IDENTIFICATION OF TRIGGERED EARTHQUAKES IN THE GULF OF GÖKOVA

by

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ABSTRACT

IDENTIFICATION OF TRIGGERED EARTHQUAKES IN THE GULF OF GÖKOVA

Triggering mechanism is described as the consequence of the stress transfer on the faults. Triggered earthquakes can either be close to the triggering one as well as far from it. The study done in this thesis is to search for the static and dynamic triggering effects in the Gulf of Gökova using three different catalogues. One of the catalogues is prepared within this study (July, 2007) and the others are from NEMC and USGS (2001-2009). Triggering effect due to local, regional and global earthquakes are considered. First, a high accuracy analysis is carried out using the catalogue produced within the framework of this thesis for a period of two months. Mean values of stress amplitudes with their variances and standard deviations are evaluated for each seismic station (BLCB, BODT, CETI, DALT, DAT, ELL, MLSB, OREN, TURG and YER). These stress variations in the Gulf are compared with the seismicity rates. Teleseismic events are seen to give signs of triggering. A second analysis is made using a catalogue of longer duration (2001-2009) but lower resolution (completeness, M>2.5). This second analysis also shows evidences for triggering by teleseismic events.

ÖZET

GÖKOVA KÖRFEZİ’NDEKİ TETİKLENMİŞ DEPREMLERİN BELİRLENMESİ

Tetikleme mekanizması, faylardaki gerilme aktarımının bir sonucu olarak tanımlanır. Tetiklenmiş depremler hem tetikleyen depreme yakın hem de ondan uzak olabilirler. Bu tezde yapılan çalışma Gökova Körfezi’ndeki statik ve dinamik tetikleme etkilerini üç farklı katalog kullanarak araştırmaktır. Kataloglardan biri bu çalışma kapsamında hazırlanmış (Temmuz, 2007) ve diğerleri UDIM (Ulusal Deprem İzleme Merkezi) ile USGS’den alınmıştır (2001-2009). Yerel, bölgesel ve global tetikleme etkileri ayrı ayrı dikkate alınmıştır. Önce iki aylık bir süre için bu tez çerçevesinde üretilen katalog kullanılarak yüksek hassasiyetli bir analiz gerçekleştirilmiştir. Gerilim genliklerinin ortalama değerleri, varyansları ve standart sapmalarıyla her sismik istasyon için hesaplanmıştır (BLCB, BODT, CETI, DALT, DAT, ELL, MLSB, OREN, TURG and YER). Körfezdeki bu gerilim değişimleri sismisite değerleri ile karşılaştırılmıştır. Uzak depremlerin tetikleme belirtisi verdiği görülmüştür. İkinci bir analiz daha uzun süreli (2001-2009) ama daha düşük çözümlü (geçerliliği; M>2.5) bir katalog kullanılarak yapılmıştır. Bu ikinci analiz de birincisine benzer şekilde uzak depremler tarafından tetiklenme yönünde kanıtlar gösterir.

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LIST OF SYMBOLS / ABBREVIATIONS

σ_c	Coulomb stress failure
τ	Shear stress
μ	Coefficient of friction
σ_n	Normal stress
p	Pore fluid pressure
$\Delta\sigma_c$	Coulomb stress change
$\Delta\tau$	Shear stress change
$\Delta\sigma_n$	Normal stress change
Δp	Pore fluid pressure change
μ'	Effective coefficient of friction
B	Skempton's coefficient
μ_o	Friction constant
V	Slip rate
V_o	Reference velocity
a	Constant in the rate and state friction equation
b	Constant in the rate and state friction equation
D_c	Critical slip distance
Θ	State variable
$d\,\Theta / dt$	Dietrich-Ruina law
N	Number of events
a	Constant for magnitude-frequency relation
b	Constant for magnitude-frequency relation
M	magnitude minimum
μ/β	shear modulus / shear velocity
$>$	Greater-than
$<$	Less-than
$\leq$	Less-than or equal to
$\geq$	Greater-than or equal to
$^\circ$	Degree sign

~	Tilde
VI	6
VIII	8
IX	9
X	10
M	Magnitude
M_L	Local magnitude
M_d	Duration magnitude
M_S	Surface wave magnitude
M_w	Moment magnitude
BLCB	Balçova seismic station
BODT	Bodrum seismic station
CETI	Çetibeli seismic station
DAT	Datça seismic station
DALT	Dalyan seismic station
ELL	Elmalı seismic station
MLSB	Milas seismic station
OREN	Ören seismic station
OZCA	Özcan seismic station
TURG	Turgutlu seismic station
YER	Yerkesik seismic station
h	hour
s	second
min	minute
amp	amplitude
dist	distance
km	kilometer
hz	hertz
cm / s	centimeter / second
Pa	Pascal
MPa	Mega Pascal
J / m^2	Joule / square meter
m / s	meter / second

kg / m^3	kilogram / cubic meter
i.e.	id est
E-W	East-West
N-E	North-East
N-S	North-South
S-W	South-West
NW-SE	North West-South East
ENE-WSW	East North East-West South West
NNW-SSE	North North West-South South East
WNW-ESE	West North West-East South East
WSW-WNE	West South West-West North East
AD	Anno Domini
BC	Before Christ
GPS	Global Positioning System
KOERI	Kandilli Observatory and Earthquake Research Institute
NAF	North Anatolian Fault
NEMC	National Earthquake Monitoring Centre
PQL II	PASSCAL Quick Look II
TUBITAK	The Scientific and Technological Research Council of Turkey
USGS	U.S. Geological Survey

1. INTRODUCTION

An earthquake which occurred in any region of the world can be followed by subsequent events in other regions which are near or remote. This process is a consequence of stress changes and it is called as triggering (Freed, 2005). For two different stress changes which are named as static and dynamic, triggering mechanism is divided into two titles: Static Triggering and Dynamic Triggering (Freed, 2005). Generally, static triggering can be explained by Coulomb stress failure (Jaeger and Cook, 1979, Scholz, 1990) which points to a slow varying process (Freed, 2005). Dynamic triggering can be explained by the same approach, but it is mostly associated with the passage of seismic surface waves rather than static deformation (Love and Rayleigh waves) (Freed, 2005).

The Gulf of Gökova is a place in the Aegean Region of Anatolia. It is surrounded by Bodrum in the north, Datça in the south and the island of Cos in the west. The Gökova area has an N-S regional extensional tectonic system and seismically active grabens (Uluğ *et al.,* 2005a). All geological, geophysical and seismological studies until today have proved those features and numerous major earthquakes or swarm activities have occurred in the area since the ancient times (Aktar *et al.,* 2006, Dewey and Sengör, 1979, Görür *et al.,* 1995, Gürer and Yılmaz, 2002, Kurt *et al.,* 1999, Uluğ *et al.,* 2005a, Uluğ *et al.,* 2005b, Yılmaz *et al.,* 2000). Furthermore, Gökova is not the only location in western Turkey which is prone to earthquakes. All Aegean Region in fact Anatolia has a potential to produce earthquakes at the same magnitudes or larger than the ones occurred in Gökova. For this reason, stress transfer may play a role in the seismicity of the Gulf of Gökova and around. Recently broadband stations are installed in the area and they make it possible to detect events down to magnitude 2.0 and even lower. It therefore makes a suitable location for studying the triggering mechanism.

The purpose of this thesis is to investigate the triggering effect in the Gulf of Gökova by the use of the distribution of seismicity. Local, regional and global earthquake catalogues have been compared and their stress amplitudes are calculated to determine any triggering event in the study area. Both signatures of static and dynamic triggering processes are searched and a general evaluation is given without implementing more involved statistical methods.

A brief description of static and dynamic triggering of earthquakes is given in Chapter 2 together with examples from the literature. Gökova Bay is described in Chapter 3 in the context of geology, faults and earthquake activity through the years. Chapter 4 describes the data processing, how location and magnitude determination of earthquakes are made and the magnitude-frequency relations are explained. Then in Chapter 5, observations of triggering are explained by the stress variations, seismicity rates and earthquake activity in recent time. The final conclusion is given in Chapter 6.

2. TRIGGERING OF EARTHQUAKES

2.1. Stress Changes

Stress changes (static and dynamic) on the faults due to tectonic processes are the main reason of earthquake occurrences and also the most important factor of triggering. The function of Coulomb stress failure is a fundamental definition to indicate static stress changes and it is given by (Jaeger and Cook, 1979, Scholz, 1990),

$$\sigma_c = \tau - \mu(\sigma_n - p) \qquad (2.1)$$

where τ is shear stress, σ_n is normal stress, p is pore fluid pressure, and μ is the coefficient of friction. It is understood that a failure may happen if there is an increase in the shear stress or a decrease in the effective normal stress (σ_n – p) (or vice versa). Absolute stress values cannot be accurately known, but Coulomb stress change can be calculated by the formula,

$$\Delta\sigma_c = \Delta\tau - \mu(\Delta\sigma_n - \Delta p) \qquad (2.2)$$

Cocco and Rice (2002) consider that pore fluid pressure values are proportional to normal stress changes, so they joined the effective coefficient of friction,

$$\mu' = \mu(1 - B) \qquad (2.3)$$

where B is Skempton’s coefficient, which ranges between 0.5 and 0.9 for rocks (*e.g.*, Roeloffs, 1996). Thus, coseismic Coulomb stress changes can be formed as (Reasenberg and Simpson, 1992)

$$\Delta\sigma_c = \Delta\tau - \mu'\Delta\sigma_n \qquad (2.4)$$

2.2. Static Triggering

Static stress changes after the slip of a fault influence nearby faults, so for ruptures of those faults stress changes can be calculated by using the Coulomb failure criterion. In Figure 2.1, these calculations are illustrated for a right-lateral fault and as shown the shear stress change and normal stress change generate the Coulomb stress change. Warm colours represent positive Coulomb stress change and regions are about to failure. Cool colours represent negative Coulomb stress change and regions are away from rupture. Commonly, negative changes are called to be stress shadow zones (Harris and Simpson, 1996) and only a reduced aftershocks or triggered earthquakes activity is expected to occur in these locations. Lienkaemper *et al.*, (2001) estimate the creep rates along the Hayward Fault and find that stress shadows of the 1989 Loma Prieta Earthquake caused low creep rates on the fault.

Some particular earthquakes together with the aftershock sequence that follows constitute a very good example for the mechanism of Coulomb stress changes and the associated triggering. The 1983 M = 6.5 Coalinga and M = 6.0 Nuñez earthquakes are few of them. Toda and Stein (2002) explain that these two earthquakes delayed the regular sequence of Parkfield earthquakes more than expected (expected one: 2004, M = 6) because of the Coulomb stress decrease on the locked segment of the San Andreas Fault.

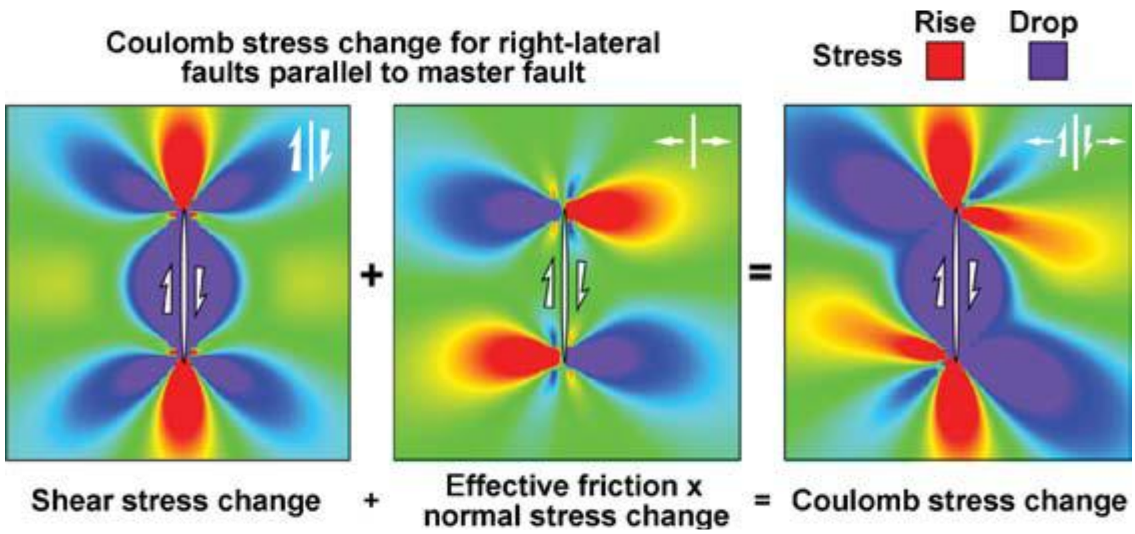

Figure 2.1. Illustration of a Coulomb stress change calculation (Equation 2.4). The panels show a map view of a vertical strike-slip fault embedded in an elastic halfspace, with imposed slip that tapers toward the fault ends. Stress changes are depicted by graded colours; green represents no change in stress (From King *et al.*, 1994)

Some aftershock distributions may not be explained by Coulomb failure. For example, after the Chi-Chi, Taiwan Earthquake, in two stress shadow regions aftershocks occurred intensively (Wang and Chen, 2001) and Marsan (2003) points that some stress shadow zones after the Loma Prieta, Landers, and Northridge earthquakes started to have activity after a period about 100 days.

Background seismicity which is usually taken as the standard level of seismicity for determining activity increase or decrease plays an important role for stress shadows, Coulomb theory and postseismic rates. The activity at any given time may not represent the standard level since it may be due to a previous earthquake and therefore shifted away from its usual interseismic level. Toda and Stein (2003) reveal that the Kagoshima (Japan) earthquakes which occurred subsequently in March and May, 1997 are in a situation opposite to each other. The March event caused important aftershock activity whereas the May event decreased the Coulomb stress in the same region (Figure 2.2).

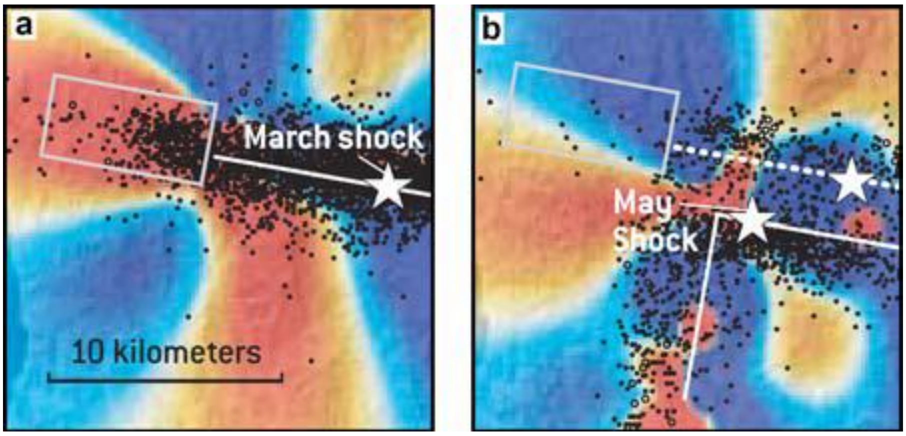

Figure 2.2. The earthquakes of Kagoshima, Japan in 1997. (*a*) A March 1997 earthquake in Kagoshima, Japan increased Coulomb stress and seismicity to the west of the ruptured fault. (*b*) A second earthquake in May decreased stress in that region, leading to a sharp decrease in seismicity (From Toda and Stein, 2003 and Stein, 2003)

Rate and state friction is a kind of equation that is observed in a laboratory and expresses the friction of a fault as a function of slip, time and slip-velocity (Dieterich, 1979, 1981). The equation adopted from Ruina (1983) can be expressed by the formula (Marone, 1998),

$$\mu = \mu_o + a \ln[V/V_o] + b \ln[V_o \Theta / D_c] \tag{2.5}$$

where μ is friction, μ_o is a friction constant, V is the slip rate, V_o is a reference velocity, a and b are constants based on observations, D_c is a critical slip distance for evolution of fault surface contacts, (Dieterich, 1994). Θ is a state variable (Ruina, 1983) which must be combined with Dieterich's law in Equation 2.5 and its behaviour is governed by the dynamics given by equation below, named as Dieterich-Ruina law,

$$d\Theta/dt = 1 - V\Theta/D_c \tag{2.6}$$

Rate and state friction equations should be analysed seismologically. Because of the logarithmic dependence of friction on rate and state, a sudden increase in load on the fault, even of very small amplitude, can rapidly reduce fault friction and bring about an acceleration to failure (Gomberg *et al.*, 1998). In other words, the stress increase has a different triggering effect on different parts of the earthquake cycle of a fault. Therefore, Gomberg *et al.,* (1998) believe that Coulomb stress failure ($\Delta\sigma_c$) needs to be considered in combination with the state of fault respect to its earthquake cycle, in order to evaluate its triggering potential. This means that Coulomb stress failure is best used in combination with rate and state equations. Thus, rate and state friction predicts that the sudden stressing of a population of faults, in which the time of each fault’s earthquake cycle is randomly distributed, will lead to a clustering of triggered seismicity (*e.g*., Gomberg *et al.*, 1998).

Earthquake triggering studies help to reveal whether other events will occur after a mainshock. Stein *et al.,* (1997) study the earthquakes along the North Anatolian Fault (NAF) in Turkey. The period they make calculations is about 60 years and after these calculations, historic events show that all of them are contributed by the previous events. However, they indicate that there have been a high risk for events with major magnitudes on the two sections of the NAF and one of them is the M = 7.4 Izmit Earthquake.

2.3. Dynamic Triggering

Besides static stress changes, dynamic stress changes may also account for triggering. Thus, dynamic (transient) stresses may trigger both near-field and far-field earthquakes. After a large earthquake, more earthquakes begin to happen and the seismicity becomes higher than the previous state. These sequential events usually have some time-delay. However, delay times can range from seconds to years. California is a considerably fine example for it: The M_w = 6.1 Joshua Tree Earthquake in April, 1992 was followed by the 1992 M_w = 7.3 Landers Earthquake in June and then the M_w = 6.3 Big Bear Earthquake came only after 3 1⁄2 h (Hauksson *et al.*, 1993). With a delay time of seven years, the M_w = 7.1 Hector Mine Earthquake (Dreger and Kaverina, 2000) occurred in October, 1999. The best way for explaining dynamic triggering is to correlate the distribution of triggered events and rupture directivity (Freed, 2005).

Hill *et al.,* (1993) indicate that some of the dynamically triggered activity after the Landers Earthquake was crucially related to geothermal activity or volcanic activity, for instance, the Geysers geothermal area in northern California, Coso Hot Springs and Long Valley Caldera. Anderson *et al.,* (1994) believe that around many active volcanoes there are some triggered earthquakes following Landers, but only a small section of the triggered events associate with volcanism.

2.3.1. Far-Field Dynamic Triggering

After the 1992 Landers Earthquake, seismic activity increased rapidly at remote distances outside the aftershock zone which are more than 1000 km from the Landers epicentre (Hill *et al.,* 1993, Anderson *et al.,* 1994). It contains many clusters along the boundary between the Basin and Range and the Sierra Nevada (at distances up to 650 km), the Geysers geothermal field (750 km away), the southern Cascade range (900 km away), western Idaho (1100 km away), and Yellowstone National Park (more than 1250 km away) (Hill *et al.,* 1993). Majority of these triggered events had magnitudes from 1 to 3, but there were a number of M>4 events and M_s = 5.6 earthquake at Little Skull Mountain in southern Nevada (240 km from Landers) only 22 h later following the Landers Earthquake (Anderson *et al.,* 1994, Gomberg and Bodin, 1994) and is considered to be the largest earthquake in southern Basin and Range since 1868 (Hill *et al.,* 1993).

To compare the seismicity rates before and after the Landers Earthquake, Hill *et al.,* (1993) plot the cumulative numbers of earthquakes for the selected sites in the western United States in 1992 (Figure 2.3). Day 180 shows a significant increase with the Landers Earthquake and this increase can be a proof for remote dynamic triggering. Stark and Davis (1996) note that the 1989 M = 7.1 Loma Prieta and the M = 7.1 Petrolia earthquakes did not trigger extensively, but they triggered microearthquakes ($M \leq 2$) at the Geysers geothermal region in northern California's Coast Ranges.

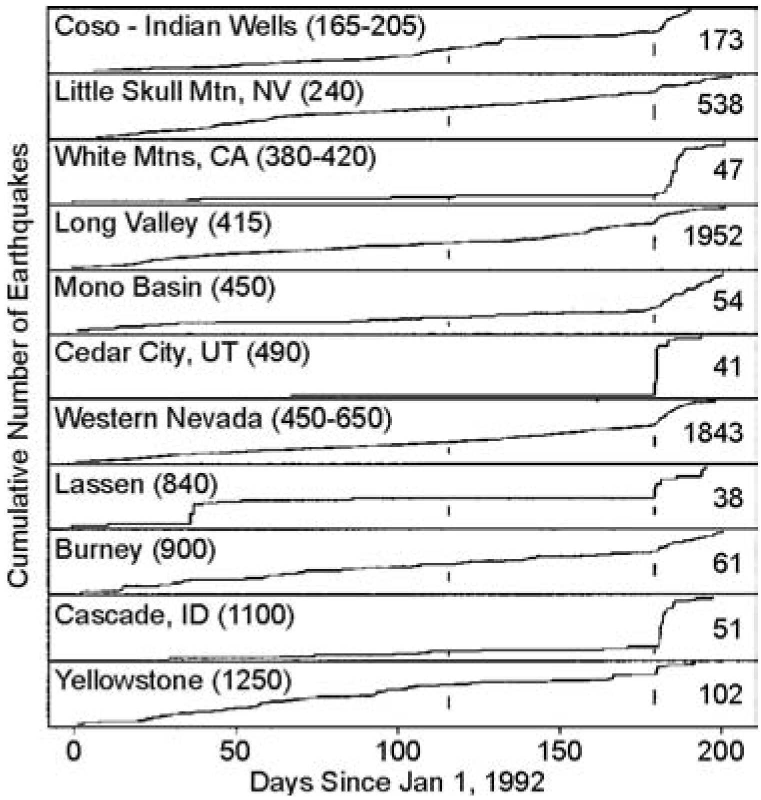

Figure 2.3. Cumulative number of locatable earthquakes in selected zones, beginning January 1, 1992. Numbers in parentheses are distances (in kilometers) from the Landers epicentre. Total number of earthquakes in each zone is shown at right. Vertical lines mark times of the M = 7.1 Petrolia (Cape Mendocino) and M = 7.3 Landers earthquakes (From Hill *et al.,* 1993)

1999 M_w = 7.1 Hector Mine Earthquake also caused an increased seismicity along the southern California and into northern Mexico (Gomberg *et al.,* 2001). Triggered events included two moderate earthquakes near the Salton Sea had magnitudes of 4.7 and 4.4 and they were triggered 30 s and 10 min later, respectively (Hough and Kanamori, 2002). Before the Hector Mine Earthquake, on August 17, 1999, the M_w = 7.4 Izmit Earthquake occurred. However, Brodsky *et al.,* (2000) focused on Greece for increased small earthquakes following the Izmit Earthquake. In Figure 2.4, recordings of the network of the Department of Geophysics of the University of Thessaloniki are shown. They include the period between January 1, 1999 and October 9, 1999. On the day of the mainshock of Izmit (Julian day 229), an increased activity can be observable and this seismicity is approximately 400 km away from the epicentre of the quake. To indicate that more effectively, the seismicity was mapped as seen in Figure 2.5 and those maps also show a relation between the earthquakes in Greece and the Izmit Earthquake.

