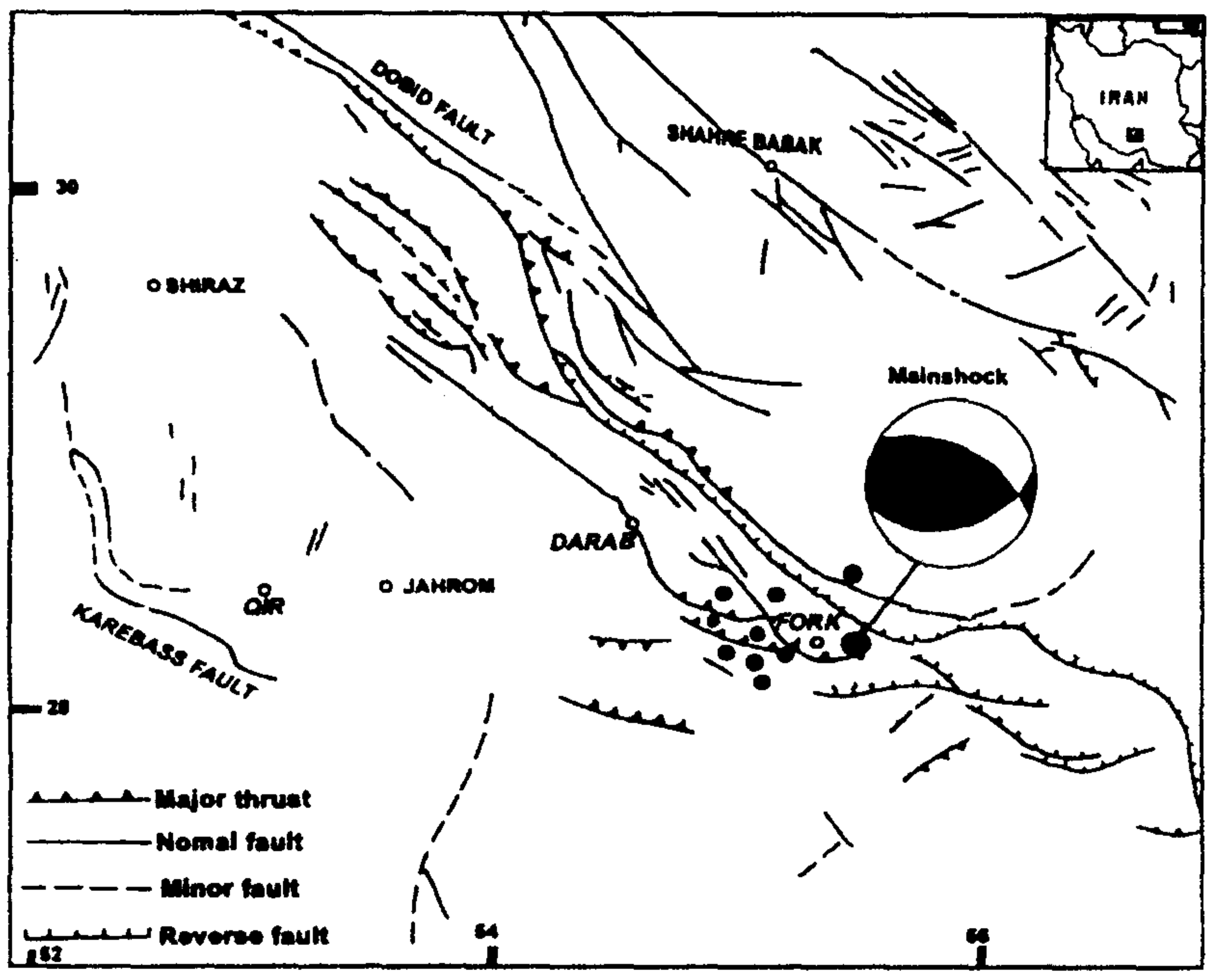

Figure 1. A simplified tectonic map based on the works done by Berberian (1976) and Jackson and McKenzie (1984), and the mechanism and the location of mainshock as well as the epicenter of strong aftershocks reported by ISC.

