Spatial variation of aftershock activity across the rupture zone of the 17 August 1999 Izmit earthquake, Turkey

M. Aktar a,b,*, S. Özalaybey b, M. Ergin b, H. Karabulut a, M.-P. Bouin c, C. Tapirdamaz b, F. Biçmen b, A. Yörük b, M. Bouchon d

aBoğaziçi University, Kandilli Observatory and Earthquake Research Institute, Çengelköy, 81220 Istanbul, Turkey
bTÜBİTAK Marmara Research Center, Earth and Marine Sciences Research Institute, P.O. Box 21, Gebze, 41470 Kocaeli, Turkey
cInstitut du Physique du Globe, Université Paris VI Pierre et Marie Curie, 4 Place Jussieu, 75005 Paris, France
dLaboratoire de Géophysique Interne et Tectonophysique, Université Joseph Fourier, P.O. Box 53, 38041 Grenoble, France

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Abstract

We examined the spatial variation in the aftershock activity from the 17 August 1999 Izmit, Turkey earthquake. We found that this aftershock sequence is non-uniform both in space and time, aspects that need to be taken into account in any further statistical analysis. Other aspects of this aftershock sequence are similar to other aftershock sequences, namely low $b$-values and a high degree of spatial variation. We have detected three zones of relatively high $b$-values, two of which coincide with asperities revealed by previous slip inversion studies. The third zone with an anomalous $b$-value is located beyond the fault rupture and indicates a weakened fractured zone in the Yalova-Tuzla area. This $b$-value analysis provided no evidence for any significant difference that may exist between the two sides of the mainshock fault plane.

Keywords: Seismicity; Parameters; Aftershocks; Slip distribution; North Anatolian Fault

1. Introduction

Extensive studies in recent years have shown that earthquake rupture zones have very high spatial heterogeneity in terms of fracture properties, particularly the slip distribution. As more information becomes available about the high-frequency behaviour of the rupture process, detailed models can be developed for describing complex slip histories (e.g., Hartzell et al., 1996, Kamae and Irikura, 1998, Bouchon et al., 2002). It is now clear that three basic properties of coseismic slip are variable: the total amount of slip (static slip), the rupture velocity and the slip velocity. Among those, the static slip is often the most highly constrained and can therefore be mapped in detail. Inversion of both seismological and geodetic data, often used simultaneously, reveals
highly complex patterns of slip variations across fault zones (e.g., Delouis et al., 2002). However, there is little agreement on why exactly and under what conditions these slip patterns are formed. The physical meaning of slip heterogeneity is a matter of major concern for the characterisation of future earthquakes and for making realistic hazard assessments. It is commonly believed that the slip at each point on the rupture depends on the physical conditions at that point. Various physical parameters can be measured in order to determine what may possibly be correlated with the slip heterogeneity (e.g., Öncel and Wyss, 2000). Recently, consistent observations have been reported indicating that the seismicity rate in after-shock sequences is directly correlated with the slip in the mainshock (Wiemer and Weiss, 1997; Wiemer and Katsumata, 1999).

Generally speaking, seismicity is assumed to obey the Richter law; in other words, it shows a power-law distribution. This behaviour is characterized by the slope of the cumulative occurrence against magnitude curve, typically named as the $b$-value. Physically, the $b$-value is an indicator of whether the bulk of the seismic energy is released in a large number of smaller events, or oppositely through a small number of larger events. A high $b$-value means an abundance of smaller events with respect to larger ones. As explicitly proven in laboratory experiments, it is not difficult to imagine that a weak and less resistant environment under stress would produce high number of small events, leading to a high $b$-value. Conversely, a more compact and resistant material under stress would not fail so easily and will thus generate relatively larger and fewer events, leading to a low $b$-value. However, the physical conditions that classify any environment as resistant or weak are more difficult to establish. We identify two distinct factors that determine the strength of a material: the level of the stress to which the material is subjected (Sholtz, 1968) and the inherent physical properties of the constituents of the material itself, such as its heterogeneity (Mogi, 1962). Both have an important role regarding how the material fails when it is subject to shear stress. The state of stress has also been shown to play a major role in determining the character of the magnitude–frequency distribution (e.g., Wyss, 1973; Urbancic et al., 1992; Mori and Abercrombie, 1997; Toda et al., 1998). The complexity of the fault trace (Stirling et al., 1996), and the extent of creep (Amelung and King, 1997), are among other factors that influence the shape of the magnitude–frequency distribution. In this study, we investigate the spatial variation of the $b$-value for the aftershock sequence of the Izmit earthquake (17 August 1999, $M=7.6$). We interpret our results in terms of the conditions that may be responsible for the observed spatial variations. On the one hand, we correlate observed $b$-value patterns with slip distribution models obtained through the inversion of seismological and geodetic data. Like previous studies, we find high $b$-values at asperities, which indicates that the crustal material has been severely crushed due to high slip during the mainshock rupture. In this situation, the intense fracturing and subsequent increase in $b$-value is directly associated with the rupture process itself. On the other hand, we also show that high $b$-values may also indicate reactivation of highly fractured zones, which existed prior to the mainshock. In this situation, the generation of small aftershocks is not directly associated to the mainshock slip itself, but relates instead to a redistribution of the stress field, and possibly also indicates the effect of fluids trapped in small fractures.

