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Full Length Research Paper

A seismological view to Gökova region at southwestern Turkey

Doğan KALAFAT[1]* and Gündüz HORASAN[2]

[1]National Earthquake Monitoring Center, Kandilli Observatory and Earthquake Research Institute, Boğaziçi University, 34680 Çengelköy, Istanbul, Turkey.
[2]Department of Geophysical Engineering, Engineering Faculty, Sakarya University, 54040, Sakarya, Turkey.

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Recent seismic activity in Gökova region can be characterized by earthquake swarms, which mostly occured during 2004 and 2005. This activity was continued for seven months and 1558 seismic events were recorded at this period. The b-value of Guthenberg-Richter relation is investigated for this earthquake swarm and a high b-value is found as 1.73±0.08 using the maximum likelihood method. In this study, seismogram and spectrum characteristics of the earthquake events in Gokova region are analyzed. Accordingly, low frequency waveform and spectrum are obtained for shallow events whereas deeper events are observed to be characterized by high frequency waveform and spectrum. Apart from direct P and S waves, we noticed the presence of strong reflection phase on the seismograms. These reflected phases come from ~17 to 18 km depth in Gökova region.

Key words: Gökova region, reflected phase, earthquake swarm.

INTRODUCTION

The Gökova province is located in the southeast Aegean Sea along the coast of southwest Anatolia which is a region including the major horst and graben systems such as Gediz, Büyük Menderes, Gökova, Burdur and Acıgöl grabens. The gulf of Gökova is surrounded by Bodrum Peninsula to the north, Datça Peninsula to the south and the island of Kos to the west (Figure 1). It has about 25 km maximum N–S width and 100 km E–W length .

During the Early–Middle Miocene period thick volcano sedimentary associations were formed within approximately NS trending fault-bounded continental basins under an E–W extensional regime (Yılmaz et al., 2000). After starting N–S extension, intracontinental plate alkaline volcanic province of western Anatolia was formed during Late Miocene to Quaternary time (Aldanmaz, 2002; Tonarini et al., 2005). Approximately E–W trending grabens and their basin-bounding active normal faults are the most prominent neotectonic features of Western Anatolia (Bozkurt, 2001).

In this region, east-west and northwest-southeast-trending rifts and related faults are the dominant neotectonic features (Şengör et al., 1984). Among these, the Gökova, Yatağan-Muğla and Milas-Ören rifts are most prominent (Figure 1; Görür et al., 1995).

The Milas-Ören and Yatağan-Muğla rifts are older than the Gökova rift. The east-west faults of the Gökova rift everywhere cut the northwest-southeast faults of the Milas-Ören and Yatağan-Muğla rifts. The structural relationships between the northwest-southeast and the East-west Gökova rifts, can be useful to explain the north-south extension of the Gökova region.

Recent studies based upon surface morphology, fault mechanism solutions, seismicity and marine seismic reflection data (Şaroğlu et al., 1995; Görür et al., 1995; Eyidoğan at al., 1996; Kurt et al., 1999; Uluğ et al., 2005) provide evidence of active normal faults in the area. A normal fault trending east-west is a prominent feature of the Gökova rift. The southern border of the Gökova graben is characterised by low-angle faults with listric type (Kurt et al., 1999).

Muğla, Bodrum, Yatağan and Gulf of Gökova are some of the most seismically active areas of the western Turkey. In the instrumental period seismic activity in the Gökova region includes the earthquakes of 23 April, 1933 (Ms = 6.4), 23 May, 1941 (Ms = 6.0), 13 December, 1941

*Corresponding author. E-mail: kalafato@boun.edu.tr.