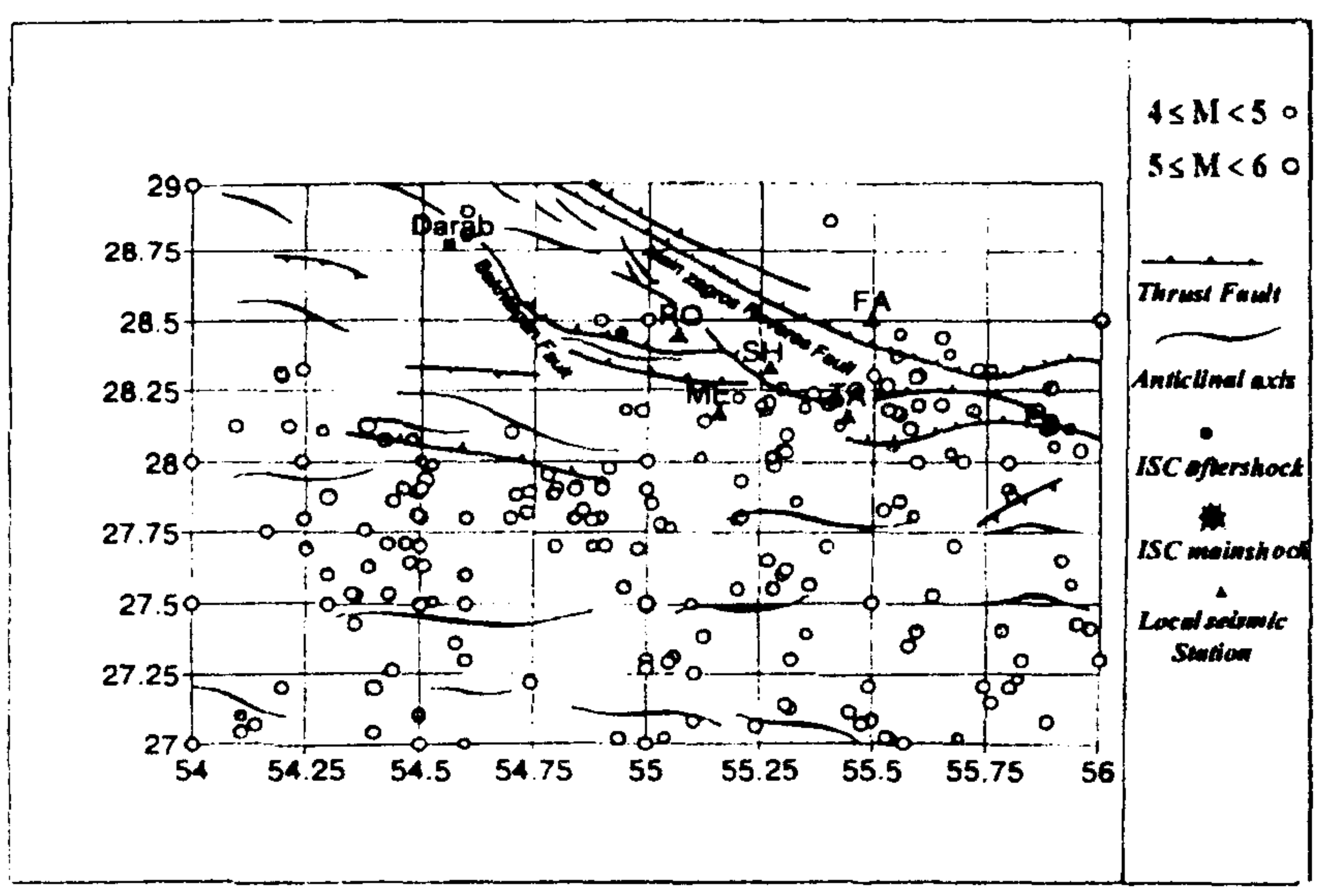

Figure 2. The epicenters of earthquakes within 150 km from the epicentral area reported by ISC during the past 100 years are overlapped in the fault map.

earthquake was reported in Fork region. However, most areas from the center to the southeast of Iran are deserts or semi-deserts, sparsely populated and isolated from the rest of country. Therefore, information on the historical seismicity is scarce. Because of the lack of local seismic station, the parameters of earthquakes in southern Iran include high uncertainties. Hence, the local events with magnitudes smaller than 4.5 either are not located or do not have reliable source parameters. The teleseismicly located seismicity of the Zagros Folded Belt is very high and characterized by a large number of moderate shocks in the magnitude range of 5 to 6 and a very small number with magnitudes equal to or slightly greater than 7 (McKenzie, 1972). Compared with the other regions in Iranian plateau, seismicity in this region has a rather regular pattern and is mainly characterized with moderate earthquakes. A map of epicenters of the earthquakes within 150 km from the location of mainshock was given in Figure 2.

3. Source parameters of mainshock

Using the inversion technique developed by Kikuchi and Kanamori (1991), the long-period body waves of the Fork earthquake recorded by GDSN stations were inverted to their sources to investigate the source mechanism. The P and SH waveforms of 12 stations with epicentral distances between 30 and 100 degrees were used for this study. The records with a duration of 40 seconds were inverted with a sampling interval of 1.0 second. The S-wave seismograms were rotated in order to obtain the transverse component for the SH analysis. Both the observed and synthetic Green's functions for all the stations were equalized to GDSN seismograms with the same gain. Then, considering the quality of the observed records, we applied station weighting factors as 1.0 for vertical P waves and 0.5 for SH wave (Gheitanchi, 2002).