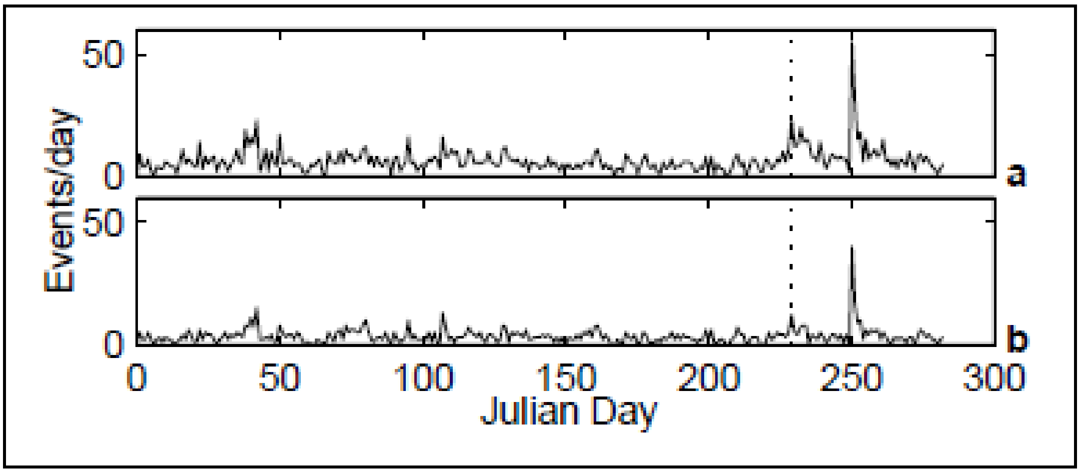

Figure 2.4. Number of earthquakes per day recorded by Thessaloniki network west of longitude 25°. (a) Events of M ≥ 2. (b) Events of M ≥ 3.5. The dotted line indicates the date of the Izmit Earthquake. The peak on day 250 is caused by the Athens M_w = 5.8 event and its aftershocks (From Brodsky *et al.*, 2000)

For the Thessaloniki catalogue, Brodsky *et al.*, (2000) determine a magnitude threshold and they do not take into account the events below M = 3.5. They make some statistical calculations by using different periods or magnitude intervals. Thus, they find that the twelve or more events with M ≥ 3.5 on day 229 have the probability of occurrence related to the Izmit earthquake at ~3 per cent level. For a large catalogue of 1988 to 1998 or whole events with M ≥ 2, they again calculate statistics and it results in the way that the increase on day 229 is not a coincidence since the probability is at 95 per cent.

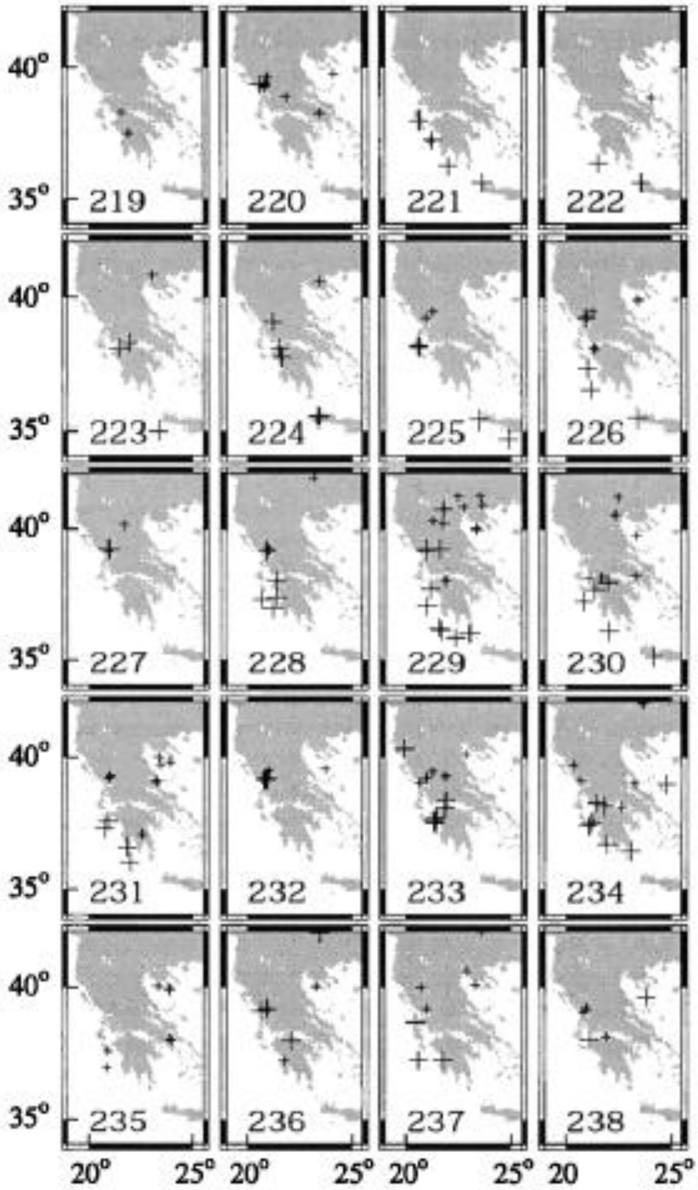

Figure 2.5. Map of events recorded by the Thessaloniki network west of 25° during the days surrounding the Izmit Earthquake (Julian day 229). The large crosses are events with M ≥ 3.5 and the small crosses indicate 3.5>M ≥ 2 (From Brodsky *et al.*, 2000)

As already mentioned, dynamically triggered activity is commonly observed in geothermal and volcanic areas (Hill *et al.,* 1993) and in regions which already have an active background of seismicity (Anderson *et al.,* 1994). Brodsky *et al.,* (2000) relocate the earthquakes in the Thessaloniki catalogue and they are shown on the map in Figure 2.6. The map also shows the background activity between the years of 1961 and 1998. The earthquakes are taken from the catalogue of the Council of National Seismic System and the events with $M \geq 4$ are chosen.

Instead of making statistical estimations by using the distribution of earthquakes Brodsky *et al.,* (2000) concentrate on two significant clusters and they are the ones which are near Arta and Pirgos. These areas are grabens and also have thermal springs (Waring, 1965). The Arta cluster region labelled with ‘a’ in Figure 2.6 does not have much activity before, so in this example triggered earthquakes do not occur in the most active places. In this case, Brodsky *et al.,* (2000) change the assumption of Anderson *et al.,* (1994). On the other hand, Pirgos cluster region labelled with ‘b’ had an earlier seismicity as observed in all previous cases.

The 2002 $M = 7.9$ Denali, Alaska Earthquake is another important event for remote triggering after the Landers Earthquake. The triggered far-field seismicity of the Denali Earthquake was located from British Columbia to southern California (Figure 2.7) (Eberhart-Phillips *et al.*, 2003, Gomberg *et al.,* 2004, Prejean *et al.*, 2004, Pankow *et al.,* 2004) and as soon as seismic waves passed triggered earthquakes began to occur in many areas, lasting from several minutes to several days (Eberhart-Phillips *et al.,* 2003, Gomberg *et al.,* 2004, Prejean *et al.*, 2004).

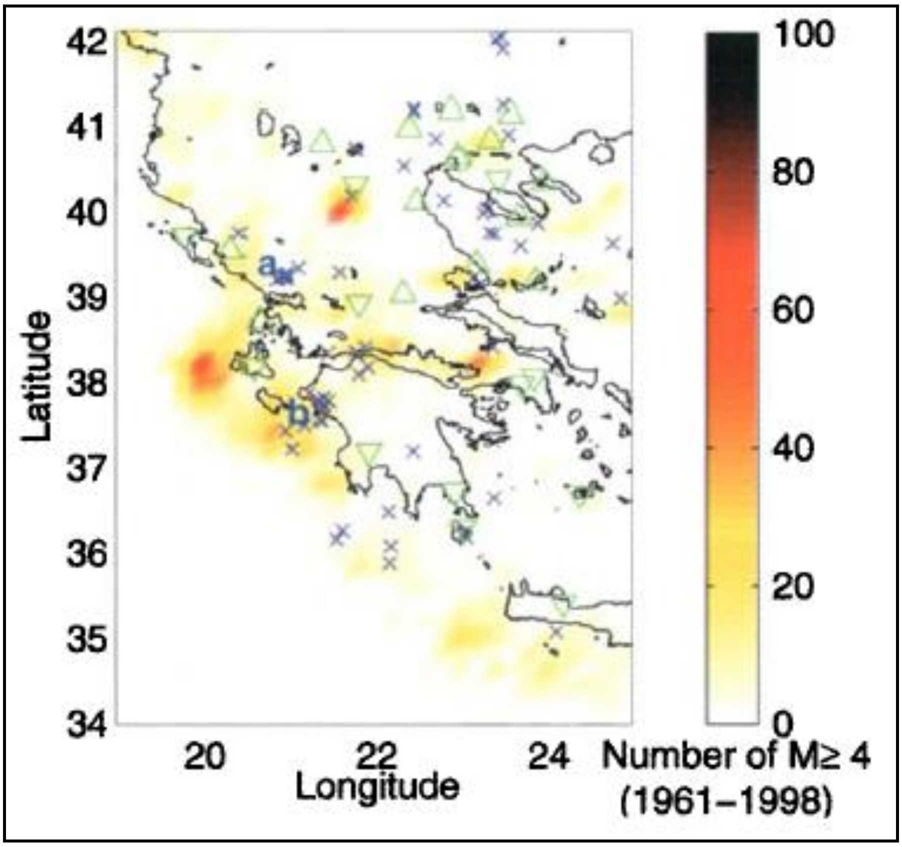

Figure 2.6. Plot of relocated events with M ≥ 2 from 8/17/99-8/22/99 inclusive (crosses) and background activity (colorbar). The background seismicity is binned into 0.2° x 0.2° cells and then smoothed by linear interpolation between cells. Seismic network stations are plotted in green. Thessaloniki stations are upward pointing triangles, Athens stations are downward pointing triangles and station 7905 is a circle. Groups of events are labelled (a) Arta cluster (b) Pirgos cluster (From Brodsky *et al.*, 2000)

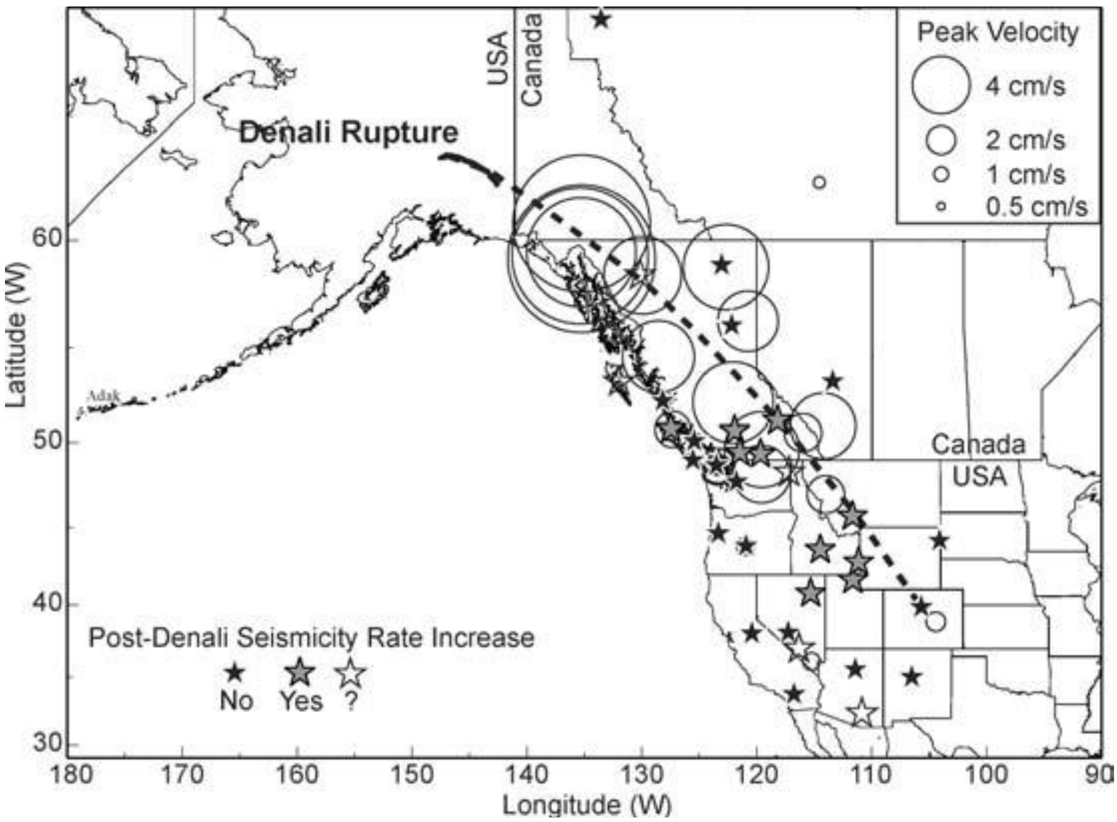

Figure 2.7. Map of the distribution of Denali-related seismicity rate changes and measured peak Denali seismic ground velocities. The latter are proportional to the dynamic strains at these distances. Circle centres show locations of recording sites with radii proportional to the measured peak velocity (scale in inset). The velocities decrease away from the direction of theoretically expected maximum seismic radiation (*dashed line segment of a great circle*). Notably, the sites of triggered rate increases lie roughly within the same azimuthal span as is defined by the maximum measured velocities and theoretically expected dynamic deformations (From Gomberg *et al.*, 2004)

2.3.2. Near-Field Dynamic Triggering

Only dynamic stress changes are responsible for remotely triggered earthquakes, but for the near-field earthquakes both of dynamic and static stress changes play an important role in triggering. Kilb *et al.*, (2000, 2002) point out that there should be a relation between the increased seismicity and peak transient stresses. Thus, it can be proved if dynamic stresses cause aftershocks or not.

The models of seismicity rate change originated by 15 earthquakes with the dynamic deformation fields are compared by Gomberg *et al.,* (2003). Eight earthquakes with unilateral rupture have greater near-field seismicity rate increases in the rupture direction. Since only dynamic deformations should be due to rupture directivity, this result is compatible with the theory that dynamic stresses are important for the distribution of aftershocks within a fault length. In an equivalent study, Kilb *et al.*, (2002b) suggest that large peak dynamic stresses could have been effective in the triggering of the Hector Mine Earthquake after the Landers Earthquake.

As it is clearly seen from examples given above, the triggering mechanism is extremely diverse and complex. The changing stress conditions, summarized by the Coulomb failure criteria and the rate and state equations give only a general theoretical framework where triggering mechanism can start to be interpreted. However, many details still remain obscure. For example, the Coulomb criterion does not give an explanation why the triggering does not occur at locations within close distance to the rupture plane where stress increases are largest, but away from it. The examples above show that triggering mechanism occurs at very different distances and at very different interval times. There is no single procedure which would help to determine whether an event is a triggered one or not. One often has to use a large variety of intuitive or statistical tools to determine the existence of a triggering process. In this thesis, the analysis is mostly based on the intuitive evaluation of the data for the detection of the triggering mechanism if it exists. However, in a more concise application, more objective approaches utilising statistical methods should be used.

3. STUDY AREA

3.1. The Gulf of Gökova and Surroundings

3.1.1. Geology

The Anatolian Plate is located in the Alpine-Himalayan orogenic belt and together with Aegean Sea, western Anatolia is a tectonically significant region of Turkey. General interpretation of the Aegean Sea is being an active extensional region developed as a result of an active subduction along the Hellenic Arc. The collision of the Arabian-African and Eurasian plates creates the Bitlis-Zagros Suture Zone and combined with the slab pull on the west causes the movement of the Anatolian Plate to the west (Dewey and Şengör, 1979). However, some other researchers believe that the eastern Mediterranean subduction is the only cause for the N-S extension (Figure 3.1) (Le Pichon and Angelier, 1981). Recent Aegean deformation is explained by the relative motions of four micro plates and the presence of these micro plates is proved by comparing the observed and the model generated GPS data (Nyst and Thatcher, 2004). Hence, results show that Aegean region has the most important deformations in very narrow zones which are from 10 to 100 kilometers (Nyst and Thatcher, 2004).

The Gulf of Gökova is located in a region which includes many rifts and E-W trending grabens such as Gökova, Büyük Menderes, Küçük Menderes, Gediz and Simav (Uluğ, 2005a). The oldest basin in the Gökova region is Kale-Tavas Basin (Şengör and Yılmaz, 1981) which is orientated ENE-WSW (Gürer and Yılmaz, 2002). The youngest basin is the Gökova Graben and between these basins, approximately N-S trending basins are developed (Gürer and Yılmaz, 2002). The Kale-Tavas Basin is mostly observed from northeast of Denizli in the east, to the Gökova Graben in the west (Figure 3.2) (Gürer and Yılmaz, 2002). The Gökova Graben is one of the most significant one of western Anatolia which is 150 km-long, mostly offshore and forming the Gulf of Gökova (Gürer and Yılmaz, 2002). The Ören area is in the north of the Gökova Graben and has large Neogene outcrops (Gürer and Yılmaz, 2002). It is called the Ören Basin, located between the Gulf of Gökova and Milas (Figure 3.2) (Gürer and Yılmaz, 2002). The Ören and Yatağan basins trend approximately NNW-SSE and their evolution mostly consists of oblique-slip fault systems (Gürer and Yılmaz, 2002). The basin sediments and their sublayers are separated by E-W trending normal faults nearby the Gulf of Gökova (Gürer and Yılmaz, 2002).

In the Gökova region, two rift systems developed through complex tectonic processes concerned with the evolution of southwestern Anatolia (Görür *et al*., 1995). The first system consists of NW-SE orientated rifts and small grabens and the second system which is described as the large E-W trending Gökova rift cuts across the first one (Görür *et al*., 1995). Continental sedimentation is formed in the rifts and grabens of the first system, but marine sediments are also formed in the younger system (Figure 3.3) (Görür *et al*., 1995).

The fault systems of western and southwest Anatolia were developed in conjunction with the major graben and rift structures. The multi-channel seismic reflection study of Kurt *et al*., (1999) argue that the major fault of the area is an E-W orientated listric fault, located south of the Gulf, dipping north and parallel to the southern coastline. They also point out the existence of south dipping antithetic faults on the northern part of the Gulf. These faults are also active and some of them reach the surface on the land although their deep extension is offshore (Figure 3.3) (Kurt *et al*., 1999). In the southwest of the gulf, there is another fault in the sea which is not active and that fault is a normal listric fault called as Datça Fault (Kurt *et al*., 1999).

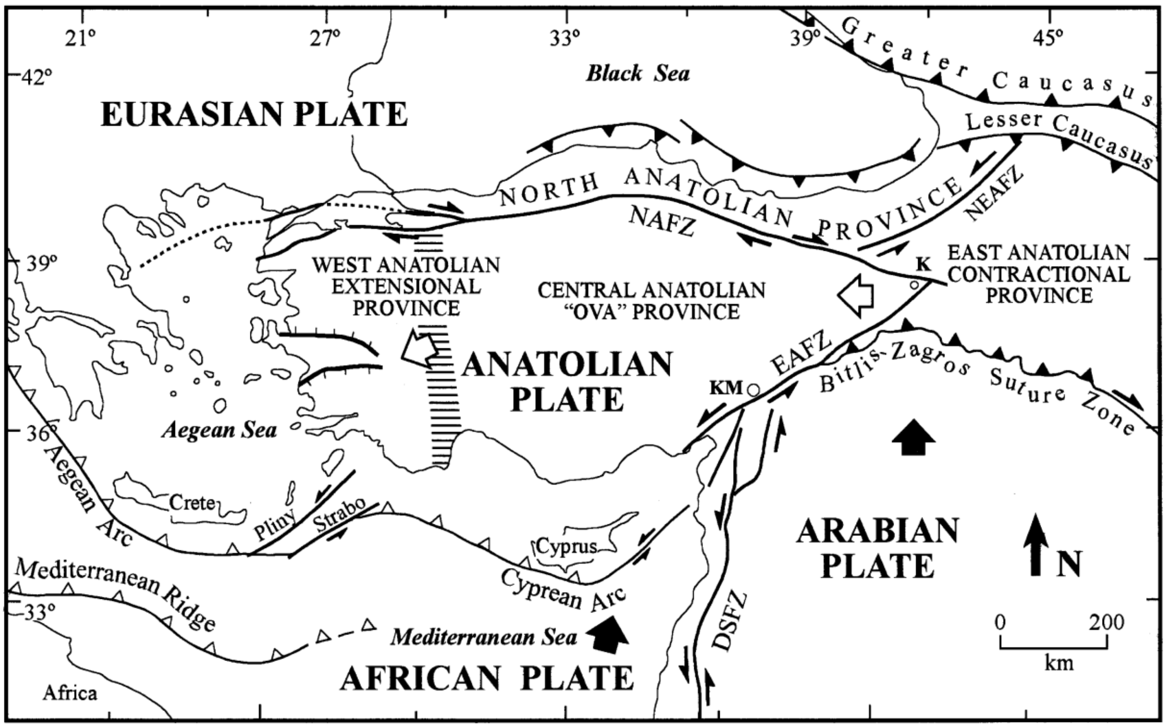

Figure 3.1. Simplified tectonic map of Turkey showing major neotectonic structures and neotectonic provinces (from Şengör *et al.*, 1985, Barka, 1992). K – Karlıova, KM – Kahramanmaraş, DSFZ – Dead Sea Fault Zone, EAFZ – East Anatolian Fault Zone, NAFZ – North Anatolian Fault Zone, NEAFZ – Northeast Anatolian Fault Zone. Heavy lines with half arrows are strike-slip faults with arrows showing relative movement sense. Heavy lines with filled triangles shows major fold and thrust belt: small triangles indicate direction of vergence. Heavy lines with open triangles indicate an active subduction zone, its polarity indicated by the tip of small triangles. The heavy lines with hachure show normal faults: hachures indicate down-thrown side. Bold filled arrows indicate relative movement direction of African and Arabian plates; open arrows, relative motion of Anatolian Plate. Short arrows show the sense of plate motion, half arrows the relative motion senses on strike-slip Faults. The hatched area shows the transition zone between the western Anatolian extensional province and the central Anatolian '*ova*' province from Şengör *et al.,* 1985

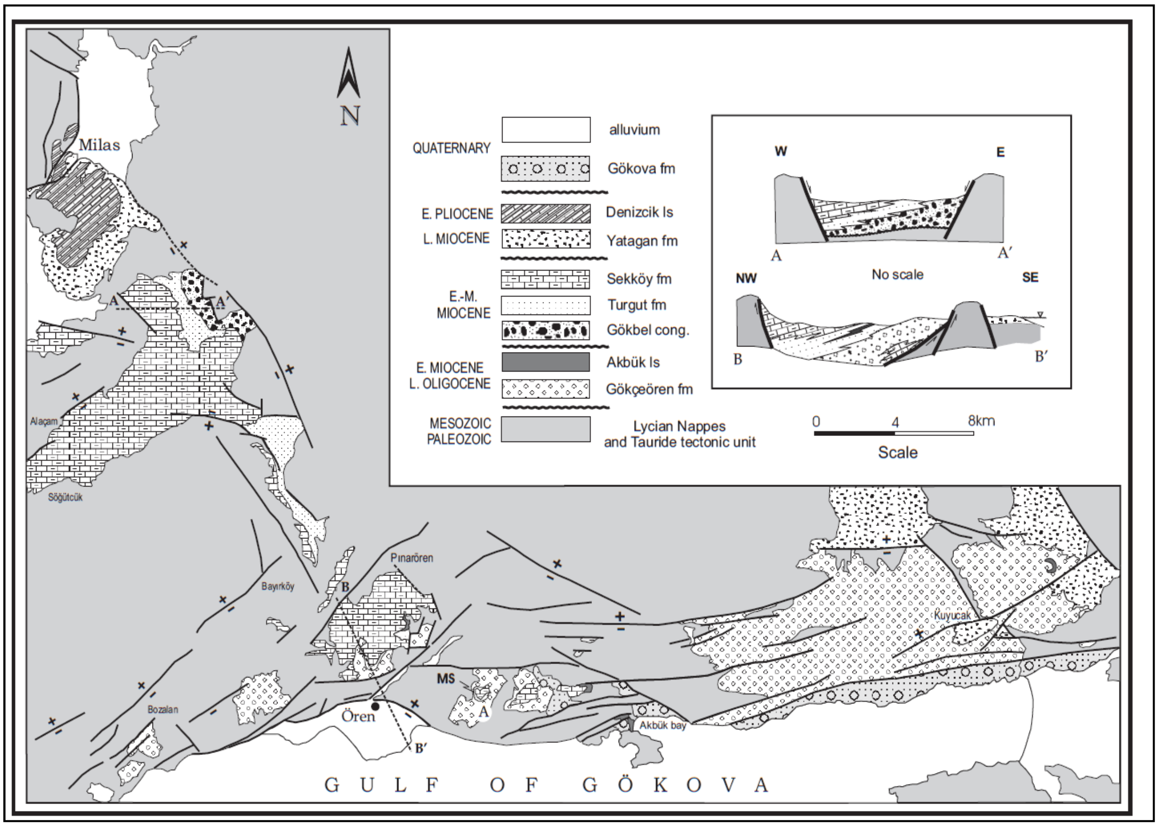

Figure 3.2. Simplified geological map of the Ören graben (Yılmaz *et al*., 2000) and associated cross sections

