

**AN EXAMPLE OF DETERMINATION OF S WAVE VELOCITY
USING MICROTREMORS AND SPATIAL
AUTOCORRELATION METHOD**

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TABLE OF CONTENTS

ABBREVIATIONS	v
LIST OF FIGURES	vi
LIST OF SYMBOLS	ix
SUMMARY	x
ÖZET	xi
1. INTRODUCTION	1
2. SURFACE WAVES, MICROTREMOR AND SURVEY METHODS	4
2.1. Physical properties of Rayleigh waves and dispersion	4
2.2. Properties of Microtremors	9
2.3. Temporal and Spatial Variation of Microtremors	12
2.4. Microtremor Survey Method (MSM)	19
2.4.1. Basic Concepts of SPAC (Spatial Autocorrelation Method)	20
2.4.1.1. Spectral Representation of Microtremors in Polar Coordinate System	22
2.4.2. Other microtremor survey methods	29
2.4.2.1. Time domain beamformer	30
2.4.2.2. Conventional frequency domain beamformer (FDBF)	31
2.4.2.3. Minimum variance distortionless look (MVDL)	31
3. GEOLOGY, TECTONICS AND GEOTECHNICAL CHARACTERISTICS OF THE ADAPAZARI AREA	33
3.1. Geology	33
3.1.1. İstanbul Zone	39
3.1.1.1. Soğuksu Formation	39
3.1.1.2. Karadere Formation	41
3.1.1.3. Kocatöngel Formation	41
3.1.1.4. Bakacak, Kurtköy and Aydos formations	41
3.1.1.5. Kartal formation	41
3.1.1.6. Yılanlı formation	41
3.1.1.7. Çakraz formation	42
3.1.2. Sakarya Zone	42
3.1.2.1. Pamukova Metamorphics	42
3.1.2.2. İznik Metamorphics	42
3.1.3. Late Cretaceous and younger units	42
3.1.3.1. Akveren formation	43
3.1.3.2. Yığılca - Çaycuma Formations	43
3.1.3.3. Karapürçek – Kanlıçay Formations	43
3.1.4. Quaternary units	44
3.1.4.1. Basin margin deposits	45
3.1.4.2. Basin deposits	45

3.2. Geotechnical characteristics of Adapazarı basin	52
3.3. Tectonic setting of the region	54
4. DATA ANALYSIS AND DETERMINATION OF V_s PROFILES	58
4.1. Data acquisition and analysis procedure for SPAC method	58
4.1.1. Data analysis by Spatial Autocorrelation (SPAC) Method	59
4.1.1.1. Standardization of the spatial autocorrelation function	61
4.1.1.2. The size of sample N and the analysis interval T_r	62
4.1.1.2. Calculation of the phase velocity	64
4.2. Field study	64
4.2.1. Instruments	66
4.3. Analysis of microtremors and determination of subsoil structure	68
4.4. Inversion procedure and inverted V_s profiles.	73
5. DISCUSSION OF THE RESULTS AND CONCLUSION	77
REFERENCES	80
APPENDIX A	85
RESUME	95

ABBREVIATIONS

ASF	: Array Smoothing Function.
CTX	: Cyclic Triaxial
DRM	: Disaster Risk Management
DSİ	: Genaral Directorate of State Hydraulic Works
EIES	: Eurasia Institute of Earth Sciences
FDBF	: Frequency Domain Beam Former
GPS	: Global positioning system
HD	: Hersek Delta
HKD	: Microtremor Measurement Station Name
İL	: İznik Lake
KH	: General Directorate of Rural Services
KOERI	: Kandilli Observatory and Earthquake Research Institute
LASA	: Large Aperture Seismic Array
MIS	: Microtremor Measurement Station Name
MSM	: Microtremor Survey Method
MTA	: General directorate of Mineral Research and Exploration
MVDL	: Minumum Variance Distortionless Look
NAF	: North Anatolian Fault
ODTÜ	: Middle East Technical University
PSD	: Power Spectral Density
RC	: Rosanant Column
SL	: Sapanca Lake
SMO	: Sapporo Meteorological Observatory
SPAC	: Spatial Autocorrelation
TS	: Turkish Standart
TÜBİTAK	: The Scientific and Technical Research Council of Turkey
TYT	: Toyota Automobile Factory
UBC	: Unified Building Code
UL	: Ulubathı Lake

LIST OF FIGURES

	<u>Page Numbers</u>
Figure 2.1: Deformation produced by surface waves: (a) Rayleigh wave; (b) Love wave (Bolt, 1993).....	4
Figure 2.2: Relation between Poisson's ratio and velocity of propagation of compression (P), shear (S) and Rayleigh (R) waves in a linear elastic homogeneous half space (Kramer, 1996).	5
Figure 2.3: Particle motion on the surface during the passage of a Rayleigh waves in an elastic homogeneous halfspace.(Foti, 2000)	6
Figure 2.4: Amplitude ratio vs. dimensionless depth for Rayleigh wave in a homogenous half-space (Richart et al., 1970).....	6
Figure 2.5: Geometrical dispersion in layered media (Rix, 1988).....	7
Figure 2.6: Examples of non-dispersive (homogeneous half-space), normally dispersive and inversely dispersive profiles (Rix, 1988).	8
Figure 2.7: (a)Group and phase velocities. (b) Arrival of a dispersive wave at different geophones (Sheriff and Geldart, 1995).	9
Figure 2.8: Seventy-five permanent seismic observatories in the world, which record common vibration of the earth including microtremors. (Peterson, 1993)	12
Figure 2.9: The acceleration power spectra of common vibration of the earth recorded at the seventy-five seismic observatories in the world (Peterson, 1993)	12
Figure 2.10: Power spectra of the vertical (U-D) and east west (E-W) components of microtremors at HKD and MIS observed in the 15 day period between 19 th Nov. and 03 Dec., 1997 the spectra are calculated and overplotted for 120 10-minutes blocks (Okada, 2003).	14
Figure 2.11: Day-time and night-time averages of power spectra of the vertical (U-D) and east-west (E-W) components of microtremors at HKD and MIS, observed in the 15-day period between 19 th Nov. and 03 Dec., 1997 (Okada, 2003).	15
Figure 2.12: Running power spectra of the E-W component of microtremors observed over the 29-day period between 5 th Nov. and Dec 3 (HKD) and the 15-day period between 19 Nov and 3 Dec, 1997 (MIS). Atmospheric pressure at SMO for this period is shown on the top (Okada, 2003).....	16
Figure 2.13: Power spectra of 10 stations within the close proximity for 45-minute records at 6 different times (Okada, 2003).	18
Figure 2.14: A schematic graph of spatial autocorrelation coefficient (Okada, 2003).	28
Figure 2.15: Flow of data analysis by the f-k method (Okada, 2003).	30
Figure 2.16: A schematic representation of time domain beamformer method.	30

Figure 3.1: Geological units of the Marmara region (after Okay 1986).....	33
Figure 3.2: Tectonic setting of the eastern Marmara region (from Görür et al., 1997).	34
Figure 3.3: Stratigraphic section of the İstanbul zone (After Akyüz, unpublished).....	35
Figure 3.4: Map showing the distribution of Neogene and Quaternary sediments in the eastern Marmara region. HD: Hersek Delta, SL: Sapanca Lake, İL: İznik Lake, UL: Ulubat Lake (After Emre et al., 1998)	36
Figure 3.5: Columnar stratigraphic sections of Eastern Marmara Region (not in scale) (Emre et al., 1998)	37
Figure 3.6: Neotectonic period faults in the Eastern Marmara Region (Emre et al., 1998).....	38
Figure 3.7: Geological map of the Sakarya basin and close vicinity. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999)	40
Figure 3.8: Simplified geological map that shows borehole locations. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999)	44
Figure 3.9: Vertical cross-section of boreholes in north of Adapazarı Basin. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999)	46
Figure 3.10: Vertical cross-section of boreholes in south of Adapazarı Basin. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).....	47
Figure 3.11: Vertical cross-section of boreholes in center and east of Adapazarı Basin. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999)	49
Figure 3.12: Vertical cross-section of boreholes in northwest of Adapazarı Basin. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999)	50
Figure 3.13: Vertical cross-section of boreholes in north, south, west and the centre of the Adapazarı Basin. (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).....	51
Figure 3.14: Compression of the structural models suggested for the Marmara region ([A] Pınar,1943 [B] Pfannenstiel et al.,1944 [C] Crampin et al.,1986 [D] Şengör,1987 [E] Barka et al.,1988 [F] Wong et al.,1995 Ergun et al.,1995)	54
Figure 3.15: Active fault map of the region. (Şaroğlu et al,1992)	55
Figure 3.16: The recent high resolution bathymetric map obtained from the survey of the Ifremer RV Le Suroit vessel that indicates a single, thoroughgoing strike slip fault system.(Le Pichon et al.,2001)	56
Figure 3.17: Tectonic map of Marmara region compiled from various studies (Yaltırak, 2002)	56
Figure 4.1: Simple array geometry for SPAC that is composed of four geophone (Kudo et al., 2002).	59
Figure 4.2: Flow of data analysis by the SPAC method (Okada, 2003).	60
Figure 4.3: Some photos from the field work. (a) Arrangement of the stations of smallest array (2 m). (b) Arrangement of the signal conditioner and digitizer unit for the smallest array measurement	65
Figure 4.4: The coherences of instrumental responses used in this study.	66
Figure 4.5: Some photos from the field instrument. (a) Triaxial accelerometer (b) Signal conditioner and data logger (c) General view of	

accelerometer, signal conditioner, data logger, computer and their connection.	67
Figure 4.6: Microtremor recordings at site TYT for 60 min total continuous time duration	69
Figure 4.7: Microtremor recordings at site TYT selected time window (81.92 sec.)	69
Figure 4.8: (a)Cross power spectrum of the microtremors recorded at the four stations, (b) Frequency coherence functions for pairs of stations corresponding to the distance of 116 m.	71
Figure 4.9: (a) SPAC coefficients of every pair of stations with inter-distance 116m, (b) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 116m.....	72
Figure 4.10: Observed SPAC coefficients (circles) and Bessel functions fitted to them (continuous line) for representative frequencies 1.5, 2.0, 2.1, 2.4, 3.4, 4.5, 4.6 and 4.8 Hz.	74
Figure 4.11: Fitting between the observed dispersion curve (circles) and the theoretical one (solid line) of phase velocity of the Rayleigh wave recorded in the microtremor measurements at the site Toyota.	76
Figure 4.12: Final inverted Vs profile calculated by the experimental dispersion curve and initial artificial model.	76
Figure A.1: Microtremor recordings at site TYT for 35 min total continuous time duration ($r = 5$ m)	86
Figure A.2: Microtremor recordings at site TYT selected (20.48 sec) time window ($r = 5$ m)	86
Figure A.3: Cross power spectrum of the microtremors recorded at the four stations ($r = 5$ m)	87
Figure A.4: (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 5 m.(c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 5 m.....	88
Figure A.5: Microtremor recordings at site TYT for 30 min total continuous time duration ($r = 20$ m.)	89
Figure A.6: Microtremor recordings at site TYT selected (20.48 sec.) time window ($r = 20$ m)	89
Figure A.7: Cross power spectrum of the microtremors recorded at the four stations ($r = 20$ m)	90
Figure A.8: (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 20 m.(c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 20 m.....	91
Figure A.9: Microtremor recordings at site TYT for 60 min total continuous time duration ($r = 58$ m.)	92
Figure A.10: Microtremor recordings at site TYT selected (40.96 sec.) time window ($r = 58$ m)	92
Figure A.11: Cross power spectrum of the microtremors recorded at the four stations ($r = 58$ m)	93
Figure A.12: (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 58 m.(c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 58 m.	94

LIST OF SYMBOLS

B_e	: Resolution bandwidth
$c(f)$	: Phase velocity
f	: Frequency
f_N	: Highest expected frequency
G_k	: Smoothed estimate value of the spectra
g	: Spatial covariance function
h	: Damping value
H	: Hermitian transpose
$h_0(w)$	: Power spectral density function
J_0	: Bessel function of the first kind of zero order
k	: Wavenumber vector
k_x, k_y	: Components of wavenumber vector
m	: Maximum lag number
N	: Sample size
n	: Degree of freedom
r	: Radius
S_0	: Spatial autocorrelation function
S_A	: Soil classification indice in UBC system for hard rock
S_B	: Soil classification indice in UBC system for rock
S_C	: Soil classification indice in UBC system for soft rock
S_D	: Soil classification indice in UBC system for stiff soil
S_E	: Soil classification indice in UBC system for soft soil
T_r	: Minimum record length
t	: Time
V_s	: Shear wave velocity
V_R	: Rayleigh wave velocity
w	: Angular frequency
X	: Sample function of recorded microtremor
y_i	: Signal plus noise measured i^{th} sensor
$z(t)$	: beamformer output for an assumed vector
ζ	: Stationary stochastic process
ϕ	: Angle between the plane containing the radius vector and z axis and the plane containing the x axis and z axis, or the longitude of the particle
θ	: Angle between the radius vector and the z axis, or the latitude of the particle
ε	: Standard error of the power spectrum
λ	: Wavelength
ξ	: Position vector
$\rho(f,r)$	: Spatial autocorrelation coefficient
Δ_i	: Time delay
Δt	: Time increment

AN EXAMPLE OF DETERMINATION OF S-WAVE VELOCITY USING MICROTREMORS AND SPATIAL AUTOCORRELATION METHOD

SUMMARY

It is widely recognized that the determination of S-wave velocity structure is important in accurately predicting strong ground motion. Such information is required for aseismic design and mitigating the damage distribution caused by earthquakes. Microtremor array measurement has been recognized as one of the attractive and convenient exploration method for determining the S-wave velocity with less practical restriction than the other geophysical and geotechnical methods, especially in highly populated urban areas where much consideration must be given to the vibrotational environment.

Microtremor survey method (MSM) has not been developed as an application or extension of the conventional seismic method, but it is a totally new method based on totally new idea. The microtremor method is derived from several assumptions and conditions in its basic theory, which impose a limitation in applicability, so that it needs additional observation. For this reason, the microtremor method is, at present, regarded as a method not suitable for detailed survey in any field of application.

The applicability of the microtremor array measurement is, at least theoretically, not limited to any field. It is a survey method for delineating subsurface structure, not only for urban ground but also for exploration of oil, gas and other resources. In this sense it is no different from the conventional seismic method (Okada, 2003).

In this study, array measurement of microtremors at Toyota (TYT) automobile factory site to the south of the city of Adapazarı were performed to estimate the S-wave velocity of soil formation. Adapazarı is a city where strong earthquakes repeatedly occurred and many buildings were completely destroyed during the instrumental and historical period. It is located at one of the most hazardous regions in Turkey, very close to North Anatolian Fault (NAF) zone.

The Spatial Autocorrelation (SPAC) method was used to determine phase velocity dispersion curve in the frequency range from 0.9 to 4.5 Hz. An inversion technique was subsequently applied to determine the S-wave velocity profile at the examined site using by SURF96 surface waves analysis software developed by Hermann in 2002. The determination of S-wave velocity profiles reached a depth of 400 meters.

The evaluation of obtained S wave velocity profile and associated information was performed using by Uniform Building Code (UBC, 1997). According to this, soil classification for Toyota site can be summarized as follows; The top 19 m was classified soft soil (S_E). From 19 m to 34 m was classified stiff soil (S_D), from 34 m to 94 m was classified very dense soil and soft rock (S_C) and deeper than 94 m was classified rock (S_B).

MİKROTREMOR VE UZAYSAL OTOKORELASYON YÖNTEMİ KULLANILARAK S DALGASI HIZININ BELİRLENMESİNE BİR ÖRNEK

ÖZET

S dalgası hız yapısının belirlenmesi, kuvvetli yer hareketinin doğru tahmin edilmesinde son derece önemli bir kavramdır. Bu belirlemeler aynı zamanda depreme dayanıklı yapı tasarımı ve depremden kaynaklanan hasar dağılımının değerlendirilmesi için de ayrı bir önem taşımaktadır. Özellikle çevresel titreşimlerin önem kazandığı yerleşim ve nüfus açısından çok yoğun kentsel bölgelerde, diğer jeofizik ve jeoteknik yöntemlere nazaran uygulamalarda daha az kısıtlamalar içermesi nedeniyle Mikrotremor Ağ Ölçümleri S dalgası hız yapısının belirlenmesinde oldukça ilgi çekici ve amacına uygun bir yöntem olarak kabul edilmektedir.

Mikrotremor Ölçüm yöntemi (MSM) bilinen klasik sismik yöntemlerin bir uygulaması yada uzantısı olarak geliştirilmiş bir yöntem değil tamamıyla farklı bir fikre dayanan yeni bir yöntemdir. Bu yöntem uygulamada çeşitli sınırlamalar getiren ana teorisinden çeşitli kabuller ve koşullar getirilerek türetilmiş ve üzerinde araştırmaların devam ettiği bir yöntemdir. Bu sebepten en azından şimdilik detaylı bir arazi araştırması uygulaması için uygun bir yöntem değildir.

Mikrotremor Ağ Ölçümlerinin uygulanabilirliği, en azından teorik olarak, herhangi bir arazi ile sınırlı değildir. Bu yöntem, yalnızca kentsel bölgelerin zemin yapısının belirlenmesinde değil aynı zamanda petrol, doğalgaz ve diğer doğal kaynakların araştırılmasında da kullanılabilen bir araştırma yöntemidir ve bu yönüyle klasik sismik yöntemlerden farkı yoktur (Okada, 2003).