In calculating the synthetic wavelet for a point dislocation we used the Jeffreys-Bullen A model (Jeffreys and Bullen, 1958). First, a source time function of trapezoid shape having rise time of 3 seconds and process time of 4 seconds was best fitted. Then, with the fixed source time function, the data was inverted for several source depths. The residual error was minimized for the depth of 5-10 kilometers. This suggested that the centroid depth was not deeper than 10 km. In next stage, by a point source approximation, we obtained the mechanism solution. Finally, for a fixed fault plane, the spatio-temporal distribution of fault slip was determined by the waveform inversion procedure, in which, the slip direction was allowed to vary. The comparison of the observed and synthetic seismograms after the first iteration is given in Figure 3. This figure indicates that the fit of observed and synthetic waveforms is acceptable for the first 20 seconds. The iteration was repeated two times; no significant decrease in the residual error was found after two iterations. This suggested that there were two main fault slip during the source process of the mainshock. The largest slip took place during the first 10 seconds while the next slip initiated after 20 seconds. Out of two possible fault planes, the one striking NW-SE gave a much better variance reduction and was in agreement with the strike of geological faults in the region. The mechanism solution for the total source was obtained as striking N73W, dipping 42 NE, and having rake angle 119. The

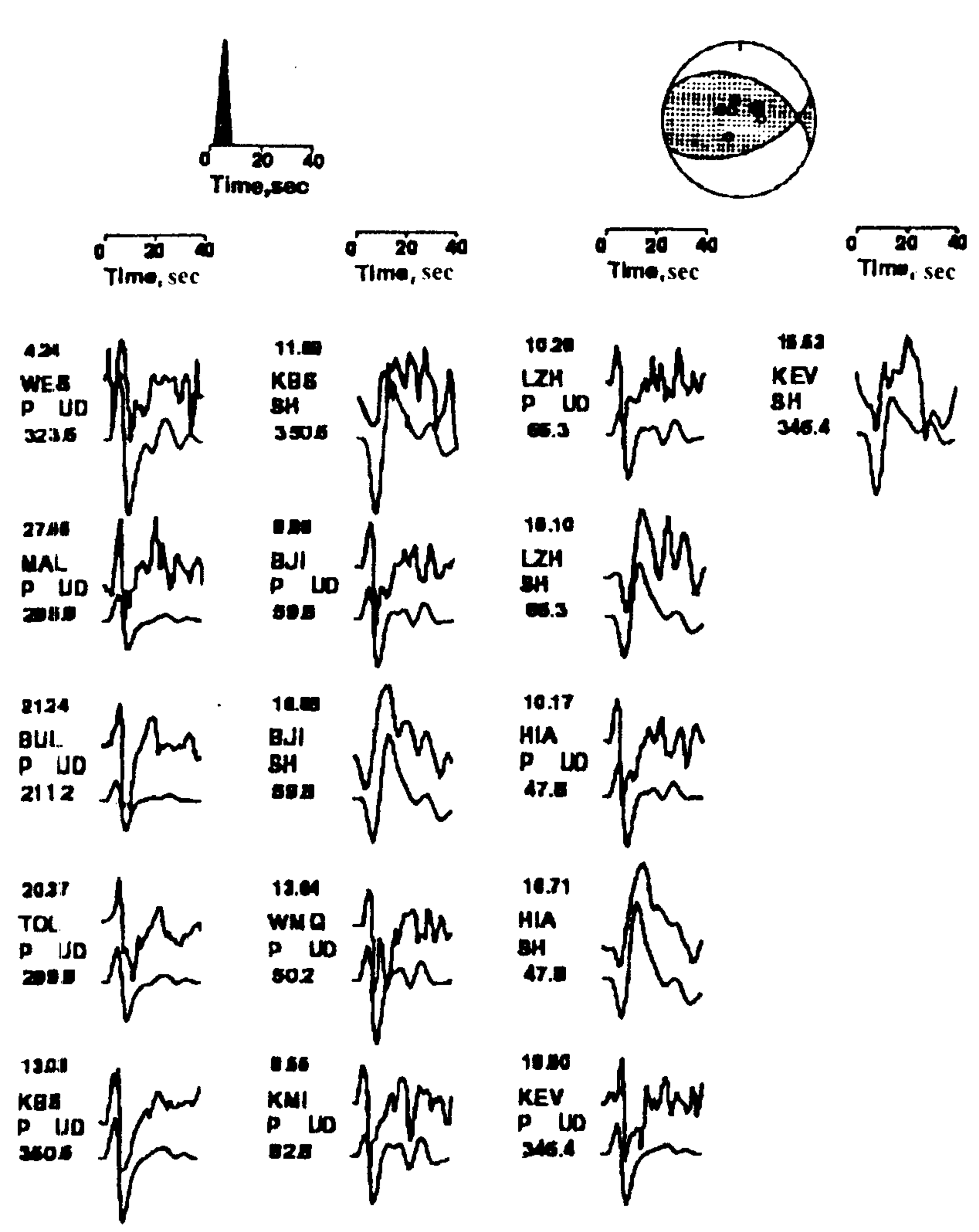

Figure 3. The source time function, the focal mechanism, and the ray directions of the stations used in this analysis as well as the comparison of the observed (top) and synthetic (bottom) waveforms after the first iteration for the 1990 Fork earthquake. The correlation coefficient, the name, component and azimuth of station are given on the left side of each waveform.