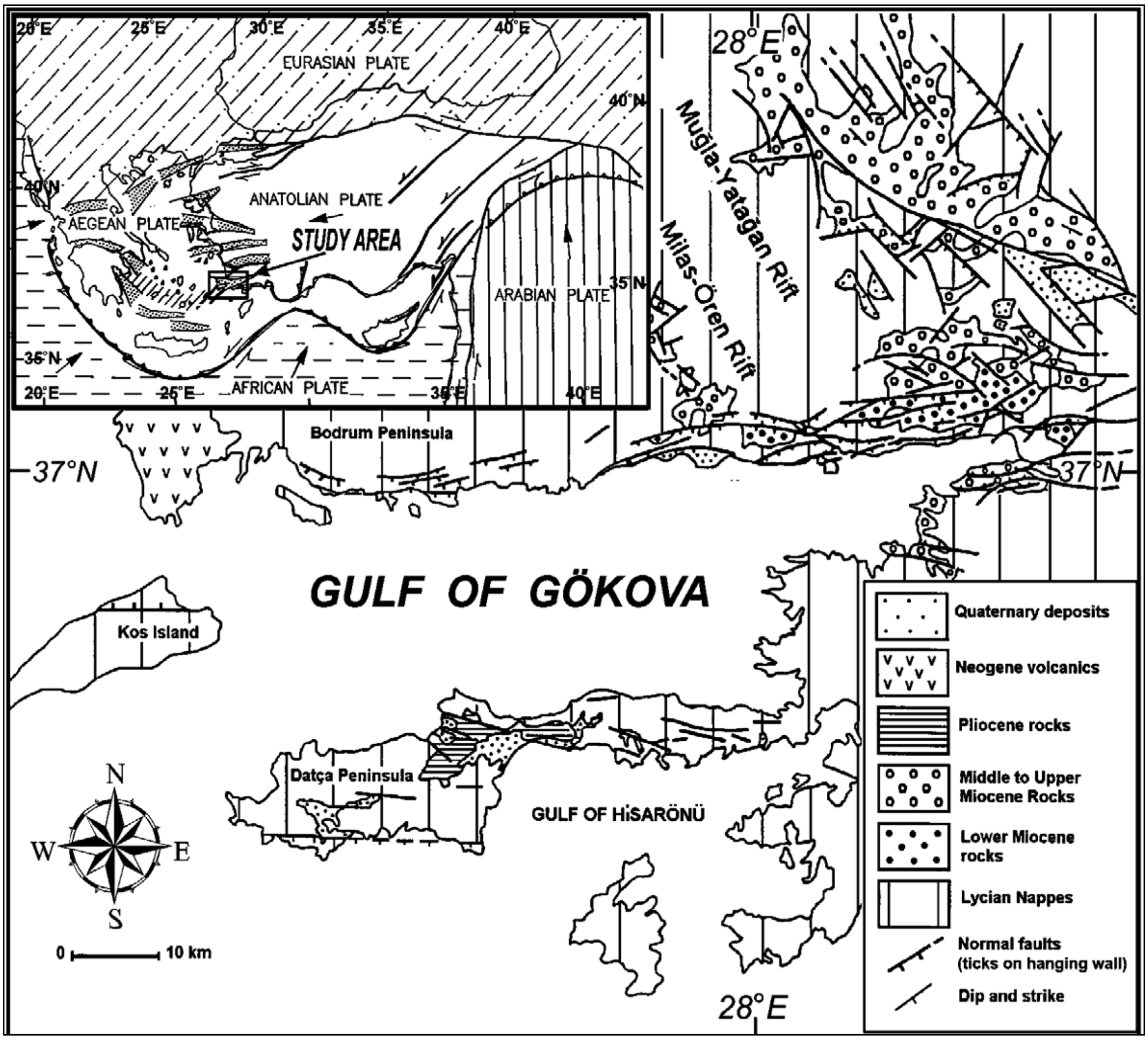

Figure 3.3. Land geology map of the Gökova province (from Uluğ *et al.,* 2005a) (modified from Görür *et al.,* 1995). Inset shows the location of the study area and tectonic framework of the Aegean Sea and surrounding regions, compiled from Şengör and Yılmaz (1981), Hancock and Barka (1981) and Dewey *et al.*, (1986)

Another study was made by using shallow seismic reflection data and it is suggested that faulting of the Gulf of Gökova is mainly related to the E-W oriented, Datça Fault (Uluğ *et al*., 2005a). Secondly, Uluğ *et al.*, (2005a) observe WNW-ESE trending subgrabens in the centre of the gulf and indicate a major WSW-WNE normal fault in the northwest part. Finally, they propose a younger and N-E directed Gökova Transfer Fault, located in the centre of the gulf. Its motion is strike-slip and parallel to the collision direction of the Aegean-Anatolian and African plates (Uluğ *et al*., 2005a).

3.1.2. Earthquake Activity

The graben systems of western Anatolia produced great number of earthquakes along history and many of them caused serious damage (Figure 3.4). When it is searched for catalogues of historical and instrumental periods, it is apparent that major earthquakes happened quite often.

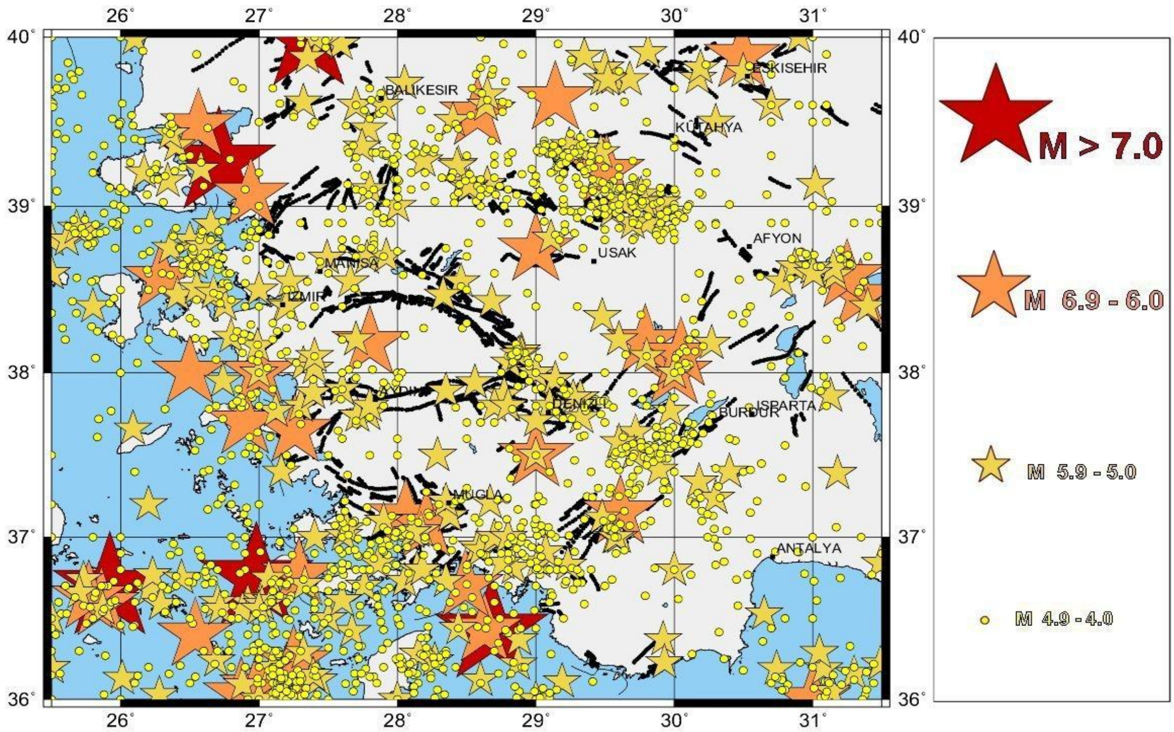

Figure 3.4. Distribution of earthquakes in Western Anatolia (KOERI)

In the Hellenic-Roman Period, from 411 BC to 155 AD there are mentions of many earthquakes in the historical records (Table 3.1). However, from 1304 until 1900, there is a well kept record of intensive earthquake activity around the Gulf of Gökova, causing great damage (Table 3.1). The settlements on island of Kos were destroyed many times and inhabitants had to immigrate to mainland or other islands. It is very likely that the western extension of the normal faults in the Gulf of Gökova were the major cause of these historical disasters. There are also many earthquakes in the 20th Century as shown on Figure 3.5 and listed in Table 3.1.

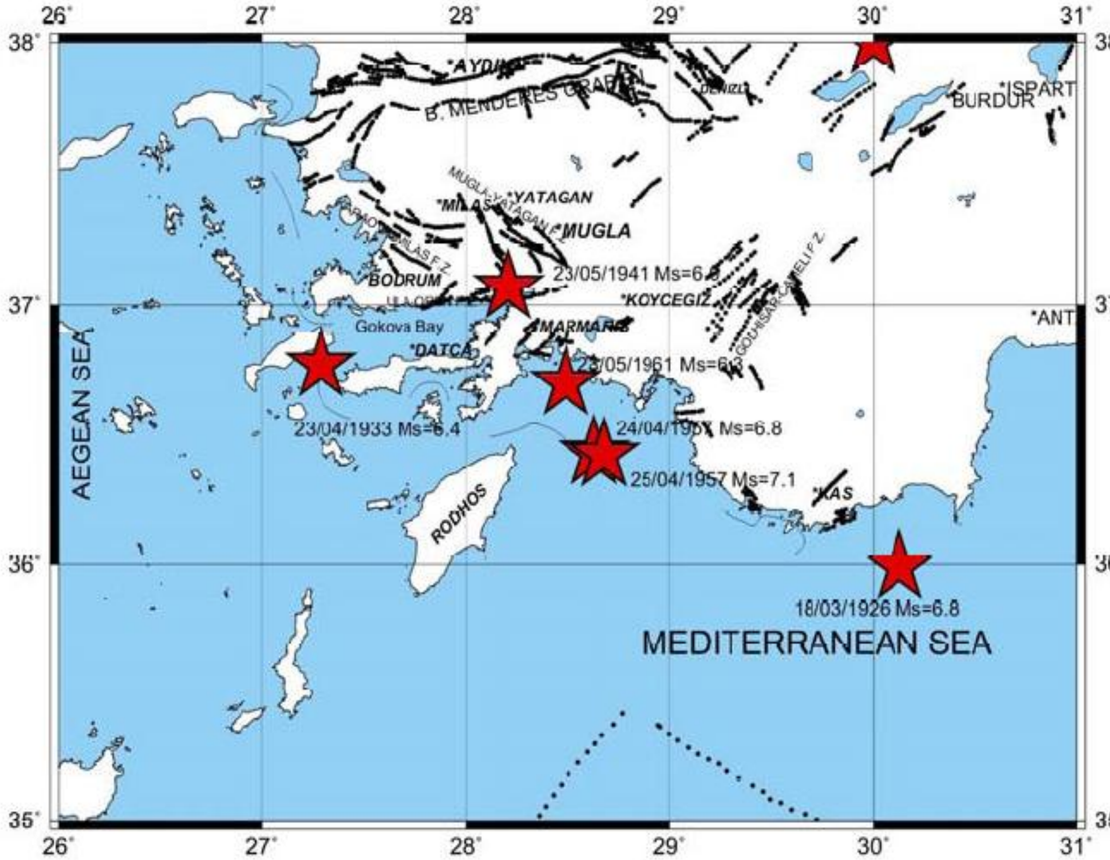

Figure 3.5. Some of major earthquakes around Gulf of Gökova in the 20th century (KOERI)

In 2004, earthquake activity in the Gulf of Gökova began to increase intensively. Beginning from 2 August, 2004 many earthquakes which their magnitudes are between 3.0 and 5.5 were recorded. This earthquake sequence occurred between S-W of Bodrum and 3-5 km offshore south of Yalıçiftlik. It is clear that the area has the potential of producing large destructive earthquakes that constitute a treat to the rapidly increasing population of Bodrum, Akyaka and Ören. Medium size events occur nearly every decade in addition to swarm activities which continue for long times, so it is clear that the crustal stress is constantly at a subcritical level, which makes Gulf of Gökova an ideal location for studying triggering activity.

Table 3.1. Earthquakes occurred around the Gulf of Gökova during historical times

DATE	**LATITUDE**	**LONGITUDE**	**INTENSITY**	**M**	**LOCATION**	**REFERENCE**
411 BC	_	_	_	M_S = 7.0	Gulf of Güllük	Uluğ *et al.*, 2005b
227 BC	_	_	_	_	Near Gökova	Guidoboni *et al.*, 1994
222 BC	36.50	28.00	X	_	Rhodes-Cyprus	KOERI
199 BC	_	_	_	_	Near Gökova	Guidoboni *et al.*, 1994
198 BC	_	_	_	_	Near Gökova	Guidoboni *et al.*, 1994
197 BC	_	_	_	_	Near Gökova	Uluğ *et al.*, 2005b
185 BC	36.00	28.00	IX	_	Rhodes-Cyprus	KOERI
183 BC	_	_	_	_	Near Gökova	Uluğ *et al.*, 2005b
142	_	_	_	_	Near Gökova	Guidoboni *et al.*, 1994
144	_	_	_	_	Near Gökova	Guidoboni *et al.*, 1994
155	36.30	28.00	X	_	Rhodes-Muğla-Fethiye	KOERI
08.08.1304	36.50	27.50	X	_	Rhodes-Crete-Cyprus	KOERI
03.10.1481	36.00	28.00	IX	_	Rhodes-SW Anatolia	KOERI
18.08.1493	36.75	27.00	IX	_	Cos Island	KOERI
1570	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1571	_	_	_	_	Cos Island	Ambraseys and Finkel, 2006
1616	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1660	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1673	_	_	_	_	Cos Island	Ambraseys and Finkel, 2006
1685	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1686	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1714	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1741	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
1776	_	_	_	_	Rhodes	Ambraseys and Finkel, 2006
18.10.1843	36.25	27.50	IX	_	Rhodes-Aegean Sea	KOERI
28.02.1851	36.50	29.10	IX	– _	Fethiye-Muğla-Rhodes	KOERI
12.10.1856	36.25	28.00	X	_	Rhodes-Karpathos-Crete	KOERI
13.11.1856	38.25	26.25	IX	_	Rhodes-Aegean Sea	KOERI
22.04.1863	36.50	28.00	IX	_	Rhodes	KOERI
1865	_	_	_	_	Near Gökova	Uluğ *et al.*, 2005b
1869	_	_	_	_	Near Gökova	Uluğ *et al.*, 2005b
29.02.1885	37.20	27.20	IX	_	Aegean Sea	KOERI
27.08.1886	_	_	_	_	Bodrum	Uluğ *et al.*, 2005b

1896	_	_	_	_	Near Gökova	Uluğ *et al.*, 2005b
26.06.1926	36.54	27.33	_	M_s = 7.7	Cos Island-Rhodes	KOERI
23.04.1933	36.77	27.29	_	M_s = 6.4	Gökova-Cos Island	KOERI
23.05.1941	37.07	28.21	VIII	M_s = 6.0	Muğla	KOERI
13.12.1941	37.13	28,06	VIII	M_s = 6.5	Muğla	KOERI
24.04.1957	36.43	28.63	IX	M_s = 6.8	Fethiye-Rhodes (Muğla)	KOERI
25.04.1957	36.42	28.68	VIII	M_s = 7.1	Fethiye-Rhodes (Muğla)	KOERI
25.04.1959	36.94	28.58	VIII	M_s = 5.9	Köyceğiz (Muğla)	KOERI
23.05.1961	36.70	28.49	VIII	M_s = 6.3	Fethiye-Rhodes (Muğla)	KOERI
05.10.1999	36.75	28.74	VI	M_d = 5.2	Marmaris	KOERI

4. DATA PROCESSING AND SEISMICITY OF GÖKOVA

4.1. Data and Solutions of Earthquakes

The data used for this thesis are obtained from eleven broadband stations near Gulf of Gökova. Some of them are temporary stations (CETI, OREN, OZCA, TURG) installed within the framework, a TUBITAK project (Aktar *et al*., 2006) and others are the stations of NEMC, Kandilli Observatory (BLCB, BODT, DALT, DAT, ELL, MLSB, YER). BLCB and ELL do not take place on the map but they are used for some solutions (Figure 4.1). At all temporary stations, the types of seismometers are GURALP CMG-6T and their frequency band is between 0.03-25 Hz. GURALP CMG-3T ESP typed seismometers are used at the permanent stations and their frequency band is between 0.012-25 Hz (Aktar *et al*., 2006).

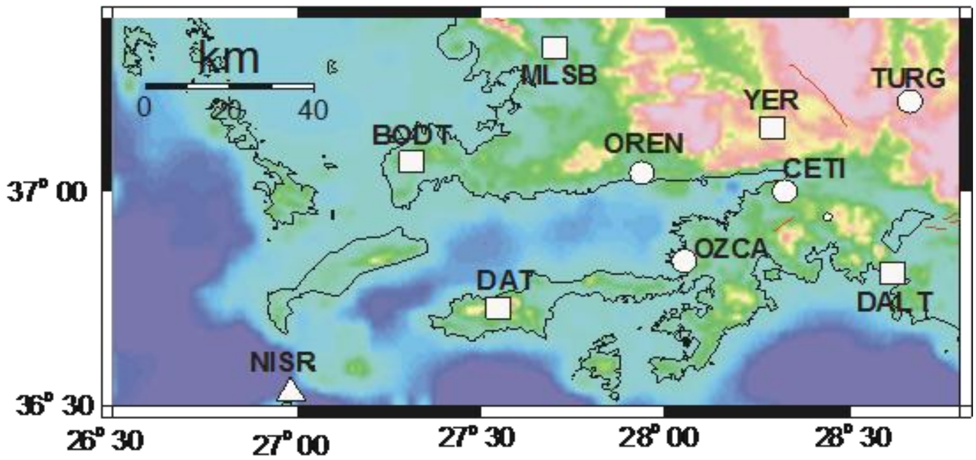

Figure 4.1. Locations of the stations in the Gulf of Gökova. Squares are for the stable stations, circles are for the temporary stations and the triangle is for a station belongs to Greece, but not used in this study (From Aktar *et al*., 2006)

The data set includes the waveforms recorded during the one month period of July, 2007. The stations DAT, CETI and OZCA have a higher background noise. The waveforms of OZCA are not clear to monitor the S-phases clearly, so these phases at this station are rarely used in the locations. DAT has timing problems and it is not included in locations generally. Its contribution is only to the estimations of magnitudes.

P and S phases of all waveforms are picked in software called PQL II (PASSCAL Quick Look II). While picking, a bandpass filter with cutoff frequencies between 2-12 Hz are applied to the waveforms. Then the events are located by the help of these pickings and for these locations the HYPO program is used. After all, the local magnitudes (M_L) of events are estimated by the following formula, which is an approximate expression for the Richter relation (SEISAN Manual),

$$M_L = Log\ (amp) + 1.11\ Log\ (dist) + 0{,}00189\ dist - 2.09 \tag{4.1}$$

4.1.1. Determining Explosions

After the events have been analysed, it can be thought that most of those may be explosions, so three parts which are too intensive from the region are selected and these parts are shown on Figure 4.2.

The event numbers are calculated for every interval of one hour for all clusters (A, B, C), so the total number of them is 777 (Table 4.1). Figure 4.3, 4.4, 4.5 and 4.6 also show their numbers graphically. It is observed that at some given time of the day, the event number increase. This is a sign that these are artificial events which are probably quarry blasts or other mining explosions operate during the daytime. The hours which their numbers are higher than others in all clusters are selected and these hours are eliminated (namely, 08:00, 09:00, 11:00, 12:00, 14:00 and 15:00). The last view of earthquakes after the elimination is shown on the map in Figure 4.10.

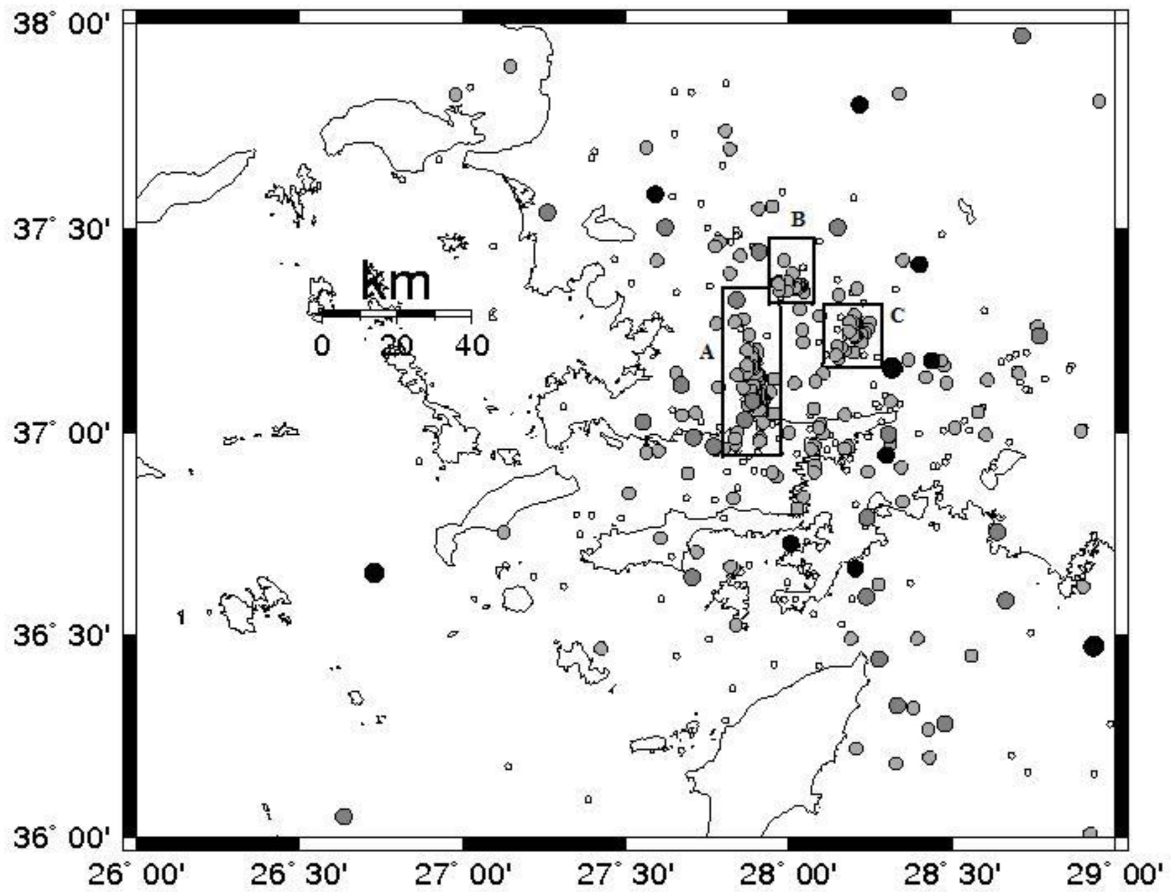

Figure 4.2. Locations of the earthquakes from this study (July, 2007). The coordinates for (A): *36.95°-37.25°, 27.50°-28.00°*, (B): *37.20°-37.30°, 27.85°-28.05°*, (C): *37.15°-37.25°, 28.05°-28.20°*

In all selected parts (A, B, C) from the map (Figure 4.2), it can still be seen a clustering although the artificial events are eliminated. The reason of this case is the existence of some earthquake activity within the mining activity region. An example of waveform of such an event with mining region is shown in Figure 4.7.