Bu çalışmada Adapazarı şehrinin güneyindeki Toyota (TYT) otomobil fabrikası alanı içerisindeki zeminin S dalgası hız yapısının belirlenmesi için Mikrotremor Ağ Ölçümleri yapılmıştır. Adapazarı tarih boyunca güçlü depremlerle sarsılmış ve bu depremlerde pekçok bina yıkılmıştır. Şehir Türkiye'nin deprem açısından en tehlikeli yerleşim alanlarından biridir ve Kuzey Anadolu Fay Zonuna çok yakındır.

0.9-4.5 Hz aralığında faz hızı dispersiyon eğrisinin elde edilmesi için Uzaysal Otokorelasyon (SAPC) Yöntemi kullanılmış ve daha sonra çalışılan sahadaki S dalgası hız profilinin belirlenmesi için Hermann tarafından 2002 yılında geliştirilmiş olan SURF96 yazılımı yardımıyla faz hızı dispersiyon eğrisi verilerine ters çözüm uygulanmıştır. Belirlenen S dalgası hız profilinin penetrasyon derinliği 400 metreye kadar ulaşmaktadır.

Analizler sonucu elde edilen S dalgası hız profili ve diğer bulgular Amerikan şartnamesi (Uniform Building Code, UBC, 1997) kullanılarak değerlendirilmiştir. Buna göre Toyota otomobil fabrikası alanı içerisindeki zeminin sınıflaması kısaca şöyle özetlenebilir. İlk 19 metre yumuşak zemin (S_E), 19 metre ile 34 metre arası sert zemin (S_D), 34 metre ile 94 metre çok sıkı zemin ve yumuşak kaya (S_C), son olarak 94 metreden sonrası kaya (S_B).

1. INTRODUCTION

The knowledge of seismic site transfer function is important, both in seismic source studies, in order to separate site effects from the earthquake spectrum, and for assessing the seismic hazard. This knowledge can particularly be high in areas geologically characterised by soft sediments, which may produce amplification of the ground motion at low frequencies (Aki, 1988). The distribution of damages produced by strong earthquakes is in fact highly controlled by the local geology (Anderson et al., 1986, Singh et al., 1983, Hough et al., 1990), and the amplification of earthquake ground motion by local site effects has important implications for urban planning and development.

The geometry of the subsoil structure, the soil types and the variation of their properties with depth, the lateral discontinuities and the surface topography are at the origin of large amplification of ground motion and hence of intensive damage during destructive earthquakes. For this reason, the accurate knowledge of the geometry and the shear wave velocities of the alluvial-diluvial deposits and the basement are the key parameters controlling the amplification of seismic motion.

The shear wave velocity can usually be determined in the field by using conventional seismic methods (e.g. reflection, refraction, borehole) and in the laboratory through dynamic/cyclic tests (RC, CTX), on intact and remolded soil samples. The use of conventional exploration seismic methods to determine the shear wave velocity presents many difficulties, when deep sedimentary structures need to be investigated. For example, in reflection and refraction test, the use of artificial sources such as explosives or vibrators is inevitable, which is very difficult in urban areas. Furthermore, the dimensions of the required arrays are large according to the desired penetration depth and therefore it is difficult to be deployed in densely populated

areas. Moreover the cost of large scale deep geophysical prospecting is very high and therefore in most cases in site effect studies the depth of the seismic basement is limited to a layer with shear wave velocity larger than 1000 m/s and not the real very deep substructure reflector of the incident waves. Additionally, the cost for implementing deep borehole seismics is also very high while the reliability of these seismic methods, such as crosshole and downhole, at large depths is often questionable, because of practical limitations (e.g. energetic sources, equipment management etc.).

The motivations for using surface waves for soil characterization derive by the inherent nature of this kind of waves and by some specific properties of them. Indeed such waves travel along a free surface, so that it is relatively easy to measure the motion associated to them, and they carry important information about the mechanical properties of the medium.

The specific features of surface wave propagation in stratified medium make them very interesting also in the field of Earth's constitutive materials characterization. Applications started to be developed for earthquake seismology with the attempts to infer the characteristics of the rocks through which earthquake generated waves travel (Dorman and Ewing, 1962).

The wave theoretical approach based on modelling the wave propagation to fit the observed phase velocities or amplitudes spread out starting from the early sixties with the advent of digital computers. This kind of research founded the basis of all the tools needed for soil characterization: Mathematical models of layered systems, techniques to obtain wave propagation parameters from point measurements of motion and inversion algorithms (Aki and Richard, 1980, Ewing et.al., 1957, Doyle 1995).

Applications to site characterization at a geotechnical scale started at the end of fifties. The implementation of site characterization using by surface wave requires essentially three separate steps; in field testing an arbitrary characteristic of particle motion associated to wave propagation is experimentally measured; subsequently a signal analysis procedure is applied to extract from the records the experimental dispersion curve; finally using an inversion algorithm based on an appropriate model, the mechanical properties of the soil are obtained.

Two methods are currently used to extract surface waves from microtremors by array measurements, the frequency (f) wave number (k) power spectral density (f-k) method and the spatial autocorrelation (SPAC) method. The SPAC method is probably more convenient method compared to the f-k power spectral density method, because it gives equally well results with the f-k method by using less recording stations and arrays of shorter dimensions (Okada, 2003).

Aki (1957) gave a theoretical basis of the SPAC coefficient defined for ambient noise and developed a method to estimate the phase velocity dispersion of surface waves contained in microtremors using a specially designed circular array. Okada et al. (1990) extended it to an exploration method that is called the SPAC method.

The SPAC method is based on the theory of stationary random functions according to which, microtremor is considered as stationary stochastic process both temporally and spatially. The SPAC method is an innovating method, which has so far only limited application in Japan. The main purpose of this thesis is to implement the SPAC method in an area to validate its reliability and to examine its possibilities for soil and site characterization in case of site effect studies.

The accuracy of the V_s profile, the ability to reach large penetration depth in densely populated urban areas and its low cost compared to conventional geophysical prospecting, make microtremor survey method (MSM) very attractive and useful for microzonation and site effect studies.

Between 10th and 20th of the September 2003, array observations of microtremors at 25 different sites in and around the city of Adapazarı were carried out. In this thesis, microtremors records that were obtained from Toyota car factory, one of the observation site, were analyzed to determine V_s velocity structure. A inversion technique was subsequently applied to determine V_s profile.

This thesis composed of five chapters. In chapter two, basic theoretical explanations related to surface wave, microtremors and array signal processing methods was given. In chapter three, geological, geotechnical and tectonic structures of observation site and city of Adapazarı was summarized. In chapter four, microtremor records were analyzed using by SPAC method and than V_s profile was obtained. Finally, discussion of the result was given in chapter five.

2. SURFACE WAVES, MICROTREMOR AND SURVEY METHODS

2.1. Physical properties of Rayleigh waves and dispersion

The earth is not an infinite body; it is a very large sphere with an outer surface on which stresses does not exist. For near surface engineering problems, the earth is often idealized as a semi infinite body with a planar free surface (the effects of the earth's curvature are neglected). The boundary conditions associated with the free surface allow solutions obtained from the equations of motion that describe waves whose motion is concentrated in a shallow zone near the free surface. These are surface waves (Figure 2.1).

Two types of surface waves are of primary importance in seismology and relevant fields. One, the Rayleigh waves are always generated when a free surface exist in a continuous body. The other surface wave, the Love waves, can exist only in presence of a soft superficial layer over a stiffer half space and they are produced by energy trapping in the softer layer for multiple reflections.

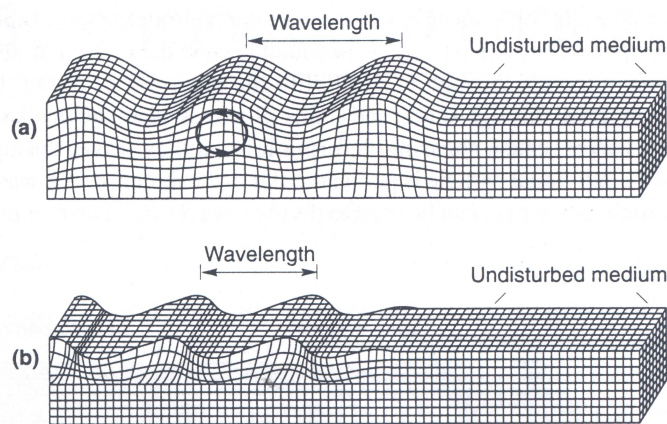


Figure 2.1. Deformation produced by surface waves: (a) Rayleigh wave; (b) Love wave (Bolt, 1993).

Since geotechnical earthquake engineering is concerned with the effects of earthquakes on humans and their environment, which are located on or very near the earth's surface, and since they attenuate with distance more slowly than body waves, surface waves are very important.

Waves that exist near the surface of homogeneous elastic half space were first investigated by the Rayleigh (1885) and are known to this date as Rayleigh waves. These waves in a homogeneous isotropic linear elastic half space are not dispersive, their velocity of propagation is a function of the mechanical properties of the medium, but it is not a function of frequency

In a homogenous elastic medium, Rayleigh wave is characterized by a unique value of wave velocity. In Figure 2.2, it is evident that the difference between shear wave velocity and Rayleigh wave velocity is very limited, being the latter slightly smaller than the former. In particular, the exact range of variation is given by:

$$0.87 < V_R / V_S < 0.96 \quad (2.1)$$

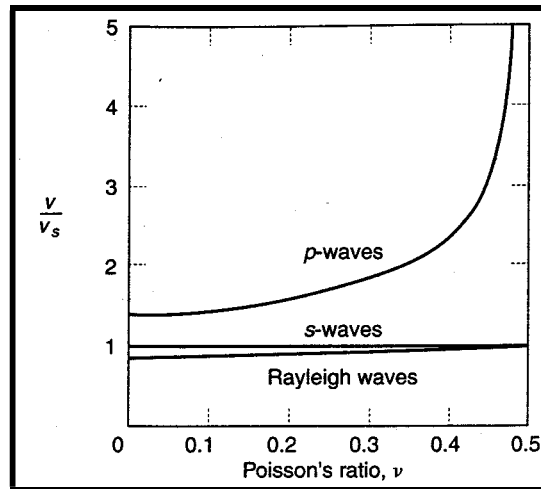


Figure 2.2. Relation between Poisson's ratio and velocity of propagation of compression (P), shear (S) and Rayleigh (R) waves in a linear elastic homogeneous half space (Kramer, 1996).

The horizontal and vertical components of Rayleigh waves are out of phase of exactly 90° one with the other, with the vertical component bigger in amplitude than the horizontal one, hence the resulting particle motion is an ellipse. On the ground surface the ellipse is retrograde (e.g. counterclockwise if the motion is propagating from left to right as shown in Figure 2.3), but going into depth the ellipse is reversed at a depth equal to about $1/2 \pi$ of the wavelength.

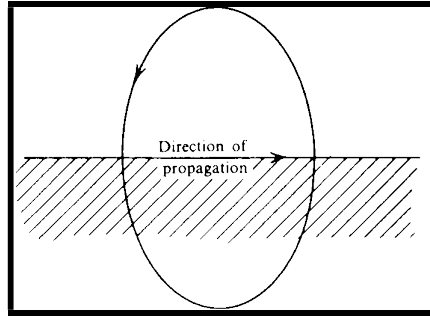


Figure 2.3. Particle motion on the surface during the passage of a Rayleigh waves in an elastic homogeneous halfspace (Foti, 2000).

Another important remark is that being the decrease with depth exponential, the particle motion amplitude becomes rapidly negligible with depth. For this reason it can be assessed that the wave propagation affects a confined superficial zone (see Figure 2.4), hence it is not influenced by mechanical characteristics of layers deeper than about a wavelength.

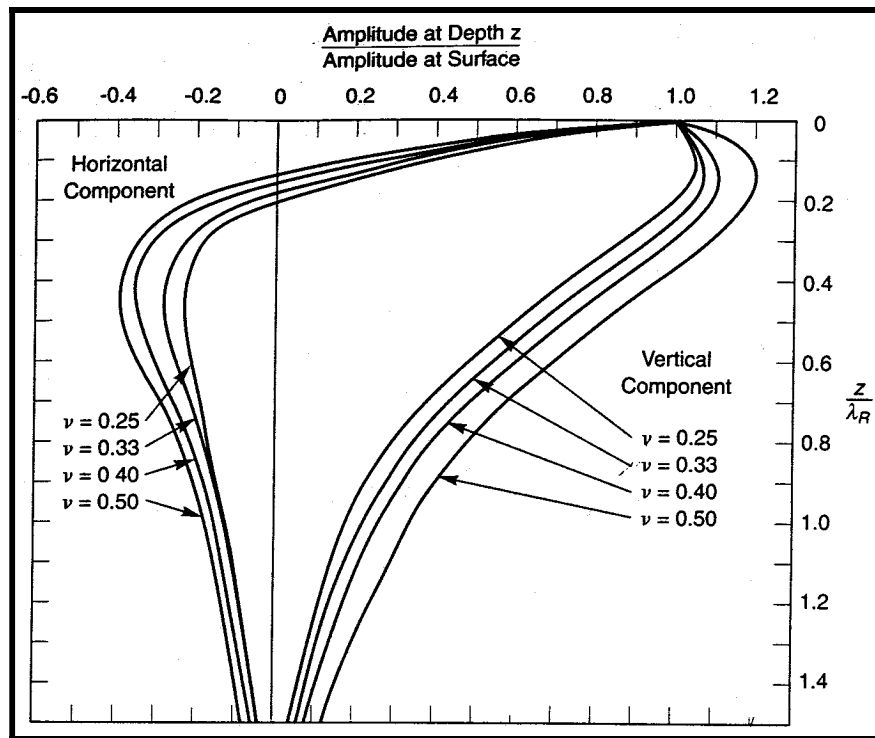


Figure 2.4. Amplitude ratio vs. dimensionless depth for Rayleigh wave in a homogeneous half-space (Richart et al., 1970).

In stratified media the phenomenon of geometrical dispersion arises, for which the phase velocity of Rayleigh wave is a function of frequency. The geometrical dispersion can easily be explained recalling the characteristics of shallowness of this wave. For a homogeneous linear elastic half space, the exponential decay of particle

motion with depth is such that the portion of the medium, that is affected by the wave propagation is equal to about one wavelength. Since the wavelength λ_R is related to the frequency f by the following relation;

$$\lambda_R = V_R / f \quad (2.2)$$

It is clear that low frequency waves will penetrate more into the ground surface. In the case of a vertically heterogeneous medium, Rayleigh waves at different frequency will involve their propagation in different layers and consequently the phase velocity will be related to a combination of their mechanical properties. Consequently the Rayleigh wave velocity will be a function of frequency. The above concept is summarized in Figure 2.5, where the vertical displacements wave field in depth at two different frequencies is presented for a layered medium.

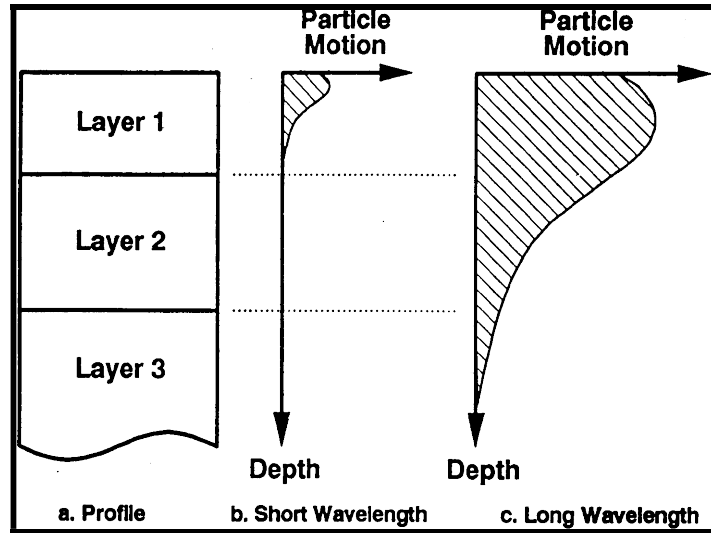


Figure 2.5. Dispersion in layered media (Rix, 1988).

It is important to remark that the shape of the dispersion curve (Rayleigh phase velocity vs. frequency or wavelength) is strongly related to the variation of stiffness with depth. Usually a distinction is made between a layered system for which the stiffness is monotonously increasing with depth and another one in which there is the presence of stiffer layers over softer ones. The first case is indicated as normally dispersive profile, the latter one as inversely dispersive profile. An example is presented in Figure 2.6 where the shape of the dispersion curve is presented in the phase velocity-wavelength plane. This representation is often used since it gives a clear picture of the variation of stiffness with depth. Obviously, in real media the alternation of stiff and soft layers can be much more complex if compared to the

above cases, still Figure 2.6 gives an idea of the relation existing between the stiffness profile and the dispersion curve.

