Analysis of Izmit aftershocks 25 days before the November 12th 1999 Düzce earthquake, Turkey

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**Abstract**

We investigate spatial clustering of 2414 aftershocks along the Izmit Mw = 7.4 August 17, 1999 earthquake rupture zone. 25 days prior to the Düzce earthquake Mw = 7.2 (November 12, 1999), we analyze two spatial clusters, namely Sakarya (SC) and Karadere–Düzce (KDC). We determine the earthquake frequency–magnitude distribution (b-value) for both clusters. We find two high b-value zones in SC and one high b-value zone in KDC which are in agreement with large coseismic surface displacements along the Izmit rupture. The b-values are significantly lower at the eastern end of the Izmit rupture where the Düzce mainshock occurred. These low b-values at depth are correlated with low postseismic slip rate and positive Coulomb stress change along KDC. Since low b-values are hypothesized with high stress levels, we propose that at the depth of the Düzce hypocenter (12.5 km), earthquakes are triggered at higher stresses compared to shallower crustal earthquake. The decrease in b-value from the Karadere segment towards the Düzce Basin supports this low b-value high stress hypothesis at the eastern end of the Izmit rupture. Consequently, we detect three asperity regions which are correlated with high b-value zones along the Izmit rupture. According to aftershock distribution the half of the Düzce fault segment was active before the 12 November 1999 Düzce mainshock. This part is correlated with low b-values which mean high stress concentration in the Düzce Basin. This high density aftershock activity presumably helped to trigger the Düzce event (Mw = 7.2) after the Izmit Mw = 7.4 mainshock.

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**1. Introduction**

The North Anatolian Fault Zone (NAFZ) is one of the seismically most active strike–slip faults world-wide extending 1600 km from eastern Anatolia to the north Aegean Sea (Fig. 1a). The direction of slip corresponds well with Global Positioning System (GPS) derived 20 to 25 mm yr

-1 westward motion of the Anatolian block with respect to Eurasia (Mcclusky et al., 2000; Reilinger et al., 2006). The Izmit earthquake of 17 August 1999 Mw = 7.4 exhibits a maximum surface displacement of 5.2 m at the Sapanca–Akızy segment (Barka et al., 2002, Fig. 1b). Average coseismic slip obtained from teleseismic waveform inversion is 2.5 m (Tibi et al., 2001) and 2.9 m from strong motion records (Bouchon et al., 2002). Synthetic Aperture Radar interferometry (InSAR) data inversion (Wright et al., 2001) shows a maximum displacement of approximately 5 m near the mainshock and a total coseismic moment of 2.6 × 10

20 Nm. GPS data by Reilinger et al. (2002) indicate a geodetic coseismic moment of M0 = 1.7 × 10

20 Nm and a maximum displacement at the Izmit segment of about 5.7 m. Delouis et al. (2002) identified four segments along the Izmit rupture by using combined GPS, SAR, teleseismic and strong motion data.

Lay and Kanamori (1980) studied body waves and surface waves of large earthquakes in the Solomon Islands region in an attempt to determine the stress distribution on the thrust plane. They found that relatively short-period seismic body waves are radiated from only small parts of the entire rupture plane, which generates longer-period surface waves and over which the aftershocks occur. They interpreted the results in terms of an asperity model. This asperity model is an outgrowth of laboratory experiments on rock friction. Byerlee (1970), Scholz and Engelder (1976) suggested that two sides of a fault are held together by areas of high strength, which they termed asperities. Extending this model to earthquake faults, Lay and Kanamori (1981) called the areas on the fault plane from where relatively short-period seismic body waves are radiated the fault asperities; it is assumed that the stronger spots are responsible for high-frequency seismic radiation.