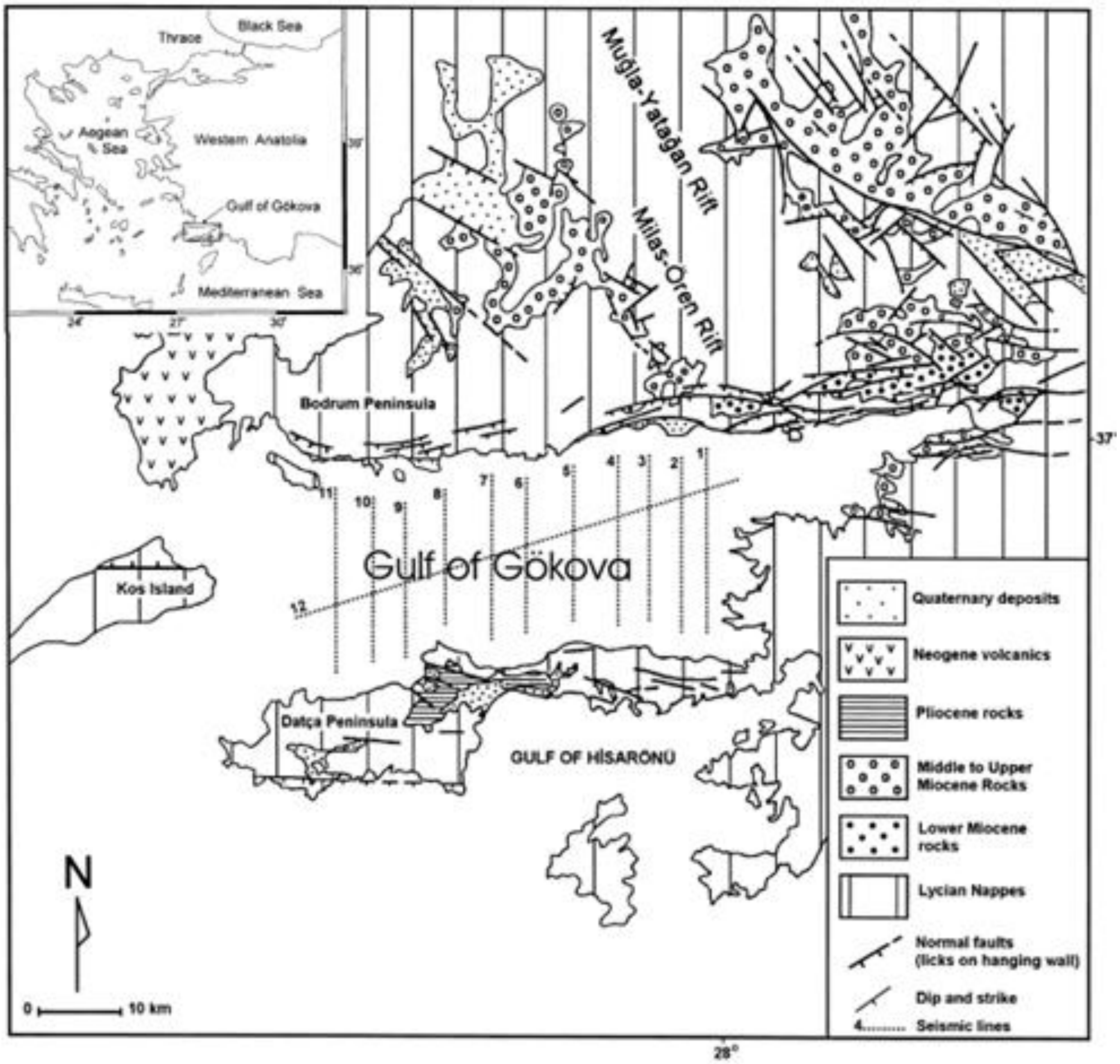

Figure 1. Land geology map of the Gökova province (modified from Görür et al., 1995; the original map also includes inferred submarine faults). Notice the E–W-oriented new graben system (the Gulf of Gökova and its margins) and NW–SE-oriented older graben systems (Muğla–Yatağan Rift and Milas–Ören Rift) (from Kurt et al., 1999).

(Ms = 6.5), 25 April, 1959 (Ms = 5.9) and 5 October, 1999 (Ms = 5.2) events. In August 2004 series of earthquakes (3 August M = 5.0; 4 August M = 5.4; 4 August M = 5.0) occured in Gökova Gulf. Another earthquake sequence continued from 20 December, 2004 (Mw = 5.3) to 10 January, 2005 (Mw = 5.4). In Figure 2 we can clearly see that the rate of seismicity increased in time after August 2004.

The released energy is predominantly in swarm type in this region (Figures 3 and 4). Earthquake swarms occur in many regions of the world like Long Valley caldera, USA; Yellowstone, USA; West Bohemian, Germany.

During the individual swarms, numerous events occurred consecutively as multiplets, that is event occur at nearly the same position and have vary similar source mechanisms (Horálek et al., 2000). The occurance of multiplets is a phenomenon obivously observed in geothermal or volcanic regions (Lees, 1998). It has been suggested that earthquake swarms occured because of stress perturbations associated with the migration of magmatic or hydrothermal fluids through new or previously formed crustal inhomogeneities including crustal fractures (Hill, 1977; Toda et al., 2002; Waite and Smith, 2002). There is a correlation between higher b values and the location of hydrothermal features in the western half of Yellowstone (Farrell, 2009). This would indicate that the high b-values may be due to both the highly fractured (heterogeneous) crust and the high temperatures as well as high pore pressures that allow hydrothermal fluid flow. Therefore, the high b-values

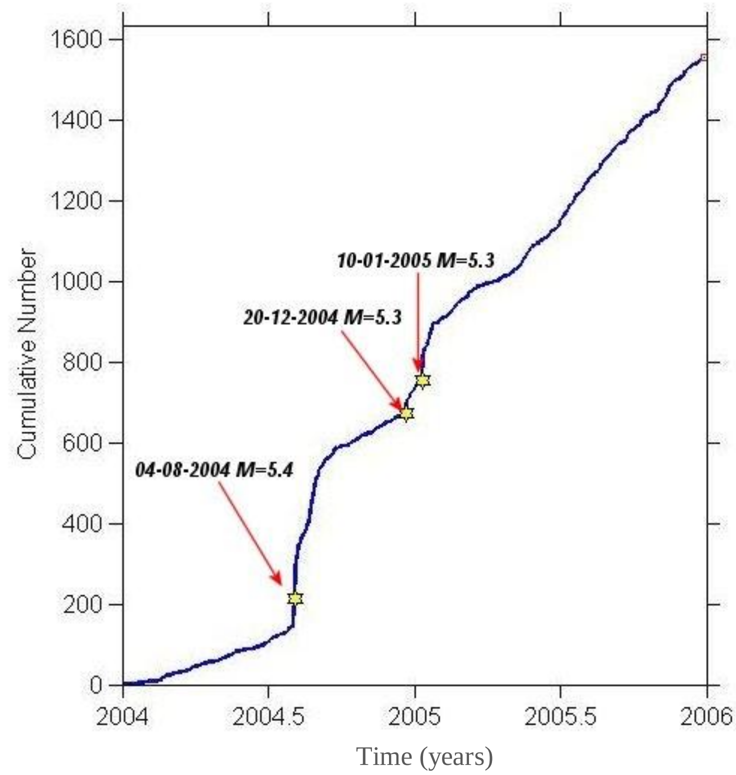

Figure 2. Cumulative number of earthquakes recorded at Gökova, as a function of time, January 2004 through December 2005. Stars represent the 4 August 2004, M = 5.4; the 20 December 2004, M = 5.3 and the 10 January 2005, M = 5.3 earthquakes.

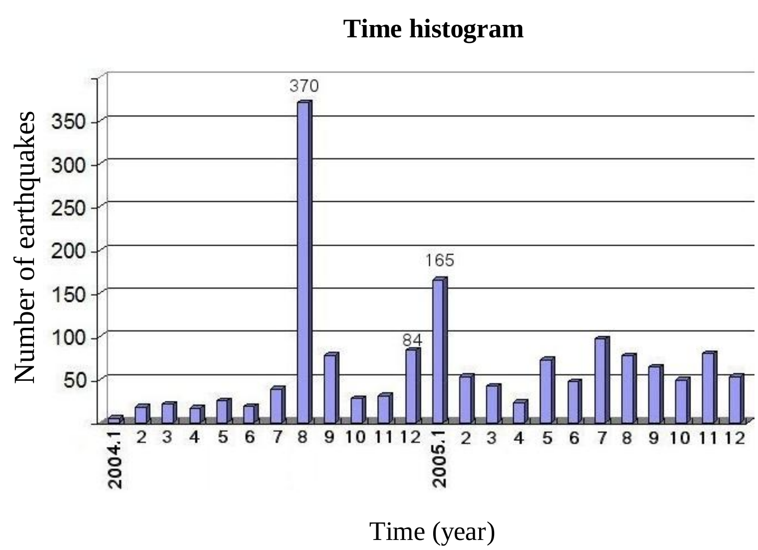

Figure 3. Plot of the number of events located for two years (2004-2006) in Gökova region.

could be an indication of the highly fractured crust that facilitates the movement of hot, hydrothermal fluids.