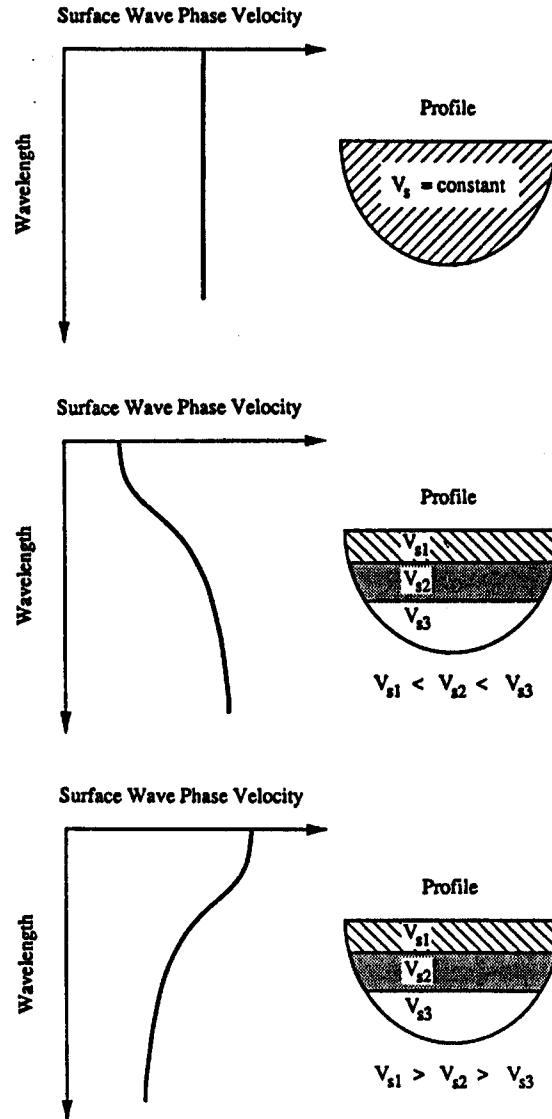


Figure 2.6. Examples of non-dispersive (homogeneous half-space), normally dispersive and inversely dispersive profiles (Rix, 1988).

An important consequence of dispersive Rayleigh wave behavior in layered media is the existence of two different kinds of wave velocity. The phase velocity that is the velocity of a wave front (locus of constant phase points), such as a peak or a trough. For a dispersive medium, this is not the same as the velocity of a pulse of energy, which was named as group velocity. Indeed, the latter can be seen (Fourier analysis) as composed of several single frequency signals, each one traveling with its own velocity because of dispersion. Figure 2.7 clarifies this concept. The velocity of the wave train, i.e. the velocity of the envelope is indicated as group velocity, in contrast

with that of the carrier that is the phase velocity. Obviously, for a non-dispersive medium group velocity and phase velocity coincide.

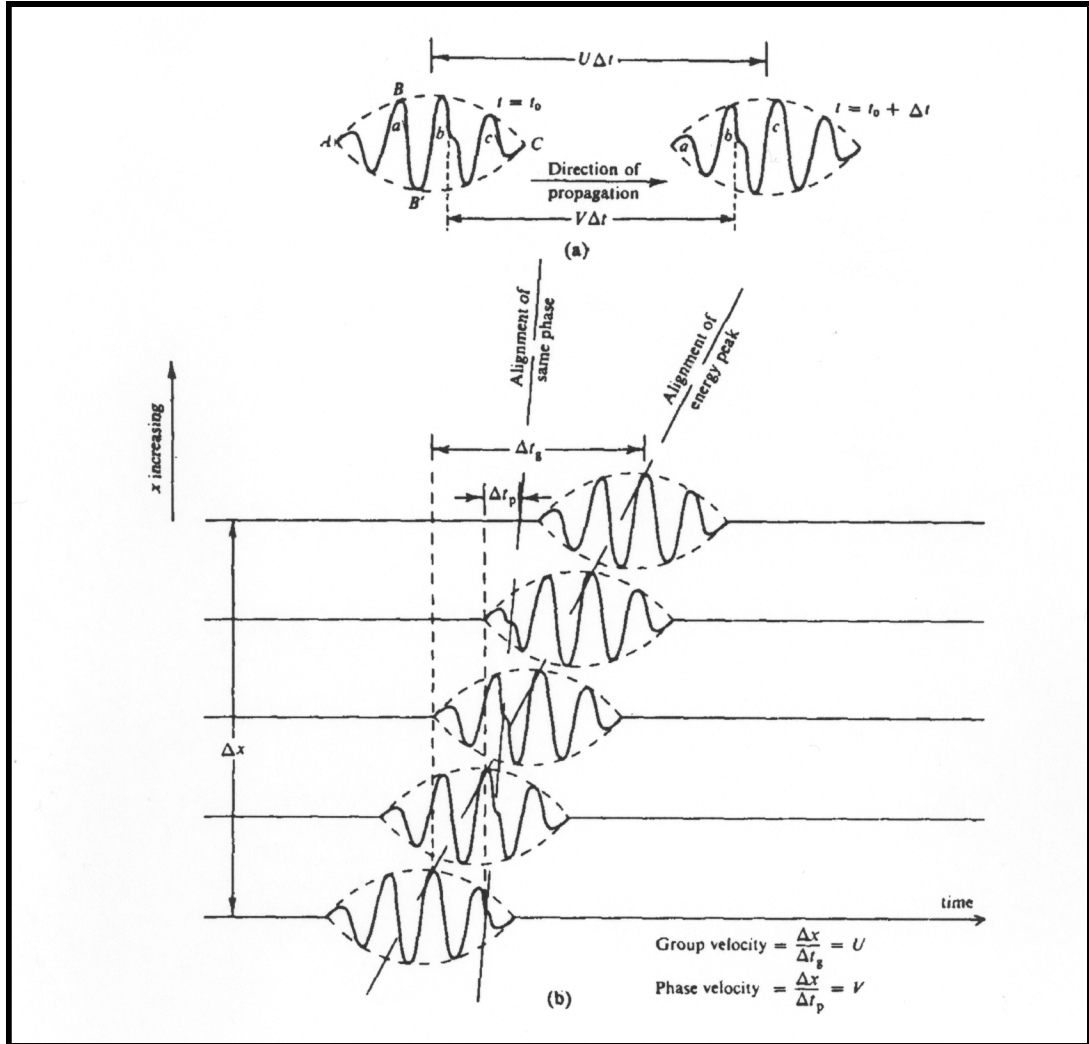


Figure 2.7. (a) Group and phase velocities. (b) Arrival of a dispersive wave at different geophones (Sheriff and Geldart, 1995).

2.2. Properties of Microtremors

There are usually various vibrations, at small amplitude level of about several micrometers, which appear on surrounding ground surface. The constant vibrations of earth's surface are called microseism or microtremors. The term "microtremor(s)" is more commonly used in the field of earthquake engineering.

The range of vibration period of such microtremors is from 1/10 of a second to 10 seconds. The amplitude of these microtremors is, with some extreme exceptions, generally very small. Displacements are in the order of 10^{-4} to 10^{-2} mm far below human sensing. Although they are very weak, they represent a source of noise to

researchers of the earthquake seismology; if the amplifier gain is increased in order to record earthquake signals from a distant source, the amplitude of microtremors proportionally increases and the desired earthquake signal is buried in the “noise” of microtremors. Elimination of this background noise is technically extremely difficult or impossible to achieve. Therefore, the earthquake researchers call microtremors “seismic noise” or simply “noise”.

It was not until the late nineteenth century that, seismologists could employ seismometers to observe the movements of the earth’s surface. Since then, microtremors have been a focus of their strong interest, as is evidenced by the large number of research papers published on the subject. Much of the research concerns the source of the vibration and variation of the character of the vibration depending on time and location. From this research, we now know that the microtremors are caused by daily human activities such as movement of machinery in factories, motor cars and people walking; and natural phenomena such as flow of water in rivers, rain, wind, variation of atmospheric pressure, and ocean waves. Thus, the source of microtremors constitutes not only natural phenomenon but also human activities. However, microtremors are now not regarded as “noise” but rather a useful “signal”. In this sense, they are sometimes referred to as “uncontrolled signal”.

Due to close relation between the nature of microtremors and the fundamental dynamic behavior of the surface soil layer, these small vibrations are well-known and useful in the field of earthquake engineering. Thus, in the nature of this ground vibrations, an important relation between surface ground structures and vibrations has been suggested. Kanai has originally introduced a theoretical interpretation and practical engineering application of microtremors, especially as convenient tool for evaluating frequency properties of surface ground (Okada, 2003).

Both human activity and natural phenomena (such as climate and oceanic conditions) vary with time. Accordingly, microtremor activity varies over time. This variation is very complex and irregular, and not repeatable.

When microtremors are observed simultaneously at several spatially separated stations, it is noted that these tremors are not completely random and that some coherent waves are contained in the records. In other words, microtremors are an assemblage of waves traveling in various directions. In fact, Toksöz and Lacoss

(1968) clearly demonstrated from the data of large-aperture seismic array (LASA) that microtremors are an assemblage of body waves and surface waves.

The microtremors originating from human activities are dominated by the components with periods shorter than one second, or higher than 1 Hz in frequency (Kulhanek, 1990 for example), and have clear diurnal variation in both amplitude and period.

On the other hand, due to natural phenomena such as climatic and oceanic condition the microtremors have dominant periods greater than one second (frequency lower than 1 Hz), with associated amplitude and period variations corresponding to the vagaries of the respective natural phenomenon.

A detailed analysis reveals that microtremors vary depending upon location. The microtremor survey method has been devised to focus on this variation.

In 1960s, explanations of power spectra were proposed to the effect that the high power level of the low frequency components (below 1.0 Hz) and the two characteristic spectral peaks originated from the ocean, while the higher frequency components (above 1.0 Hz) were attributed to human activity and climatic conditions.

One of the origins of microtremors as being natural phenomena such as climatic and marine conditions. Can microtremors be observed in the middle of a continent? This is a natural question when one considers the limits to application of the microtremor method.

The answer is given in Figure 2.9 (Peterson, 1993). The dataset was collected at 75 permanent seismic stations all over the world (see Figure 2.8) at a time of no seismic event due to earthquake or nuclear explosion. These power spectra are calculated for the vertical component of background vibration of the earth including microtremors, wave motions and even earth tides. They may contain power components whose source is not known to us, hence not yet investigated, perhaps including some originating in the interior of the earth. Regardless of their sources, these spectra show that the energy of the earth's background seismic vibration exists in abundance at all locations on the earth's surface, and that with an appropriate observation instrument one can record the microtremor signals.

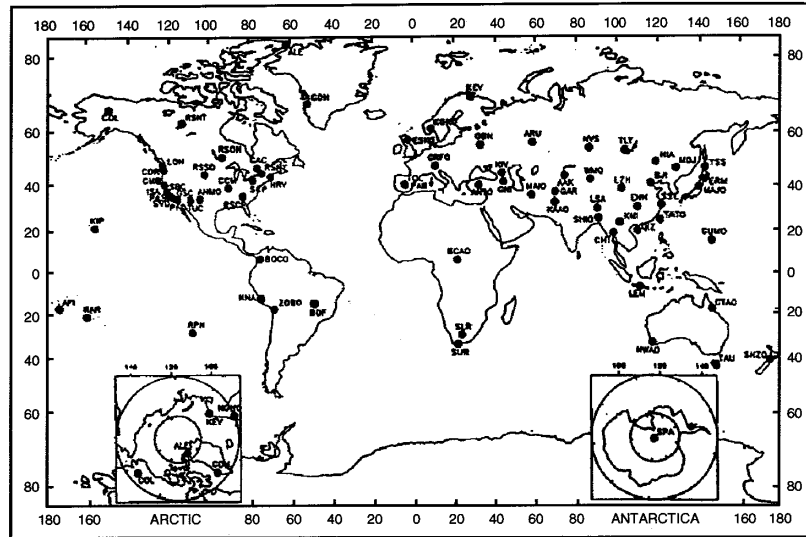


Figure 2.8. Seventy-five permanent seismic observatories in the world, which record common vibration of the earth including microtremors (Peterson, 1993).

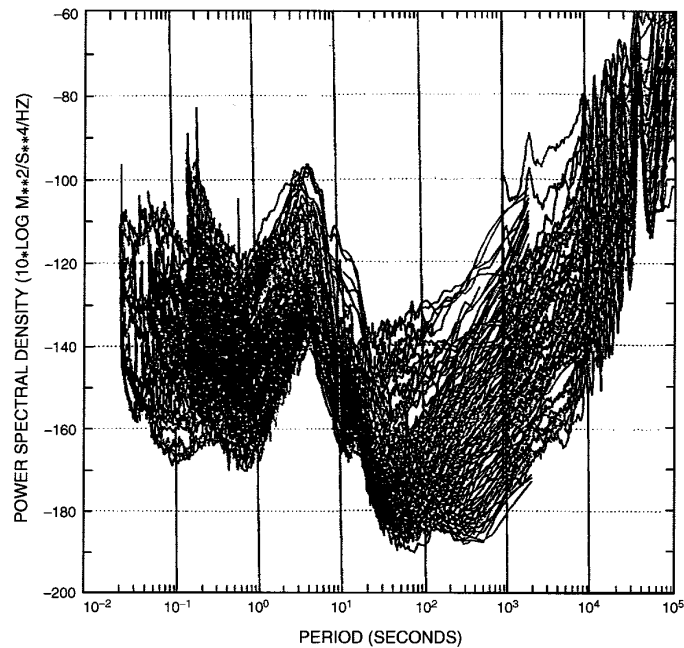


Figure 2.9. The acceleration power spectra of common vibration of the earth recorded at the seventy-five seismic observatories in the world (Peterson, 1993).

2.3. Temporal and Spatial Variation of Microtremors

Microtremors are a phenomenon which varies both spatially and temporally. The microtremor survey method assumes both spatial and temporal stationarity of microtremors, this variation is integral to the microtremor survey method (MSM). The spectrum of the signal determines the lower limit of the depth that can be explored by MSM. When information on deeper structure is desired, data should be acquired at a time when the lower frequency component is dominant.

A sample of 120 blocks (20 hours) of vertical and east-west components of microtremor data were taken from the record collected at two different stations, Hokkaido (HKD) and Misumai (MIS), in the fifteen-day period from 19th November to 3rd December 1997 (Okada, 2003). Power spectra were calculated for each ten-minute block. Figure 2.10 shows the power spectra of the microtremor records for each time block over-plotted on one axis (Okada, 2003). In this explanation “day-time” refers to 9, 12, 15 and 18 hrs, while “night” refers to 21, 0, 3 and 6 hrs. Figure 2.11 compares the average spectra of day and night. The data shown here are collected at two stations around Sapporo, the capital city of Hokkaido, in different times:

1. A station situated in the campus of Hokkaido University (HKD), amid the metropolitan area of Sapporo on a sedimentary basin; collected during the one month period from 4th November to 4th December 1997; similarly sampled; the seismometer was set on a concrete block in a pre-fabricated storage building.
2. A station set up in Toyama, nicknamed Misumai (MIS), in a suburb of Sapporo during the same period as HKD; directly on the basement rock; similarly sampled; the seismometer placed on a concrete floor in an RC building.

The type of seismometer used is compact long-period, PELS Model 73 with a natural period of 8 seconds (Project Team for the Development of the Small-Size Long-Period Seismometer; Matsumoto and Takahashi, 1977). The recorders used at HKD and MIS were of type DATAMARK LS-8000SH by Hakusan Industries, and one long-duration recorder designed by project team.

From Figures 2.10 and 2.11, it is clear that the overall shape of the spectra does not vary in the fifteen-day period, although the power level of the record at the urban HKD station is generally higher than that of suburban MIS. Closer observation reveals temporal variation differences of the power spectra, between these frequencies lower than 1Hz and those frequencies higher than 1Hz. For the frequencies under 1Hz (or periods longer than 1 second), the difference in power between day-time and night is minimal. On the other hand, that difference is significant above 1Hz (period shorter than 1 second) by an order of 1 to 3. It is more apparent in the urban HKD station. It is concluded that the frequency components under 1 Hz are little affected by human activities, while frequency components over 1Hz are significantly affected.