From rock-deformation experiments in the laboratory three stages in acoustic emission fingerprints characterizing the failure of asperities are observed (Lei, 2003). During the first loading stage, the event rate which increases is documented by the inverse Omori law (Utsu, 1961). Simultaneously the b-value sharply decreases at the edge of the asperity. In the second stage prior to the mainshock, large magnitude events appear at the edge of the asperity (very close to the first stage....

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During the third stage, significant increase and subsequent gradual decrease of \( b \)-value is observed along with a decreasing event rate (Zang et al., 1998). In microscale and macroscale environments the \( b \)-value seems to rely on crack or rupture densities, which themselves result from the amount of applied stress or pressure, type of material, and rupture dynamics. Wiemer and Katsumata (1999) presented a "rupture mechanics approach" that relates \( b \)-values, coseismic slip and stress drop. Prior to an earthquake they suggest that areas subjected to high stress show low \( b \)-values. According to this approach high slip regions correspond to regions of high stress drop and high \( b \)-values of aftershocks. In this approach calculated slip distributions based on seismological and geodetic records of the mainshock are related to the aftershock seismicity parameters \( b \)-values. Sobiesiak et al. (2007) suggested that inhomogeneities in seismogenic fault areas can be mapped using \( b \)-value. These areas can be attributed to potential asperities. Wiemer and Wyss (1997) described that areas of low \( b \)-values with a size range of 5–15 km correlate with asperities in the Parkfield and Morgan Hill sections of the San Andreas Fault (SAF). This hypothesis that low \( b \)-values indicate asperities on the fault prior to a large earthquake is supported by the study of Wyss et al. (2004) showing that \( b \)-values on locked patches of the SAF near Parkfield are systematically lower than \( b \)-values on creeping patches. This observation is a direct consequence of the inverse relationship of \( b \)-value to the applied shear stress. Amitrano (2003) and Schorlemmer et al. (2005) show that the high-stress environment of locked fault patches is more likely to support future large earthquake occurrence. Westerhaus et al. (2002) analyzed SABONET (Sapanca–Boğuş NETwork) data before the İzmit earthquake. The lowest \( b \)-values (~0.8) were located at the fault bend which runs through the epicenter of the 1999 İzmit mainshock. At the bend, a localized stress concentration is expected from numerical models of seismicity along asperities. The site of lowest \( b \)-values had been considered to be the most likely place for a major earthquake, a conclusion that was confirmed by the İzmit earthquake, with epicenter located about 13 km from the anticipated site. Aktar et al. (2004) who analyzed the first 45 days of the İzmit aftershock sequence detected three zones of relatively high \( b \)-values, two of which coincide with asperities revealed by Bouchon et al. (2002). Aktar et al. (2004) mainly focused on the area between the Yalova and Karadere segments (see Fig. 1) and their analysis ended 42 days before the Düzce earthquake occurred. In this study we focus on the eastern part of the İzmit rupture (Sakarya, Karadere and Düzce regions). We discuss the İzmit aftershock sequence between October 18, 1999 and November 12, 1999 (i.e. the 25 days prior to the Düzce mainshock \( M_{w}=7.2 \)). We analyze spatial variations of 2414 aftershocks with errors <5 km (horizontal and vertical). Variations in \( b \)-values along the rupture and with depth are

**Fig. 1.** a) Topographic map of the North Anatolian Fault Zone (NAFZ) region (Saroğlu et al., 1992). Black arrows indicate the GPS-derived surface displacement rate (McClusky et al., 2000; Reilinger et al., 2006). b) İzmit–Düzce segment of the NAFZ. Red lines indicate the İzmıt 1999 surface rupture as mapped by Barka et al. (2002) and blue line indicates the Düzce 1999 surface rupture trace as mapped by Pucci et al. (2006). Hypocenters of the 1999 İzmıt (blue) and Düzce (red) earthquakes are indicated by stars. Maximum surface displacement observed (5.2 m) is located in the Sapanca–Akyazi segment. In the İzmıt–Sapanca Lake segment, the maximum surface displacement observed is 3.5 m and in the Karadere segment 1.5 m. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
compared with stress level change and postseismic slip distribution, and are discussed in the context with the preparation process of the Düzce earthquake.