West of Datça (Cnidus) peninsula lies near the active volcanic centres of Nisyros and Yali (Figure 5), from which ash deposits crop out in patches around the Datça area. Major eruptive activity has occurred on Nisyros in

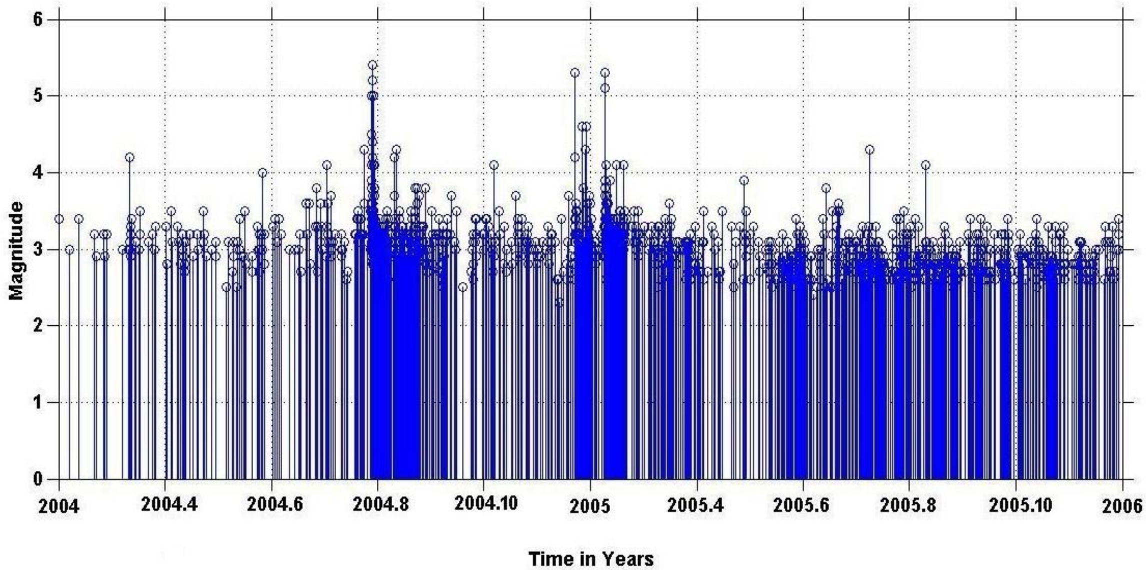

Figure 4. Plot of earthquake magnitudes as a function of time for an earthquake swarm that occured in August 2004, and from December 2004 to January 2005.

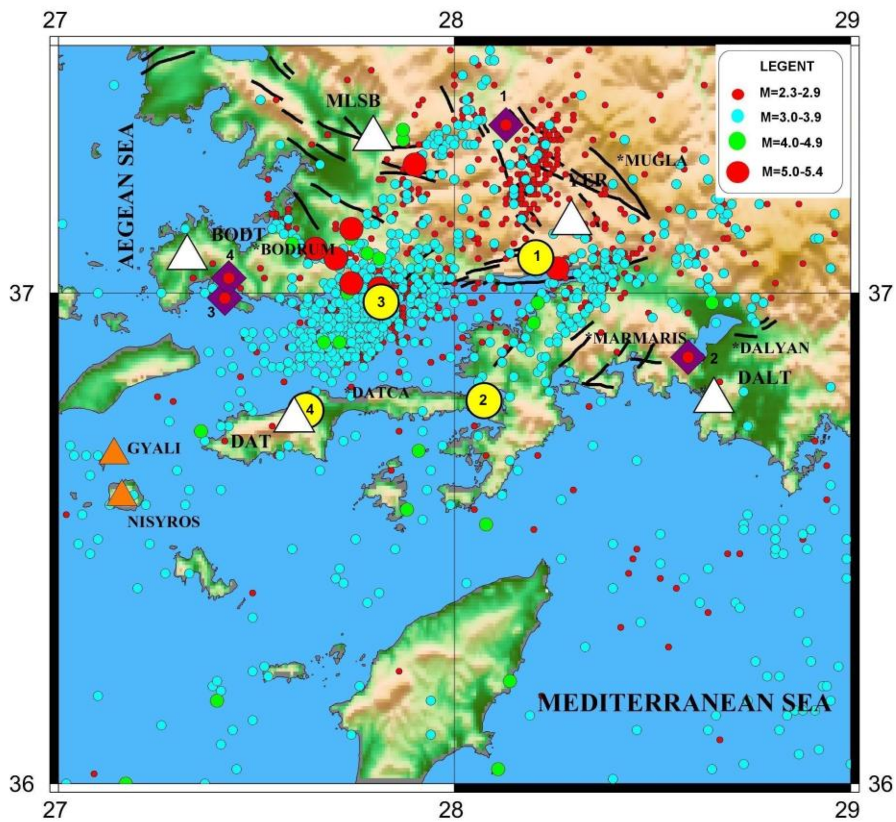

Figure 5. This map shows the epicentre distribution of earthquakes located by the seismic network of KOERI during a period from the begining of 2004 to the end of 2005 and location of seismic events investigated in the article (filled big yellow circles) and seismic stations (filled white triangles). Purple diamonds (1-4) indicate the following hot water springs in the area (Çağlar et al., 2000); 1. Bozhöyük (Yatağan), 2. Sultaniye (Köyceğiz), 3. Karaada (Bodrum), 4. Tavşanburnu (Bodrum). The orange triangles indicate Nysros and Yali volcanoes.