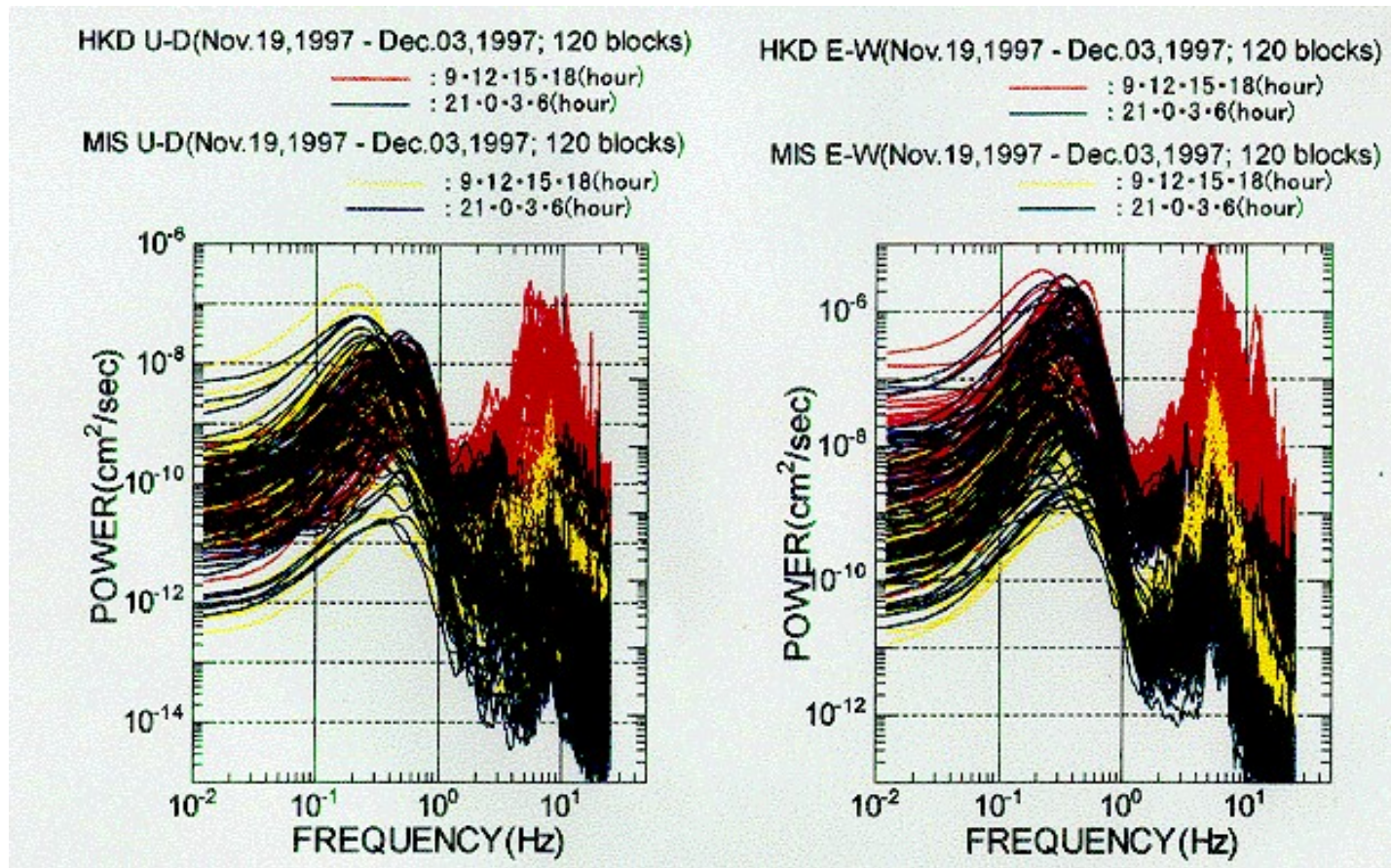


Figure 2.10. Power spectra of the vertical (U-D) and east west (E-W) components of microtremors at HKD and MIS observed in the 15 day period between 19th Nov. and 03 Dec., 1997 the spectra are calculated and overplotted for 120 10-minutes blocks (Okada, 2003).

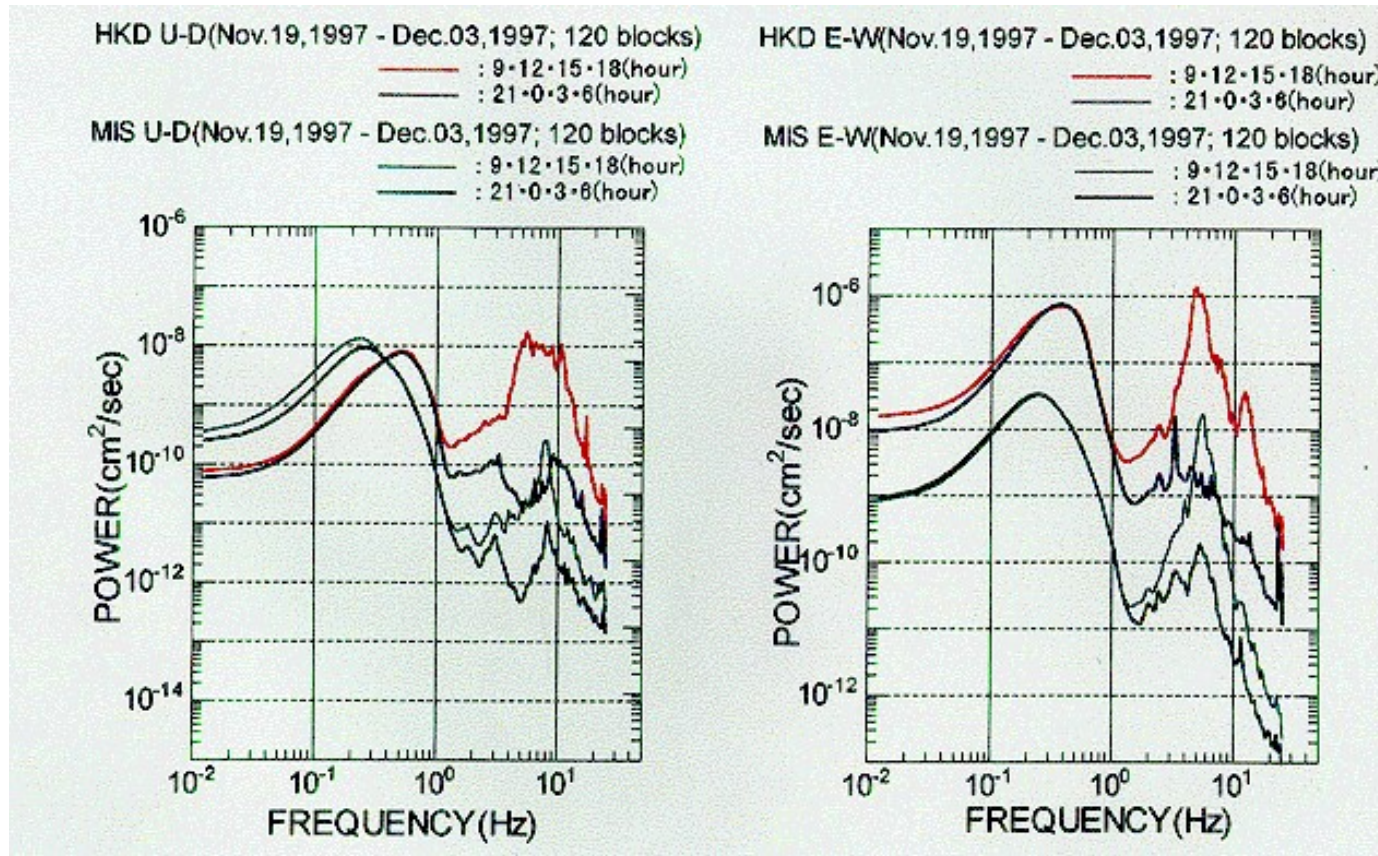


Figure 2.11. Day-time and night-time averages of power spectra of the vertical (U-D) and east-west (E-W) components of microtremors at HKD and MIS, observed in the 15-day period between 19th Nov. and 03 Dec., 1997 (Okada, 2003).

Figure 2.12 is the running power spectra of the east west (E-W) component of microtremors obtained from the observations at HKD and MIS. These data were sampled for 10 minutes simultaneously, every three hours from midnight. The data at HKD are from the 29-day period between 5th November and 3rd December 1997, and those of MIS are from the 15-day period from 19th November to 3rd December 1997. The curve above is the atmospheric pressure data at Sapporo Meteorological Observatory (SMO) for the corresponding period. It shows variation between 980 and 1040 hPa. As seen in the graph, the character of spectra of microtremors differ between the ranges higher and lower than 1Hz.

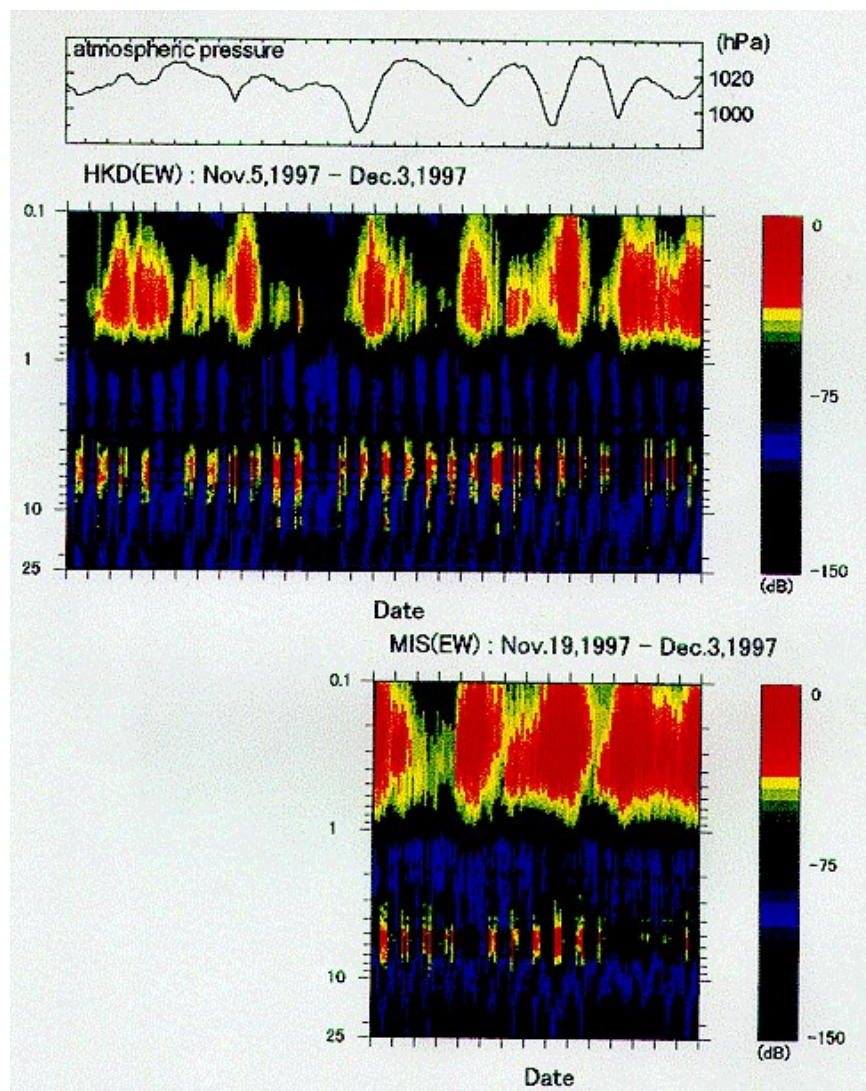


Figure 2.12. Running power spectra of the E-W component of microtremors observed over the 29-day period between 5th Nov. and Dec 3 (HKD) and the 15-day period between 19 Nov and 3 Dec, 1997 (MIS). Atmospheric pressure at SMO for this period is shown on the top (Okada, 2003).

The temporal variation of microtremors has a good inverse relationship with the variation of atmospheric pressure: power of microtremors increases with lower atmospheric pressure and it presents a maximum soon after the trough in atmospheric pressure. On the other hand, the power decreases when atmospheric pressure increases. This correlation was almost without exception for the two stations 14.8 km apart, for the duration of the recordings. In quantitative comparison of the three-component data of HKD and MIS with atmospheric pressure, Okada (2003) calculated cross-correlation coefficients, and found the maxima of power of the low-frequency (long period) component of microtremors lag the minima of atmospheric pressure by 3 to 15 hours.

As mentioned, the power spectra of higher-frequency microtremors are affected by human activity. This is clear in Figure 2.12. The power spectra of the frequencies between 4 and 7 Hz clearly have diurnal variation. More precise observation may reveal that the power of microtremors diminishes relatively around lunchtime and also on Sundays and public holidays.

Another important point is spatial variation of the microtremors. By examining the power spectra in Figure 2.10, one can observe that the power spectra of both HKD and MIS are affected by atmospheric pressure. However the peak level and its frequency differs between the stations. It varies by time and the pattern is not consistent. This is seen in the average of the day-time and night-time power spectra of the shown in Figure 2.11.

The peak frequencies of HKD are about 0.5 Hz for the U-D component and 0.35 Hz for E-W component, while MIS has a peak at approximately the same frequency of 0.25 Hz for both the U-D and E-W components. The peak level of the U-D component is similar between stations, but the peak level of the E-W component at HKD is about 30 times higher than at MIS. In other words, low-frequency (long-period) E-W component was more easily induced at HKD than MIS.

This demonstrates that while the variation of atmospheric pressure is considered to be a common phenomenon over a large area, including HKD and MIS and over a relatively long time, it causes different microtremor spectrum structures at different places; that is output responses to the same input differ with location.

The variation of atmospheric pressure is known to affect the low-frequency (long period) component of microtremors over a relatively large area and long time. However, for these high-frequency (short period) components, it is not clear in the power spectra of Figure 2.10 whether the same input causes a different response depending on the location. The extent of “spatial stationarity”, which is important for the microtremor survey method, is not obvious.

The spectra in Figure 2.13 are of microtremors simultaneously recorded at ten stations scattered within a 1.5 km radius of suburban Obihiro (see map in Figure 2.13). At each station, six discrete forty-five minute recordings were made, two hours apart, from 16 00hr on 21st July 1985 to 02 00hr on the following day (Okada, 2003).

The temporal variation in the records is considered to be influenced by the temporal variation of the source of microtremors. On the other hand, the cause of the characteristic spatial differences observed between the numerous locations is attributed to the fact that the microtremors are not satisfying spatial stationarity at these points. However, very little difference is observed in the longer-period spectral components (under 1 Hz) at all of the locations. This leads to the belief that spatial stationarity of the microtremor energy is satisfied within this frequency band.

This example shows that the spectral structure of microtremors is largely stationary over the spatial extent of a 1.5 km radius and a temporal extent of 45 minutes.

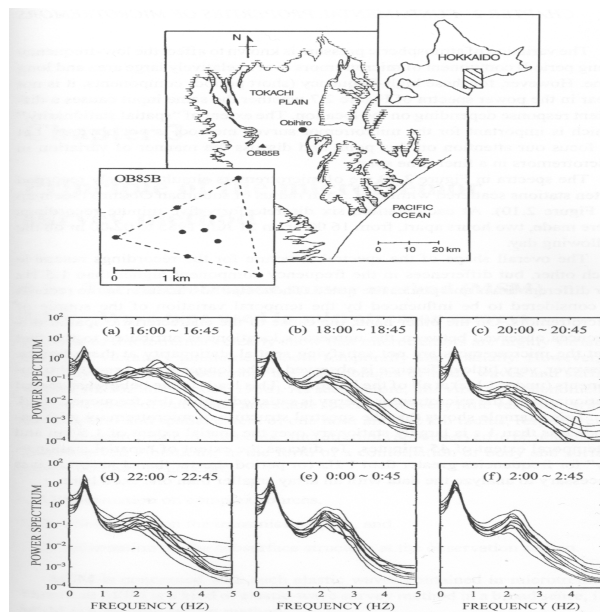


Figure 2.13. Power spectra of 10 stations within the close proximity for 45-minute records at 6 different times (Okada, 2003).

2.4. Microtremor Survey Method (MSM)

Records of microtremors clearly show that microtremors are highly variable, irregular, vibratory phenomenon both temporally and spatially. However, in elasticity theory, microtremors are assemblages of body waves and surface waves (Toksöz and Lacoss, 1968). As seen before, this vibratory phenomenon is constituted by a set of stationary and stable spectra with very little variation, within the temporal and spatial extent of one hour and 1-2 km radius.

With appropriate instruments, microtremors are ubiquitously observed. The observed vibration, in the form of a combination of body and surface waves, generally contain:

- Information on complex sources;
- Information on the transmission path; and
- Information on the sub-surface structure at the observation station.

The Microtremor Survey Method (MSM) is concerned with such elastic waves contained in microtremors. Therefore the MSM is a kind of elastic wave survey method in a broad sense, i.e. MSM is a seismic survey method.

However, unlike the conventional reflection and refraction seismic methods, which use artificial source, and treats waves as controlled in the phase domain, the MSM utilizes uncontrolled natural phenomenon as the source, and treats the various aspects of microtremor spectra (temporal and spatial properties) according to the theory of stochastic process.

Many of the sources of microtremors can be identified as acting on the Earth's surface or the sea floor. Therefore surface waves are naturally considered to be the dominant component of the microtremors, over body waves. The MSM utilizes this dominant surface wave mode of propagation.

As is well known as dispersion, the velocity of surface waves varies depending on the frequency (or period). Since dispersion is a function of sub-surface structure, the sub-surface structure can, in theory, be estimated from the dispersion. The MSM is basically a method to estimate this dispersion of surface waves contained within microtremors.

At present, our level of theoretical knowledge can only solve the characteristics of dispersion of surface wave for parallel, isotropic and homogeneous layers. Therefore the sub-surface structures estimated by the MSM are approximated by parallel, isotropic and homogeneous layers. Consequently, precision of the estimated structure obtained from the MSM is lower than that from conventional seismic methods. This is the reason why the MSM is regarded as a reconnaissance method. However, the MSM yields the physical properties of the sub-surface structure by exploiting the S-wave velocity, which is difficult to measure with conventional engineering-scale seismic surveys. Analysis of the response to earthquake movements identifies the S-wave velocity as more important than P-wave velocity in determining sub-surface structure. In this regard, the MSM is suitable for the estimation of sub-surface structure, such as required by the field of earthquake engineering.

The MSM deals with surface waves and it applies surface-wave theory to detect useable signals. There are some problems remaining to be solved in the MSM, for example:

- Standardizing the observation system including field procedure and instrumentation.
- The inversion problem; from phase velocity dispersion curves, to velocity structure.

However the basic scheme of the MSM for estimating sub-surface structure has been more or less completed.