2. Data and method

We focus on recordings obtained by a 15-station seismic network covering the Sakarya, Karadere and Düzce segments of Izmit rupture zone (Fig. 2a). A long-term seismic network consisting of 15 stations (Fig. 2a, yellow triangles) has been in operation since 1996 [SABONET, (SAPanca–BOlu NETwork, Milkereit et al., 2000; Westerhaus et al., 2002)]. Fig. 2a and b show 2414 of aftershock epicenters and hypocenters along the rupture zone, respectively (Bohnhoff et al., 2007).

Local magnitudes \( (M_l) \) were calculated following Baumbach et al. (2003) who developed a procedure to determine \( M_l \) for NW Turkey that was refined by Bindi et al. (2007) based on an updated attenuation curve and refined station corrections. The procedure involves an automatic estimation of \( M_l \) for all events at each station deconvolving the traces to synthetic Wood–Anderson torsion seismograms using instrument response and maximum horizontal peak amplitudes.

The frequency–magnitude distribution [Eq. (1)] describes the relationship between the frequency of occurrence and the magnitude of earthquakes (Ishimoto and Iida, 1939; Gutenberg and Richter, 1944):

\[
\log_{10} N(M) = a - bM, 
\]

where \( N(M) \) refers to the frequency of earthquakes with magnitudes larger or equal than \( M \). In a semi-logarithmic plot, the constants \( a \) (zero-offset) and \( b \) (slope) can be determined. The slope \( b \) has been interpreted to indicate the presence of asperity, stress and material heterogeneities along the fault plane (Mogi, 1962; Scholz, 1968; Amitrano, 2003; Schorlemmer et al., 2005). The \( b \)-value is calculated

\[
b = \frac{\log_{10}(e)}{\langle M \rangle - (M_c - \Delta M_{\text{bin}}/2)}
\]

Here \(\langle M \rangle\) is the mean of the binned magnitudes, \(M_c\) is the magnitude of completeness and \(\Delta M_{\text{bin}}\) is the binning width of the catalogue [Eq. (2)].

An important issue in assessing data quality is the magnitude of completeness, \(M_c\). Fig. 3a depicts how the \(M_c\) level varies with time for the entire catalog. The completeness level ranges from 0.4 to 1.0. We compute the \(M_c\) threshold automatically for each nodal point, using the change of the slope of the magnitude curve (Wiemer, 2001). To ensure a robust estimation of \(b\)-values, we also restrict the catalog to events above the fixed level of \(M_c\) value (1.5) for \(b\)-values over the entire range of space and time. We observe that \(b\)-values do not change significantly in this case.

Fig. 4. a) Surface map view of standard deviations in \(b\)-values for the catalog using bootstrap approach. b) Map view of \(b\)-values for SC and KDC. Black crosses represent Izmit aftershocks. c) Frequency–magnitude distributions for the marked regions in frame (b).
The \( b \)-value variations are calculated using the public software package ZMAP by Wiemer (2001). For sampling of earthquakes, we use cylindrical volumes perpendicular to the cross-sectional plane and centered at the nodes spaced at 1 km \( \times \) 1 km. The length of these cylindrical volumes is defined by the width of the volume along the cross section, which is 5 km. In each sample, we also require a minimum number of events with \( M \geq M_c \), \( N_{\text{min}} \), in order to determine reliable \( b \)-value. For samples containing fewer than \( N_{\text{min}} \) events, we do not compute \( b \)-values. Here, we arbitrarily set \( N_{\text{min}} = 50 \), because below this value the uncertainty in \( b \)-value increases rapidly (Wiemer and Wyss, 2002; Schorlemmer et al., 2004). The \( b \)-values are calculated in map view (Fig. 4b) and as depth section (Fig. 5b).