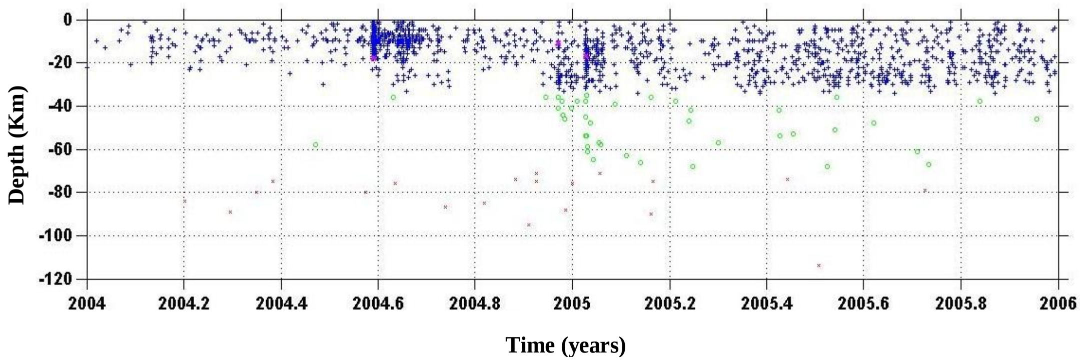

Figure 6. Time-depth plot of earthquakes recorded at Gökova region between 2004 and 2005.

recent times (AD, 1887, 1873 and possibly around 1422) and these violent volcanic events may have been associated with intense seismic activity (Stiros, 2000).

West of Bodrum peninsula is covered by volcanic rock outcrops (Ercan et al., 1984; Robert et al., 1992; Kurt and Arslan, 2001; Çubukçu, 2002; Genç, 2001; Güleç and Hilton, 2006). The volcanic rocks of the Bodrum Peninsula, in SW Turkey and NE of the Hellenic Arc, outcrop over an area of 138 km^2 (Ulusoy et al*.,* 2004). Ulusoy et al. (2004) indicated that a monzonitic intrusion is exposed in the western part of Bodrum peninsula. They investigated the structure of the Bodrum caldera using "Satellite Pour Observer la Terre" (SPOT) image, digital elevation model (DEM), aerial photographs as well as field data.

There are numerous hot water springs in the SW of Turkey (Figure 5) such as Sultaniye (Köyceğiz), Karaada, Tavşanburnu (Bodrum) and Bozhöyük (Yatağan) (Çağlar et al., 2000).

Gökova region is suitable for the occureance of swarm activity. We can tell these characteristics as active faults, crustal fractures, volcanic rock outcrops, and hydro-thermal features. Our main aim is to determine the temporal features (frequency-magnitude distribution) of the earthquakes, to analyze waveform and spectrum characteristics of earthquakes in Gökova region and to give tectonic implications using these waveform data.

DATA AND METHODOLOGY

The seismicity in Gökova region is continously monitored using broadband stations operated by Kandilli Observatory Research Institute since 2004. Kalafat et al. (2005) installed a network of broad band stations called as Blue net to determine precise earthquake locations in that region. Before this date there was only a one component short period station (YER) in the region. This station was replaced with a broadband in July 2006.

The epicenter distribution of earthquakes located by the seismic network of KOERI (Kandilli Observatory and Earthquake Research Institute) during a period from the begining of 2004 to the end of 2005 are shown in Figure 5.

The locations of seismic events were determined by the program hypo71pc (Lee and Valdes, 1989) with a 1-D velocity depth model (Kalafat et al., 1987).

In August 2004 earthquake series (3 August ML = 5.0; 4 August ML = 5.4; 4 August ML = 5.0) occured in Gulf of Gökova. Another earthquake sequence continued from 20 December, 2004 (Mw = 5.3) to 10 January, 2005 (Mw = 5.4). This activity continued for six months. 1558 seismic events were recorded at this period. Magnitudes range between 2.2 to 5.4 for these events. The most of the micro-earthquakes occured at depths between 1 and 30 km. Depth distribution of the events are seen in Figure 6.

The earthquakes are recorded digitally after the installation of MLSB (Milas) (installed in September 2003), DALT (Dalyan) (installed in August 2004), BODT (Bodrum) (installed in February 2005), DAT (Datça) (installed in October 2005) and YER (Yerkesik) (installed in July 2006) stations. All stations (MLS, DALT, BODT, DAT and YER) are located on limestone.

Sampling rate of the digital data is 20 samples per second before December 2005 and 50 sps after that.

The most common characterization of earthquake populations is the cumulative frequency–magnitude distribution that can be described by the Gutenberg–Richter relation (Gutenberg and Richter, 1956):

$$\log_{10} N = a - b \cdot M \quad (1)$$

where N is the absolute number of earthquakes with magnitudes greater than or equal to a magnitude M. Frequency-magnitude distribution of the recent 2004 to 2005 earthquake swarms (1558 earthquakes) in Gökova region is examined in this study. We used the ZMAP software package for this examination (Wiemer, 2001). We also examined the digital waveforms and spectrum chracteristics of the earhquakes given in Table 1.

We used the fast Fourier transform (FFT) method of PITSA program (Programmable Interactive Toolbox for Seismological Analysis) (Scherbaum and Johnson, 1992) to calculate the normalized amplitude velocity spectra of seismograms.

RESULTS AND DISCUSSION

Waveforms and spectrum of the four events are seen in Figures 7, 8, 9 and 10 respectively. We gave the information for these four events in Table 1.

We found different behaviors when examined the waveforms and spectrum characteristics of the events occured in the Gökova region. We observed low frequencies on seismogram and spectrum for event 1 in Figure 7. Spectrum of this event has the frequency content restricted in a narrow band between 1 and 2.5 Hz.

Table 1. Location parameters used for waveform and spectral analyses of seismic events given in Figure 5 (filled big yellow circle).