It consists of three steps:

- Observation by seismometer network (array) arranged on the ground surface;
- Estimation of dispersion of surface wave as a response to the sub-surface structure directly below the array; and
- Estimation of the sub-surface structure causing the dispersion by means of inversion.

2.4.1. Basic Concepts of SPAC (Spatial Autocorrelation Method)

Microtremors contain surface waves in abundance. These surface waves are generated randomly, both in a temporal and spatial sense, by a variety of

mechanisms, and travel through a wide-range of geological conditions. Naturally, microtremors become very complex assemblage of elastic waves containing not only body waves and surface waves but also scattered and diffracted waves.

To separate surface waves from this assemblage, researchers once tried various methods. For example, when microtremors were recorded by analog instruments, they were recorded at three points forming a tripartite and wavelets resembling each other were identified. By reading the wavelets, it was possible to derive apparent periods and velocities more quantitatively, a band-pass filter was applied to microtremors recorded by tripartite to derive apparent periods by digital processing, the autocorrelation of similar wavelets in tripartite records were used to derive periods and phase velocities quantitatively.

Since the development of the digital recording of microtremors, the analysis method has advanced to understanding microtremors as a kind of stochastic process. At present, one of the important methods used to detect surface waves are known as SPAC (Spatial Autocorrelation) method. (Okada, 2003).

The SPAC method is generally an application of the circular array observation and a data analysis method used for understanding the transmission properties of a variety of waves based on the theory of stochastic process developed by Aki in 1957. He attempted to estimate the subsurface structure from microtremor records assuming the microtremors are isotropic waves coming from all directions.

The significance of the Aki's theory is its treatment of the complex wave phenomenon as a stochastic process in temporal and spatial dimensions. This complex wave phenomenon, of course, includes the microtremors. The idea forms an important foundation when considering a method of drawing necessary information from the complex "noise".

The basic principles of the SPAC method are:

- Assume the complex wave motion of microtremors to be a stochastic process in time and space;
- A spatial autocorrelation coefficient for microtremor data, observed by circular array, can be defined when the waves composing the microtremors are dispersive like surface waves;

- The spatial autocorrelation coefficient is a function of phase velocity and frequency.

Aki (1957), as an example of the application of his theory, tried to estimate the subsurface structure from a record of short period (<1 second) microtremors. At his time, the digital data recording system was not available and the result of his experiment cannot be regarded as “excellent”. However, the idea of understanding the ambient noise as a signal and suggesting a new direction in geophysical prospecting for meaningful subsurface structure, using that signal, is highly respected (Okada, 2003).

Hidaka, Okada, Ferrazzini, Hough, Malagnini, Ling and Matsuoka were published later applications of the SPAC method and they offer many theoretical and practical results (Okada, 2003).

There are not many reports on the application of the spatial autocorrelation method (SPAC). However, it has two advantages over the other array signal processing method:

It requires fewer stations and a smaller array than the other array signal processing method to achieve a similar result. The size of the array in the microtremor observation is very important, because:

- A large array increases the field effort and decreases field efficiency;
- A large array may affect the assumption of the microtremor method that the layers are sub-parallel under the array (Okada, 2003).

2.4.1.1. Spectral Representation of Microtremors in Polar Coordinate System

In analyzing microtremor data by the spatial autocorrelation method, it is convenient to use the polar coordinate system for the microtremor spectra. By using polar coordinate relations:

$$\xi = r(\cos \theta, \sin \theta) \quad \text{and} \quad \mathbf{k} = k(\cos \phi, \sin \phi), \quad (2.3)$$

Sample function for recorded microtremor is written as:

$$X(t, r, \theta) = \int_{-\infty}^{\infty} \int_0^{\infty} \int_0^{2\pi} \exp\{i\omega t + irk \cos(\theta - \phi)\} d\zeta(\omega, k, \phi) \quad (2.4)$$

where:

$$\begin{aligned} d\zeta(\omega, \mathbf{k}, \phi) &= kd\zeta'(\omega, k, \phi) \\ &= dZ'(\omega, k) \text{ in (2.3)} \end{aligned} \quad (2.5)$$

From (2.4) the microtremors as a stationary stochastic process can be expressed as a continuous sum of the waves of various angular frequency ω and wavenumber k independently (*i.e.* without correlation) arriving from various directions ϕ .

The $\zeta(\omega, k, \phi)$ of (2.4) satisfies the following relationships:

$$E[d\zeta(\omega, k, \phi)] = 0 \text{ (for all } \omega, k, \phi) \quad (2.6)$$

$$E\left[\left| d\zeta(\omega, k, \phi) \right|^2 \right] = dH(\omega, k, \phi) \quad \text{(for all } \omega, k, \phi) \quad (2.7)$$

for any two distinct ω , and ω' (where $\omega \neq \omega'$), and two distinct sets (k, ϕ) and (k', ϕ') (where $k \neq k'$ and $\phi \neq \phi'$),

$$E[d\zeta^*(\omega, k, \phi)d\zeta(\omega', k', \phi')] = 0 \quad (2.8)$$

where * denotes the complex conjugate.

Now we make two assumptions:

Assumption 1: The microtremors consist mainly of surface waves and one of their modes (often the fundamental mode) is dominant. Hence,

Assumption 2: ω and k are related as a function of each other, and Z or ζ , namely the microtremors expressed as a stochastic process, have significance only on a curve $[\omega, k(\omega)]$ (Okada, 2003).

When we discuss the vertical component of the microtremors, the surface wave referred to is the Rayleigh wave. In this case, the spectral representation of (2.4) becomes:

$$X(t, r, \theta) = \int_{-\infty}^{\infty} \int_0^{2\pi} \exp\{i\omega t + irk(\omega)\cos(\theta - \phi)\} d\zeta(\omega, \phi) \quad (2.9)$$

In general, as the spectra of the microtremors are considered to be continuous and differentiable with respect to frequency and direction, the stochastic process $\zeta(\omega, \phi)$ satisfies the following relationship:

$$\begin{aligned} E\left[\left|d\zeta(\omega, \phi)\right|^2\right] &= dH(\omega, \phi) \\ &= h(\omega, \phi)d\omega d\phi \end{aligned} \quad (2.10)$$

where $h(\omega, \phi)$ may be called “frequency-direction spectral density”, and $h(\omega, \phi)d\omega d\phi$ represents the average contribution to the total power from the components of the waves coming from the directions between ϕ and $\phi+d\phi$ with angular frequencies between ω and $\omega+d\omega$.

When this average contribution is summed up (*i.e.* integrated), in the respect of all the directions, the power spectral density function (or simply “power spectrum”) of microtremors at one station $h_0(\omega)$ is obtained as: (Kudo et al., 2002)

$$h_0(\omega) = \int_0^{2\pi} h(\omega, \phi) d\phi \quad (2.11)$$

2.4.1.2. Spatial Autocorrelation Function and Spatial Covariance Function

Suppose there are two microtremor observation stations A and B, the distance between which is r . Let the A be the origin of the coordinate system (0,0), then the coordinates of station B are (r, θ) .

From (2.9), the microtremor record at station A can be represented as:

$$X(t, 0, 0) = \int_{-\infty}^{\infty} \int_0^{2\pi} \exp(i\omega t) d\zeta(\omega, \phi) \quad (2.12)$$

and the record at B as:

$$X(t, r, \theta) = \int_{-\infty}^{\infty} \int_0^{2\pi} \exp\{i\omega t + irk \cos(\theta - \phi)\} d\zeta(\omega, \phi) \quad (2.13)$$

Defining the spatial autocorrelation function between A and B as:

$$S(r, \theta) = E[X^*(t, 0, 0)X(t, r, \theta)]$$

$$\begin{aligned}
&= \lim_{T \rightarrow \infty} \frac{1}{2T} \int_{-T}^T X^*(t, 0, 0) X(t, r, \theta) dt \\
&= \int_{-\infty}^{\infty} \int_0^{2\pi} \int_{-\infty}^{\infty} \int_0^{2\pi} \exp\{i(\omega' - \omega)t + irk' \cos(\theta - \phi')\} \\
&\quad \cdot E[d\zeta^*(\omega, \phi) d\zeta(\omega', \phi')] \cdot
\end{aligned} \tag{2.14}$$

This equation is reduced by (2.7), (2.8), (2.9) and (2.10) to:

$$S(r, \theta) = \int_{-\infty}^{\infty} \left[\int_0^{2\pi} \exp\{irk \cos(\theta - \phi)\} h(\omega, \phi) d\phi \right] d\omega \tag{2.15}$$

$$= \int_{-\infty}^{\infty} g(\omega, r, \theta) d\omega \tag{2.16}$$

where,

$$g(\omega, r, \theta) = \int_0^{2\pi} \exp\{irk \cos(\theta - \phi)\} h(\omega, \phi) d\phi \tag{2.17}$$

is called the spatial covariance function of the microtremors at the angular frequency ω (Henstridge, 1979).

This equation at the origin (0,0) is evaluated as:

$$\begin{aligned}
g(\omega, 0, 0) &= \int_0^{2\pi} h(\omega, \phi) d\phi \\
&= h_0(\omega)
\end{aligned} \tag{2.18}$$

and gives the power spectrum of (2.11). In the same way, the spatial auto correlation function at the origin is:

$$\begin{aligned}
S_0 &\equiv S(0, 0) = E[|X(t, 0, 0)|^2] \\
&= \int_{-\infty}^{\infty} h_0(\omega) d\omega
\end{aligned} \tag{2.19}$$

Here, $h_0(\omega)d\omega$ is the average contribution to the total power from the components of microtremors $X(t, r, \theta)$ with angular frequency between ω and $\omega+d\omega$ observed at one station (A or B) within the space where the array is spread. Therefore, S_0 in the left hand side of the equation (2.19) gives the total power of the stochastic process (*i.e.* microtremor record) at one station within the array space.

2.4.1.3. Spatial Autocorrelation Coefficient of Circular Array and Phase Velocity

Supposed that an array consisting of several stations on a circle with a radius r around the origin A, was laid. Spatial averaged autocorrelation function and spatial autocorrelation coefficient for the microtremors observed by the circular array can be defined. For simplicity, consideration is given to one component with a particular frequency ω .

Now, let $g(\omega, r, \theta)$ be the spatial covariance function between the center and one point on the circumference of the circular array at the frequency ω . We can then define the *directional average* of the spatial covariance function by averaging $g(\omega, r, \theta)$ over all the directions:

$$\bar{g}(\omega, r) = \frac{1}{2\pi} \int_0^{2\pi} g(\omega, r, \theta) d\theta \quad (2.20)$$

By substituting the integrand with (2.17):

$$\bar{g}(\omega, r) = \frac{1}{2\pi} \int_0^{2\pi} \int_0^{2\pi} \exp\{irk \cos(\theta - \phi)\} h(\omega, \phi) d\phi d\theta \quad (2.21)$$

In (2.21), as the integral along θ is:

$$\frac{1}{2\pi} \int_0^{2\pi} \exp\{irk \cos(\theta - \phi)\} d\theta = J_0(rk) \quad (2.22)$$

which is the Bessel function of the first kind of zero order with the variable rk .

Therefore (2.21) is simply:

$$\bar{g}(\omega, r) = \int_0^{2\pi} J_0(rk) h(\omega, \phi) d\phi$$

$$= J_0(rk) \int_0^{2\pi} h(\omega, \phi) d\phi \quad (2.23)$$

By (2.11) or (2.18), this equation is further simplified to:

$$\bar{g}(\omega, r) = h_0(\omega) J_0(rk) \quad (2.24)$$

or :

$$\bar{g}(\omega, r) = g(\omega, 0, 0) J_0(rk) . \quad (2.25)$$

Similarly, the directional average of the spatial autocorrelation function defined by (2.14) is reduced to:

$$\bar{S}(r) = \int_{-\infty}^{\infty} h_0(\omega) J_0(rk) d\omega . \quad (2.26)$$

The terms to be integrated in the equation (2.24) or (2.26) means that for the total power of the microtremors within the array space, the average contribution of the components of the waves between ω and $\omega+d\omega$ is always weighted by the coefficient $J_0(rk)$, which is related to the sub-surface structure. Therefore, $\bar{S}(r)$, the left-hand side of (2.26), gives the total power of all the microtremors propagating under the influence of the sub-surface structure directly beneath the circular array of radius r .

Now we define the “spatial autocorrelation coefficient at the angular frequency ω ”, $\rho(\omega, r)$, or simply “spatial autocorrelation coefficient”, $\rho(f, r)$, as the power spectra of microtremors at one station within the array space (the center of the circle, *i.e.* origin, for instance) normalized to $h_0(\omega)$:

$$\rho(\omega, r) \equiv \bar{g}(\omega, r) / h_0(\omega) . \quad (2.27)$$

By (2.24):

$$\rho(\omega, r) = J_0(rk) , \quad (2.28)$$

or from $k=\omega/c(\omega)$, (where $c(\omega)$ is the phase velocity):

$$\rho(\omega, r) = J_0(r\omega / c(\omega)) \quad (2.29)$$

and from $\omega=2\pi f$:

$$\rho(f, r) = J_0(2\pi f r / c(f)). \quad (2.30)$$

Therefore the spatial autocorrelation coefficient at the frequency f is related to the phase velocity $c(f)$ via the Bessel function of the first kind of zero order. Figure 2.14 is a schematic diagram of the spatial autocorrelation coefficient controlled by the two variables f and r (Kudo et al., 2002).

$$\rho(f, r) = J_0(2\pi f r / c(f))$$

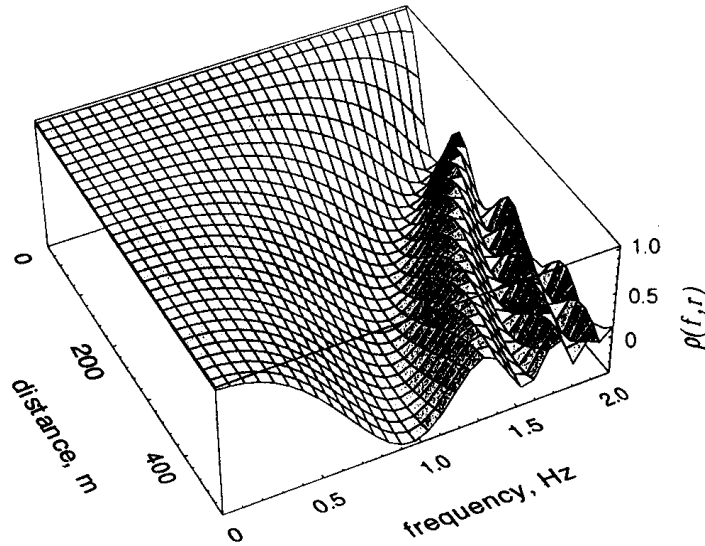


Figure 2.14. A schematic graph of spatial autocorrelation coefficient (Okada, 2003).

As clear in the above theoretical derivation, the phase velocity of a certain frequency can be calculated by the spatial autocorrelation coefficient of the component of the wave whose frequency is ω , from the microtremors recorded with a circular array of radius r .

The spatial autocorrelation coefficient defined above is a unique quantity to the location of the array, and it reflects the subsurface structure directly below the array.

2.4.2. Other microtremor survey methods

As stated previously, the method to detect surface waves and to produce dispersion curve from the microtremor record is critical to the microtremor survey method. To date, two methods have been developed: the frequency-wavenumber method and spatial autocorrelation method (Okada, 2003).

The properties common to both methods are: the microtremors are regarded as a stochastic process and their spectra form the basis of analysis, and both methods observe the vertical component of microtremors to extract the Rayleigh wave, a type of surface wave.

Of the methods fundamentally based on these common grounds, there seem to be more case histories reported using frequency-wavenumber method. Lots of earth scientist studied related to frequency-wavenumber method including Asten, Henstridge, Horike, Matsushima, Ohshima and Okada, Tokimatsu *etc.* (Okada, 2003).

The frequency-wavenumber method acquires microtremor data using an array whose size is appropriate to the target depth, then calculates the frequency-wavenumber power spectral density function (f-k spectra). The surface wave contained in the microtremors is detected as a function of phase velocity and frequency (or period). In figure 2.15, data processing flow chart for the frequency-wavenumber method was represented. The method detecting surface waves by the parameter of f-k spectra is, called the f-k method. The f-k method owes its origin to the research for LASA by LaCoss et al. (1969) and Capon (1969).

More flexibility is allowed in array design for the f-k method than for the SPAC method. The ideal array spread should have a large lateral extent with varied distance between seismometers. In general, topography and other constraints on seismometer location dictate the shape and size of the array.

A larger the number of observation stations provides greater precision in the f-k spectra. For example the gigantic LASA in America deploys 525 seismometers in the array with a diameter of about 200 kilometres. However, considering restrictions in station setting and analysis time, around ten seismometers may be a practical number for the purpose of exploring sub-surface structure.