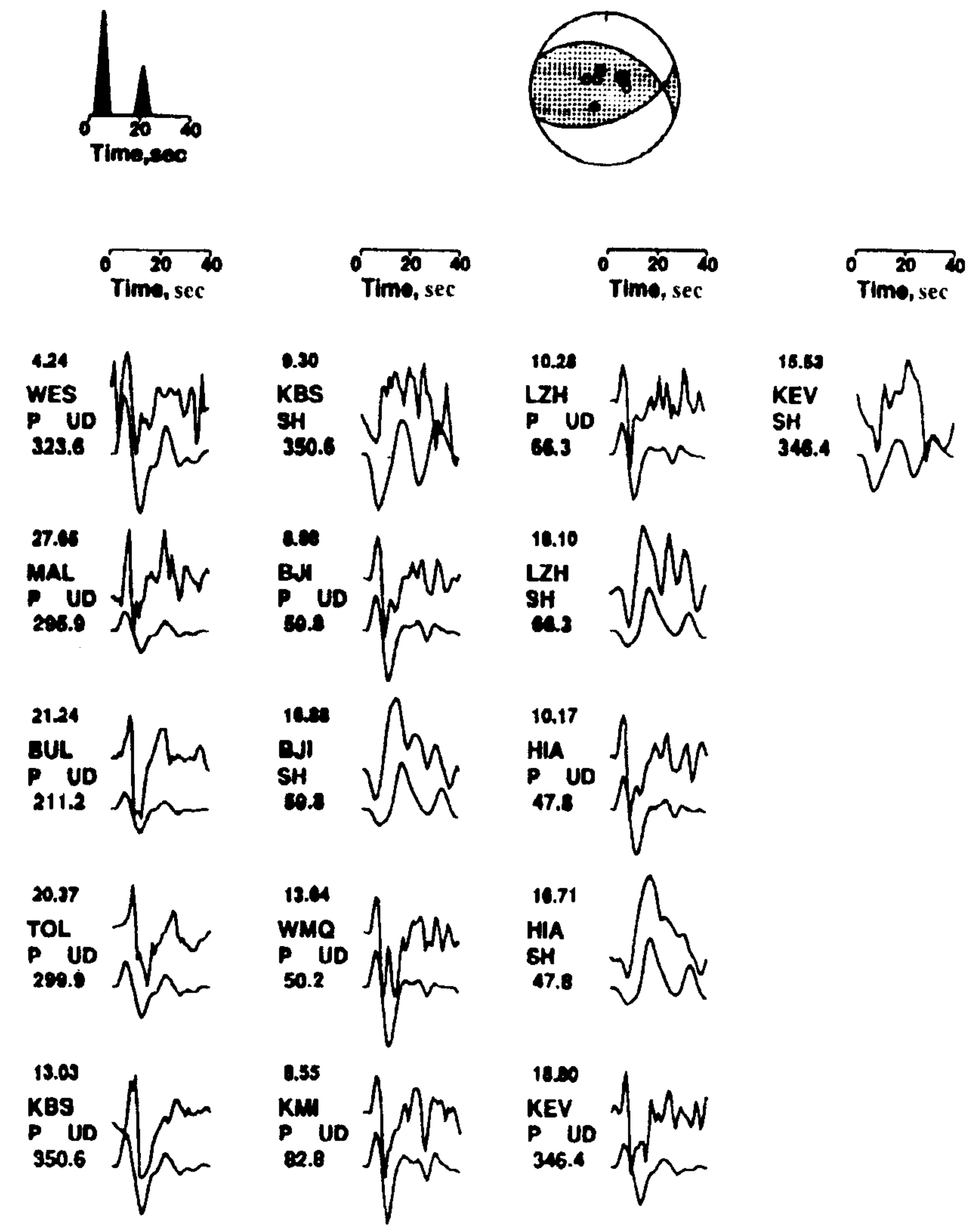

Figure 4. The source time function, the focal mechanism, and the ray directions of the stations used in this analysis as well as the comparison of the observed (top) and synthetic (bottom) waveforms for the final solutions of the 1990 Fork earthquake. The correlation coefficient, the name, component and azimuth of station are given on the left side of each waveform.

fault slip was consistent with the geological evidences such as folding and thrust type faulting in the region. The total seismic moment was calculated to be $M_0= 3.1\times10^{25}$ dyne cm. The calculated maximum dislocation was about 50 cm and the obtained moment magnitude in this analysis was Mw = 6.2 while the estimated rupture velocity was 3.0 km/s. Using the relation $\Delta\sigma= 2.5M_0/(S)^{3/2}$ and approximating the rupture area, S, by L×(L/2), where L=40 km was the fault length which was estimated by the extension of aftershock activity, thus the average stress drop, $\Delta\sigma$, could be estimated (Gheitanchi et al., 1993). In this study, following the same relation, the average stress drop, $\Delta\sigma$, was estimated to be about 25 bar. Using the relation $M_0 = \mu DS$, where $\mu = 3 \times 10^{11}$ dyne cm^{-2} was the rigidity and S the fault area, the average dislocation, D, was calculated to be 25 cm. Examples of the observed and synthetic waveforms, the focal mechanism and the ray directions of the stations used in this analysis are given in Figure 4.