We use a bootstrap approach to estimate errors in \( b \) and \( M_c \) (Chernick, 1999). At every grid node we draw its detected number of events from its population, allowing any events to be selected more than once. From these events we compute \( M_c \) and \( b \) for \( M \geq M_c \) and repeat this for every grid node 500 times. We then estimate the errors the standard deviation of \( b \)-values, taking into account the imperfection of the local frequency–magnitude distribution map view (Fig. 4a) and depth section (Fig. 5a) (Schorlemmer et al., 2003; Woessner and Wiemer, 2005).

3. Results

3.1. Sakarya cluster

Fig. 4b shows \( b \)-values in map view varying from 0.6 to 1.4 along the Izmit rupture zone. The \( b \)-value distribution (Fig. 4b) shows strong spatial distributions in the Sakarya Cluster (SC). Different anomalous patches along this cluster can be distinguished. At region A in Fig. 4b, high \( b \)-values (1.31) correlate with high coseismic displacements from field observations by Barka et al. (2002). At region B in Fig. 4b, we find very low \( b \)-values at the intersection of Karadere–Düzce Cluster (KDC) and SC. We compare the \( b \)-values for both regions and find a factor of two difference (Fig. 4c, left). The low \( b \)-value zone (region B in Fig. 4b) extends from northeast of SC to Karadere segment. This very low \( b \)-value (0.65) zone may be locked or high stress concentrated. At SC we identify a volume with high \( b \)-values extending to a depth of 12 km (Fig. 5b). The frequency–magnitude distributions based on the catalog for asperities in the SC illustrate the large \( b \)-value contrast located in two volumes which are in the middle and southeast of SC. In these regions, the \( b \)-values vary from 1.2 to 1.4. For the catalog we obtain errors of 0.05–0.25. In the shallow and active parts, we observed errors of about 0.2–0.25 units, decreasing with depth and having a maximum in the high \( b \)-value region (Figs. 4a and 5a).

The average \( b \)-value of aftershocks as a function of depth between October 17 and November 12, 1999 indicate \( b = 1.0 \) at 4 km depth and decreases gradually to \( b = 0.6 \) at ~6 km depth. At greater depth \( b \)-value increases to \( b = 1.1 \) at 9 km depth (Fig. 6a). A second minimum of \( b \)-value (\( b = 0.9 \)) is evident at along 10 km depth (Fig. 6a).

3.2. Karadere–Düzce cluster

Map of the spatial distribution of the \( b \)-values is computed using Izmit aftershocks for period 25 days before the Düzce mainshock.
(November 12, 1999 $M_w = 7.2$). The high density of aftershocks within the Karadere–Düzce Cluster (KDC) is observed east of KDC (Düzce Basin) ($30.95^\circ$–$31.15^\circ$ in Fig. 4b). The Karadere segment ($30.78^\circ$–$30.94^\circ$ in Fig. 4b) has less activity than the Düzce Basin. Within the Izmit aftershock sequence, these two regions of high activity are revealed by SABONET data (Fig. 2a and b), particularly at the eastern end of the Izmit rupture in the Düzce Basin. The highest $b$-values (region C in Fig. 4b) are found in the Karadere segment. The lowest $b$-values are found in the Düzce Basin (Fig. 4b, region D). A comparison of $b$-values for regions C and D are shown in Fig. 4c (right). The difference in $b$-value is significant ($0.27$). Region C coincides with 1.5 m surface slip (Barka et al., 2002) during the Izmit mainshock. No surface slip was observed in Düzce Basin before the Düzce mainshock on 12 November 1999. According to our results low $b$-values can be related to high Coulomb stress increase in this region (Utkucu et al., 2003). The Düzce Basin is characterized by low $b$-values (0.6) (see Fig. 5b cross-section). For comparison we plot the postseismic slip inversion derived by GPS (Bürgmann et al., 2002) 25 days before the Düzce earthquake (Fig. 5c). 25 days prior to the Düzce mainshock low postseismic slip values in Düzce Basin correlate with low $b$-values (Fig. 5b and c). In Düzce Basin we observed standard deviations of about 0.05–0.15 units for $b$-values (Fig. 5a).