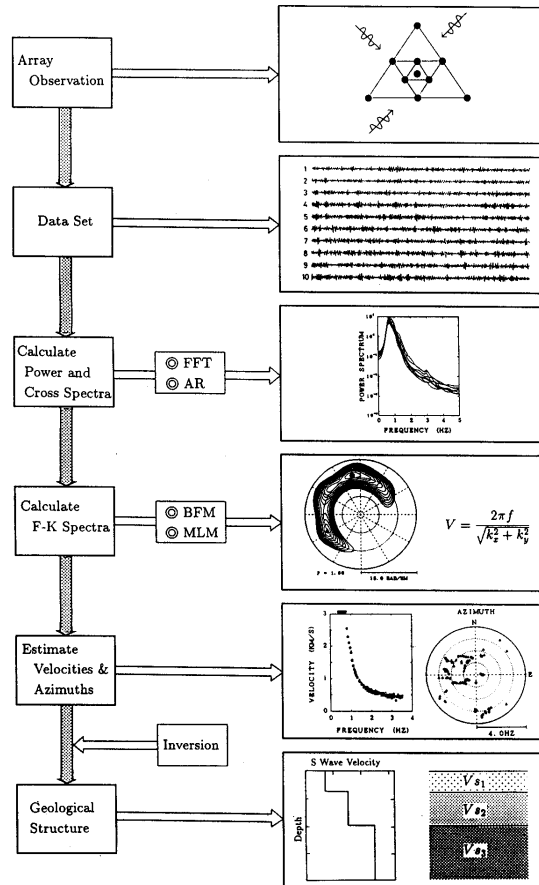


Figure 2.15. Flow of data analysis by the f-k method (Okada, 2003).

2.4.2.1. Time domain beamformer

The oldest array signal processing algorithm for f-k method is the time domain beamformer, or delay and sum beamformer. If a signal propagates across an array and is present in all the sensor outputs, delaying the output of each sensor appropriately and summing the results will reinforce the underlying signal, while diminishing the effect of noise. Choosing the delays optimally will focus the array on energy propagating from a particular direction (Johnson and Dudgeon, 1993).

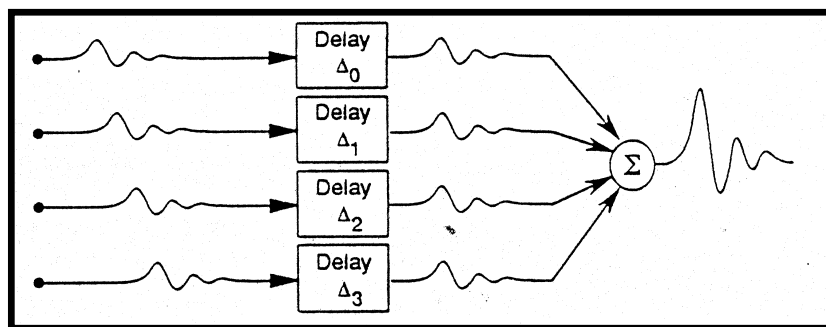


Figure 2.16. A schematic representation of time domain beamformer method.

Allowing the sensors to be weighted, the delay and sum beamformer equals;

$$z(t) = \sum_{i=1}^S w_i y_i(t - \Delta_i) \quad (2.31)$$

Where, $z(t)$ = the beamformer output for an assumed vector velocity, y_i = the signal plus noise measured in the i^{th} sensor, and Δ_i equals the assumed time delay for the i^{th} sensor, which tries to align the signal in each of the sensors (Johnson and Dudgeon, 1993).

2.4.2.2. Conventional frequency domain beamformer (FDBF)

Since seismic surface wave propagation tends to be dispersive and contain multiple modes, the frequency domain problem is more appropriate and well defined. The term beamforming refers to an array and signal processing algorithm's ability to focus on a particular direction, and mainlobe of an array smoothing function (ASF) is called a beam (Johnson and Dudgeon, 1993).

Conventional FDBF uses uniform weighting of all sensors, $W_m=1$ for $m=1$ to S . Since the weights are fixed, the ASF has a fixed mainlobe and sidelobe structure, and bears remarkable similarity to the one dimensional power spectral density (PSD) estimation problem with a rectangular data window.

The power in particular f-k pairs is estimated by multiplying the measured spatospectral matrix by the phase shift vector and summing the total power over all sensors. In vector notation, the power estimate is quadratic form

$$P_{FDBF}(k, w) = e^H(k)R(w)e(k) \quad (2.32)$$

Where H indicates the Hermitian transpose. The phase shift vectors try to align the array with plane waves propagating from a given direction with a given phase velocity, and if it is successful, a peak occurs in the f-k spectrum estimate.

2.4.2.3. Minimum variance distortionless look (MVDL)

The fixed ASF structure is the primary disadvantage of the FDBF. Capon's MVDL method (Capon, 1969) adaptively alters the weights of the sensors to optimize the characteristics of the f-k smoothing kernel at each frequency and wave-number pair.

The MVDL method, contained in a larger class of constrained optimization methods, attempts to pass a plane wave with a given f_0 k_0 undistorted (i.e. with unity gain) while minimizing the beamformer power output (Zywicki, 1999).

To simplify the problem, the independent variable will be assumed real valued. The constrained optimization problem is formulated as

$$\min_w w(w)^H R(w)w(w) \quad \text{subject to} \quad w(w)^H e(k_0) = 1 \quad (2.33)$$

3. GEOLOGY, TECTONICS AND GEOTECHNICAL CHARACTERISTICS OF THE ADAPAZARI AREA

3.1. Geology

According to Okay (1986), Marmara region is tectonically divided into three parts namely, Istranca massif, İstanbul zone and Sakarya zone (Fig. 3.1). The study area is on the İstanbul zone and is close to the border between the İstanbul and Sakarya zones.

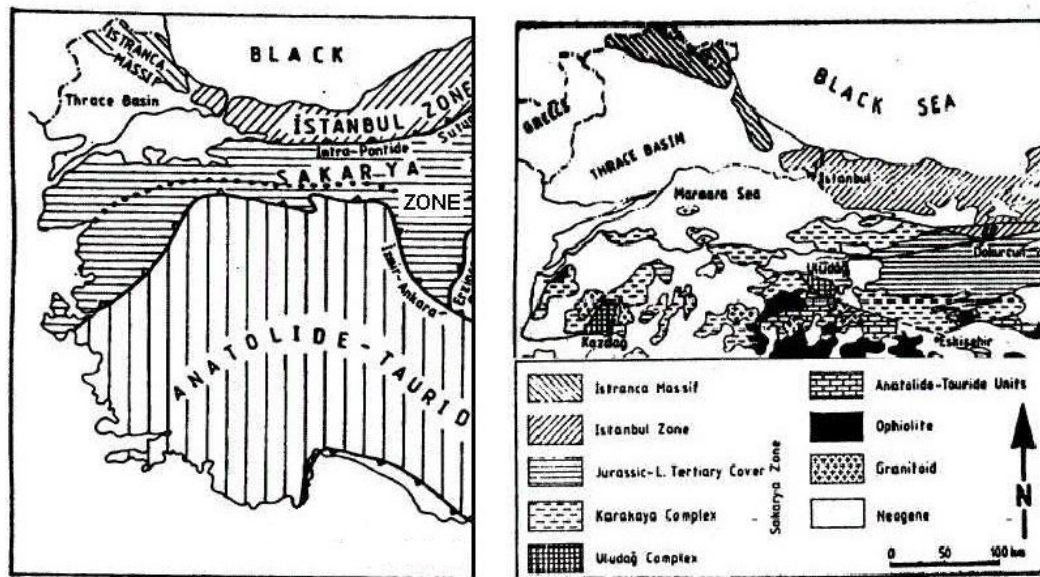


Figure 3.1. Geological units of the Marmara region (after Okay 1986).

The İstanbul zone is very distinctive from the neighbouring tectonic units in its stratigraphy, absence of metamorphism and lack of penetrative deformation. The İstanbul zone is characterised by a well developed, unmetamorphosed and slightly deformed Palaeozoic sedimentary succession, extending from Ordovician to the

Carboniferous which is overlain with a major unconformity by latest Permian to lowermost Triassic continental red beds (Fig. 3.3).

In marked contrast to the İstanbul zone, the Sakarya zone is characterized by a variably metamorphosed and strongly deformed Triassic basement called the Karakaya complex. This basement is unconformably overlain by Liassic conglomerates and sandstones which passes up to Middle Jurassic-Lower Cretaceous limestone and Upper Cretaceous flysch. Karakaya complex of Triassic age made up of strongly deformed and metamorphosed basic volcanic rocks, limestone and greywacke with ophiolite and limestone olistoliths.

These two zones juxtaposed along the Intra-Pontide suture which formed as a result of the closure of the Intra-Pontide Ocean in late Cretaceous-early Tertiary where the NAF is present now (Şengör and Yılmaz, 1981; Okay, 1989; Okay and Görür, 1995). According to geological data, the collision between the İstanbul and Sakarya zones occurred in the study area and its vicinity during the latest Cretaceous. After the collision of these two different continental blocks, a new period of sedimentation started during the Maastrichtian and lasted until the Middle Eocene.

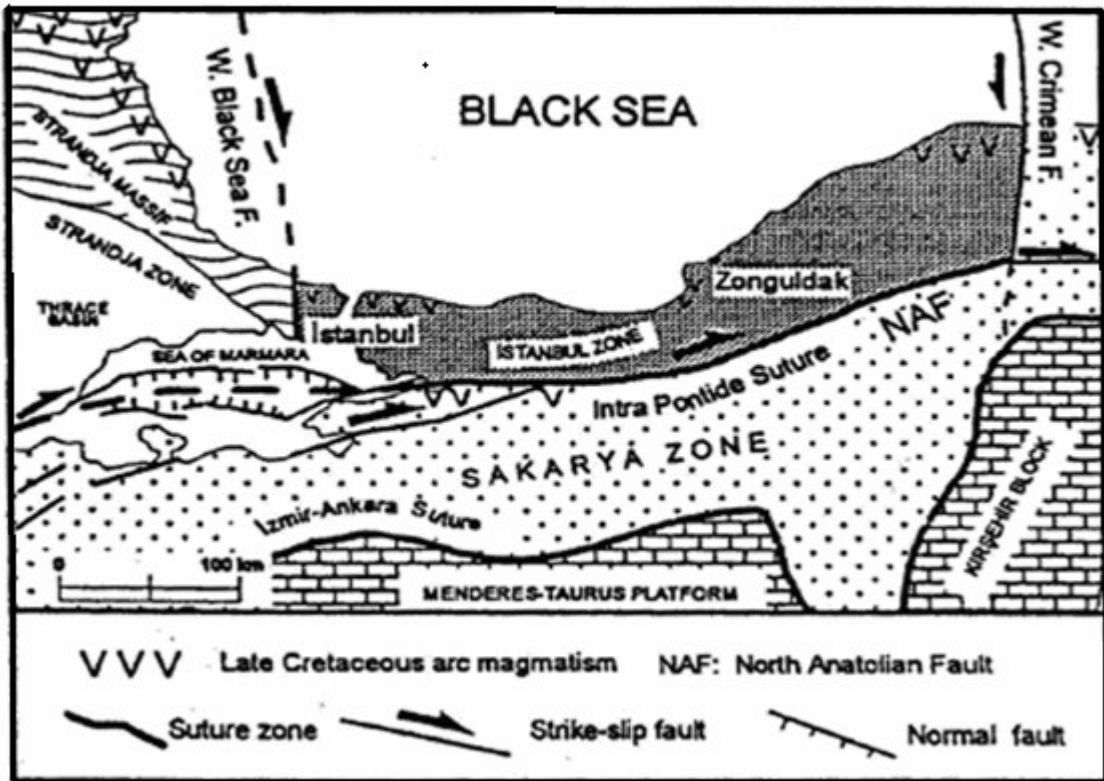


Figure 3.2. Tectonic setting of the eastern Marmara region (from Görür et al., 1997).

			LITHOLOGY	EXPLANATION
CENOZOIC	After Eocene			Eocene and younger sediments
	Paleocene Lower Eo.			
MESOZOIC	Upper Cretaceous			Clastics, limestones and volcanics
	Triassic			Clastics and limestones
PALEOZOIC	Carboniferous			Trakya Formation Sandstone-shale alternation, limestone and chert interbeds in the upper part
				Baltalimanı Formation Radiolarian chert
				Tuzla Formation Nodular limestone
	Devonian			Kartal Formation Fossiliferous shale
				Dolayoba Formation Fossiliferous limestone
	Silurian			Gözdağ Formation Laminated shale quartzite lenses in the upper part
				Aydos Formation Quartzarenite
	Ordovician			Kurtköy Formation sandstone, conglomerate and shale
Precambrian (?)				Metamorphic rocks

Figure 3.3. Stratigraphic section of the İstanbul zone (After Akyüz, unpublished).

The Pre-Neogene basement of the eastern Marmara region consist of rock assemblages of İstanbul and Sakarya zones developed in paleotectonic period (Şengör and Yılmaz, 1981; Okay, 1989; Okay and Görür, 1995; Yılmaz et al., 1995) (Fig. 3.2). Latest Cretaceous-Eocene rocks of these two different units can be compared, and they generally are regarded common cover of both tectonic units.

Eocene and older units of both Sakarya and İstanbul zones form the paleotectonic basement of the eastern Marmara region. After the middle Eocene regression an erosional period started in the region. Miocene and younger rocks deposited unconformably on the Eocene and older rocks of this paleotectonic basement.

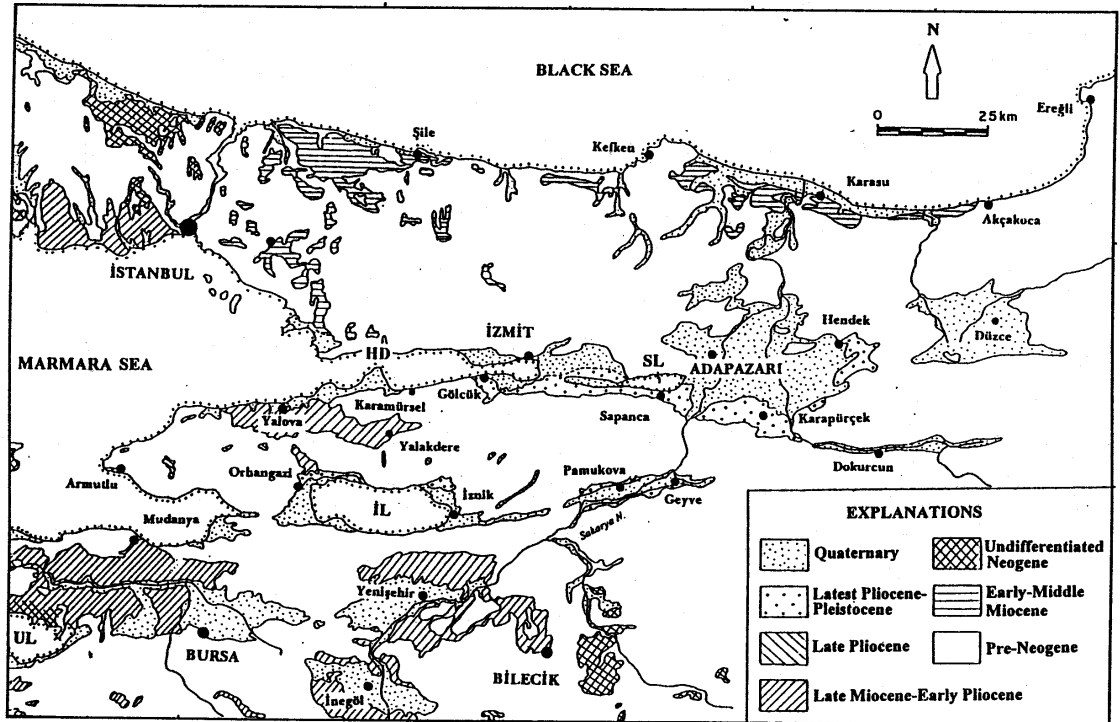


Figure 3.4. Map showing the distribution of Neogene and Quaternary sediments in the eastern Marmara region. HD: Hersek Delta, SL: Sapanca Lake, İL: İznik Lake, UL: Ulubat Lake (After Emre et al., 1998).

The Neogene-Quaternary aged rock assemblages in the region make up three different sedimentary sequences developed in different time and facies (Figs. 3.4 and 3.5). They can clearly be distinguished by angular unconformities. Of these, the oldest one, early-middle Miocene aged rocks are seen in Kocaeli peninsula, the late Miocene- Pliocene aged ones are observed in Armutlu peninsula and in Bilecik-Bursa regions, and the latest Pliocene-Recent aged ones are seen in the basins/corridors situated along the NAF zone (Figs. 3.4 and 3.6).