4. Locally recorded aftershock sequence

The mainshock was followed by several strong aftershocks, which caused additional damage and destruction in the affected areas. However, in this region where the precision of depth calculation was very poor and the epicentral locations even for large shocks could be in error by several tens of kilometers (Ambraseys 1978; Berberian 1979; Jackson, 1980). The aftershock study carried out by a local temporary seismic network after the occurrence of mainshock had a very important role and could be the main tool to investigate details about the characteristics of the mainshock. In order to study the aftershock activity in details, the seismology Division Institute of Geophysics, University of Tehran deployed a temporary seismic network in the affected area three days after the occurrence of mainshock and monitored the activities from the 9th November 1990 for two months. The recording instruments were five portable Sprengnethers MEQ-800, which were provided and operated by the same Institute. The seismic stations were installed on the slope of mountains on bed rock, where clear and sharp wave traces could be obtained. The coordinates of recording stations are given in Table 1. The coordinates of the temporary stations were allocated by using the topographic maps with scale 1:50000. The accuracy of time in each station was checked daily and if necessary adjusted by using the radio signals. The speed of recording drum in each station was 2 mm per second. P and S arrival times were read from the records of the local network and for the computation of the hypocentral parameters of the earthquakes under study, the modified version based on the Hypo71PC program of Lee and Valdes (1985) was used. Considering the geological and other seismological evidences, several crustal models were examined for hypocentral determination and the crustal model that had minimum residual errors was selected for the computations. The minimum value of RMS was obtained for a velocity of 5.7 km/s. Our best model was made of a thin layer of 8 km with a velocity Vp = 5.7 km/s, over a half-space with Vp^{*}= 6.0 km/s. The best locations were selected on the basis of RMS smaller than 0.5. We located the whole set of aftershocks both, in a half-space model (Vp = 5.7 km/s), and in the selected two-layer model. The results did not differ significantly. The epicenters differed by no more than 200m on the average, and 83 per cent change by less than 500m. The depths, which were more sensitive to model perturbations, differed by about 700m on the average, 70 per cent of them changing by less than 1 km. Hence, we believed that the epicerters given by HYPO71 were quite realistic (Madooliat, 1995). The temporary seismic network recorded many aftershocks but the parameters of 400 well-located aftershocks were used for this analysis. The locations of seismic stations and the epicenter of mainshock and well-located aftershocks are overlapped in the fault map and are shown in Figure 5. This figure shows that aftershocks extend over a zone approximately 40 km long with a general east-west elongation and the seismicity appears to be more scattered than along a fault zone. The aftershock activity appeared to be close to the macroseismic epicenter having focal depth deep to 30 km. However, the vertical cross-sections that are shown in figure 6, indicate that the majority of aftershocks took place at a depth range 10-20 km.

5. Empirical relation for the rate of aftershock decay

Omori suggested an empirical relation for the rate of aftershock decay (Utsu, 1961). This relation states that the frequency of aftershocks n(t) per unit time t, following the mainshock, is represented by $n(t)=K/t^{-c}$, while K and c are constants and should be determined for each region. The data obtained by temporary seismic network in epicentral region was used in order to obtain the constants of the empirical relation for the rate of aftershock decay in Fork region. The constants K and c are calculated as 72.8 and 0.5 and the result is given in Figure 7. This model of aftershock activity could be acceptable in the Fork region though for the first three days no data was available.

Table 1. Source parameters of the earthquake with magnitude greater tha 5 around the affected area during 1964-1999.

Date	Origin Time	Lat. N	Lon. E	Dep.	mb	Ms
Y m d	h m s	Deg	Deg	km		-
911106	184553.0	28.25	55.47	16	6.1	6.6
901106	193020.0	28.20	55.37	18	5.3	5.7
901106	195003.5	28.30	55.44	33	4.0	-
901107	040900.8	28.20	55.20	33	4.2	-
901108	051014.0	28.20	55.20	68	4.0	-
901108	231826.0	28.10	55.40	40	4.0	-
901111	191137.0	28.40	55.65	72	3.8	-
901111	205641.4	28.19	55.57	67	4.1	-
901111	221947.0	28.10	56.00	33	3.9	-
901111	224507.0	28.20	55.25	67	4.3	-
901121	034236.0	28.34	55.54	57	4.6	4.1
901208	212248..0	28.20	55.31	54	4.5	-

Table 2. The coordinates of temporary seismic stations in the Fork region. The recording strated on the 9th November 1990 and monitored the aftershocks for two months.

Number	St. code	Loc	Long E	Lat. N	Lat (m)
1	RO	Rostaq	55°-04.4'	28°-26.1'	1300
2	TA	Tashkuyeh	55°-26.4'	28°-09.3'	800
3	ZA	Zakareya	55°-15.2'	28°-18.3'	950
4	FA	Fakyrabad	55°-29.4'	28°-29.6'	1500
5	ME	Mehrabad	55°-08.0'	28°-09.5'	800

6. Damage to houses and structures

The damaged area was remote, isolated and sparsely populated due to the rough topography and the lack of water. The low casualty rate (23 people killed) was mainly due to the fact that the area was sparsely populated. There was no major city around the earthquake area. There were small villages within several kilometers from each other in the earthquake region. Eighteen villages were damaged in the earthquake and the two villages of Qareh-No and Mohammad-Abad in Fork region were totally demolished. The report of damage to villages which were confirmed by the local government is given in Table 2. Most of the buildings in the earthquake zone were of traditional adobe or stone masonry construction. The great destruction in all the affected villages was due to poor lateral supports and weak seismic resistance and