2. Data acquisition and processing

The data used in our analysis covers the first 45 days of the aftershock sequence. Two different institutions were independently involved in this data collection process: The Earth Science Research Institute of TUBITAK (Özalaybey et al., 2002), and a combined effort by LGIT (Grenoble) and BÜ Kandilli Observatory and Earthquake Research Institute (Karabulut et al., 2002). As is often the case in aftershock studies, during the first few hours following the mainshock, coverage was rather limited (Wiemer and Katsumata, 1999). About 12 short-period permanent seismograph stations were in operation during this initial period, using L4-C sensors and MIDAS digitizers (Aktar and Bicimen, 1989), recording on a triggered basis. Following the first 10 h after the mainshock, new stations began to be installed. By the end of the second day, this temporary network was completed, with a total of 54 digital stations: 30 vertical-component and 10 three-
component short period sensors; and 14 three-component broadband sensors; all recording through 24-bit Reftek digitizers, using GPS timing and recording continuously. This continuous mode of recording lowered the detection threshold significantly, particularly within the central part of the rupture zone, and enabled the location of events down to $M_l=0.1$. More than 5000 events were recorded and located within the observation period of 45 days. Following this period, the configuration of the temporary network was modified considerably to extend the coverage to a wider area, including the Sea of Marmara Sea and the epicentral area of the Düzce earthquake of November 12. Considering the high decay rate of the aftershock activity, we concluded that around 90% of the total aftershock activity occurred during the initial 45 days and that this period was sufficient to carry out our $b$-value analysis. Further extension of the observation period would not change the overall picture significantly, but would instead introduce non-uniformity in the catalogue.

The uncertainty associated with event location was higher at the initial stage when station coverage was limited. Nonetheless, we note that location errors were never more than 5 km and were generally ~3 km; sufficient accuracy for the purposes of the present study. Local magnitude, $M_L$, is used throughout our analysis to quantify the size of aftershocks and is computed automatically by an offline process that follows the manual epicentre location procedure. Waveform data from all stations with three-component recording was used. To determine $M_L$, synthetic Wood-Anderson seismograms were calculated by deconvolving the instrumental response of each record and convolving the resulting signal with the standard Wood-Anderson torsion seismograph response ($T_0=0.8$ s, damping constant, $h=0.8$ and static magnification, $V=2800$). The peak response at each station is then converted to the local magnitude using the calibration curve of Richter (1958). $M_L$ is then determined by averaging the magnitudes from all stations. An identical procedure is applied routinely by the Earth Science Research Institute of TUBITAK during observation of background seismicity using the permanent stations in this region. We filtered the data to eliminate events with very high location errors. However, we took special care to not exclude any aftershocks with relatively high magnitudes ($M_l>3.5$), by improving their locations using additional data from other networks. The final set of epicentral locations and the stations used are shown in Fig. 1.

### 3. Estimation of $b$-values

Spatial analysis of $b$-values has followed the approach of Wiemer and Katsumata (1999), using the software package (ZMAP) developed by the same author (Wiemer and Zúñiga, 1994). The methodology involves establishing a two-dimensional grid over the plane on which the $b$-value variation is to be mapped. This plane can be chosen to be either horizontal, giving a map view of the $b$-value variation, or vertical, giving a cross-section across the fault plane itself. In this study, the distance between the grid points was chosen as 1 km. The effective resolution of the technique depends on the density of earthquakes surrounding each grid node, which varies with position. A map view of the resolution across the aftershock zone shows that it varies from 3 km, around the highly active Yalova and Akyazi clusters, to 13 km in the quieter zone of the Karadere segment farther east.

Maximum likelihood estimation is used for determining $b$-values. This criterion often gives slightly lower values compared with the least squares approach, but it is found to be more stable. We also computed the differences between the results of both types of estimation methods, which was found to be nearly uniform and never exceeding 0.1. The $b$-value at each node is estimated from the ensemble of the 150 nearest earthquakes, from which a minimum of 45 must be above the completeness threshold.