Event number	Date (d:m:y)	Origin time (h:m:s)	Latitude (°N)	Longitude (°E)	Depth (km)	Magnitude
1	22.12.2004	20:29:16.8	37.06	28.19	3.0	3.6
2	25.09.2005	19:09:25.4	36.77	28.06	67.0	3.4
3	31.01.2007	23:13:46.0	36.97	27.80	11.0	3.7
4	21.05.2007	07:30:52.9	36.75	27.61	5.5	3.8

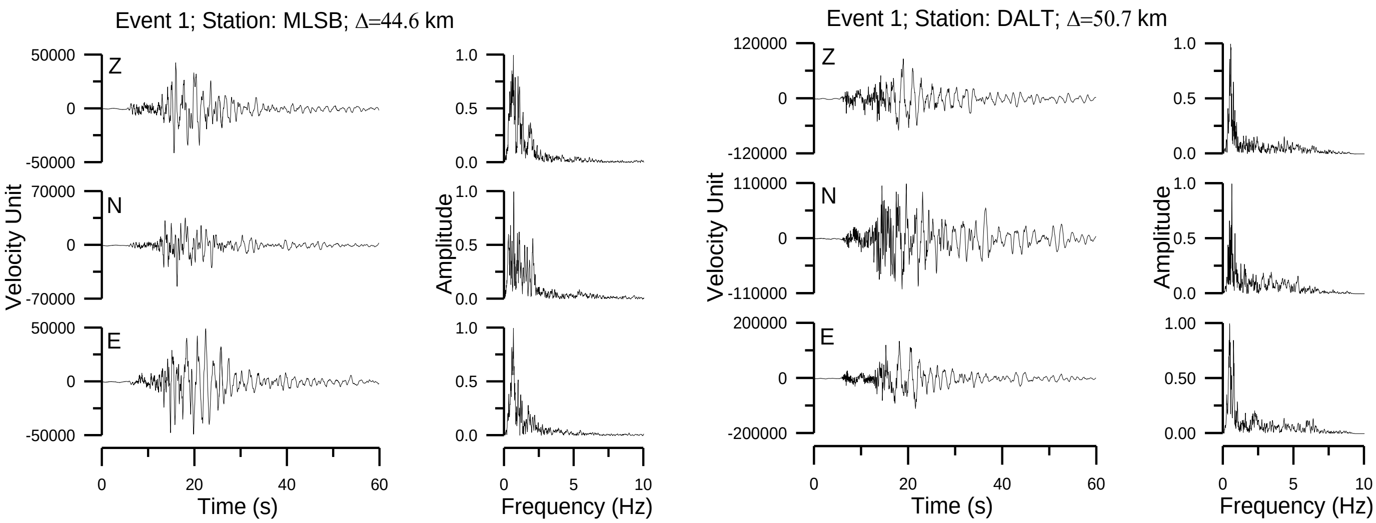

Figure 7. Three components of the velocity seismograms (MLSB, DALT) with a time window length of 60 s and its normalized amplitude spectrums for Event 1.

Source depth is 3 km for this event.

Whereas, we observed high frequencies on waveform and spectrum for events 2, 3 and 4 in Figures 8, 9 and 10. Event 2 has a 67 km focal depth. Hypocenter locations also give 60 to 70 km focal depth in this part of the region because of the Anatolian and Agean lithospheric border. The spectrum of MLSB and DALT for events 3 and 4 have lower frequency content compared to others (DAT, BODT and YER).

The frequencies of shallow earthquakes decrease when the seismic waves travel through attenuative medium. The fact that hot springs and volcanic rocks which attenuates high frequencies are observed in the area (Figure 5) can be linked to resulting low frequency spectrum found in this study.

Volcanic regions, particularly ones where shallow magma bodies and/or hydrothermal systems are present, frequently exhibit seismic swarm activity. Long Valley caldera (Hill et al., 2003), Campi Flegrei (Aster et al., 1992), Yellowstone (Waite and Smith, 2002), and the Socorro Magma Body have all experienced recent seismic swarm activity associated with vertical deformation.

The *b*-value parameter itself is often useful in understanding the causes of an earthquake swarm. For most tectonic regions of the Earth, $b \leq 1.0$ (Minakami, 1990). However, active volcanic areas can have much larger *b*-values, often with $b \geq 1.5$, because of increased crack density and/or high pore pressure. Examples include Mount Pinatubo, Philippines (Sánchez et al., 2004), Ito, Japan (Wyss et al., 1997), and Etna, Italy (Vinciguerra, 2002).

We utilize the maximum likelihood method (Weichert, 1980) to compute *a, b* value for this earthquake sequence (Figure 11). The *b*-value for this earthquake sequence is found as 1.73 ± 0.08 (Figure 11). This high *b*-value is attributed to the presence of a high thermal gradient due to the emplacement of magmatic fluids, existence of hot springs and/or highly fractured heterogeneous media.

The b-value distribution for the Yellowstone volcanic region was determined as 1.5 ± 0.05. This high value associated with the youthful 150,000-year old Mallard Lake resurgent dome (Farrel et al., 2009). Sánchez et al. (2004) obtained the frequency–magnitude distribution of earthquakes at Mount Pinatubo, Philippines measured by the *b*-value. They found that *b*-values are higher than normal ($b = 1.0$) and range between $b = 1.0$ and $b = 1.8$. This high *b*-value anomaly infered as increased crack density, and/or high pore pressure, related to the presence of nearby magma bodies.