We calculate $b$-value variations as a function of depth at KDC (Fig. 6b) using vertically sliding windows with 50 events. Horizontal error bars indicate the standard deviation while vertical bars indicate the width of the sliding window. The $b$-value is $\approx 0.8$ at 4.5 km depth. Below 7 km depth the $b$-value gradually decreases from $b \approx 1.0$ to $b \approx 0.7$ up to 12 km depth. Below 12 km depth again $b$-value decreases from 1.0 to 0.8. Again, the double minimum of the depth variation of $b$-value as evident from SC (Fig. 6a) is also visible at KDC (Fig. 6b). Hypocentral depth of the Düzce earthquake is 12.5 km. This depth seems to correlate with $b$-value decrease (Fig. 6b).

**4. Discussion**

The $b$-values along the Izmit rupture of the NAFZ vary significantly. High $b$-values at the middle of SC coincide with the maximum surface displacement observed (Figs. 1b and 4b). Southeast of SC, $b$-values are high since this region spreads into the Mudurnu Valley fault where the 1967 earthquake ($M_w = 7.2$) occurred (Barka, 1996). There is one region found with high $b$-values at KDC. This region is associated with the Karadere segment (1.5 m surface displacement).

When mapping spatial variability in $b$-value one has to balance available resolution, with uncertainty in estimate and size of the seismotectonic feature in question. Schorlemmer et al. (2004) investigated $b$-value using a sampling radius ranging from 2 to 20 km. They suggested that sampling with radius $>5$ km mixes populations of dissimilar frequency–magnitude distributions and thus cannot resolve the apparent real structure. For radius $<5$ km no significant additional heterogeneity of $b$-values is detected. We conclude that a radius in the range of 4–5 km is the best choice for this fault segment and data set for identifying and resolving contrasts in $b$-values.

Schorlemmer and Wiemer (2005) found that prior to an earthquake low $b$-values correlate with high stress and low slip of unruptured areas. Taken together this suggests that drastic temporal changes in $b$-values may occur locally around asperities. Low $b$-values in some areas prior to a major seismic event can be followed by high $b$-values after the rupture occurred. The change in $b$-value indicates a coseismic stress drop if $b$-value is used as a stress-meter as suggested by Zang et al. (1998), Lei (2003), Schorlemmer and Wiemer (2005).

The exact spatial extension of the Izmit rupture to either end, i.e. below the Sea of Marmara and in the Düzce Basin is still a matter of debate. GPS results indicate that the western end of the Izmit rupture extended well into the Izmit Bay up to 20 km west of the Hersek Delta...
and may have ended SE of the Princes Islands (Karabulut et al., 2002). At the eastern end of the rupture it remains unclear if the rupture below the surface stopped on the Karadere fault or extended into the Düzce Basin (Reilinger et al., 2000; Wright et al., 2001; Bürgmann et al., 2002; Delouis et al., 2002; Pucci et al., 2006). The hypocenter catalog of Izmit aftershocks presented here indicates a sharp termination of activity towards the eastern end of the rupture (Fig. 2b) where the Düzce earthquake initiated 87 days later (Bohnhoff et al., 2007; Bulut et al., 2007). This boundary correlates with low $b$-values (Fig. 5b).