Table 4.1. Numbers of events due to hours (July, 2007)

TIME	**All Events**	**Part A**	**Part B**	**Part C**
00:00	22	8	2	0
01:00	31	11	4	3
02:00	20	6	1	2
03:00	17	5	0	2
04:00	12	4	3	1
05:00	16	8	6	1
06:00	18	6	3	0
07:00	26	7	7	4
08:00	48	32	26	2
09:00	106	67	54	9
10:00	26	16	10	2
11:00	47	34	27	2
12:00	41	29	27	3
13:00	26	15	9	4
14:00	49	36	30	2
15:00	87	32	26	16
16:00	23	9	7	0
17:00	27	8	2	3
18:00	15	2	1	2
19:00	23	10	3	4
20:00	18	1	0	2
21:00	21	7	1	1
22:00	35	14	7	4
23:00	23	7	3	3
TOTAL	777	374	259	72

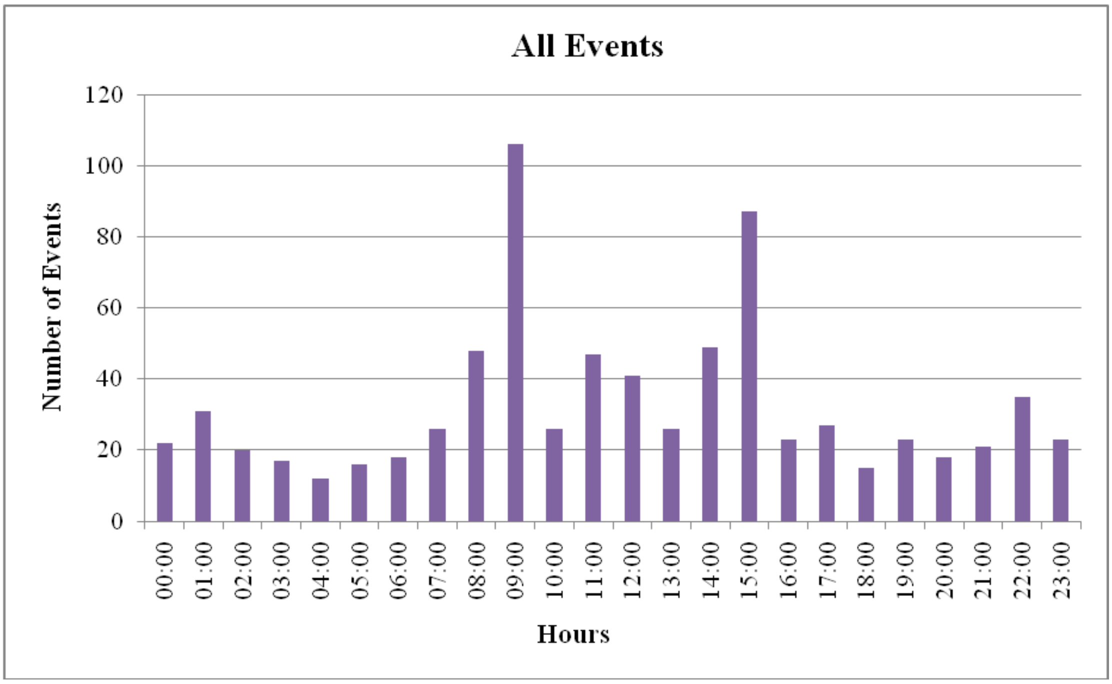

Figure 4.3. Numbers of events due to hours on Figure 4.2 (July, 2007)

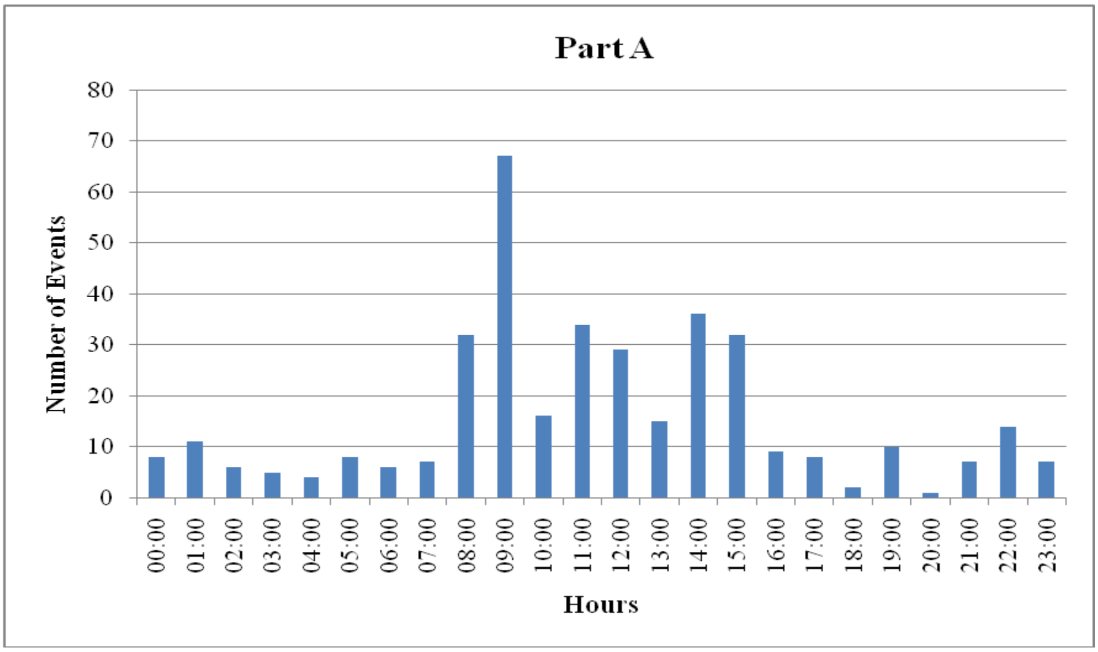

Figure 4.4. Numbers of events due to hours in Part A on Figure 4.2

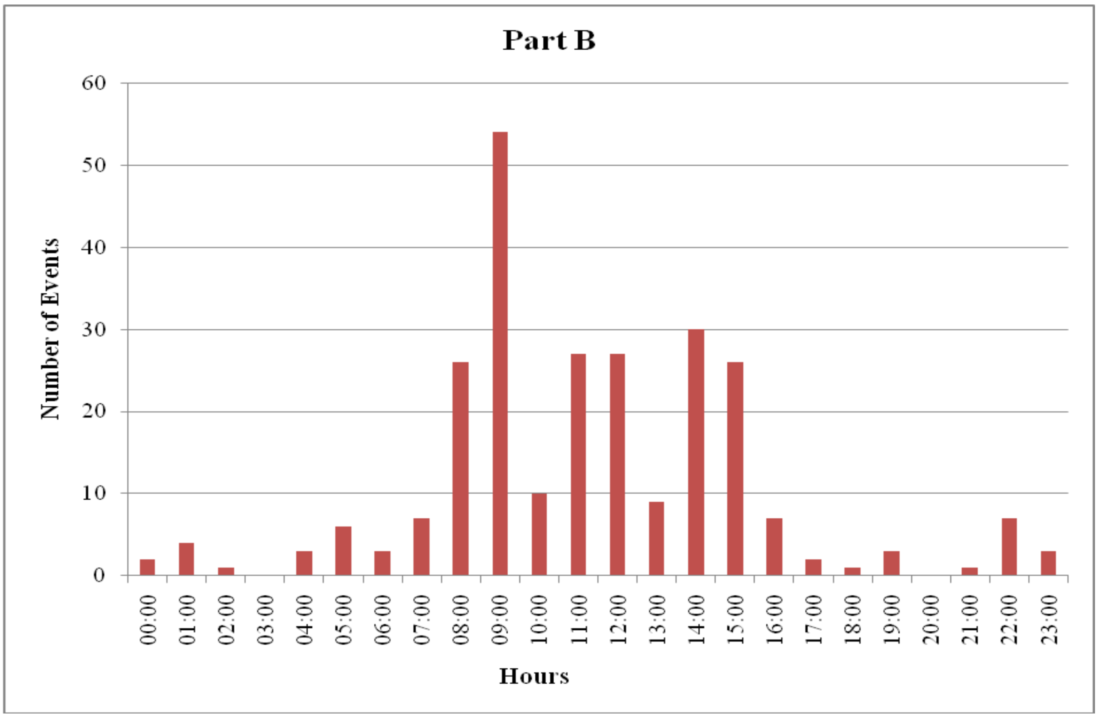

Figure 4.5. Numbers of events due to hours in Part B on Figure 4.2

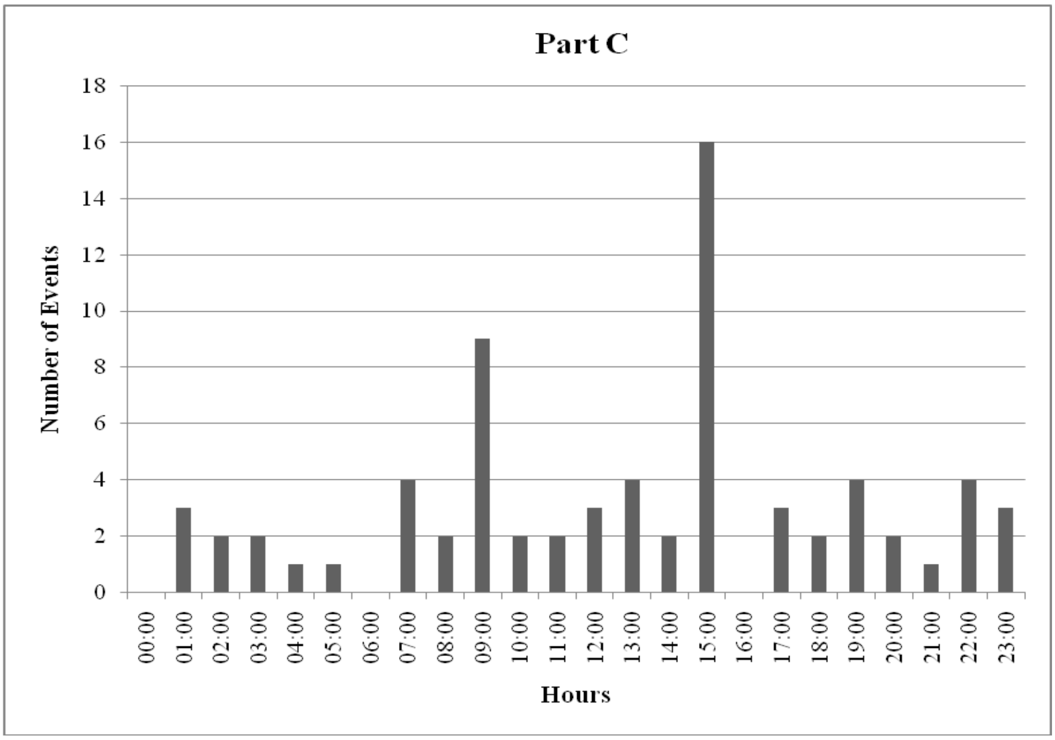

Figure 4.6. Numbers of events due to hours in Part C on Figure 4.2

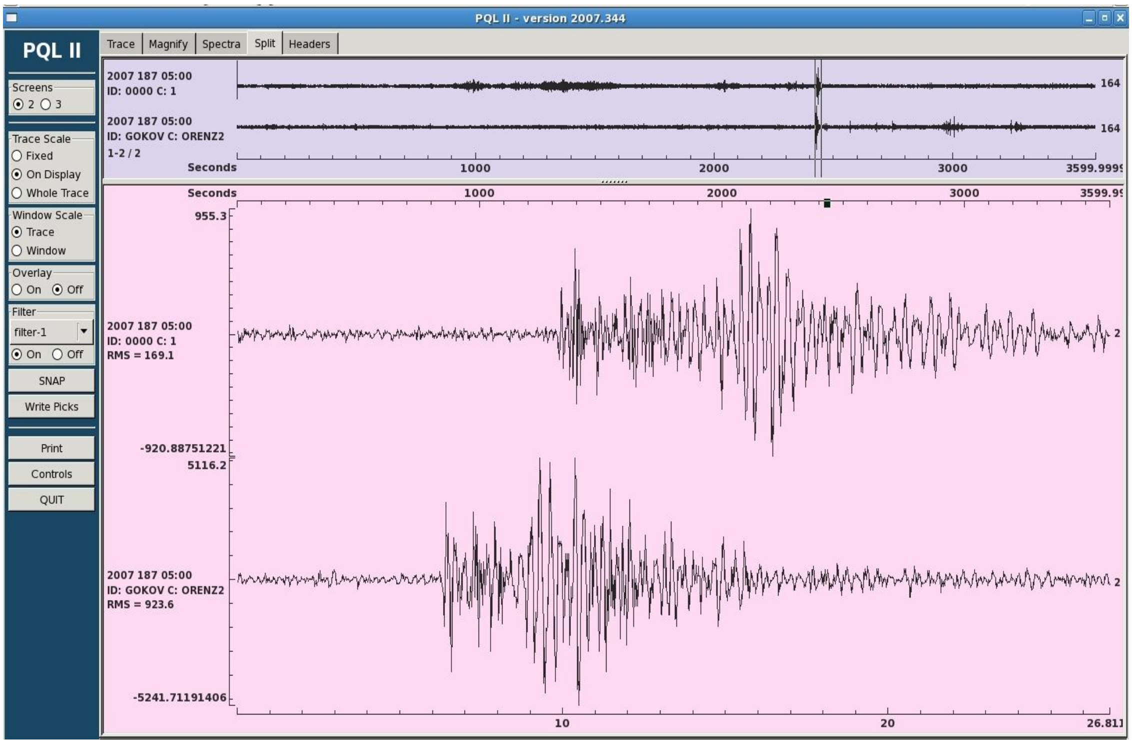

Figure 4.7. Waveform of an earthquake (06.07.2007, 05:40, M_L = 1.4) from Part A on Figure 4.2 (recorded at OREN and YER, occurred among quarry blasts)

4.1.2. Comparison of Events Located In July, 2007 with the NEMC Catalogue

The earthquakes of July, 2007 in catalogue of NEMC and the locations made in this thesis are compared according to their numbers, magnitudes, depths and latitudes-longitudes. The histogram in Figure 4.8 shows the numbers of earthquakes for two different catalogues. It also indicates that how many earthquakes occurred at various magnitudes between M = 0.1 and 4.1.

Other parameter for comparison is the depth. Figure 4.9 and Figure 4.10 show the differences of the present relocation with that of NEMC.

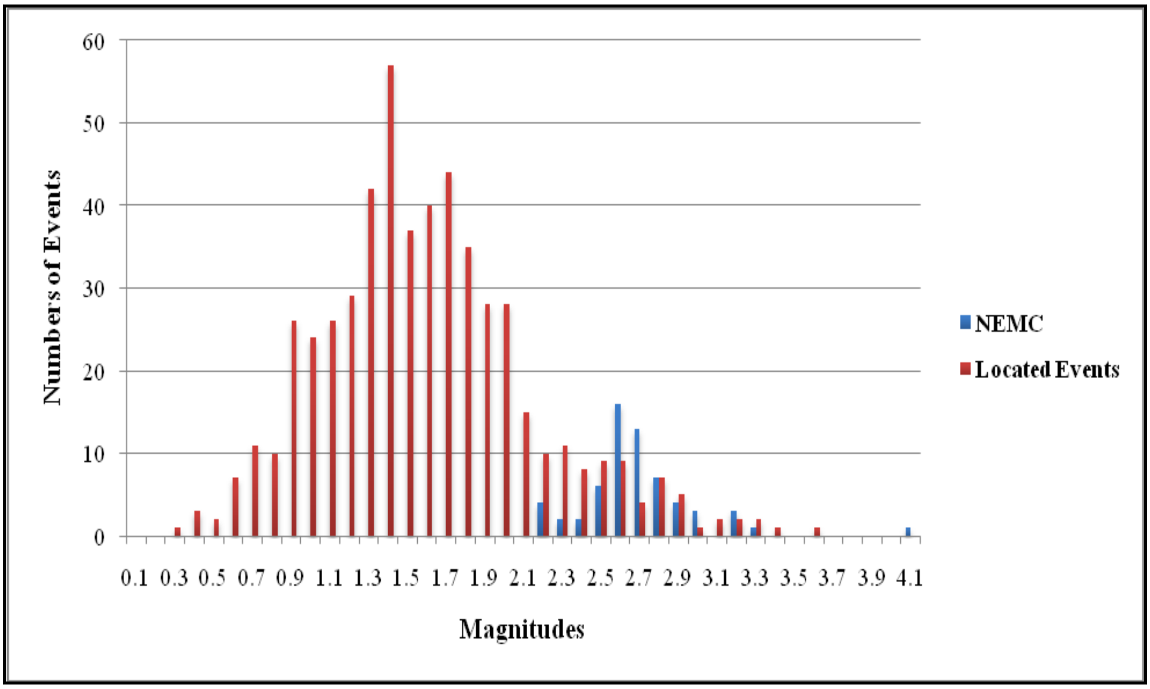

Figure 4.8. The numbers of earthquakes due to their magnitudes located by NEMC and in this study (July, 2007)

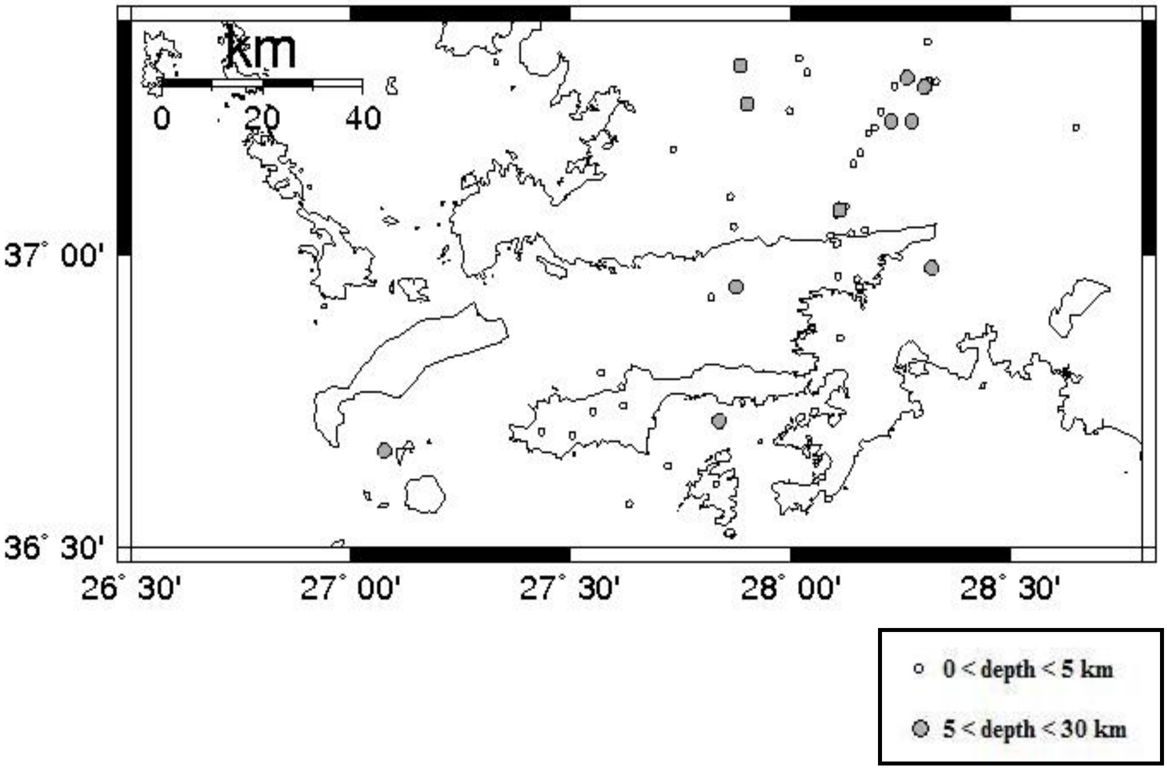

Figure 4.9. Locations of NEMC for July, 2007

The map in Figure 4.11 indicates the similarities and differences of locations, magnitudes and depths for the 15 earthquakes shown in Table 4.2.

The two data sets are generally same with each others. However, catalogue of NEMC have earthquakes which are less deep and also magnitude is the parameter that represents the most important change between two catalogues. The reasons of the difference between the two magnitudes are the fact that all the ones are estimated during this thesis are based on a single method which is the local magnitude as defined by Richter while the ones from NEMC are often based on different methods (mostly duration method in addition to local). The magnitude estimations in this thesis are also based on recordings from higher number of stations (with the addition of four temporary stations) and therefore are expected to be more reliable.