We observed secondary phases on seismograms

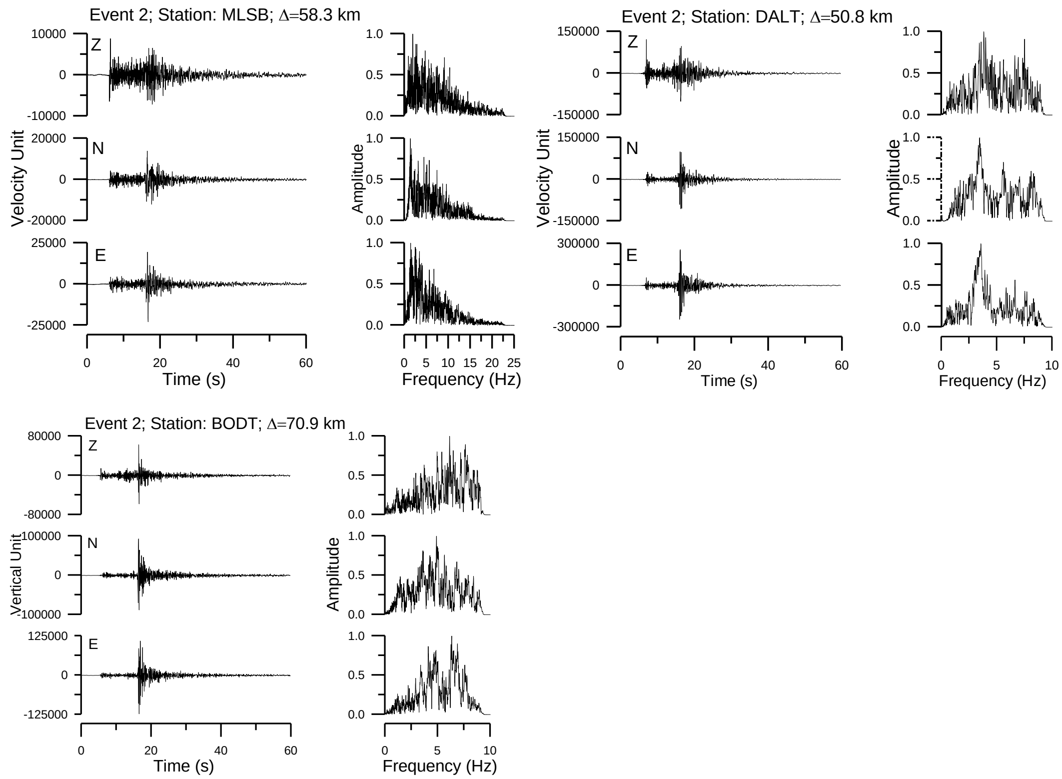

Figure 8. There components of the velocity seismograms (MLSB, DALT and BODT) with a time window length of 60 s and its normalized amplitude spectrum for Event 2.

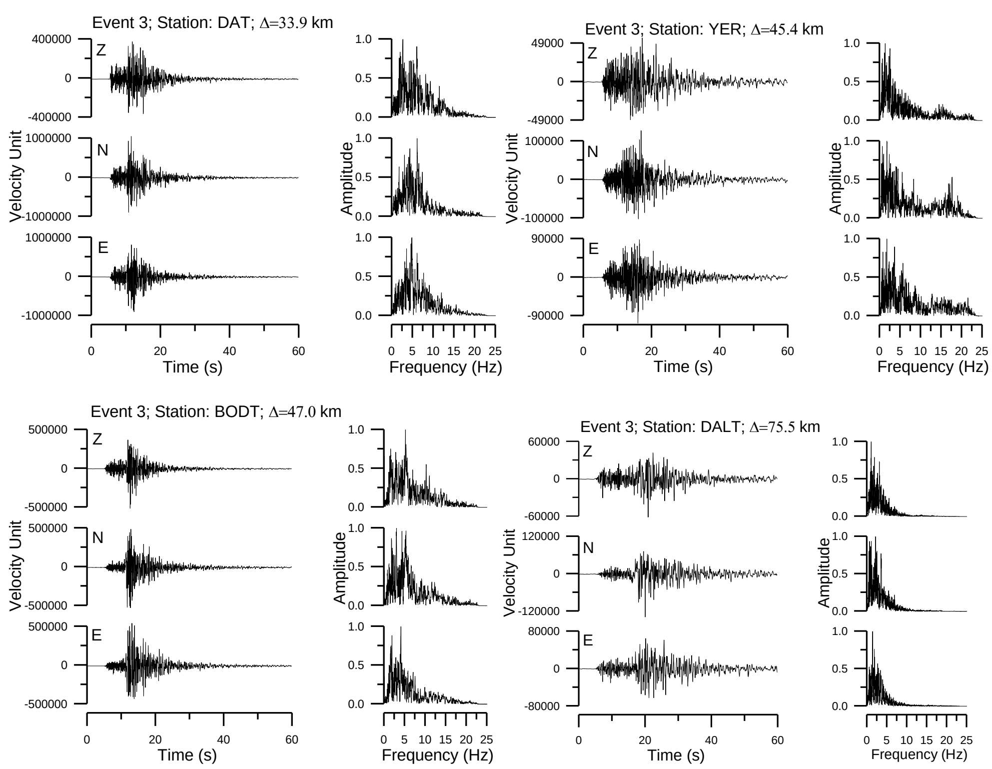

Figure 9. Three components of the velocity seismograms (DAT, MLSB, YER, BODT and DALT) with a time window length of 60 s and its normalized amplitude spectrum for Event 3.