In laboratory studies (Scholz, 1968; Zang et al., 1998; Lei, 2003) and a variety of tectonic regimes (Wiemer and Katsumata, 1999; Schorlemmer et al., 2005) the frequency–magnitude distribution has been shown to be perturbed by stress and material heterogeneity. Both parameters are very heterogeneous along the Izmit rupture zone. It is conceivable that near the largest slip release the applied shear stress drops significantly during the mainshock, favoring a higher $b$-value for the aftershocks (Wiemer and Katsumata, 1999). For the Izmit aftershock sequence, we observe high $b$-values (=1.31) near large surface displacements.
Areas subject to high applied shear stresses are consistent with the observed low b-values. This is also compatible with observations of lower b-values at highly stressed asperities (Öncel and Wyss, 2000; Westerhaus et al., 2002; Schorlemmer and Wiemer, 2005). We observe low b-values in the Düzce Basin prior to the Düzce mainshock. This observation indicates that the Düzce epicenter area is consistent with the model of Wiemer and Katsumata (1999) which would predict low b-values.

The most likely interpretation of our observation of low b-values at the eastern rim of the Izmit aftershock activity is based on the argument that the Izmit earthquake already ruptured parts of the Düzce fault segment. Comparing the location of our Izmit aftershocks at the eastern end of the rupture (Fig. 7) with the observed surface rupture of the Düzce earthquake (Fig. 7, blue line, from Pucci et al., 2006), it is apparent that half of the Düzce rupture was seismically active already in the Izmit aftershocks sequence.

For Düzce Basin, we find that areas showing low b-values are related to reduced postseismic slip (Fig. 5c) within the uppermost 20 km of the Düzce Basin. According to time dependent aftershot studies, slip migrates from the Izmit mainshock area to the east (Reilinger et al., 2000; Bürgmann et al., 2002). We observe that b-values are gradually decreasing from 7 to 11.5 km depth (Fig. 6b). The low b-values demonstrated here to occur between 7 and 11.5 km depth beneath the KDC suggest a stress concentrator. The depth of the Düzce mainshock (12.5 km) corresponds to within 1 km this stress concentrator. Slip is very low (≈0.0 m) above 20 km (Fig. 5c). At Düzce Basin low b-values at depth correlate with low postseismic slip.

The surface rupture associated with the Düzce earthquake follows the southern boundary of Düzce Basin and re-ruptured half of the part ruptured by the Izmit mainshock. The time span between the first Izmit aftershocks and the Düzce earthquake was 87 days and incorporate the preparation stage of Düzce mainshock. The rupture plane of Düzce earthquake experienced a positive Coulomb stress change and this is clearly consistent with Coulomb triggering (Utkucu et al., 2003). Small aftershocks are important role in preparing a fault zone for failure (Steacy and McCloskey, 1998). Late Izmit aftershocks are found to be pioneer earthquakes of the Düzce rupture. Analysis of b-value at Düzce Basin reveals that low b-values and low postseismic slips coincide with the sharp boundary of aftershocks determined 25 days prior to the Düzce mainshock.

5. Conclusions

By investigating the spatial variations of 2414 aftershocks of the 1999 Izmit earthquake using recordings from a 15-station seismic network covering the Sakarya, Karadere and Düzce regions we found the following conclusions.

Analyzing b-values in space, we identify three asperity regions along the Izmit segment of the North Anatolian Fault zone (Fig. 8). High b-values (≥1.2) were found in the region middle (Asperity 1) and southeast (Asperity 2) of Sakarya Cluster, as well as in the region at the Karadere (Asperity 3) fault. Our observation is in agreement with the Wiemer and Katsumata (1999) rupture model where high slip corresponds to high b-value in the asperity region.

Low b-values in the Düzce Basin correlate with low postseismic slip derived from GPS. The change of b-value with depth in Sakarya Cluster is different from Karadere–Düzce Cluster. Lowest b-value in SC is b=0.7 at 5.5 km depth. Lowest b-value in KDC is b=0.7 at 11.5 km depth.

Pioneer earthquakes (late Izmit aftershocks) at the eastern end of the Izmit rupture indicate that half of the rupture plane activated during the Düzce earthquake was already seismically active after the Izmit earthquake.

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