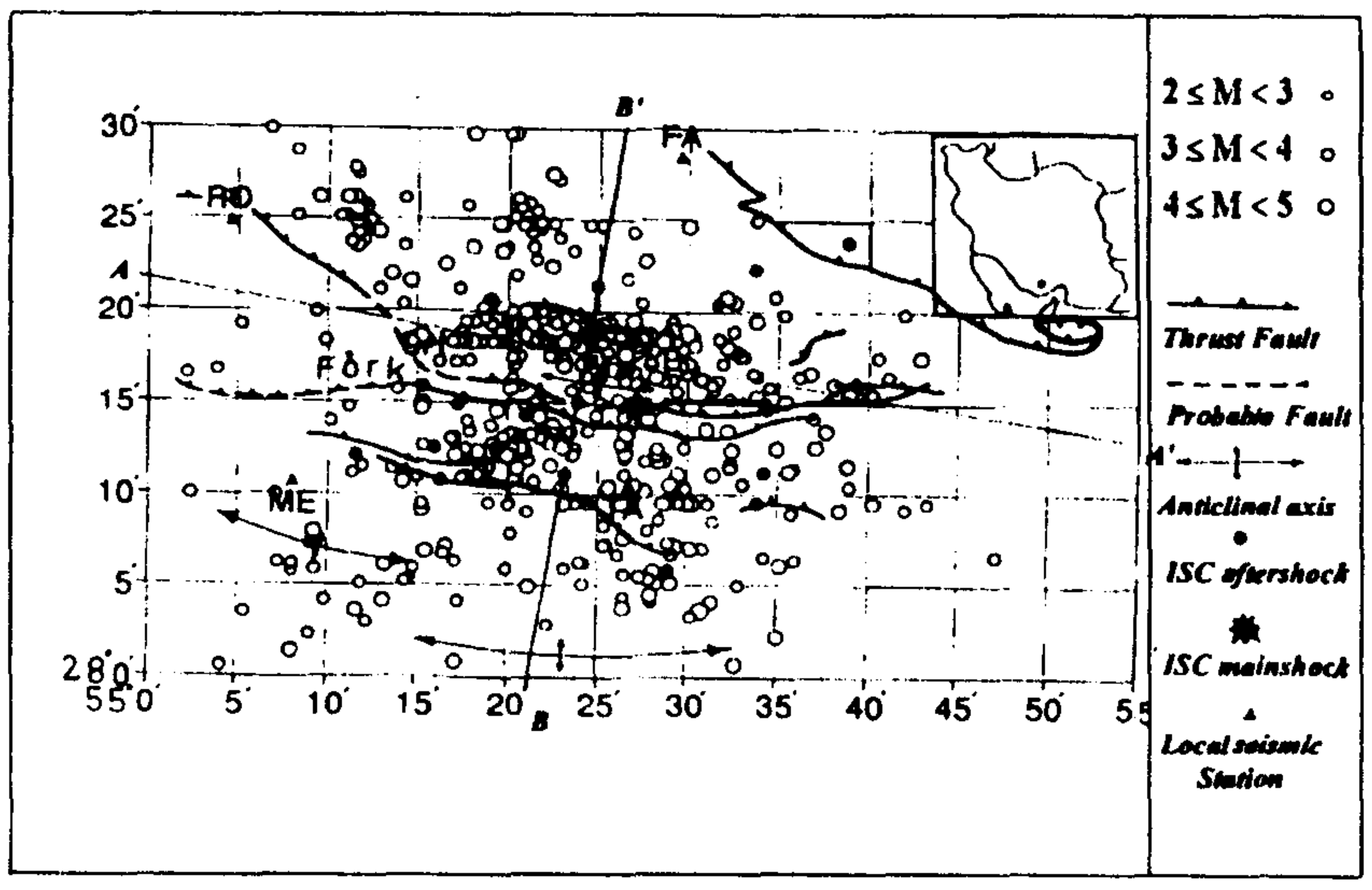

Figure 5. The locations of temprary seismic stations shown by soild triangles, the epicenter of mainshock indicated by soild star, the teleseismic aftershocks reported by ISC shown by solid circules, and well-located aftershocks given by open circales are overlapped in the fault map. The lines AA' and BB' are the trends of vertical cross-sections along and across the main fault that are shown in figure 6.