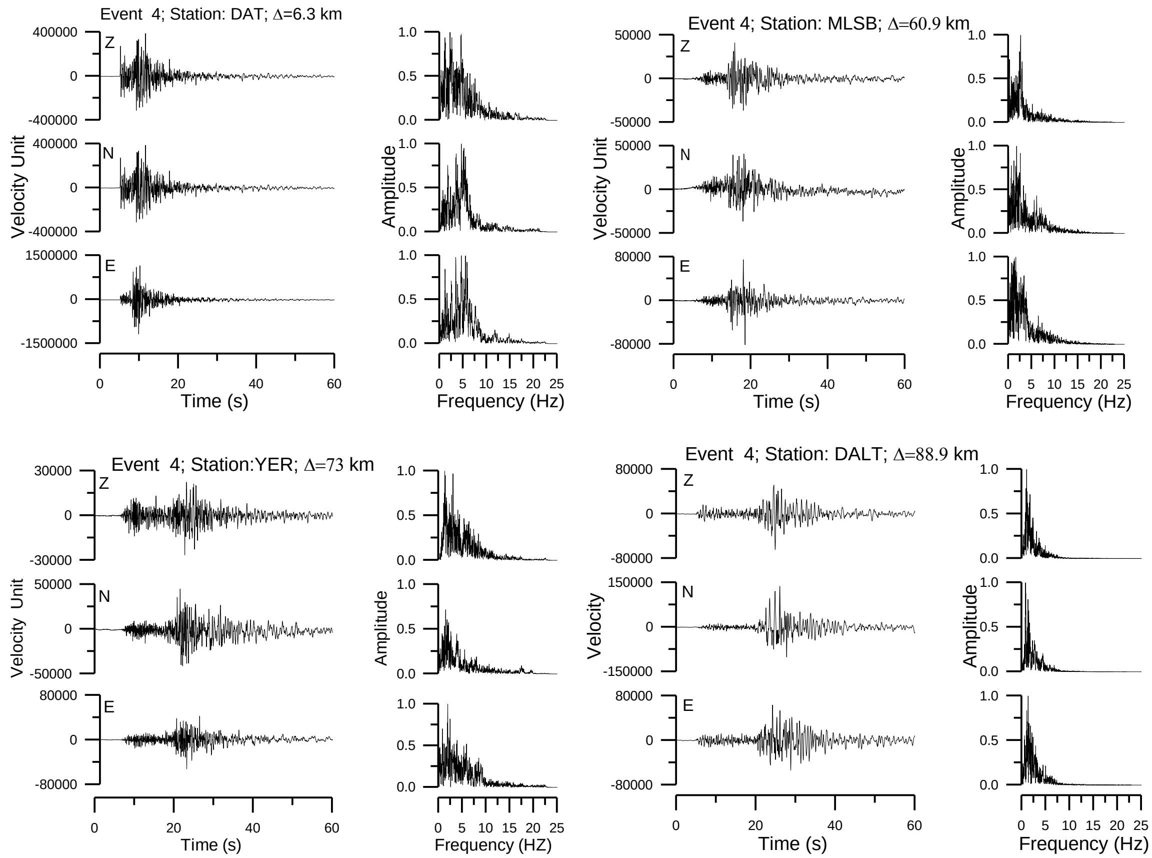

Figure 10. Three components of the velocity seismograms (DAT, MLSB, YER and DALT) with a time window length of 60 s and its normalized amplitude spectrum for Event 4. The event is not recorded at BODT station.

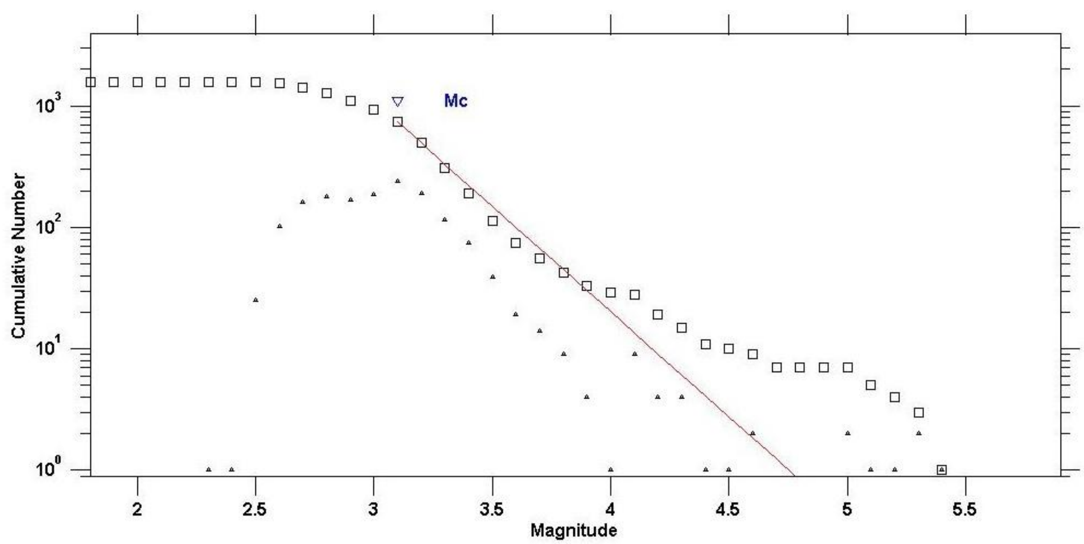

Figure 11. Plot of cumulative frequency of earthquakes as a function of magnitude for the Gökova earthquake sequence. The computed b value (line) obtained using the maximum likelihood method (Weichert, 1980) is 1.73±0.08. White triangle indicates data completeness magnitude, *M* 3.1.