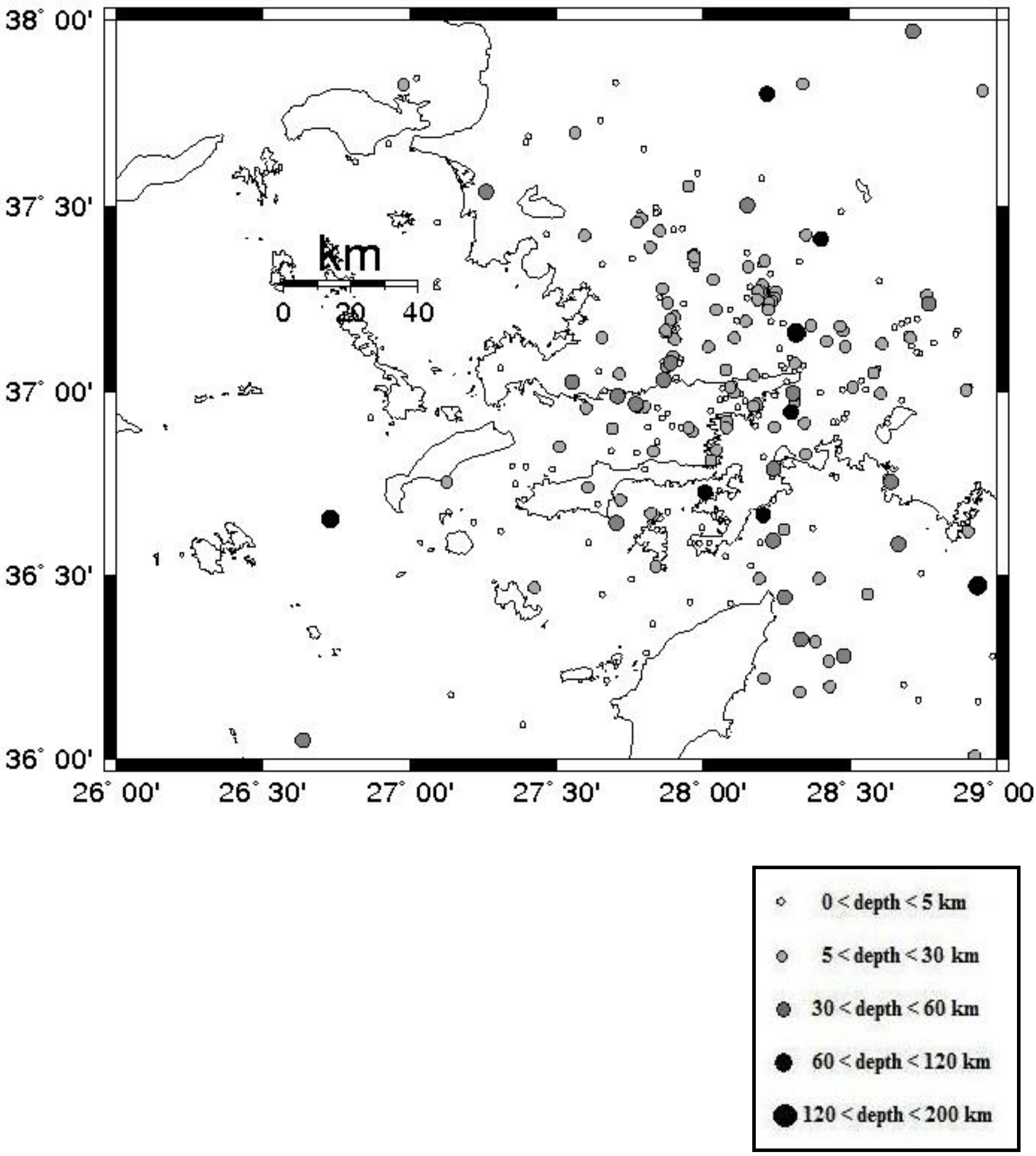

Figure 4.10. Earthquakes of July, 2007 after the elimination of quarry blasts

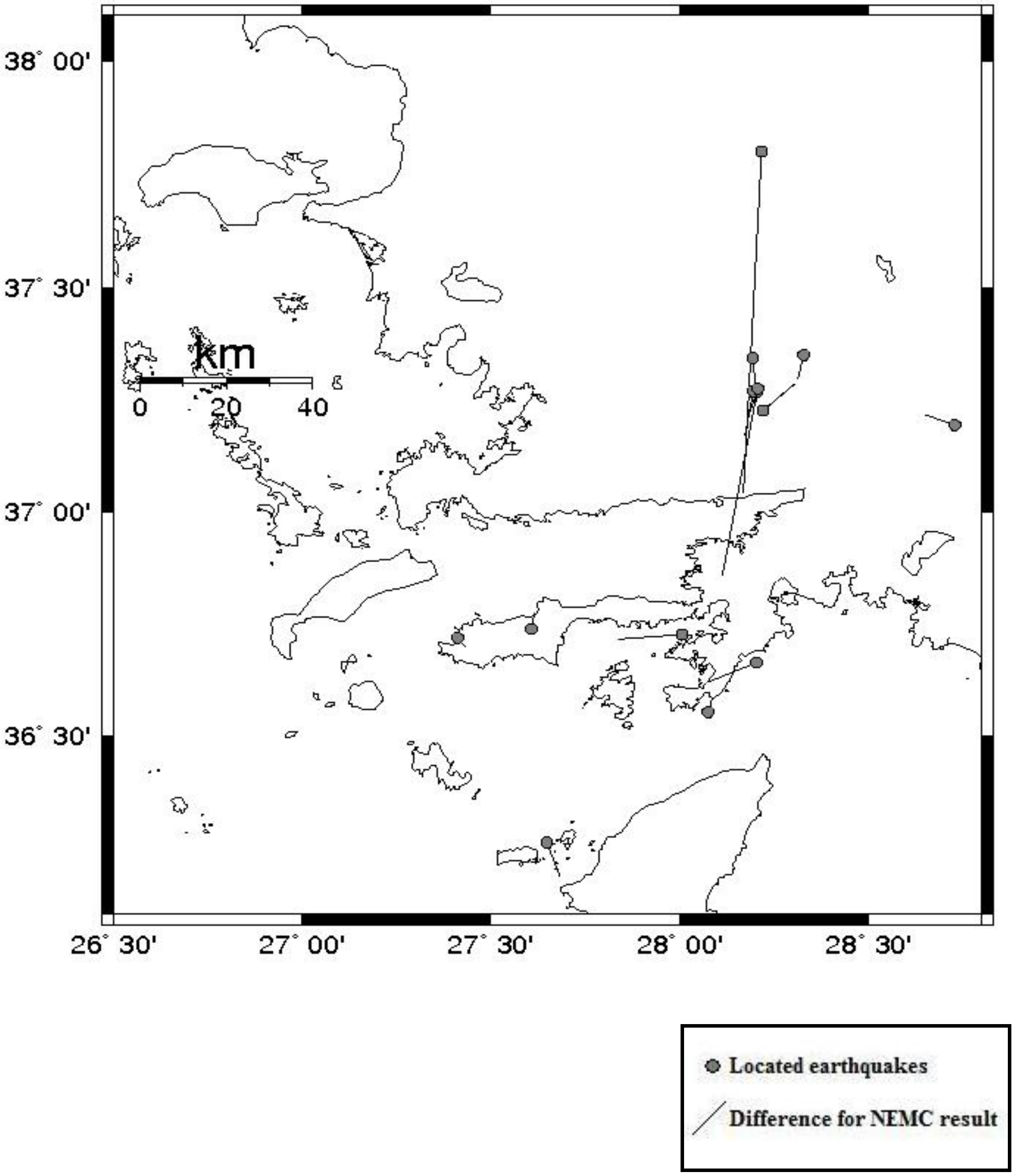

Figure 4.11. Locations of common earthquakes in July, 2007 for comparing NEMC and this study

Table 4.2. Common earthquakes from the two catalogues (this study and NEMC) (July, 2007)

	DATE	**TIME**	**LATITUDE**	**LONGITUDE**	**M_L**	**DEPTH**
NEMC	04.07.2007	09:16:37	37.042	28.169	2.7	7.2
	04.07.2007	09:16:28	37.800	28.218	2.3	60.4
NEMC	04.07.2007	23:58:39	36.716	27.838	3.2	29.3
	04.07.2007	23:58:40	36.726	28.007	3.0	67.5
NEMC	07.07.2007	17:24:15	36.619	28.072	2.6	5.0
	07.07.2007	17:24:15	36.664	28.204	2.6	62.1
NEMC	09.07.2007	13:45:33	36.183	27.684	4.1	7.4
	09.07.2007	13:45:33	36.259	27.649	3.4	0.0
NEMC	09.07.2007	16:26:24	36.774	27.616	2.9	5.0
	09.07.2007	16:26:24	36.739	27.608	2.3	11.9
NEMC	12.07.2007	09:04:16	37.286	28.306	2.5	19.5
	12.07.2007	09:04:19	37.226	28.221	1.8	3.9
NEMC	12.07.2007	09:04:17	37.286	28.305	2.5	18.0
	12.07.2007	09:04:19	37.226	28.221	1.8	3.9
NEMC	21.07.2007	04:37:24	37.297	28.313	2.6	2.2
	21.07.2007	04:37:24	37.350	28.329	2.0	3.8
NEMC	22.07.2007	12:06:10	37.207	28.178	2.8	9.0
	22.07.2007	12:06:09	37.267	28.205	2.1	13.4
NEMC	22.07.2007	16:32:05	36.697	27.435	2.3	5.0
	22.07.2007	16:32:06	36.719	27.413	1.7	0.0
NEMC	23.07.2007	03:39:51	36.582	28.087	2.7	5.0
	23.07.2007	03:39:50	36.553	28.076	2.0	0.1
NEMC	23.07.2007	15:05:45	37.244	28.206	2.6	7.3
	23.07.2007	15:05:43	37.343	28.194	1.9	0.6
NEMC	28.07.2007	15:09:37	37.170	28.173	2.7	15.8
	28.07.2007	15:09:37	37.269	28.196	2.1	13.6
NEMC	29.07.2007	01:07:56	37.217	28.650	3.2	5.0
	29.07.2007	01:07:55	37.194	28.729	2.9	3.7
NEMC	30.07.2007	09:18:22	36.858	28.114	2.6	5.7
	30.07.2007	09:18:25	37.275	28.207	1.8	0.8

4.2. Estimating Magnitude–Frequency Relation

The relationship between the magnitude and total number of earthquakes in any given region and time period of at least that magnitude is expressed by the Gutenberg and Richter law (1954):

$$Log\ (N) = a\text{-}bM \qquad (4.2)$$

where N is the number of events in a given magnitude range and M is a magnitude minimum, a and b are constants to be determined. Linear least-square solutions are used to estimate a and b:

$$\mathrm{a} = [\sum_{i=1}^{n} M_i{}^2 \sum_{i=1}^{n} \log N_i - \sum_{i=1}^{n} M_i \sum_{i=1}^{n} M_i \log N_i][\mathrm{n}\sum_{i=1}^{n} M_i{}^2 - (\sum_{i=1}^{n} M_i)^2]^{-1} \qquad (4.3)$$

$$\mathrm{b} = [\sum_{i=1}^{n} M_i \sum_{i=1}^{n} \log N_i - \mathrm{n}\sum_{i=1}^{n} M_i \log N_i][\mathrm{n}\sum_{i=1}^{n} M_i{}^2 - (\sum_{i=1}^{n} M_i)^2]^{-1} \qquad (4.4)$$

The critical issue in this estimation is to determine the lower limit of the data range which is taken into account for the estimation process, in other word the completeness threshold of the catalogue.

The magnitude interval is taken as 0.1 and then a and b-values of the data in July, 2007 are calculated to compare for both data sets of NEMC and this thesis (Table 4.3 and 4.4). The completeness level is determined by using eye inspection of the histogram and the value is selected at the flexion point of the curve. In general, quantitative methods need to be used for determining the completeness level and more objective values can then be determined. However, since in this thesis only approximate evaluation is needed, any of the quantitative approaches are not used. The completeness level for the data in this thesis and for the catalogue of NEMC (July, 2007 or total 2007) is chosen as 1.4 and 2.7, respectively. There is a difference of magnitude of one unit between the completeness of the two catalogues which are due to both of the higher number of station used and also by more careful picking of the phases.

The estimation for the linear relation gives respectively, 7.3679 and 4.3228 for a-values, and 2.1794 and 1.1356 for b-values. Yet, b-values for NEMC data are too high and are not realistic. Calculating one month data and therefore not having sufficient number of earthquakes may be the factors of this discrepancy. Thus, data for the whole year of 2007 solved by NEMC are used to find the magnitude-frequency relation (Table 4.5) and the b-values have become lower than before. They are 5.7761 for a-value and 1.2037 for b-value. Histograms in Figure 4.12, 4.13 and 4.14 which show Magnitudes *vs.* LogN(M) are drawn by Microsoft Excel program.

It is observed that using the data for the whole year of 2007 in NEMC catalogue gives a b-value which is close to the one estimated for the one month (July, 2007) of data processed in this thesis. However, both values are slightly higher than what is generally observed. This means that higher number of small events is observed as compared to larger events. This is a sign of either existence of swarm activity or an aftershock sequence. Since both processes are some kind of triggered activity, they constitute a suitable location for the investigation of the trigger mechanism.

Table 4.3. Earthquakes from NEMC catalogue used for the magnitude–frequency estimations (July, 2007)

Magnitude	**N**	**N(M)**	**Log N(M)**	**M(i)**	**M(i)^2**	**M(i)*Log N(M)**
2.6	16	47	1.6721	2.6	6.76	4.3475
2.7	13	31	1.4914	2.7	7.29	4.0267
2.8	7	18	1.2553	2.8	7.84	3.5148
2.9	4	11	1.0414	2.9	8.41	3.0200
3	3	7	0.8451	3	9	2.5353
3.1	0	4	0.6021	3.1	9.61	1.8664
3.2	3	4	0.6021	3.2	10.24	1.9266
3.3	1	1	0	3.3	10.89	0
TOTAL	47	123	7.5093	23.6	70.04	21.2372

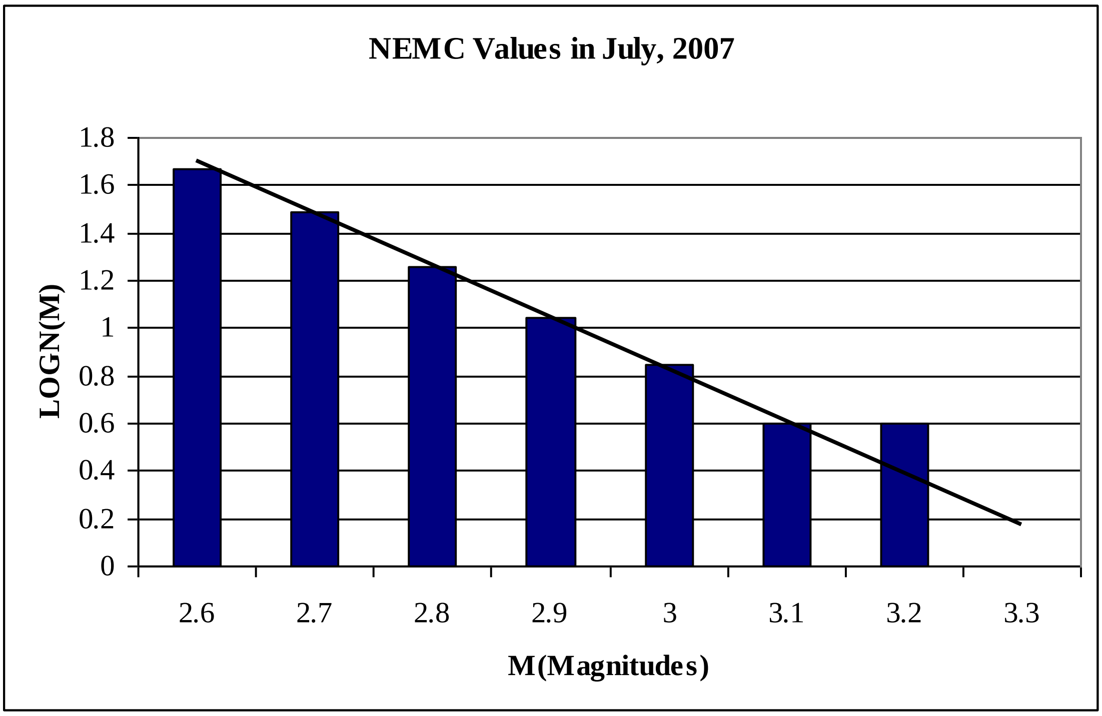

Figure 4.12. The magnitude-frequency distribution for NEMC catalogue (July, 2007)

Table 4.4. Earthquakes located in this thesis used for the magnitude–frequency estimations (July, 2007)

Magnitude	**N**	**N(M)**	**Log N(M)**	**M(i)**	**$M(i)^2$**	**M(i)*Log N(M)**
1.4	57	356	2.5514	1.4	1.96	3.5720
1.5	37	299	2.4757	1.5	2.25	3.7135
1.6	40	262	2.4183	1.6	2.56	3.8693
1.7	44	222	2.3464	1.7	2.89	3.9888
1.8	35	178	2.2504	1.8	3.24	4.0508
1.9	28	143	2.1553	1.9	3.61	4.0951
2	28	115	2.0607	2	4	4.1214
2.1	15	87	1.9395	2.1	4.41	4.0730
2.2	10	72	1.8573	2.2	4.84	4.0861
2.3	11	62	1.7924	2.3	5.29	4.1225
2.4	8	51	1.7076	2.4	5.76	4.0982
2.5	9	43	1.6335	2.5	6.25	4.0837
2.6	9	34	1.5315	2.6	6.76	3.9818
2.7	4	25	1.3979	2.7	7.29	3.7744
2.8	7	21	1.3222	2.8	7.84	3.7022
2.9	5	14	1.1461	2.9	8.41	3.3238
3	1	9	0.9542	3	9	2.8627
3.1	2	8	0.9031	3.1	9.61	2.7996
3.2	2	6	0.7782	3.2	10.24	2.4901
3.3	2	4	0.6021	3.3	10.89	1.9868
3.4	1	2	0.3010	3.4	11.56	1.0235
3.5	0	1	0	3.5	12.25	0
3.6	1	1	0	3.6	12.96	0
TOTAL	356	2015	34.1249	57.5	153.87	73.8193

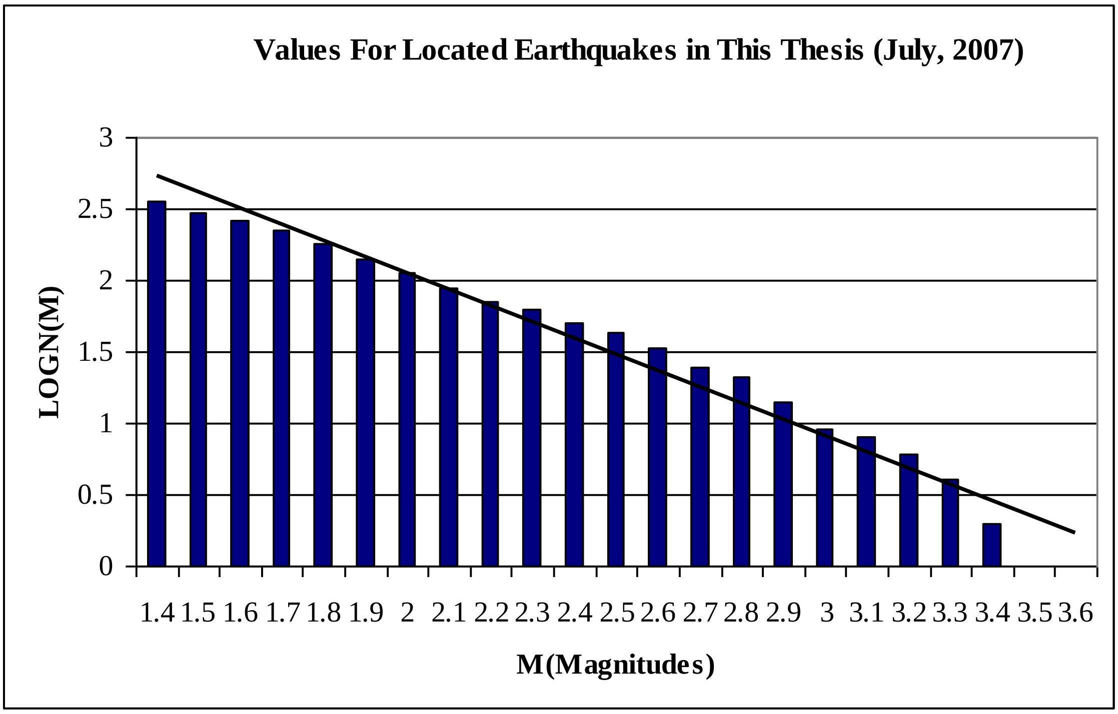

Figure 4.13. The magnitude-frequency distribution for earthquakes located in this thesis (July, 2007)

Table 4.5. Earthquakes from NEMC catalogue used for the magnitude-frequency estimations (entire 2007)

Magnitude	**N**	**N(M)**	**Log N(M)**	**M(i)**	**M(i)^2**	**M(i)*Log N(M)**
2.7	111	476	2.6776	2.7	7.29	7.2295
2.8	84	365	2.5623	2.8	7.84	7.1744
2.9	91	281	2.4487	2.9	8.41	7.1012
3	51	190	2.2788	3	9	6.8363
3.1	38	139	2.1430	3.1	9.61	6.6433
3.2	36	101	2.0043	3.2	10.24	6.4138
3.3	16	65	1.8129	3.3	10.89	5.9826
3.4	20	49	1.6902	3.4	11.56	5.7467
3.5	5	29	1.4624	3.5	12.25	5.1184
3.6	8	24	1.3802	3.6	12.96	4.9688
3.7	4	16	1.2041	3.7	13.69	4.4552
3.8	3	12	1.0792	3.8	14.44	4.1009
3.9	2	9	0.9542	3.9	15.21	3.7215
4	0	7	0.8451	4	16	3.3804
4.1	2	7	0.8451	4.1	16.81	3.4649
4.2	2	5	0.6990	4.2	17.64	2.9357
4.3	2	3	0.4771	4.3	18.49	2.0516
4.4	0	1	0	4.4	19.36	0
4.5	0	1	0	4.5	20.25	0
4.6	0	1	0	4.6	21.16	0
4.7	0	1	0	4.7	22.09	0
4.8	0	1	0	4.8	23.04	0
4,9	0	1	0	4,9	24.01	0
5	0	1	0	5	25	0
5.1	0	1	0	5.1	26.01	0
5.2	1	1	0	5.2	27.04	0
TOTAL	476	1787	26.5642	102.7	420.29	87.3253

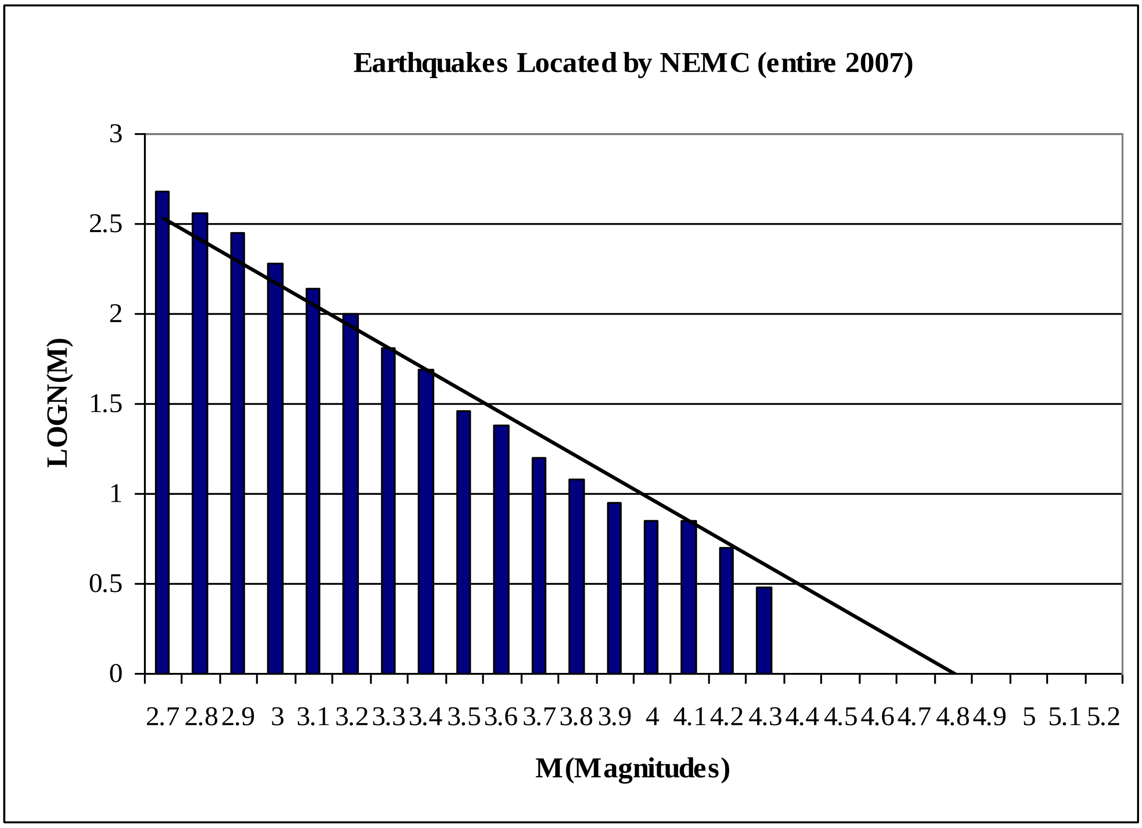

Figure 4.14. The magnitude-frequency distribution for the earthquakes located by NEMC (entire 2007)

5. OBSERVATION OF TRIGGERING IN GÖKOVA BAY

5.1. Theory For Estimation of the Stress Variation

The transient stress caused by the triggering waves can be evaluated by the amplitude which relates to the particle velocity, or by the energy density transferred from the waves (Brodsky *et al.,* 2000). There is an example in Table 5.1 which is calculated for the Landers, Hector Mine and Izmit earthquakes. Triggering in Imperial Valley, California and Greece are compared by using these amplitudes and energy densities (Brodsky *et al.,* 2000). Amplitudes represent both velocity and dynamic stress (Brodsky *et al.*, 2000).