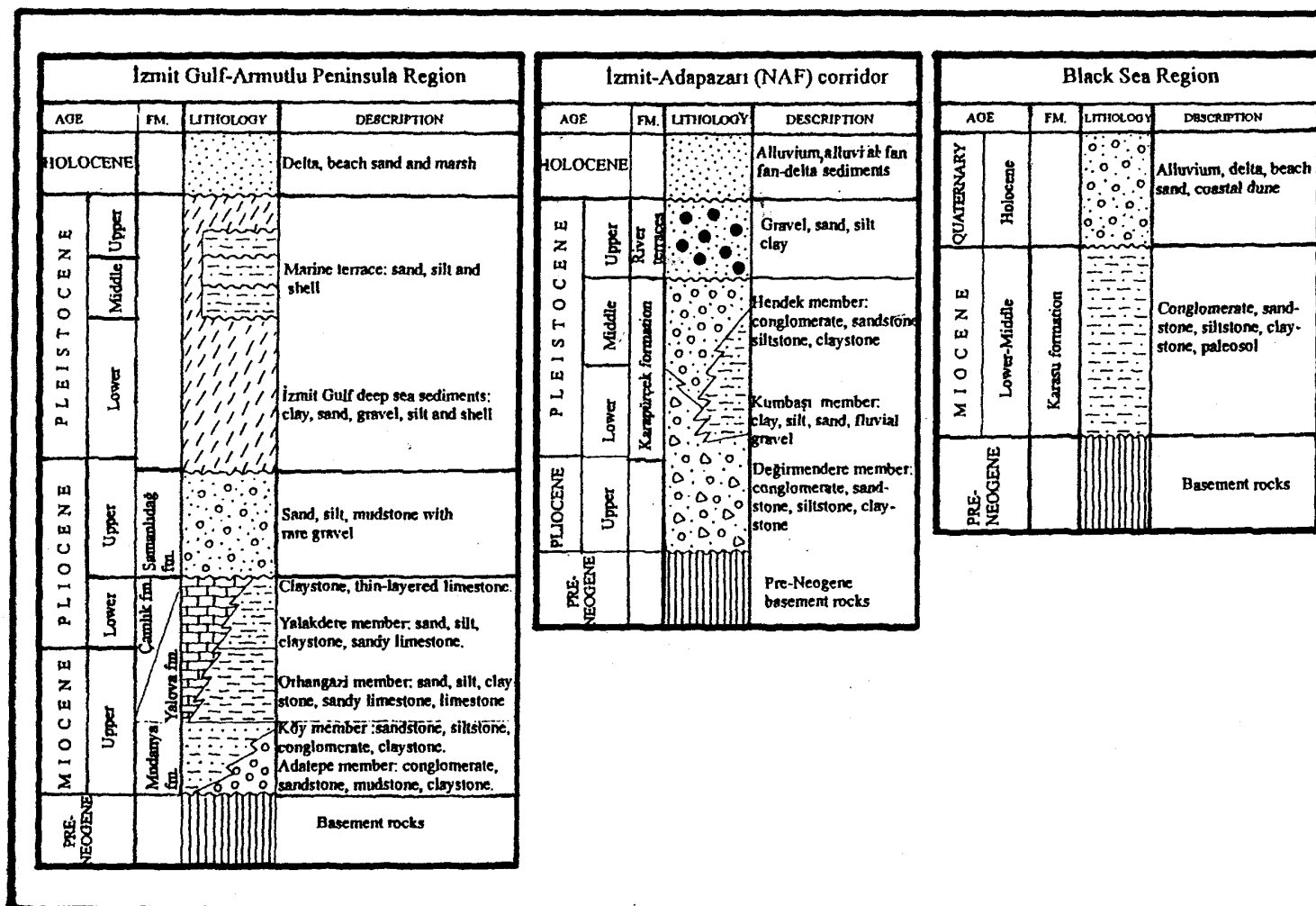


Figure 3.5. Columnar stratigraphic sections of Eastern Marmara Region (not in scale) (Emre et al., 1998).

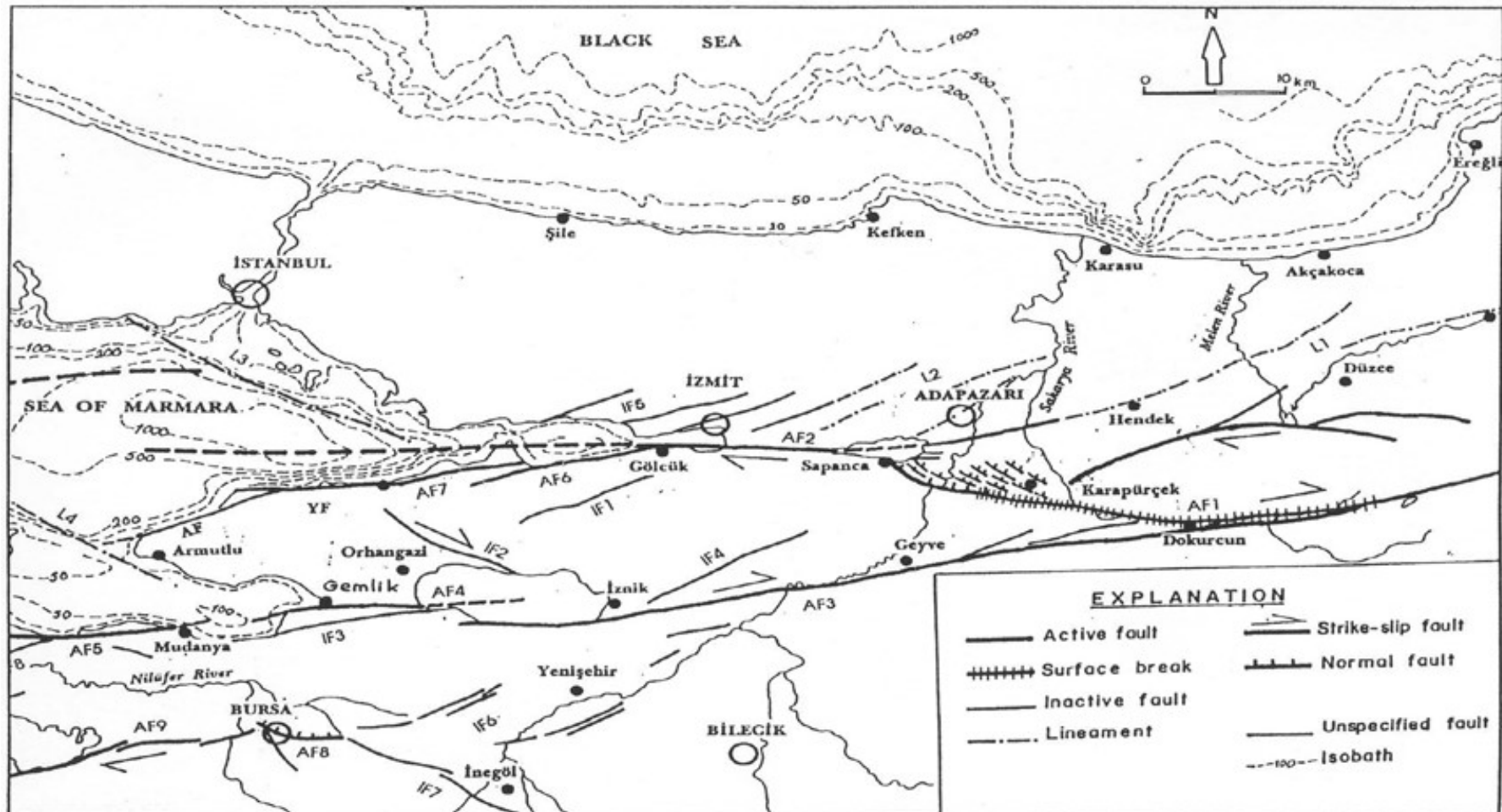


Figure 3.6. Neotectonic period faults in the Eastern Marmara Region (Emre et al., 1998).

Since no detailed geological investigations were carried out within the scope of the thesis an attempt is made here to review general geological properties of İstanbul zone, Sakarya zone, Adapazari basin and then soil properties of the Adapazari basin would be explained in detail using by some of the observations in the literature by various researchers.

Geological map of area including city of Adapazari, its close vicinity and Sakarya plain is given on Fig. 3.7. Geological units shown in Fig. 3.7 mainly consist of sedimentary rocks deposited between Lower Ordovician and Pliocene. The rocks that crop out in the Adapazari basin and around can be separated into two different groups. The first one is the rock was originate before the late Cretaceous and the second group rock and soil that was originate after the late Cretaceous.

As stated above, the basement rock that is constitute İstanbul and Sakarya zones show different characteristic and properties because of their origin and the mechanism of forming. For this reason, to provide easy comprehension these two zones should be investigated separately.

3.1.1. İstanbul Zone;

The sedimentary rocks of Ordovician-Cretaceous age that form the base of the İstanbul zone, can be explained from bottom to top;

3.1.1.1. Soğuksu Formation;

The oldest formation that crops out in İstanbul Zone is the Soğuksu formation. The lithology of the formation is comprised pink, purple and white coloured quartzite, sandstone and slate. These units were partly affected by low grade metamorphism. The formation was named by Kaya in 1982 and the dominant rock type of this formation is phyllite. The apparent thickness of the formation around eastern part of the zone is more than 1000 meters. Previous studies determined that this formation deposited in marine conditions and the age of the formation is probably Lower Ordovician or older.

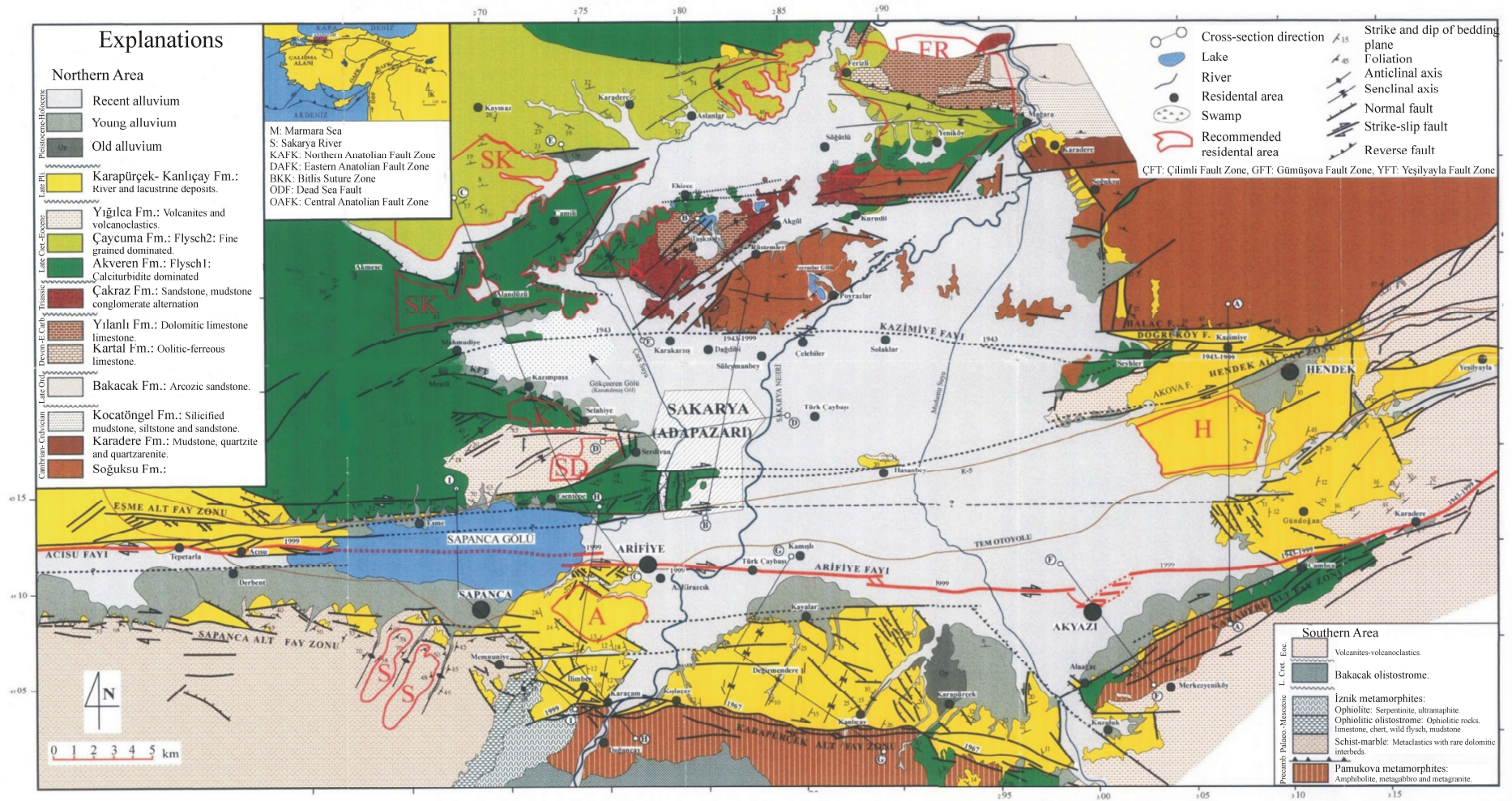


Figure 3.7. Geological map of the Sakarya basin and close vicinity (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

3.1.1.2. Karadere Formation;

This formation consists of pink, red, yellowish pink coloured quartzite, quartzarenite, that are alternating with pink-purple coloured metamudstones and metasilstones and those were affected by low grade metamorphism. Karadere formation was named by Kaya in 1982 based on its exposures around Karadere village around Araç town. The apparent thickness of the formation is 300 meters and the age of this formation is Ordovician (Kaya, 1982).

3.1.1.3. Kocatöngel Formation;

The formation made up of green coloured siltstones and mudstones and the apparent thickness of the formation is over 1000 meters. Kocatöngel formation was named by Kaya in 1982 based on regular outcrops in and around Kocatöngel village. The age of this formation is Ordovician.

3.1.1.4. Bakacak, Kurtköy and Aydos formations;

According to previous studies, this formation represents Palaeozoic sedimentary sequences (from Upper Ordovician to Silurian) in and around Adapazarı and Çamdag region. Bakacak and Kurtköy formations made up of multicoloured arkosic rocks whereas Aydos formation is represented by quartzites, quartzarenites and quartz conglomerates.

3.1.1.5. Kartal formation;

This formation crops out in the northern part of the Adapazarı basin and it is represented by oolitic limestones with oolitic fossils and grey-yellow mudstones. This formation was named by Kaya in 1982 based on the similarity of the units in İstanbul region. The thickness of the unit is around 200 meters and Lower Devonian in age (Kaya, 1982).

3.1.1.6. Yılanlı formation;

The dominant rock type of this formation is dark coloured thick and medium layered limestones and dolomitic limestones. This formation was named by Saner in 1979. The

thickness of the formation is 200 meters and the formation is aged as Middle-Upper Devonian.

3.1.1.7. Çakraz formation;

This formation is observed in the flanks of the anticlinal located between Taşlıgeçit and Akgölköy in the north of Adapazarı basin and unconformably overlies the Palaeozoic units. This formation is also seen in the southern flank of the anticlinal located between Söğütlü and Kumdil. The formation consists of a basal conglomerate overlain by sequence of red, pink, yellow and purple sandstones with red, purple mudstones. The age of the formation varies from late Permian to Triassic.

3.1.2. Sakarya Zone;

This zone comprises Pamukova and İznik tectonic units in the south and southwest of Adapazarı basin, ophiolitic rocks and various kinds of cover units from Late Cretaceous to Palaeogene.

3.1.2.1. Pamukova Metamorphics;

This formation is located in the middle and southern parts of the Armutlu Peninsula and middle part of the Almacık Mountains. It is a tectonic mixture of ophiolitic and metasedimentary rocks and it was named by Göncüoğlu in 1986.

3.1.2.2. İznik Metamorphics;

This metamorphic unit constitutes the southern and southwestern part of the Sapanca and it consists of several tectonic segments that have NNE-SSW elongation. This metamorphic basement tectonically alternates with Pamukova tectonic unit.

3.1.3. Late Cretaceous and younger units;

This group of rocks crop out inside the Adapazarı basin and along the border, that controlled by the faults, of the basin. As stated above, they deposited on both İstanbul and Sakarya zone units and form the common cover of these units. They can also be separated into two groups; the first group consists of a sedimentary sequences including Akveren, Yığılca and Çaycuma formations of Late Cretaceous to Eocene. The second

group rests unconformably on the first one and is represented by terrestrial clastics of Pliocene age, the Karapürçek and Kanlıçay formations.

3.1.3.1. Akveren formation;

This formation commonly crops out west and northwest part of the Adapazarı basin. It overlays older basement rock units with an angular unconformity at the bottom. The formation starts with terrestrial pebble stones and continues with limestones that was deposited in shallow to deep marine conditions. Akveren formation pass gradually up to Yığılca and Çaycuma formations. The age of the formation is Late Cretaceous and the thickness of it is about 250 meters.

3.1.3.2. Yığılca - Çaycuma Formations;

The Yığılca formation generally crops out in the eastern part of the Adapazarı basin and in and around of Serdivan district area that was settled western part of the Sakarya city. The age of the formation varies from Paleocene to Middle Eocene. This formation was made up of yellow-grey-brown coloured volcanic units. Çaycuma formation widespreadly crops out north and northwest part of the Sakarya basin and this formation has a lateral transition with Yığılca formation volcanic units. It is mainly represented by flysch-like regressive siliciclastics, clays and rare calciturbidites.

3.1.3.3. Karapürçek – Kanlıçay Formations;

The most widespread formations in the Adapazarı basin are Karapürçek and Kanlıçay formations. The Karapürçek formation was deposited in and around the edge of the basin by the alluvial fans and river deposits during the opening of the Adapazarı basin. This formation is made of dominantly fan type conglomerates, floodplain sediments, lagoonar limestones and organically rich claystones.

Kanlıçay formation commonly crops out in and around Kanlıçay village. This formation overlies the conglomerates of Karapürçek formation with a siltstone layer. Later on, this siltstone unit including gravel lenses alternates with sandstones. The upper part of the formation contains red, brown yellow coloured massive conglomerates. In short, this

formation starts with flood plain facies and ends with fluvial and talus facieses. The thickness of the formation together with Karapürçek formation is about 750 meters.