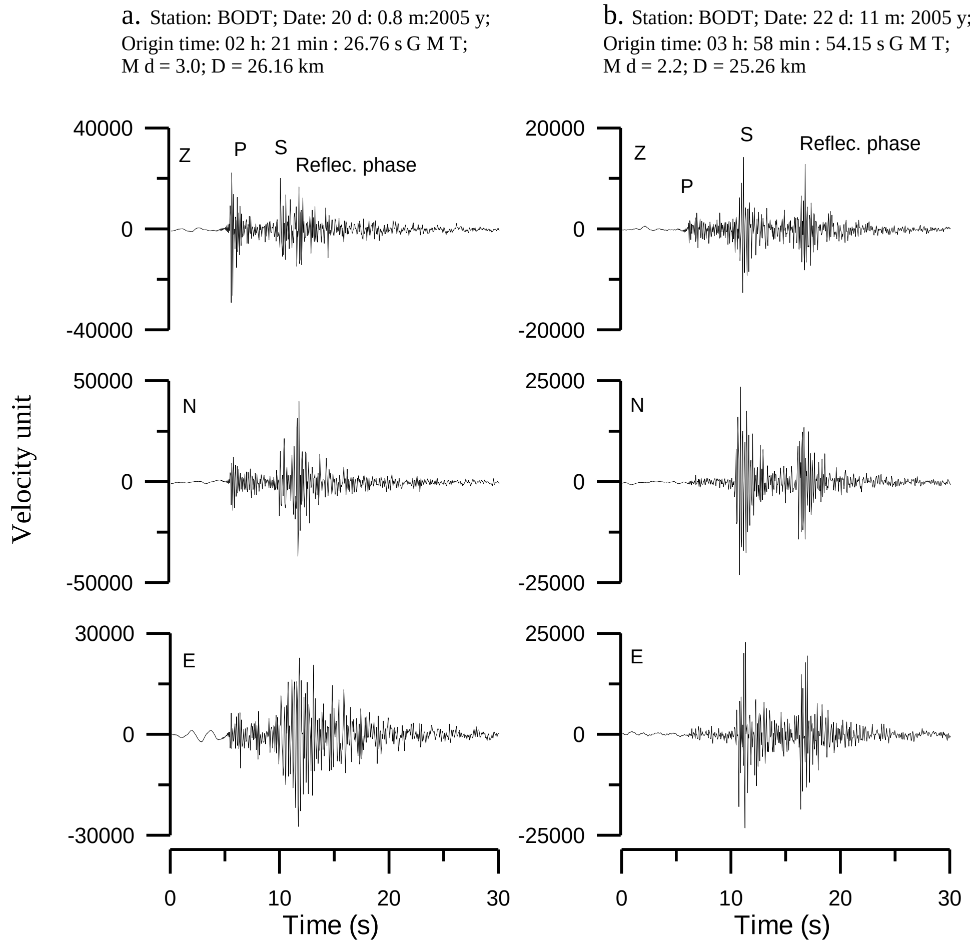

Figure 12. Microearthquake seismograms showing reflection phases. Seismograms are recorded by BODT which is a broadband station.

examined for the events in the region (Figure 12).

These phases arrive to BODT station ~ 1.9 to 7.3 s after the direct S-phase arrival. To help identify the secondary phases, we used travel time curve of events. The graph of travel times of direct and secondary phases versus distance (S-P interval) suggest that these secondary phases are reflections. In Figure 13, travel times for the two phases (direct and reflected S wave) from each earthquake are plotted with different symbols. These reflected phases are generally sharp and large amplitude in horizantal components. The large amplitudes of the reflected phases are explained by a large S-phase velocity contrast across the discontinuity and preferential downward radiation of S-wave energy from the earthquake foci (Sanford et al., 1973).

Sanford and Holmes (1961) first noted the presence of unusual secondary phases on microearthquake seismograms recorded at Socorro and suggested that the phases could be reflections. Sanford et al. (1973) attributed these phases to an interface between rigid and nonrigid crust. The fundamental characteristics of these waves are given in the paper of Sanford et al. (1973), Sanford and Long (1965). Later, Sanford et al. (1977) concluded that this interface was the top of a sill-like magma body near 19 km depth and estimated its lateral extent by calculating the reflecting positions of SzS arrivals.

When Kurt et al. (1999) made multi-channel reflection study in Gökova bay, they did not see the continuation of the Datça fault in the deeper part of seismic section 11 (Figure 1). They interpreted that the hanging wall consist of Lycian Nappes at the bottom and basin fill at the top. They said that they observed strong reflections, due to the high acoustic impedance contrast where the fault plane is in contact with the basin fill at Gökova bay.

One of possibilities for the reflector is the detachment in fault plane and basin fill. Another possibility for the reflector at these depths is magma sources beneath Gökova region. There are volcanic rocks outcroped in the region. To determine the depth of reflecting discontinuity we used the S-wave velocity for crust and upper mantle as 3.37 and 4.64 km/s respectively (Atılganoğlu, 2007).

We calculated the reflection depth as ~17 to 18 km in Gökova region. To give more detailed charactertics of

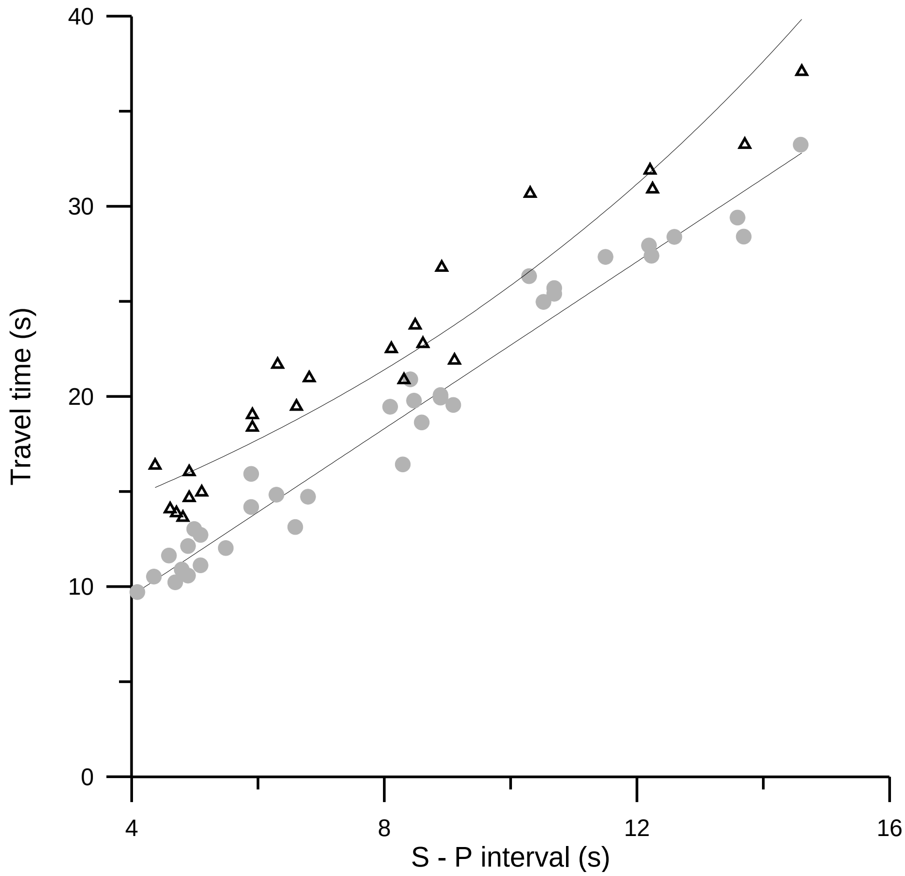

Figure 13. Direct S (filled circles) and reflectd phase (blank triangles) travel time versus S-P interval. Travel times for these phases obtained from the microearthquake events that was used in this study.

these reflected phases, further studies such as lateral velocity distribution and modelling are needed.

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