Table 5.1. Observed amplitudes and energies (adopted from Brodsky *et al.*, 2000)

Event	**Distance (km)**	**Amplitude cm/s (MPa)**	**Energy Density J/m^2**
Landers	139	6.6 (0.60)	3.5×10^5
Hector Mine	158	9.0 (0.82)	1.3×10^6
Izmit	589	1.9 (0.18)	7.8×10^4

The velocity amplitude is obtained from half the maximum peak-peak value of the horizontal velocity (Brodsky *et al.,* 2000). The stress amplitude equals the velocity amplitude multiplied by μ/β where μ is the shear modulus (3×10^{10} Pa) and β is the shear velocity (3300 m/s) (Brodsky *et al.,* 2000).

The integral over time of the velocity squared multiplied by β and the rock density (2750 kg/m^3) represents energy density (Kanamori *et al.,* 1993). This is in fact the kinetic energy which is the mass times the square of the particle motion.

The results show that amplitudes and energies for the triggering in Greece are lower than the ones in Imperial Valley. Thus, such regional-scale elevated seismicity is called as "superswarms" and they can continue for months, but they should not be confused with common mainshock-aftershock sequences for the large areas (Brodsky *et al.,* 2000).

5.2. Application

There are two ways for calculating the triggering potential of the arriving waves as mentioned in the Section 5.1. and in this application, amplitude of the transient stress is used. Firstly, large events from far distances and medium events from regional distances which occurred in July and August, 2007 have been selected from the catalogues of USGS. While selecting, the earthquakes which are seen clearly on the waveforms of the stations around Gulf of Gökova are preferred. The largest local event (occurred in Gökova Bay, M_L = 2.3) which is also the most completely recorded by all stations have also been added. The list of those earthquakes is shown in Table 5.2.

However, rather than finding the velocity amplitudes as explained in the Section 5.1. they are obtained differently. They are determined through the two horizontal components according to the highest one. Only maximum values of the waveforms are used instead of half the maximum peak-peak value. It is expected that the two approaches would not give much different results. Then the stress amplitudes are computed for every station of the earthquakes with procedure described above. They are shown in Table 5.3.

Table 5.2. List of the selected earthquakes in July and August, 2007 for estimation of stress amplitudes (MPa) (from USGS except the one in Gulf of Gökova)

YEAR	MONTH	DAY	TIME	LATITUDE	LONGITUDE	DEPTH(km)	MAGNITUDE	LOCATION
2007	7	4	23:55.32	38.80	15.20	279	5.2 M_w	Sicily
2007	7	9	16:26:24	36.77	27.61	11.9	2.3 M_L	Gulf of Gökova
2007	7	16	1:13:22	37.53	138.45	12	6.6 M_w	Honshu,Japan
2007	7	29	5:43:00	38.81	27.73	6	4.6 M_L	Manisa
2007	8	1	17:08:51	-15.60	167.68	120	7.2 M_w	Vanuatu
2007	8	8	17:05:04	-5.86	107.42	280	7.5 M_w	Java,Indonesia
2007	8	15	20:22:11	50.32	-177.55	9	6.5 M_w	Alaska
2007	8	20	22:42:28	8.04	-39.25	6	6.5 M_w	Mid Atlantic Ridge

Together with the mean values of stress amplitudes, variances and standard deviations are evaluated in order to see the stress amplitude variations between stations and they are listed in Table 5.4. Mean values of stress amplitudes with their standard deviations are plotted in Figure 5.1. For it is being dealed with earthquakes with many different magnitudes from many different origins (teleseismic, regional and local) the stress variation range is large as well as the standard deviation. This is clearly observed in Figure 5.1. When it is looked at the stress transients in Figure 5.1 again, it is noticed that there are two earthquakes which cause much larger stress peaks than the others. Surprisingly, these two stress peaks are caused by the largest (teleseismic) and smallest (local) events occurring during that time period.

The largest of the stress transients in Figure 5.1 is due to the 8 August, 2007 dated M_w = 7.5 Java Earthquake. In general, it is expected that the stress transient would have small variance for a teleseismic event since nearly the same wave arrives at all stations. However, in this example the standard deviation of the stress transient is slightly high (in particular large peak in BODT station and small peak in MLSB station, nearly a difference of 8 times). It is not possible to explain at this stage the causes of this relatively high variation unless there is an instrumental problem or confusion in the instrumental response information. Nevertheless, it is seen that the high stress transient is valid for all stations around the gulf, so if triggering occurs, it may cover the whole study area.

The second largest of the stress transients in Figure 5.1 is due to a local event of small magnitude. For a local event, since very large amplitudes are recorded at close stations the average can be very high and in particular most stations are located close to the epicentre. However, the amplitude decays very rapidly as the distance to the epicentre increases. The peak values are distributed over wide range from very small to very large, therefore the variance increases. In this example of local event in Gökova Bay, the peak stress is very high locally but the standard deviation is also very high as expected. It is therefore clear that if there is any triggering, this will concern only a small area around the epicentre. Hence, it will not affect the whole of Gökova Bay.

Table 5.3. Stress amplitudes (MPa) of earthquakes in Table 5.2 by stations

EARTHQUAKES	BLCB	BODT	CETI	DALT	DAT	ELL	MLSB	OREN	TURG	YER
Sicily	20	28.1	50.9	89.1	30	-	34.5	37.1	40	43.6
Gulf of Gökova	-	281.8	127.2	69.1	-	-	61.8	300	36.5	55.4
Honshu, Japan	-	28.5	50	-	-	40	37	34.5	42.7	35.4
Manisa	-	144.5	127.2	34.5	74.5	76.3	63.9	104.5	80	42.7
Vanuatu	13.3	26	36.5	57.8	26.4	37.3	31.5	-	73.6	42.3
Java	94.5	270	121.8	179.1	211.8	143.6	32.4	-	151.8	59.5
Alaska	11.5	26.9	46.6	53.5	28.1	48.4	32.4	-	26.5	40.1
Mid Atlantic Ridge	12.2	26.4	-	52.5	27.7	47.9	33.4	-	53.1	36.7

Table 5.4. Mean values of stress amplitudes (MPa), variances and standard deviations for the earthquakes in Table 5.2

EARTHQUAKES	MEAN	VARIANCE	STANDARD DEVIATION
Sicily	41.5	355.9	18.9
Gulf of Gökova	133.1	10650.1	103.2
Honshu, Japan	38.3	40	6.3
Manisa	83.1	1185	34.4
Vanuatu	38.3	290.1	17
Java	140.5	4910.4	70.1
Alaska	34.9	158.5	12.6
Mid Atlantic Ridge	36.2	180.1	13.4

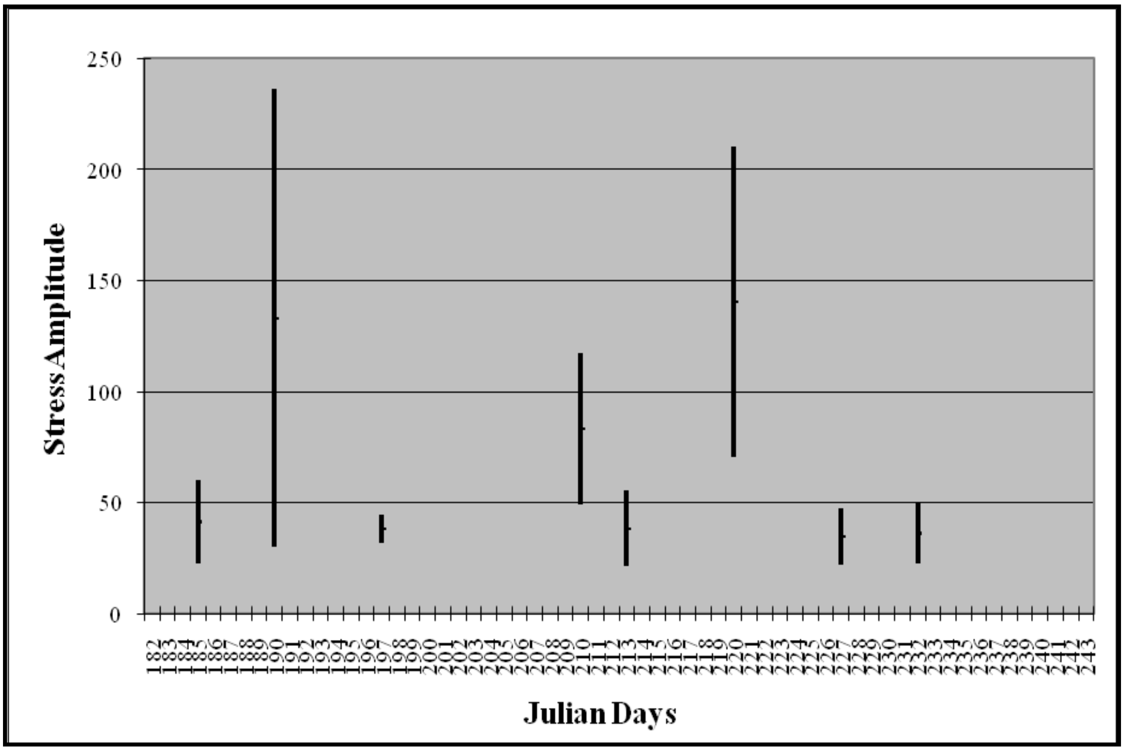

Figure 5.1. Mean values of stress amplitudes (MPa) with their standard deviations

In order to illustrate the possibility of triggering in the Gökova Bay, we use the seismicity rate in the Gulf for the period of July and August, 2007. The rate of seismicity is given by the daily numbers of earthquakes and this is shown in Figure 5.2. One can see that the seismic production is not at all uniform. There are peaks and gaps in the curve, but in particular there is a peak seismic production on the day of 224 (Julian). This peak is nearly 2.5 times the average value of seismic production. Following this peak, there is an oscillatory but decreasing trend which brings the seismicity rate to its average value in a period of about 2 weeks. Apart from the peak of 224, there is also a second peak on the day 200 (Julian), but this peak is of lesser importance. The comparison of the activity intensity with the selected triggering earthquakes will help to determine the existence of a possible triggering. In Figure 5.3, that comparison is shown by using mean values of stress amplitudes with their standard deviations.

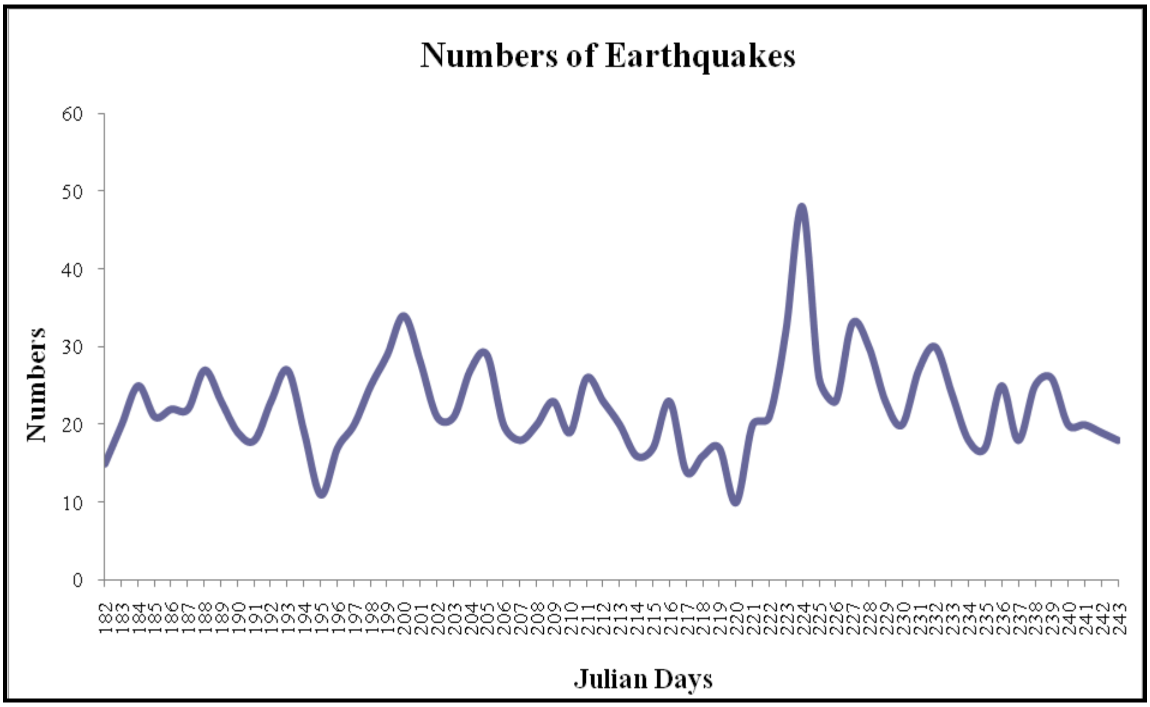

Figure 5.2. Daily variation of earthquake activity (July and August, 2007) (The data in August are taken from Coşkun, 2010)

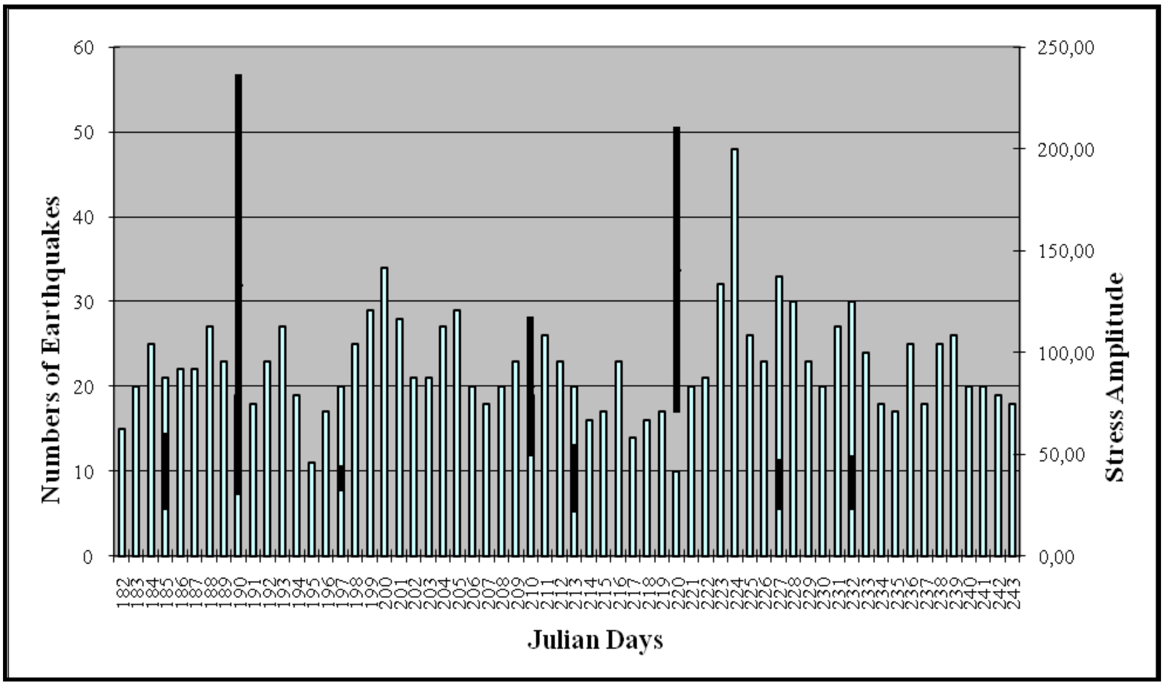

Figure 5.3. Daily variation of earthquake activity with mean values of stress amplitudes (MPa) and their standard deviations

Looking at Figure 5.3, where the stress transients and the seismicity rates are plotted simultaneously it is seen that the highest seismicity peak occurs only four days later after the highest stress peak which is due to the Java Earthquake. So, this observation may be interpreted as a possible evidence for the triggering of the microseismicity in Gökova Bay due to a teleseismic event. The second highest stress peak due to the local event is not clearly followed by a seismicity increase in the Gökova Bay and this time the triggering delay is 10 days. Considering also the fact that the event is very small and high stress transients are only observed at close distances to the epicentre and do not propagate over the entire Gökova Bay, it is considered that this does not correspond to a true triggering.

Since the main research topic in this thesis is to find if there has been any triggered activity in recent years for the Gulf of Gökova and its around, the evolution of seismic production over the years between 2001 and 2009 is also considered. In this final approach, the catalogue of NEMC is used for the locations of earthquakes occurred in Gökova region. The plotting of earthquake activity is given in Figure 5.4 as a function of occurrence time and longitude. The most striking feature of the plot is the increase in earthquake occurrence over the years. The reason of that increase may be due to the addition of seismic stations which are recently put into operation by NEMC and lowering the detection level of microseismicity. One can see that the earthquake activity is not regular both in time and space. Many aftershocks, swarms and possibly triggered clusters are expected to generate the dark patches in the Figure 5.4.

Figure 5.5 represents the same seismicity evolution in Figure 5.4, together with the regional earthquakes above 5.5 occurring around Gökova. There are 15 regional earthquakes between 2001 and 2009 of which the locations are taken from USGS catalogue and they are listed in Table 5.5. Finally, since global earthquakes also cause triggering in the Gulf of Gökova as already shown in this thesis, all global earthquakes larger than M = 7.0 have been selected from catalogues of USGS and are also plotted together with the local seismicity. These large global earthquakes are listed in Table 5.6 and their occurrence times are shown in Figure 5.7 with an arrow. Their distances to the study area are not considered, because teleseismic body waves and in particular the surface waves propagate more efficiently than local and regional ones.

Table 5.5. Regional earthquakes larger than M = 5.5 around Gökova (2001-2009) (from USGS)

YEAR	**MONTH**	**DAY**	**TIME**	**LATITUDE**	**LONGITUDE**	**DEPTH**	**MAGNITUDE**
2001	06	10	13:11:04.23	38.58	25.61	33	5.6 M_w
2001	06	23	06:52:42.96	35.55	28.16	50	5.7 M_w
2002	01	22	04:53:52.65	35.79	26.62	88	6.2 M_w
2003	04	10	00:40:15.11	38.22	26.96	10	5.8 M_w
2004	08	04	03:01:07.57	36.83	27.82	10	5.6 M_w
2004	10	07	01:05:12.89	36.43	26.80	128	5.5 M_w
2005	01	10	23:48:50.02	37.02	27.80	15	5.5 M_w
2005	01	23	22:36:05.53	35.80	29.64	10	5.8 M_w
2005	10	17	05:45:16.00	38.13	26.50	8	5.5 M_w
2005	10	17	09:46:53.90	38.20	26.50	10	5.8 M_w
2005	10	20	21:40:04.09	38.15	26.75	10	5.9 M_w
2008	03	28	00:16:19.90	34.76	25.34	45	5.6 M_w
2008	07	15	03:26:34.70	35.80	27.86	52	6.4 M_w
2009	06	19	14:04:59.03	35.36	28.45	28	5.8 M_w
2009	07	01	09:30:10.41	34.16	25.47	19	6.4 M_w

Table 5.6. Teleseismic earthquakes which are larger than M = 7.0 (from USGS)

YEAR	MONTH	DAY	TIME	LATITUDE	LONGITUDE	DEPTH	M	LOCATION
2001	01	01	06:57:04.1	6.898	126.579	33	7.5	Mindanao, Philippines
2001	01	09	16:49:28.0	-14.928	167.170	103	7.1	Vanuatu Islands
2001	01	10	16:02:44.2	57.078	-153.211	33	7.0	Kodiak Island region, Alaska
2001	01	13	17:33:32.3	13.049	-88.660	60	7.7	El Salvador
2001	01	26	03:16:40.5	23.419	70.232	16	7.7	Southern India
2001	02	13	19:28:30.2	-4.680	102.562	36	7.4	Southern Sumatera, Indonesia
2001	02	24	07:23:48.7	1.271	126.249	35	7.1	Northern Molucca Sea
2001	06	03	02:41:57.1	-29.666	-178.633	178	7.2	Kermadec Islands,
New Zealand								
2001	06	23	20:33:14.1	-16.265	-73.641	33	8.4	Near the coast of Peru
2001	07	07	09:38:43.5	-17.543	-72.077	33	7.6	Near the coast of Peru
2001	08	21	06:52:06.2	-36.813	-179.575	33	7.1	East of North Island, N.Z.
2001	10	12	15:02:16.8	12.686	144.980	37	7.0	South of the Mariana Islands
2001	10	19	03:28:44.4	-4.102	123.907	33	7.5	Banda Sea
2001	10	31	09:10:20.0	-5.912	150.196	33	7.0	New Britain region, P.N.G.
2001	11	14	09:26:10.0	35.946	90.541	10	7.8	Qinghai, China
2001	12	12	14:02:35.0	-42.813	124.688	10	7.1	South of Australia
2002	01	02	17:22:48.7	-17.600	167.856	21	7.2	Vanuatu Islands
2002	03	03	12:08:19.7	36.502	70.482	226	7.4	Hindu Kush region,
Afghanistan								
2002	03	05	21:16:09.1	6.033	124.249	31	7.5	Mindanao, Philippines
2002	03	31	06:52:50.4	24.279	122.179	33	7.1	Taiwan region
2002	04	26	16:06:07.0	13.088	144.619	86	7.1	Mariana Islands
2002	06	28	17:19:30.2	43.752	130.666	566	7.3	Jilin-Heilongjiang border
region, China								
2002	08	19	11:01:01.1	-21.696	-179.513	580	7.7	Fiji region
2002	08	19	11:08:24.3	-23.884	178.495	675	7.7	South of the Fiji Islands
2002	09	08	18:44:23.7	-3.302	142.945	13	7.6	Near the north coast
of New Guinea								
2002	10	10	10:50:20.5	-1.757	134.297	10	7.6	Near the north coast
of Irian Jaya								
2002	11	02	01:26:10.7	2.824	96.085	30	7.4	Simeulue, Indonesia
2002	11	03	22:12:41.0	63.517	-147.444	5	7.9	Central Alaska
2002	11	17	04:53:53.5	47.824	146.209	459	7.3	Northwest of the Kuril
Islands								
2003	01	20	08:43:06.0	-10.491	160.770	33	7.3	Solomon Islands
2003	01	22	02:06:34.6	18.770	-104.104	24	7.6	Colima, Mexico
2003	03	17	16:36:17.3	51.272	177.978	33	7.1	Rat Islands, Aleutian