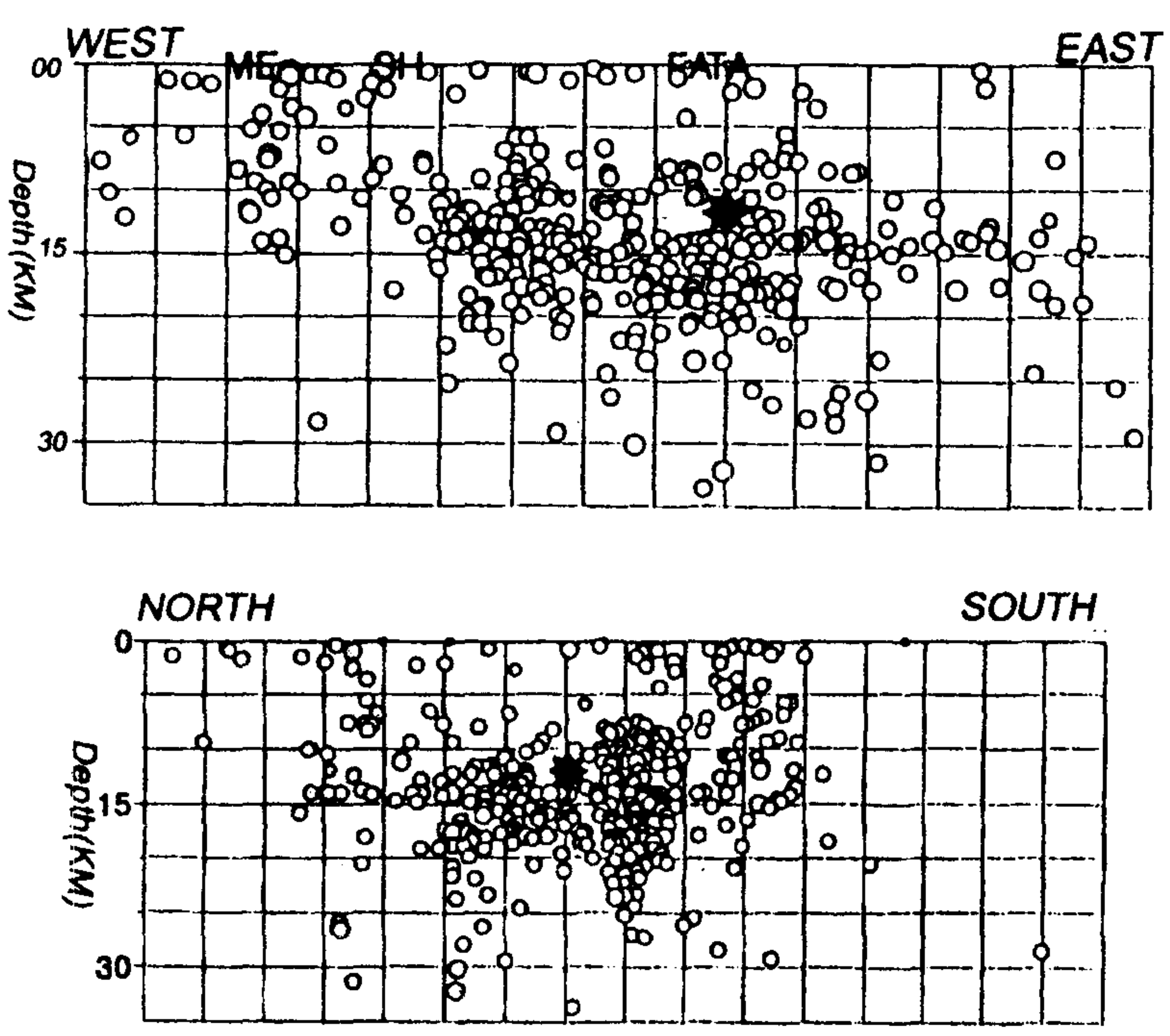

Figure 6. The vertical cross-sections along (top figure) and across (bottom figure) of the main fault indicating that the majority of aftershocks took place at a depth range 10-20 km. The trends of cross-sections are shown in figure 5.

collapse of these types of structures. Apart from the houses built in the traditional style, the area contained a few better built engineering structures. These houses near the epicentral region suffered partial damage. Regarding macroseismic evidence, the maximum intensity of the mainshock just exceeded VI on the Modified Mercalli Intensity Scale.