Completeness of the data is the most critical issue in investigating the $b$-value variation across aftershock zones. Since aftershock observations generally rely on temporary networks installed following large events, detection capability has both spatial and temporal inhomogeneity. This is particularly true for the initial hours or even days following each mainshock. The Izmit earthquake occurred in an area that has been intensively monitored for the past 20 years, so data were available from a significant number of existing permanent stations. As already noted, additional
stations were deployed along the central and western part of the rupture, and were particularly useful for decreasing the detection threshold, as well as for improving location accuracy. Later during the first week, some of the temporary stations were redeployed to more suitable sites as the overall shape of the aftershock zone became clear. This is naturally reflected in the variation of the completeness both in time and space. Fig. 2 illustrates how the completeness level varies with time in three different parts of the activity zones: its western (W), central (C) and eastern (E) parts. It is clear that the completeness level decreases over time at different rates in all three regions by 0.6–0.9 of a magnitude unit. We also note that the final completeness level is lowest in the epicentral region ($M_c=1.1$) and highest in the eastern part ($M_c=1.6$). This shows that the event catalogue includes significant non-uniformity. We have computed the completeness threshold automatically for each nodal point, using the change of the slope of the magnitude curve (Wiemer and Zuñiga, 1994). We have also double-checked our results with fixed level of completeness ($M_c=2.3$), which was set high enough to guarantee robust estimation of the $b$-values over the entire range of time and space. We have observed that the $b$-value patterns that we have identified did not change significantly in either case.

4. Results and discussions

Spatial variation of $b$-value across the Izmit earthquake rupture zone, estimated using the maximum likelihood criterion, has been mapped both in the horizontal and vertical planes (Fig. 3b and c).
Fig. 3. (a) The distribution of the aftershocks. (b, c) Map view (horizontal) and vertical cross-section along the mainshock fault plane, showing the variation of the $b$-value across the aftershock zone. The three zones of relatively high $b$-value are indicated by rectangles A, B and C. (d) The slip distribution across the fault plane as obtained from the inversion of strong motion data by Bouchon et al. (2002). Rectangles show the asperities deduced from this modelling, which coincide with two of the high $b$-value regions in (c).
Inspection reveals that the average $b$-value is fairly low and varies in the range of 0.6–1.0, which is a common characteristic observed in aftershock sequences (Wiemer and Katsumata, 1999). Areas with a relatively high $b$-values ($b$>0.8) occupy only a limited part of the total aftershock zone. From the map view of the $b$-value, we distinguish three such zones of different sizes: the Sapanca-Akyazı zone, the Gölcük zone and the Yalova zone, represented by capital letters from A, B and C in Fig. 3b, respectively, from east to west.

High-resolution slip inversion, based mainly on strong motion data, revealed that rupture in the İzmit earthquake was dominated by the breaking of two major asperities, located 23 km west and 40 km east of the nucleation zone (Fig. 3d) (Bouchon et al., 2002; Delouis et al., 2002). We observe that two of our three high $b$-value zones, namely A and B, coincide with these two inferred asperities, while the third one (C) is outside the rupture zone and thus cannot be related to asperities. The most easterly one (A), which also covers the largest area, corresponds with the eastern asperity located east of Sapanca Lake. This zone is relatively complex due to the imminent eastward branching of the North Anatolian Fault (NAF) (Taymaz et al., 1991; Langridge et al., 2002). Slip partitioning between the Karadere and Mudurnu segments of the NAF has created a highly fractured zone, possibly subject to non-uniform stress patterns, which is also reflected by the lateral spreading of the aftershock zone (Özalaybey et al., 2002; Ito et al., 2002a,b). We explain the high $b$-value in this zone, not only by coseismic shattering of the fault surface due the asperity, but also with a pre-existing highly fractured zone in the vicinity to the main rupture.

The second high $b$-value patch (B) is located at Gölcük, west of the hypocentral area. It overlaps with the second major asperity, which has been verified by nearly all inversion methodologies, including geodetic ones, because very high amounts of slip were observed even at the surface. We note that the $b$-value in zone B reaches 0.8–0.95 but only in a relatively small area. In contrast with the asperity in the east (zone A), the location of this western asperity (zone B) was completely devoid of seismic activity prior to the mainshock (Ergin et al., 1998), which was interpreted as an indication of creep or as a seismic gap. This zone B can instead now be seen to constitute a typical asperity zone: it exhibited seismic quiescence preceding the mainshock; it experienced very high coseismic slip; and this was followed by a high $b$-value during the aftershock activity. This implies a highly compact medium, showing high resistance during interseismic periods, but failing with large slip during the earthquake that, in turn, creates a narrow, highly shattered and fractured zone along the rupture.