Islands, Alaska							
2003	05	26	09:24:33.4	38.849	141.568	68	7.0 Near the East Coast of
Honshu, Japan							
2003	05	26	19:23:27.9	2.354	128.855	31	7.0 Halmahera, Indonesia
2003	06	20	06:19:38.9	-7.606	-71.722	558	7.1 Amazonas, Brazil
2003	07	15	20:27:50.5	-2.598	68.382	10	7.6 Carlsberg Ridge
2003	08	21	12:12:49.7	-45.104	167.144	28	7.2 South Island of New Zealand
2003	09	25	19:50:06.3	41.815	143.910	27	8.3 Hokkaido, Japan region
2003	09	25	21:08:00.0	41.774	143.593	33	7.4 Hokkaido, Japan region
2003	09	27	11:33:25.0	50.038	87.813	16	7.3 Southwestern Siberia,
Russia							
2003	10	31	01:06:28.2	37.812	142.619	10	7.0 Off East Coast of Honshu,
Japan							
2003	11	17	06:43:06.8	51.146	178.650	33	7.8 Rat Islands, Aleutian
Islands, Alaska							
2003	12	27	16:00:59.4	-22.015	169.766	10	7.3 Southeast of the Loyalty
Islands							
2004	01	03	16:23:21.0	-22.253	169.683	22	7.1 Southeast of the Loyalty
Islands							
2004	02	05	21:05:02.8	-3.615	135.538	17	7.0 Papua, Indonesia
2004	02	07	02:42:35.2	-4.003	135.023	10	7.3 Near the South Coast
of Papua, Indonesia							
2004	07	15	04:27:14.7	-17.656	-178.760	566	7.1 Fiji Region
2004	07	25	14:35:19.0	-2.427	103.981	582	7.3 Southern Sumatra, Indonesia
2004	09	05	10:07:07.8	33.059	136.635	14	7.2Near the South Coast
of W.Honshu, Japan							
2004	09	05	14:57:18.6	33.184	137.071	10	7.4 Near the South Coast
of Honshu, Japan							
2004	10	09	21:26:53.6	11.422	-86.665	35	7.0 Near the coast of Nicaragua
2004	11	11	21:26:41.1	-8.152	124.868	10	7.5 Kepulauan Alor, Indonesia
2004	11	15	09:06:56.5	4.695	-77.508	15	7.2 Near the West Coast
of Colombia							
2004	11	22	20:26:23.9	-46.676	164.721	10	7.1 Off West Coast
of the South Island, N.Z.							
2004	11	26	02:25:03.3	-3.609	135.404	10	7.1 Papua, Indonesia
2004	11	28	18:32:14.1	43.006	145.119	39	7.0 Hokkaido, Japan Region
2004	12	23	14:59:04.4	-49.312	161.345	10	8.1 North of Macquarie Island
2004	12	26	00:58:53.4	3.295	95.982	30	9.1 Off the West Coast of
Northern Sumatra							
2004	12	26	04:21:29.8	6.910	92.958	39	7.1 Nicobar Islands, India
2005	02	05	12:23:18.9	5.293	123.337	525	7.1 Celebes Sea
2005	03	02	10:42:12.2	-6.527	129.933	202	7.1 Banda Sea

2005	03	28	16:09:36.5	2.085	97.108	30	8.6 Northern Sumatra, Indonesia
2005	06	13	22:44:33.9	-19.987	-69.197	116	7.8 Tarapaca, Chile
2005	06	15	02:50:54.1	41.292	-125.953	16	7.2 Off the Coast of
Northern California							
2005	07	24	15:42:06.2	7.920	92.190	16	7.2 Nicobar Islands, India
2005	08	16	02:46:28.4	38.276	142.039	36	7.2 Near the East Coast of
Honshu, Japan							
2005	09	09	07:26:43.7	-4.539	153.474	90	7.6 New Ireland Region, P.N.G.
2005	09	26	01:55:37.6	-5.678	-76.398	115	7.5 Northern Peru
2005	10	08	03:50:40.8	34.539	73.588	26	7.6 Pakistan
2005	11	14	21:38:51.4	38.107	144.896	11	7.0 Off the East Coast of
Honshu, Japan							
2006	01	02	06:10:49.7	-60.957	-21.606	13	7.4 East of the South
Sandwich Islands							
2006	01	02	22:13:40.4	-19.926	-178.178	583	7.2 Fiji Region
2006	01	27	16:58:53.6	-5.473	128.131	397	7.6 Banda Sea
2006	02	22	22:19:07.8	-21.324	33.583	11	7.0 Mozambique
2006	04	20	23:25:02.1	60.949	167.089	22	7.6 Koryakia, Russia
2006	05	03	15:26:40.2	-20.187	-174.123	55	8.0 Tonga
2006	05	16	10:39:23.3	-31.810	-179.307	152	7.4 Kermadec Islands Region
2006	07	17	08:19:26.6	-9.284	107.419	20	7.7 South of Java, Indonesia
2006	08	20	03:41:48.0	-61.029	-34.371	13	7.0 Scotia Sea
2006	11	15	11:14:13.5	46.592	153.266	10	8.3 Kuril Islands
2006	12	26	12:26:21.1	21.799	120.547	10	7.1 Taiwan Region
2007	01	13	04:23:21.1	46.243	154.524	10	8.1 East of the Kuril Islands
2007	01	21	11:27:45.0	1.065	126.282	22	7.5 Molucca Sea
2007	03	25	00:40:01.6	-20.617	169.357	34	7.1 Vanuatu
2007	04	01	20:39:58.7	-8.466	157.043	24	8.1 Solomon Islands
2007	08	01	17:08:51.4	-15.595	167.680	120	7.2 Vanuatu
2007	08	08	17:05:04.9	-5.859	107.419	228	7.5 Java, Indonesia
2007	08	15	23:40:57.8	-13.386	-76.603	39	8.0 Near the Coast of Central
Peru							
2007	09	02	01:05:18.1	-11.610	165.762	35	7.2 Santa Cruz Islands
2007	09	12	11:10:26.8	-4.438	101.367	34	8.5 Southern Sumatra, Indonesia
2007	09	12	23:49:03.7	-2.625	100.841	35	7.9Kepulauan Mentawai
region, Indonesia							
2007	09	13	03:35:28.7	-2.130	99.627	22	7.0Kepulauan Mentawai Indonesia
2007	09	28	13:38:57.8	22.013	142.668	260	7.5 Volcano Islands, Japan
2007	09	30	05:23:34.0	-49.271	164.115	10	7.4 Auckland Islands, New
Zealand region							
2007	10	31	03:30:15.9	18.900	145.388	207	7.2 Pagan region, Northern
Mariana Islands							

2007	11	14	15:40:50.5	-22.247	-69.890	40	7.7	Antofagasta, Chile
2007	11	29	19:00:20.4	14.944	-61.274	156	7.4	Martinique region,
Windward Islands								
2007	12	09	07:28:20.8	-25.996	-177.514	153	7.8	South of the Fiji Islands
2007	12	19	09:30:27.9	51.350	-178.509	34	7.2	Andreanof Islands,
Aleutian Isl., Alaska								
2008	02	20	08:08:30.5	2.768	95.964	26	7.4	Simeulue, Indonesia
2008	02	25	08:36:33.0	-2.486	99.972	25	7.2	Kepulauan Mentawai
region, Indonesia								
2008	03	20	22:32:57.9	35.490	81.467	10	7.2	Xinjiang-Xizang border
2008	04	09	12:46:12.7	-20.071	168.892	33	7.3	Loyalty Islands
2008	04	12	00:30:12.6	-55.664	158.453	16	7.1	Macquarie Island region
2008	05	12	06:28:01.5	31.002	103.322	19	7.9	Eastern Sichuan, China
2008	06	30	06:17:43.0	-58.227	-22.099	8	7.0	South Sandwich Islands
2008	07	05	02:12:04.4	53.882	152.886	633	7.7	Sea of Okhotsk
2008	07	19	02:39:28.7	37.552	142.214	22	7.0	Off East Coast of Honshu,
Japan								
2008	09	29	15:19:31.5	-29.756	-177.683	36	7.0	Kermadec Islands, New
Zealand								
2008	11	16	17:02:32.7	1.271	122.091	30	7.4	Minahasa, Sulawesi, Indonesia
2008	11	24	09:02:58.7	54.203	154.322	492	7.3	Sea of Okhotsk
2009	01	03	19:43:50.6	-0.414	132.885	17	7.7	Near the North Coast of
Papua, Indonesia								
2009	01	03	22:33:40.2	-0.691	133.305	23	7.4	Near the North Coast of
Papua, Indonesia								
2009	01	15	17:49:39.0	46.857	155.154	36	7.4	East of Kuril Islands
2009	02	11	17:34:50.8	3.884	126.397	22	7.2	Kepulauan Talaud, Indonesia
2009	02	18	21:53:45.1	-27.424	-176.330	25	7.0	Kermadec Islands region
2009	03	19	18:17:40.9	-23.046	-174.659	34	7.6	Tonga Region
2009	05	28	08:24:46.5	16.731	-86.217	19	7.3	Offshore Honduras
2009	07	15	09:22:29.0	-45.762	166.562	12	7.8	Off West Coast of the
South Isl., N.Z.								
2009	08	09	10:55:55.6	33.167	137.941	297	7.1	Near the south coast of
Honshu, Japan								
2009	08	10	19:55:35.6	14.099	92.888	5	7.5	Andaman Islands, India
2009	09	02	07:55:01.1	-7.782	107.297	46	7.0	Java, Indonesia
2009	09	29	17:48:10.9	-15.489	-172.095	18	8.1	Samoa Islands region
2009	09	30	10:16:09.2	-0.720	99.867	81	7.5	Southern Sumatra, Indonesia
2009	10	07	22:03:15.9	-13.057	166.341	45	7.7	Vanuatu
2009	10	07	22:18:53.5	-12.528	166.367	55	7.8	Santa Cruz Islands
2009	10	07	23:13:48.0	-13.071	166.472	29	7.4	Vanuatu
2009	11	09	10:44:54.4	-17.211	178.411	585	7.3	Fiji

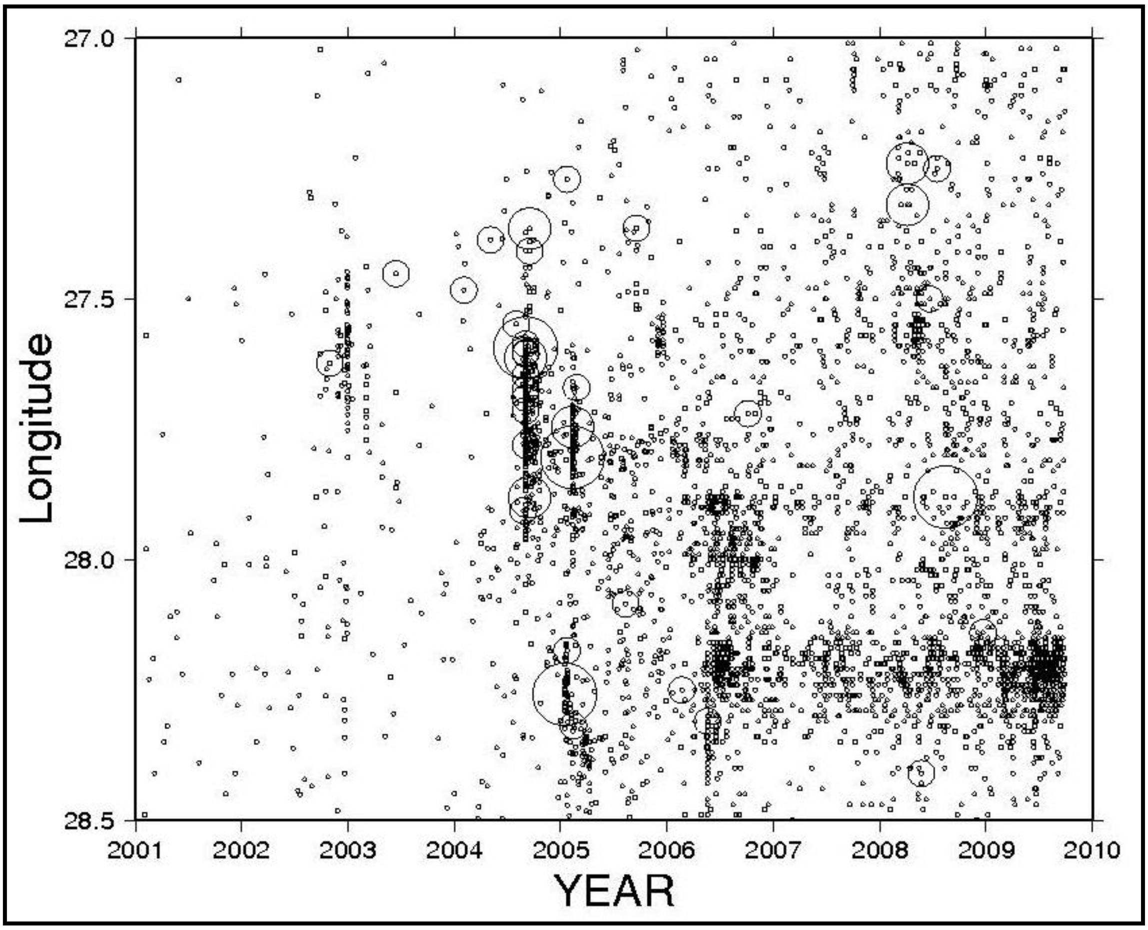

Figure 5.4. The earthquake activity for the Gulf of Gökova in the years between 2001 and 2009 (from NEMC). Circles indicate the earthquakes larger than M = 4.0

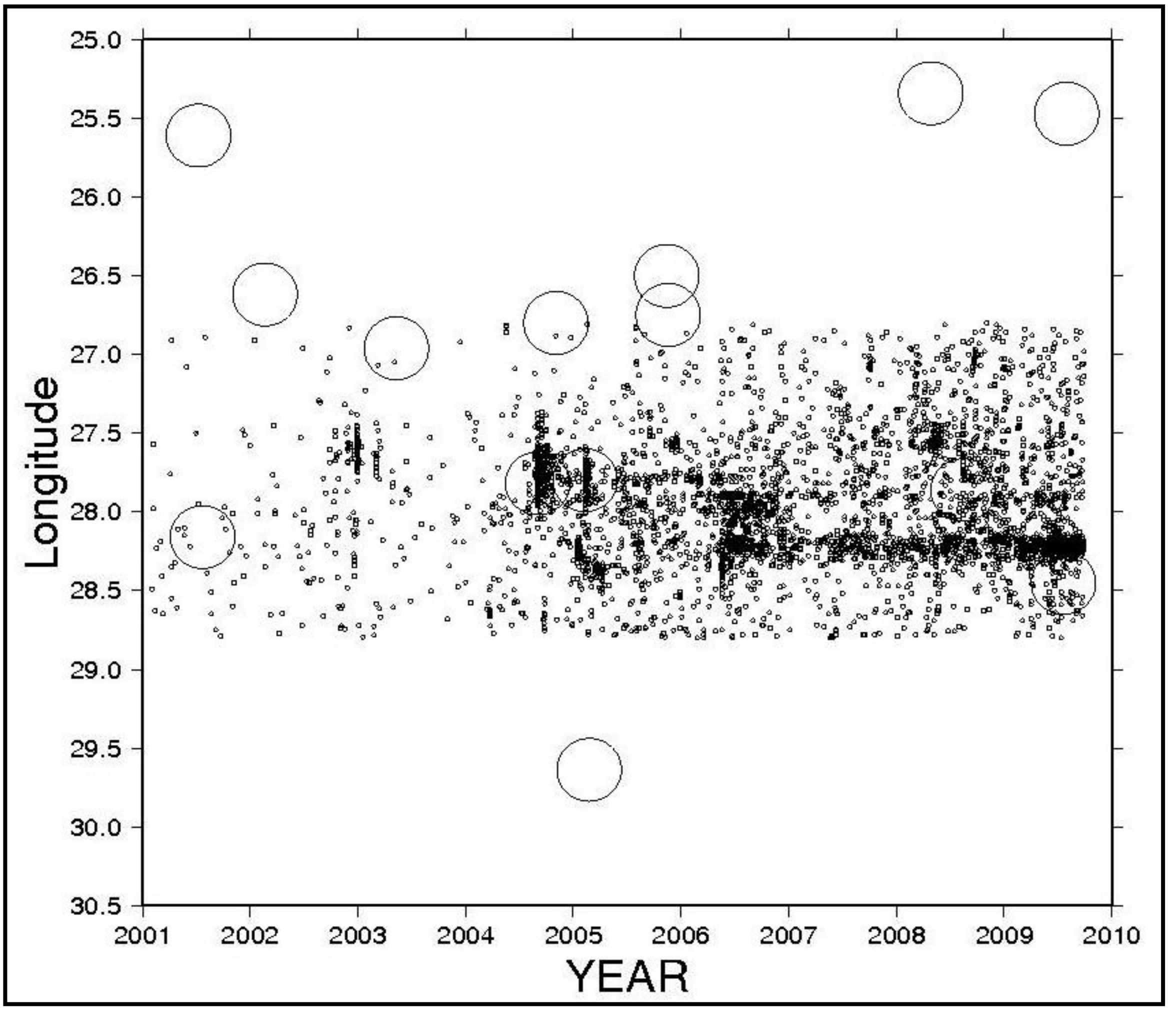

Figure 5.5. The earthquake activity for the Gulf of Gökova (from NEMC) together with regional earthquakes in the years between 2001 and 2009 (from USGS). Circles represent regional earthquakes and the coordinates for them are between 34.00° - 39.00°, 25.00° - 30.00°. The 17 October, 2005 earthquakes shown in Table 5.5 have the same longitude and so they overlapped in the plot

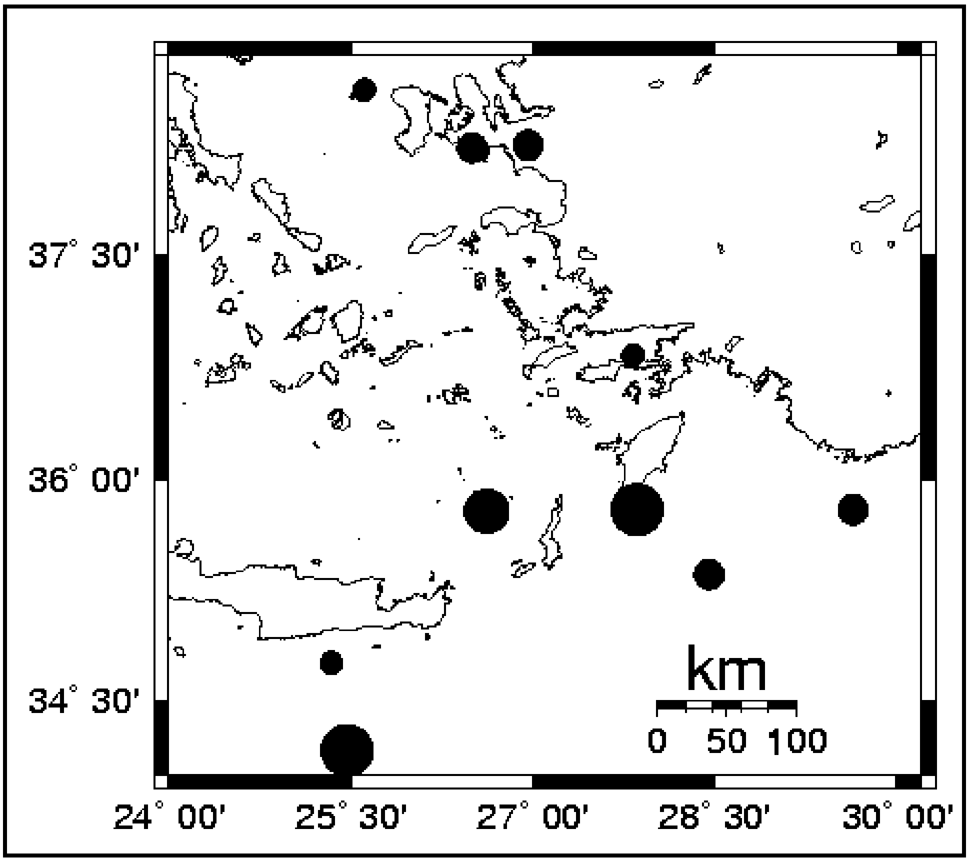

Figure 5.6. Locations of regional earthquakes larger than M = 5.5 around the Gulf of Gökova (from USGS)

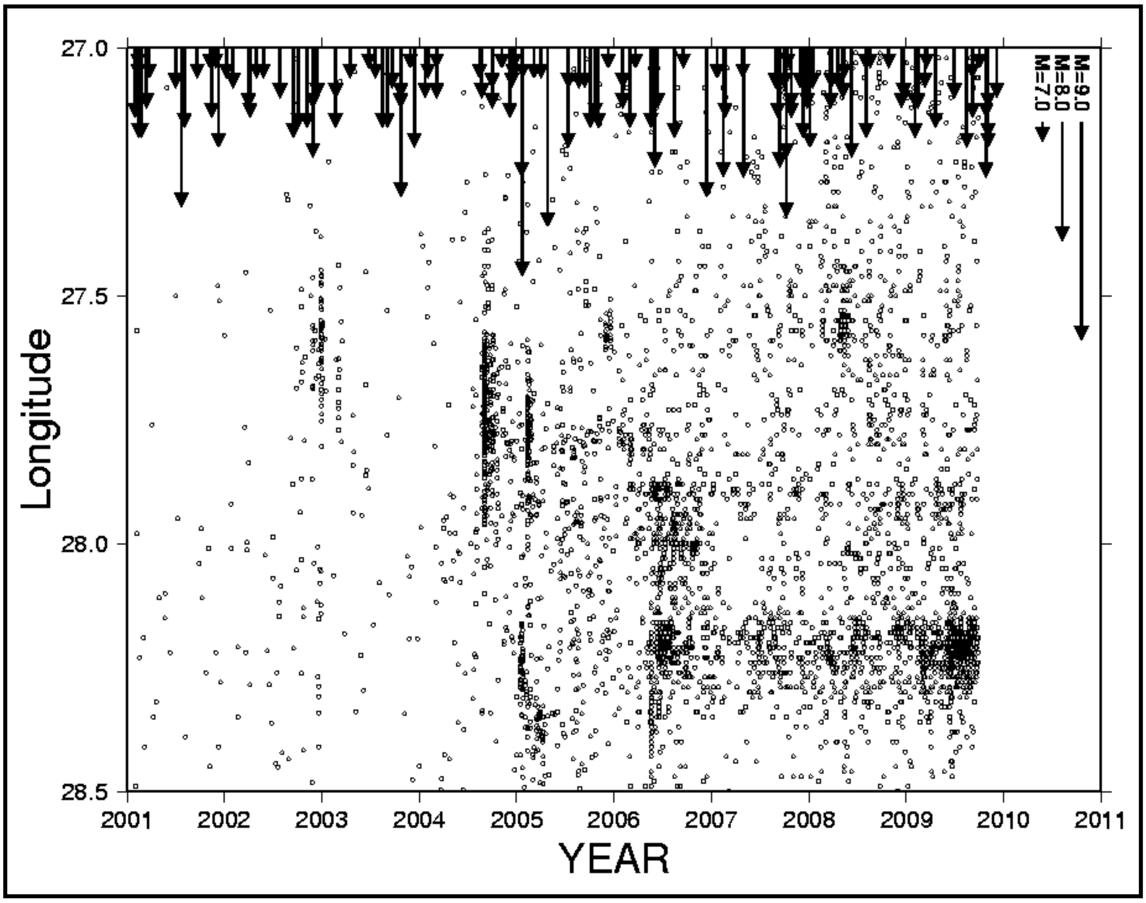

Figure 5.7. The earthquake activity for the Gulf of Gökova (from NEMC) together with teleseismic earthquakes from USGS (2001-2009, only occurrence date and magnitudes can be seen). Arrows are teleseismic earthquakes larger than M = 7.0

Those three graphs in Figure 5.4, Figure 5.5 and Figure 5.7 are analyzed respectively, in detail below.