7. Discussions

No strong motion instrument was operating in the

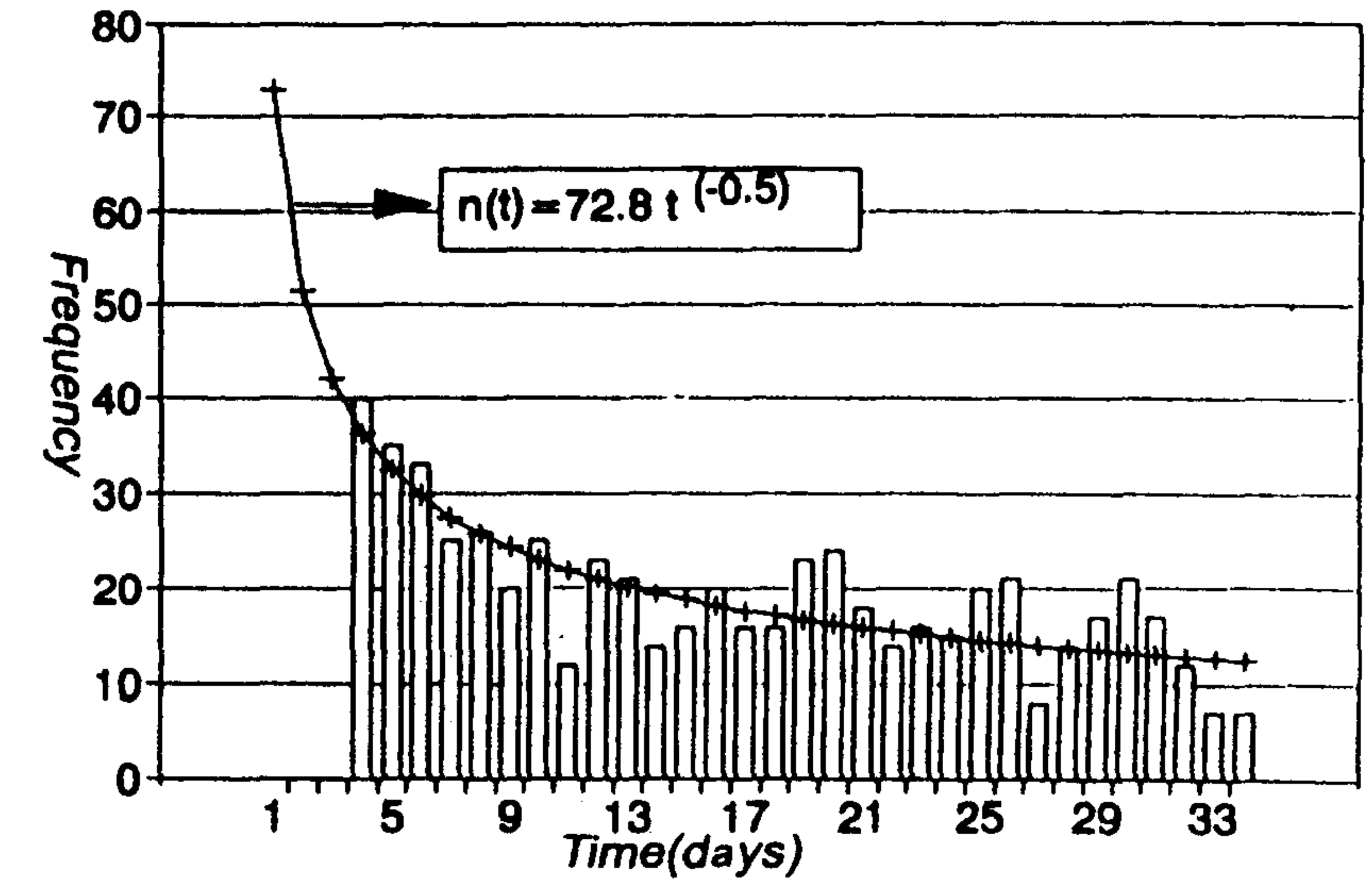

Figure 7. The Omori empirical relation for the rate of aftershock decay. The horizontal axis indicates time in term of days and the vertical axis shows the number of aftershocks. The constants are calculated as 72.8 and 0.5. The Omori model might be acceptable in the Fork region though for the first three days there is a gap of data

epicentral area during the November 6, 1990 earthquake, but macroseismic evidence, such as the observed damages in the buildings, indicated that the ground motion was intense and of rather short duration. The strong ground motion did not last too long since even small amplitude waves would have destroyed the unstable structures. The extent of aftershock activity indicates an average source dimension of about 40-50 km. The majority of aftershocks are distributed on the north side of the faults. This fact suggests an East-West strike and a north-dipping fault plane. This is consistent with the strikes of the fault plane solution obtained by waveform modeling in this study. Considering the epicenter of mainshock as the initial break, the distribution of locally recorded aftershocks indicate that the rupture should be initiated in epicentral region and extende to east and west in a bilateral manner. This fact could also be understood from the result of waveform analysis. The extent of aftershock activity indicates a range of 30-40 km source dimension, and is in agreement with the observed destruction. The vertical cross-section along the main fault indicates that the aftershocks were distributed in a depth range deep to 30 km with the highest concentration around the depth of 10-20 km. This suggests that the faulting mainly took place in the uppermost basement beneath the sedimentary cover.

Several waveform modeling of teleseismic earthquakes suggested depth of 10 km or even shallower for the earthquakes with magnitude larger than 5 in this region (Maggi et al., 2000; Jackson & Fitch, 1981). However, the depth distribution of aftershock activity in this study, highly supports the possibility of earthquakes deeper than 20 km, though the lack of a well controlled crustal velocity model might have affect on the focal depth determination. Detailed study by a dense digital seismic network could give more information about the depth of seismic activity in this region. Maggi et al. (2000) also studied this earthquake using long-period body waves. Their results are in good agreement with this study. They obtained two main subevents with very similar mechanism and the centroid depth in their study was 7 km.

8. Conclusions

The Fork earthquake was a multiple shock containing two subevents with thrust type mechanism and produced quite restricted epicentral area in which buildings were left in a highly unsafe and unstable condition. The aftershocks observation suggested an east-west trend faulting. The aftershock activity was extended to a length of about 40 km and a depth of about 30 km. The majority of aftershocks took place at a depth range 10-20 km and was scattered indicating a complex mode of faulting. The total seismic moment was calculated to be $M_0= 3.1\times10^{25}$ dyne cm. The maximum dislocation was about 50 cm and the moment magnitude in this analysis was Mw = 6.2. The average stress drop was estimated to be 25 bar and the average dislocation was 25 cm. No major engineering structures were situated in the strongly shaken area. However, such structures located within 20 km of the macroseismic epicenter, did not suffer heavy damages. Although the area had been seismicly active in historical times there was no evidence that earthquakes as severe as this earthquake had occurred in the vicinity of Fork region. From the engineering point of view, the Fork earthquake, which provided ground-motion characteristics of a rare large event in the affected area, was the controlling event for the design of structures with high safety requirements.

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