The final patch of high $b$-value (zone C) is located at the western extremity of the aftershock zone and does not coincide with any asperity. Indeed, this zone is located beyond the mainshock rupture zone, on the western extension of the NAF into the Sea of Marmara. We note that shallow activity, distinct from the main strand of the NAF, existed prior to the 17 August 1999 İzmit mainshock and was designated as the Yalova swarm (Ergin et al., 1998). Aftershock activity in this zone started 2 days after the mainshock. Many authors have associated it with secondary off-fault structures, with a variety of stress patterns including extensional ones (e.g., Özalaybey et al., 2002; Polat et al., 2002). We suggest that the presence of compressed fluids in a highly fractured zone played a major role in the development of aftershock activity at Yalova. This generation of intense aftershock activity, while the region was compressed by the right lateral slip farther east, suggests that pressured fluids trapped in a highly fractured zone lowered the normal stress, which in turn locally modified the Coulomb failure criterion for the triggering of many small events. The presence of fluid-filled fractured zones in Yalova is supported by the presence of many active hydrothermal springs in this area (Eisenlohr, 1997). Furthermore, we note that zone C extends considerably in the N–S direction, on both sides of the main fault. This may suggest the existence of a zone of weakness at upper crustal level, which spreads across the fault on both sides, from Yalova in the south to Tuzla in the north. We also note that this location corresponds to the splintering of the NAF into the northern and the southern (and possibly more) branches, before entering the Sea of Marmara (Okay et al., 2000; Örgülü and Aktar, 2001). Inversions of geodetic data (GPS and SAR) have shown that the slip during the mainshock decreased gradually westward and did not extend beyond this inferred fractured zone at Yalova-Tuzla. Whether the process of splintering of the main fault and the stopping of the mainshock rupture propagation are somehow interrelated, and
whether they relate to preexisting conditions, such as
the weak zone revealed by high $b$-values, are matters
for further research.

The vertical variation of the $b$-value across the
fault plane is shown in Fig. 3c. Like for similar
analyses of other aftershock sequences, we observed
that the $b$-value tends to decrease with depth,
particularly in asperity zones (Wiemer and Katsu-
mat, 1999). We relate this effect to a change in environ-
mental conditions rather than to the physical proper-
ies of crustal material itself. Namely, we explain this
$b$-value variation as an indication of the increase in
shear stress in deeper parts of the seismogenic zone.
The $b$-value anomaly in zone A extends deeper and
correlates well with the fact that high slip values at the
eastern asperity (A) persist deeper than the western
one (B). We also note that the hypocentral region is a
low $b$-value zone, suggesting that a nucleation in a
‘strong’ environment tends to accumulate enough
energy to sustain propagation of the mainshock
rupture. Fig. 4 illustrates the variation of the magni-
tude–frequency curve along the fault, from west to
east. Only earthquakes located inside the indicated
rectangles are used for estimating these $b$-values.
Different numbers of earthquakes were used in each
case, due both to the variations of size of the
rectangles and to the density of the seismicity in
each. Changes in the slope of the cumulative curves,
namely the $b$-values, can clearly be seen, illustrating
the heterogeneity of this aftershock activity.

Finally, we have tested if the $b$-value properties
could be used as an indicator for possible differences
that may exist between crustal constituents on the
north and south sides of the NAF, as was suggested

![Map view of the aftershock activity](image)

![Magnitude-frequency distribution and b-value of selected areas](image)

Fig. 4. (a) The distribution of the aftershocks again, now labelled to identify five rectangular regions within the aftershock area, for which magnitude–frequency cumulative curves were obtained in (b). Rectangles I, III and V coincide with the high $b$-value sections, whereas II and IV are marked by low $b$-values. Note the number of aftershocks located in each rectangle is different, resulting in different estimates for the variable $a$. (b) The resulting magnitude–frequency cumulative curves, shown with 95% confidence limits (Wiemer and Weiss, 1997).
from other observations (Ben-Zion et al., 2003). For this purpose, we have compared the $b$-values on either side of the fault rupture, for the eastern part of the epicentral area, but found no significant differences ($b_N=0.6$ to the north and $b_S=0.74$ to the south; Fig. 5).

We have observed that the most fractured zones, whether they correspond to preexisting structures in the crust, or are generated coseismically during a major rupture, are associated with high $b$-value seismicity. Analysis of $b$-values for the Izmit earthquake has
provided evidence for both these cases. Like many other aftershock studies before, our results have verified that zones of high $b$-value aftershocks correspond to asperities in mainshock rupture areas. Accordingly, we confirm the existence of a deep asperity zone to the east of Sapanca, which was not clearly revealed by geodetic slip inversion (Wright et al., 2001). Our analysis also provided a useful tool for investigating the upper-crustal properties in nonruptured zones. In this context, the Yalova-Tuzla section is identified as a pre-existing highly fractured zone filled with fluid, which was possibly reactivated by coseismic changes in stress.

References