For Figure 5.4, in the year of 2001 and 2002 the figure shows a limited activity partly due to the reduced number of stations. The last days of the year 2002 is more active than the previous days. An earthquake about M = 4.0 also occurs at the end of 2002 offshore Yalıçiftlik. In the first days of 2003, there are some earthquakes and they may be aftershocks of the earthquake of M = 4.0 in 2002 or may be swarms triggered by it. However, the following days after the first month are silent in 2003 and there is another earthquake about M = 4.0, closer to Bodrum Peninsula and there is no aftershock recorded possibly due to limited detection capability. Distinctively, in the year of 2004 there is a sharp increase for the earthquake activity and the earthquakes about M = 4.0 occurred more than the previous years. In August, 2004 there are a number of medium size events (M>5.0) which occurred offshore Yalıçiftlik and they are followed by aftershocks which last for about two months. As it is looked at the year of 2005, its first two months are very active and in different longitudes, there are earthquakes which are about M = 4.0 and larger. Whole of 2005 has more earthquakes than the previous years and these earthquakes may be aftershocks or triggered by other earthquakes occurring outside the region studied. The same explanation can be used for the cluster in longitude of between 27.50° and 27.60°. In 2006, there is an intense swarm starting approximately from the fifth month and does not seem to be related to any local event except at the eastern end where an event of about M = 4.2 was recorded. This activity is located at the eastern end of the Gulf around Akyaka. Part of this activity is due to the conversion of the YER station from short period to broadband which is expected to improve detection capability. This activity is however not an artefact of station geometry because it has produced many up and down in the following years which would not happen if it was only due to station geometry. Two more earthquakes between 4.0<M<4.5 which occurred more to the west are observed and are not seen to produce any aftershock activity. The intensive seismicity on the eastern part of the Gulf continues to the end of 2007 (longitude of between 28.20° and 28.30°), but on the other hand activity in the longitude of between 27.80° and 28.20° suddenly decreases from the beginning of 2007. After 2007, the activity in longitude of between 28.20° and 28.30° still continues in 2008. Six or seven earthquakes which are at magnitudes of M = 4.0 and larger have occurred in this year. Except the one of these earthquakes (the ones occurred in 28.20°), their aftershocks are not seen. The ones which are larger than these two may have triggered dynamically these earthquakes. Then, also in 2009 the intensive activity at Akyaka (in the longitude of between 28.20° and 28.30°) has not decreased and it seems that there has been a little increase.

It can be understood from Figure 5.4 which shows the microseismic activity along with local events with M>4.0 that nearly all events between 4.0 and 4.5 do not trigger detectable aftershock activity. Only events larger than M>5.0 trigger significant aftershocks which last for about two months. The three of them are: the sequence of August, 2004 (27.50°-28.00°) and the two in January, 2005 (27.75° and 28.30° respectively). All the remaining clustering in Figure 5.4 cannot be associated to static triggering but may possibly be due to triggering by regional or distant events. This is the issue that is tested in the next two figures.

Figure 5.5 displays the general microseismic activity together with the regional events which are larger than M>5.5. Only the longitude of the events can be seen due to the nature of the figure, therefore it is also included a map for the locations of regional events (Figure 5.6). The general appearance of the figure does not show any evidence that regional events did trigger local seismicity. There are four moderate earthquakes in 2001, 2002 and 2003. The limited detection level during these years does not show much detail. However, it looks that they do not have much effect in Gökova area. In August, 2004, the activity around 27.70° is the aftershocks of the M_w = 5.6 local earthquake offshore Yalıçiftlik and which were mentioned before. There are only two possibilities for regional event triggering. The first one is the cluster at 27.50° at the end of the year 2005 and it is possibly triggered by the earthquake sequence (M = 5.5, 5.8, 5.9) of Sığacık which started on 17 October, 2005, in the south of Çeşme Peninsula. The second possibility of regional dynamic triggering is again the cluster at the same location (*i.e.* 27.50°) which started to activate on March, 2008 and may possibly be triggered by the deep earthquake occurred on 28 March, 2008 in the south of Crete with M = 5.6. Apart from these two there seems to be no candidates for dynamic triggering at regional distance.

In Figure 5.4, the earthquake swarms from the middle of 2004 to about the second month of 2005 can be seen clearly and one of these swarm activities which start with the M_L = 5.6 earthquake in Ören, Muğla on 11 January, 2005 comes after the occurrence of 26 December, 2004 dated Sumatra, Indonesia Earthquake which is shown with the longest arrow in Figure 5.7. Therefore, most probably the Ören, Muğla Earthquake may be triggered by the Sumatra Earthquake.

6. CONCLUSIONS

In this thesis, it is asserted that triggered activity may occur in the Gulf of Gökova, for it is one of the most seismically active places in Anatolia. Actually, it is not decisively clear what conditions need to be satisfied in order to determine whether an event is a triggered one or not. Intuitive and/or statistical evaluation of the data can be used for the detection of the triggering mechanism if it exists. In this analysis, only intuitive estimations are applied for implementing a first order identification of triggering processes.

The large variations of stress amplitudes and their comparison with the seismicity rates for the same period give information about the triggering of earthquakes. In order to determine the large stress transients the maximum peaks of the seismograms occurred in July and August, 2007 are found and averaged over the whole area. It is observed that the largest stress variations are mostly due to large events from far distances and medium events from regional distances and occasionally by local events. Stress variations due to local events are seen to create a stress peak at a restricted locality but do not spread over the whole area. Therefore, although the stress average is high the standard deviation is also high which makes the major difference with the global and regional events.

It is observed that there are two earthquakes which cause much larger stress peaks than the others and these two stress peaks are caused by the largest (8 August, 2007 dated M_w = 7.5 Java Earthquake) and smallest (9 July, 2007 dated M_L = 2.3 Gulf of Gökova Earthquake) events.

The largest of the stress transients which is due to the Java Earthquake (Julian day 220) has a smaller variation as compared to the local event as expected. This means that the high stress transient is valid for all the stations around the gulf, so if triggering occurs, it should cover the whole study area. For seismicity rates, there is a peak in seismic production on the day of 224 (Julian). It is nearly 2.5 times the average value of seismic production. While analyzing the stress transients and the seismicity rates together, it is seen that the highest seismicity peak occurs only four days later after the Java Earthquake. So, this observation may be interpreted as a possible evidence for the triggering of the seismicity in Gökova Bay due to a teleseismic event.

On the other hand, the second largest of the stress transients is due to an earthquake in the Gulf of Gökova (Julian day 190) and the peak stress is very high locally, but the standard deviation is also very high as expected. If there is any triggering by this event, this will concern only a small area around the epicentre and it will not affect the whole of Gökova Bay. Accordingly, it is not observed any increase in seismicity rate following this peak.

In addition to the detailed analysis based on the stress amplitudes, the possibility of triggering due to the earthquake activity in the last decade is also analysed. Three types of events are considered for their potential of triggering: local events (M>4.0), regional events (M>5.5) and teleseismic events (M>7.0). It is generally observed that local events which are between 4.0 and 4.5 do not trigger detectable aftershock activity. Only events larger than M>5.0 trigger significant aftershocks which last for about two months. The two examples in this category are the sequence of August, 2004 (27.50°-28.00°) and the ones in January, 2005 (27.75° and 28.30° respectively). The clusters following these two local events reflect clearly the related aftershock activity. All the remaining clusters cannot directly be associated to any significant local event and therefore are not due to static triggering. They may possibly be due to dynamic triggering by regional or distant events.

In August, 2004 and 2005, there are more earthquakes occurred around Gulf of Gökova than the previous years. In 2006, there is an intense swarm starting approximately from the fifth month and does not seem to be related to any local event except at the eastern end where an event of about M = 4.2 was recorded. This activity is located at the eastern end of the Gulf around Akyaka.

There are only two possibilities for regional event triggering. The first one is the small cluster at 27.50° at the end of the year 2005. It is possibly dynamically triggered by the earthquake sequence (M = 5.5, 5.8, 5.9) of Sığacık which started on 17 October, 2005, in the south of Çeşme Peninsula. The second possibility of regional dynamic triggering is again the cluster at 27.50° which started to activate on March, 2008 and may possibly be triggered by the deep earthquake occurred on 28 March, 2008 in the south of Crete (at 25.30° with M = 5.6).

Finally for the case of triggering by teleseismic events, it is highly likely that the intense swarm activity which started with the M_L = 5.1 earthquake in Ören, Muğla on 11 January, 2005 is triggered by the occurrence of Sumatra, Indonesia Mega-Earthquake (M = 9.1, 26 December, 2004).

Although more analytical tools needs to be used for a quantitative evaluation, the present work is still sufficient to show that both static and dynamic triggering are effective in the Gulf of Gökova. This is mostly due to the fact that state of the stress along the faults in the Gulf is at a level which points to the validity of critical stability conditions. Considerable work has already been done in studying triggering mechanism, both static and dynamic, such that static triggering is on the way to enter as a standard tool for the hazard assessment process. However, there is still a long way to go for understanding fully the mechanism of the dynamic triggering and inserting it as a well established procedure for hazard evaluation. The Gökova Gulf, although much smaller in coverage as compared to examples that are studied in the literature, is likely to produce useful results and contribute to the understanding of both types of triggering.

REFERENCES

Aktar, M., H. Karabulut, D. Childs, A. Mutlu, M. Ergin, A. Yörük, V. Geçgel, F. Bulut and T. Kaya, 2006, "Active Faults and Present Seismic Activity in Gökova Bay", *TUBITAK Report*, 104Y336.

Ambraseys, N. N. and C. F. Finkel, 2006, *The Seismicity of Turkey and Adjacent Areas: A Historical Review, 1500-1800*, TUBITAK Press, Academic Series 4, pp. 167-176 [Translation into Turkish language].

Anderson, J. G., J. N. Brune, J. N. Louie, Y. H. Zeng and M. Savage, 1994, "Seismicity in the Western Great Basin Apparently Triggered by the Landers, California, Earthquake 28 June 1992", *Bulletin of the Seismological Society of America*, Vol. 84, No. 3, pp. 863-891.

Barka, A. A., 1992, "The North Anatolian Fault Zone", *Annales Tectonicae,* Vol. 6, pp. 164–195.

Boğaziçi University, Kandilli Observatory and Earthquake Research Institute, National Earthquake Monitoring Centre, http://www.koeri.boun.edu.tr/sismo

Brodsky, E. E., V. Karakostas and H. Kanamori, 2000, "A New Observation of Dynamically Triggered Regional Seismicity: Earthquakes in Greece Following the August, 1999 Izmit, Turkey Earthquake", *Geophysical. Research Letters*, Vol. 27, No. 17, pp. 2741–2744.

Cocco, M. and J. R. Rice, 2002, "Pore Pressure and Poroelasticity Effects in Coulomb Stress Analysis of Earthquake Interactions", *Journal of Geophysical Research,* Vol. 107, No. B2, pp. ESE 2-1 to ESE 2-7.

Coşkun, Z., 2010, *Rapid Determination of Earthquake Magnitude Using ElarmS*, M.S. Thesis, Boğaziçi University.

Dewey, J. F. and A. M. C. Şengör, 1979, "Aegean and Surrounding Regions Complex Multiplate and Continuum Tectonics in a Convergent Zone", *Geological Society of America Bulletin*, Vol. 90, No. 1, pp. 84-92.

Dewey, J. F., M. R. Hempton, W. S. F. Kid, F. Şaroğlu and A. M. C. Şengör, 1986, "Shortening of Continental Lithosphere; the Neotectonics of Eastern Anatolia-a Young Collision Zone" in M. P. Coward and A. C. Ries (eds.), *Collision Tectonics, Geological Society Special Publications*, Vol. 19, No. 1, pp. 3–36.

Dieterich, J. H., 1979, "Modeling of Rock Friction: 1. Experimental Results and Constitutive Equations", *Journal of Geophysical Research*, Vol. 84, pp. 2161–2168.

Dieterich, J. H., 1981, "Constitutive Properties of Faults with Simulated Gouge. In Mechanical Behavior of Crustal Rocks", *Geophysical Monograph Series*, Vol. 24, pp. 103–120, AGU, Washington, D.C.

Dieterich, J. H., 1994, "A Constitutive Law for Rate and of Earthquake Production and Its Application to Earthquake Clustering", *Journal of Geophysical Research*, Vol. 99, No. B2, pp. 2601–2618.

Dreger, D. S. and A. Kaverina, 2000, "Seismic Remote Sensing for the Earthquake Source Process and Near Source Strong Shaking; A Case Study of the October 16, 1999 Hector Mine Earthquake", *Geophysical Research Letters*, Vol. 27, No. 13, pp. 1941–1944.

Earthquake Hazards Program, http://earthquake.usgs.gov/earthquakes

Eberhart-Phillips, D., P. J. Haeussler, J. T. Freymueller, A. D. Frankel and C. M. Rubin, 2003, "The 2002 Denali Fault Earthquake, Alaska: A Large Magnitude, Slip-Partitioning Event", *Science*, Vol. 300, pp. 1113–1118.

Freed, A. M., 2005, "Earthquake Triggering by Static, Dynamic and Postseismic Stress Transfer", *Annual Reviews of Earth Planetary Sciences*, Vol. 33, pp. 335–367.

Gomberg, J. and P. Bodin, 1994, “Triggering of the Ms = 5.4 Little Skull Mountain, Nevada, Earthquake with Dynamic Strains”, *Bulletin of the Seismological Society of America*, Vol. 84, No. 3, pp. 844–853.

Gomberg, J., N. M. Beeler and M. L. Blanpied, 1998, “Earthquake Triggering by Transient and Static Deformations”, *Journal of Geophysical Research*, Vol. 103, No. B10, pp. 24411–24426.

Gomberg, J., P. Bodin and P. A. Reasenberg, 2003, “Observing Earthquakes Triggered in the Near Field by Dynamic Deformations”, *Bulletin of the Seismological Society of America*, Vol. 93, No. 1, pp. 118–138.

Gomberg, J., P. Bodin, K. Larson and H. Dragert, 2004, “Earthquake Nucleation by Transient Deformations Caused by the M = 7.9 Denali, Alaska, Earthquake”, *Nature*, Vol. 427, pp. 621–624.

Gomberg, J., P. A. Reasenberg, P. Bodin and R. A. Harris, 2001, “Earthquake Triggering by Seismic Waves Following the Landers and Hector Mine Earthquakes”, *Nature*, Vol. 411, pp. 462–466.

Görür, N., A. M. C. Şengör, M. Sakınç, O. Tüysüz, R. Akkök, E. Yiğitbaş, F. Y. Oktay, A. Barka, N. Sarıca, B. Ecevitoğlu, E. Demirbağ, Ş. Ersoy, O. Algan, C. Güneysu and A. Aykol, 1995, “Rift Formation in the Gökova Region, Southwest Anatolia: Implications for the Opening of the Aegean Sea”, *Geological Magazine*, Vol. 132, pp. 637-650.

Guidoboni, E., A. Comastri and G. Traina, 1994, “Catalogue of Ancient Earthquakes in the Mediterranean Area up to the 10th Century”, Rome, Instituto Nazionale di Geofisicap.

Gutenberg, B. and C. F. Richter, 1954, *Seismicity of the Earth and Associated Phenomena, 2nd ed.*, Princeton, N.J.: Princeton University Press, pp. 17–19 ("Frequency and Energy of Earthquakes").

Gürer, Ö. F. and Y. Yılmaz, 2002, "Geology of the Ören and Surrounding Areas, S-W Anatolia", *Turkish Journal of Earth Sciences*, Vol. 11, pp. 1-13.

Hancock, P. L. and A. A. Barka, 1981, "Oppose Shear Senses Inferred from Neotectonic Mesofractures Systems in the North Anatolian Fault Zone", *Journal of Structural Geology,* Vol. 3, No. 4, pp. 383-392.

Harris, R. A. and R. W. Simpson, 1996, "In the Shadow of 1857: The Effect of the Great Ft.Tejon Earthquake on Subsequent Earthquakes in Southern California", *Geophysical Research Letters*, Vol. 23, No. 3, pp. 229–232.

Hauksson, E., L. M. Jones, K. Hutton and D. Eberhart-Phillips, 1993, "The 1992 Landers Earthquake Sequence: Seismological Observations", *Journal of Geophysical Research*, Vol. 98, No. B11, pp. 19835–19858.

Havskov, J. and L. Ottemöller (editors), 2005, *SEISAN: The Earthquake Analysis Software for Windows, SOLARIS, Unix and MACOSX, Version 8.1 Manual*.

Hill, D. P., P. A. Reasenberg, A. J. Michael, W. J. Arabasz and G. C. Beroza, 1993, "Seismicity Remotely Triggered by the Magnitude 7.3 Landers, California Earthquake", *Science*, Vol. 260, No. 5114, pp. 1617–1623.

Hough, S. E. and H. Kanamori, 2002, "Source Properties of Earthquakes Near the Salton Sea Triggered by the 16 October 1999 M 7.1 Hector Mine, California, Earthquake", *Bulletin of the Seismological Society of America*, Vol. 92, No. 4, pp. 1281–1289.

Jaeger, J. C. and N. G. Cook, 1979, "Fundamentals of Rock Mechanics", *London: Chapman & Hall*, 3rd edition.

Kanamori, H., J. Mori, E. Hauksson, H. H. Thomas, L. K. Hutton and L. M. Jones, 1993, "Determinations of Earthquake Energy Release and M_L Using TERRAscope", *Bulletin of the Seismological Society of America*, Vol. 83, No. 2, pp. 330-346.

Kilb, D., J. Gomberg, P. Bodin, 2000, “Triggering of Earthquake Aftershocks by Dynamic Stresses”, *Nature*, Vol. 408, pp. 570–574.

Kilb, D., J. Gomberg, P. Bodin, 2002, “Aftershock Triggering by Complete Coulomb Stress Changes”, *Journal of Geophysical Research*, Vol. 107, No. B4, pp. 2060.

King, G. C. P., R. S. Stein and J. Lin, 1994, “Static Stress Changes and the Triggering of Earthquakes”, *Bulletin of the Seismological Society of America*, Vol. 84, No. 3, pp. 935–953.

Kurt, H., E. Demirbağ and İ. Kuşçu, 1999, “Investigation of the Submarine Active Tectonism in the Gulf of Gökova, Southwest Anatolia-Southeast Aegean Sea, by Multi-Channel Seismic Reflection Data”, *Tectonophysics*, Vol. 305, No. 4, pp. 477-496.

Le Pichon, X. and J. Angelier, 1981, “The Aegean Sea”, *Philosophical Transactions of the Royal Society of London*, Vol. A300, pp. 357–372.

Lienkaemper, J. J., J. S. Galehouse, and R. W. Simpson, 2001, “Long-Term Monitoring of Creep Rate Along the Hayward Fault and Evidence for a Lasting Creep Response to 1989 Loma Prieta Earthquake”, *Geophysical Research Letters*, Vol. 28, No. 11, pp. 2265–2268.

Marone, C., 1998, “Laboratory-Derived Friction Laws and Their Application to Seismic Faulting”, *Annual Reviews of Earth Planetary Sciences*, Vol. 26, pp. 643–696.

Marsan, D., 2003, “Triggering of Seismicity at Short Timescales Following Californian Earthquakes”, *Journal of Geophysical Research*, Vol. 108, No. B5, pp. 2266.

Nyst, M., and W. Thatcher, 2004, “New Constraints on the Active Tectonic Deformation of the Aegean”, *Journal of Geophysical Research*, Vol. 109, No. B11, paper 406, pp.23.

Pankow, K. L., W. J. Arabasz, J. C. Pechmann and S. J. Nava, 2004, "Triggered Seismicity in Utah from the November 3, 2002, Denali Fault Earthquake", *Bulletin of the Seismological Society of America*, Vol. 94, No. 6B, pp. S332-S347.

PASSCAL, 2009, PQL II-Program for Viewing Data, http:// www.passcal.nmt.edu/content /pql-ii-program-viewing-data

Prejean, S. G., D. P. Hill, E. E. Brodsky, S. E. Hough, M. J. S. Johnson, S. D. Malone, D. H. Oppenheimer, A. M. Pitt and K. B. Richards-Dinger, 2004, "Remotely Triggered Seismicity on the United States West Coast Following the Mw 7.9 Denali Fault Earthquake", *Bulletin of the Seismological Society of America*, Vol. 94, No. 6B, pp. S348-S359.

Reasenberg, P. A., R. W. Simpson, 1992, "Response of Regional Seismicity to the Static Stress Change Produced by the Loma Prieta Earthquake", *Science*, Vol. 255, No. 5052, pp. 1687-1690.

Roeloffs, E., 1996, "Poroelastic Techniques in the Study of Earthquake-Related Hydrological Phenomena", *Advanced in Geophysics*, Vol. 37, pp. 135–195.

Ruina, A., 1983, "Slip Instability and State Variable Friction Laws", *Journal of Geophysical Research*, Vol. 88, No. B12, pp. 10359–10370.

Scholz, C. H., 1990, "*The Mechanics of Earthquakes and Faulting*", New York: Cambridge University Press, pp. 439.

Stark, M. A. and S. D. Davis, 1996, "Remotely Triggered Microearthquakes at The Geysers Geothermal Field, California", *Geophysical Research Letters*, Vol. 23, No. 9 pp. 945–948.

Stein, R. S., 2003, "Earthquake Conversations", *Scientific American*, Vol. 288, No. 1, pp. 72–79.

Stein, R. S., A. A. Barka and J. H. Dieterich, 1997, "Progressive Failure on the Northern Anatolian Fault Since 1939 by Earthquake Stress Triggering", *Geophysical Journal International*, Vol. 128, pp. 594–604.

Şengör, A. M. C., N. Görür and F. Şaroğlu, 1985, "Strike-Slip Faulting and Related Basin Formation in Zones of Tectonic Escape: Turkey As a Case Study", in K. T. Biddle, N. Christie-Blick (eds.), *Strike-Slip Faulting and Basin Formation*, Society of Economic Paleontologist and Mineralogists, Special Publication, Vol. 37, pp. 227–264.

Şengör, A. M. C. and Y. Yılmaz, 1981, "Tethyan Evolution of Turkey: a Plate Tectonic Approach", *Tectonophysics*, Vol. 75, pp. 181-241.

Toda, S. and R. S. Stein, 2002, "Response of the San Andreas Fault to the 1983 Coalinga-Nunez Earthquakes: an Application of Interaction Based Probabilities for Parkfield", *Journal of Geophysical Research*, Vol. 107, No. B6, 10.1029/2001JB000172.

Toda, S. and R. S. Stein, 2003, "Toggling of Seismicity by the 1997 Kagoshima Earthquake Couplet: a Demonstration of Time-Dependent Stress Transfer", *Journal of Geophysical Research*, Vol. 108, No. B12, 10.1029/2003JB002527.

Uluğ, A., M. Duman, Ş. Ersoy, E. Özel and M. Avcı, 2005a, "Late Quaternary Sea-Level Change, Sedimentation and Neotectonics in the Gulf of Gökova: Southeastern Aegean Sea", *Marine Geology*, Vol. 221, pp. 381-395.

Uluğ, A., N. Kaşer and M. Duman, 2005b, "The Gulf of Gökova and Environment Anatomy and Its Earthquakes", *5th National Symposium on Coastal Engineering, Abstract Book 1st*, Volume, 23-32, Bodrum [in Turkish with English abstract].

Wang, W. H. and C. H. Chen, 2001, "Static Stress Transferred by the 1999 Chi-Chi, Taiwan, Earthquake; Effects on the Stability of the Surrounding Fault Systems and Aftershock Triggering with a 3D Fault-Slip Model", *Bulletin of the Seismological Society of America*, Vol. 91, No. 5, pp. 1041–1052.

Waring, G. A., 1965, “Thermal Springs of the United States and Other Countries of the World-a Summary”, *U.S. Geological Survey Professional Paper*, Vol. 492, pp. 1-383.

Yılmaz, Y., Ş. C. Genç, Ö. F. Gürer, M. Bozcu, K. Yılmaz, Z. Karacık, Ş. Altunkaynak and A. Elmas, 2000, “When Did the Western Anatolian Grabens Begin to Develop?”, in E. Bozkurt, J.A. Winchester and J.D.A. Piper (eds.), *Tectonics and Magmatism in Turkey and the Surrounding Area*, Geological Society, London, Special Publications, Vol. 173, pp. 353-384.