3.1.4. Quaternary units;

The city of Adapazarı and almost its all district were covered by Quaternary aged unconsolidated alluvium. Because of this, special consideration is given to the Quaternary aged units. All the geological units in Figure 3.7, except faults and folds, have been simplified and classified into three separate soil units (Figure 3.8). These are;

- Late Pliocene aged, unconsolidated soils
- Late Pliocene aged partially consolidated soils
- Quaternary aged unconsolidated soils

Quaternary aged units has further been separated into two category which is consists of coarse-grained basin margin and fine-grained basin deposits.

Basin margin sediments consist of alluvial fans reaching up 2-3 km in width. The basin sediments were deposited in meandering river, flood plain, swamp and lake environments. These zones have been shown with light colour on the map in Figure 3.8.

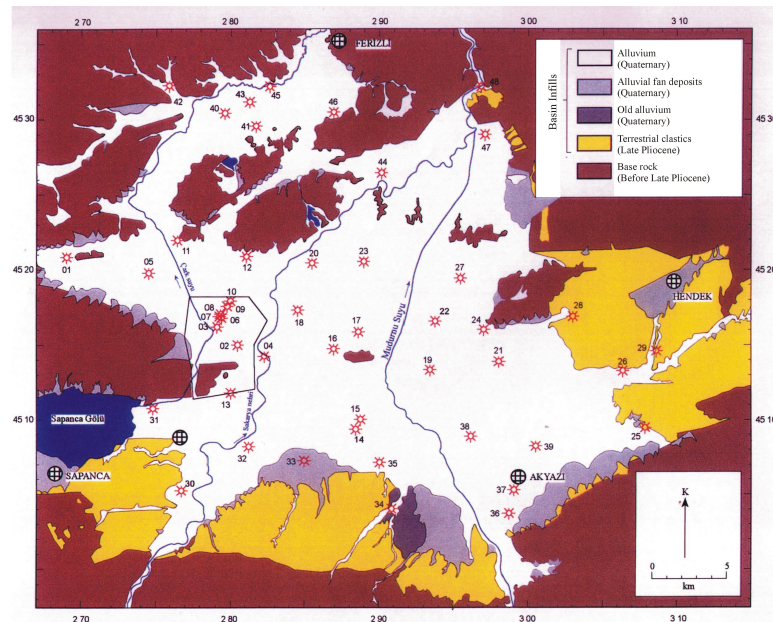


Figure 3.8. Simplified geological map that shows borehole locations (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

3.1.4.1. Basin margin deposits;

Almost all the fan, fan front and deltas were formed on the edge of Adapazarı basin which was controlled by the faults. These kinds of structures can be seen at south-east of Akyazı, Hendek, west and northwest of Sakarya, between Kayalar and Karapürçek. The dimension of the fans changes between a few hundred square meters to a few square kilometres.

Fans are conical shaped storage units that have formed with coarse materials. Those materials come from the north and west direction's rivers which losses their energy suddenly when they reach the basin. Those structures generally are formed of various size, unsorted, rounded blocks, pebble, sand, silt and these are loose materials. The dimension of grains decreases from the point of origin of the fan to its outer borders. The fan sediments in Adapazarı basin includes almost all kinds of grains belonging the rock that crops out at the edge of basin.

The conical shaped sedimentary deposits which form where the rivers reach to the sea or lakes are called deltas. The internal structure of these deltas is different and more regular than fans. In the south and northern shores of the Sapanca lake there are lots of well developed deltas and the dimensions of these deltas change between 0.2 to 5 square kilometre. For example Sapanca district is built on a delta which is formed by the İstanbul stream.

Either the fans or the fan fronts and deltas are formed by unconsolidated pebble, sand, silt, with a thickness of 20 meters to 260 meters. In distal zones they are made of mud or clay. Those sediments with a high porosity cause liquefaction and amplification in seismic waves depending on seasons due to water saturated nature.

3.1.4.2. Basin deposits;

The sediments that fills the internal parts of the Adapazarı basin is formed, fine grained clastics (silt, clay) which represents the tip of fans at basin borders, channel fills, sand bars, silt, mud and clay that were deposited in the different part of the Sakarya river which cuts through the whole basin. As the bed of the Sakarya river has changed by the time, different environments also display complex relationship with each other and show, sudden transitions from one facies to other. However it is very difficult to determine the lateral and vertical changes of facies due to the type

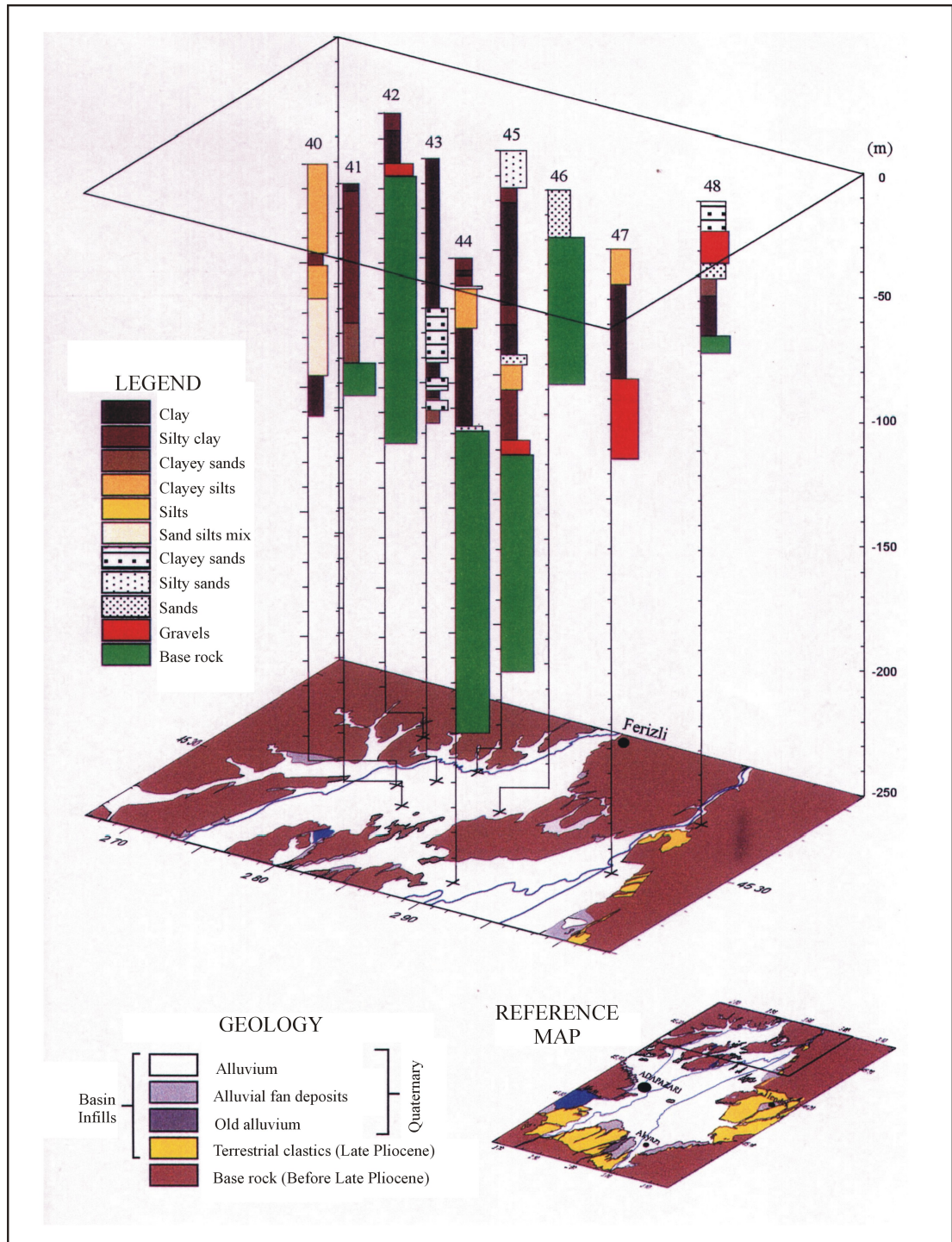


Figure 3.9. Vertical cross-section of boreholes in north of Adapazarı Basin (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

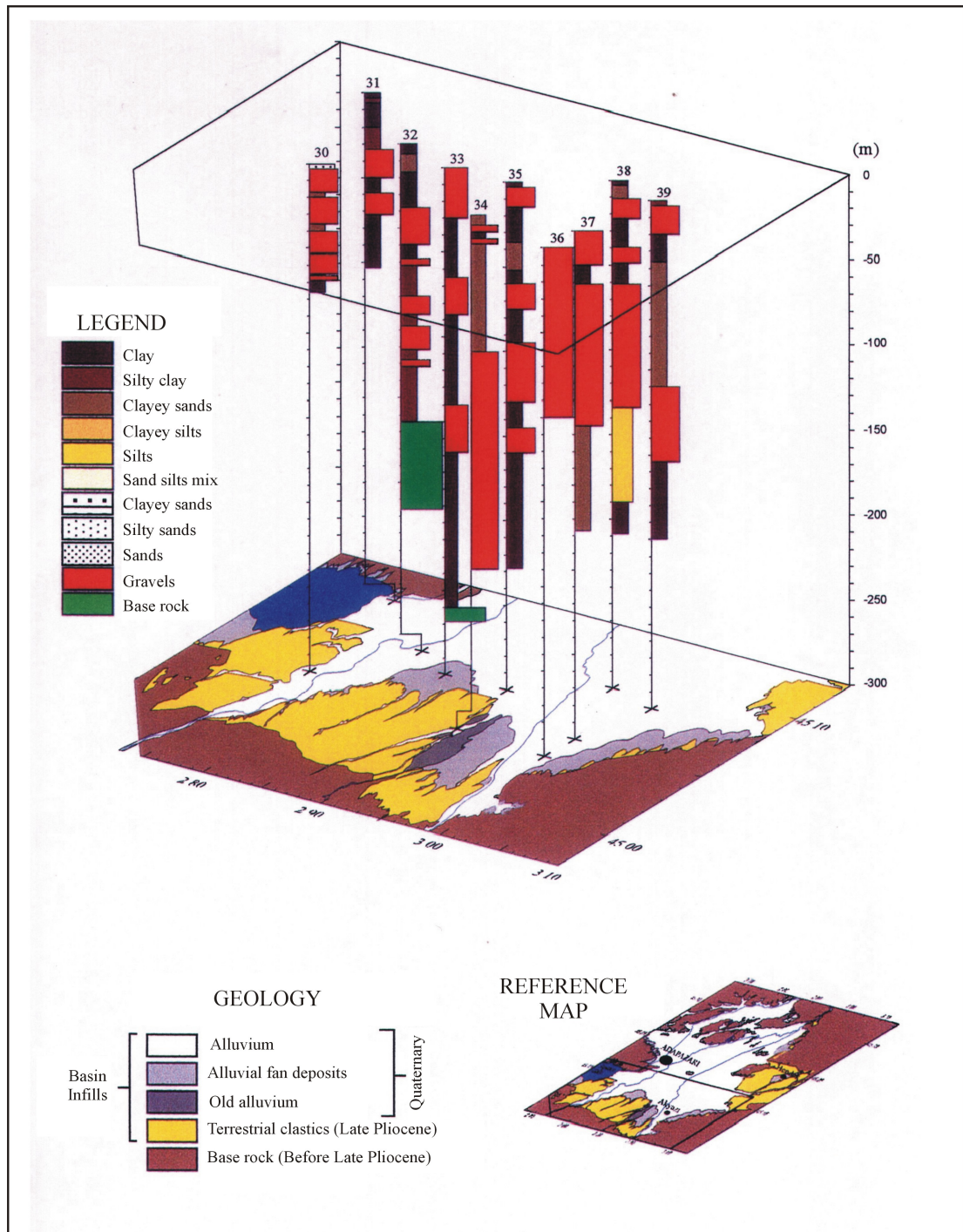


Figure 3.10. Vertical cross-section of boreholes in south of Adapazarı Basin (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

and thickness of the sediments that fills Adapazarı basin and also a great part of the basin is filled with a dense plantation and settlement.

Due to reasons that stated above, it is concluded that, the best way to determine the characteristics of the Sakarya basin is to investigate borehole logs. For these purposes 48 bore holes which are mainly used for research and watering purposes have been selected from the reports that are related to Sakarya basin. 42 of these boreholes belong to DSİ and 6 of them belong to KH (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999 - DRM report for microzonation for earthquake risk mitigation, 2003), (Fig. 3.8).

To demonstrate the detail of the sediments in the Adapazarı basin, this basin has been separated into to 4 sub zones. Sediment types and their thickness belonging to these sub zones have been given in different figures with respect to each other. Those sub zones are; north and south part of the Adapazarı basin, centre and east part of Adapazarı and northwest part of the Adapazarı basin (Figs. 3.9, 3.10, 3.11, 3.12).

Koçyiğit (1988) stated that the Quaternary aged Adapazarı basin sediments have been separated into ten facies based on the log data of the boreholes. Those are; clay, clay with silt, clay with sand, silt, silt with clay, silt with sand, sand, sand with clay, sand with silt and pebble. When the borehole logs in Figures 3.9, 3.10, 3.11, 3.12 examined with respect to each other those can be concluded about Quaternary aged sediments;

- The thickness of the alluvial fill increases from the edge of the basin to the centre.
- The sediments taken from the borehole logs shows that, %58.7 of the material inside the Adapazarı basin is composed of clay units, %32.5 pebble units and %8.8 sandy and silty units so that clay and pebble are the most dominant sediments in the Adapazarı basin (Fig. 3.13).
- The whole part of the basin, observed that sediments that fill the basin has a lateral and vertical facies variation in a short distance.

As a summary, the thickness of the sediments that filled Adapazarı basin is more than 300 meters. In this basin clay and pebble is the most dominant material and almost all of the facies shows that not only lateral but also vertical variation in a short distance (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

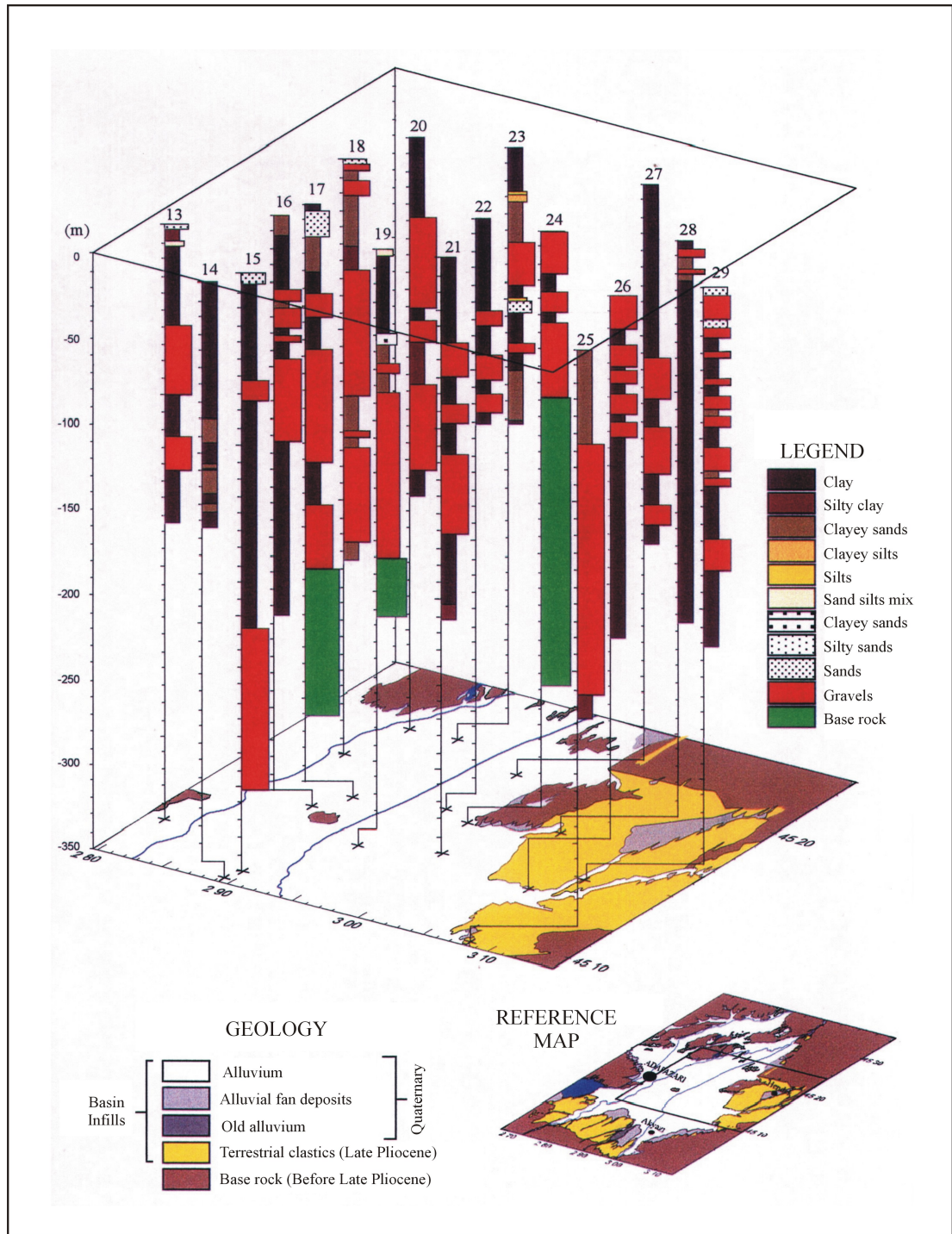


Figure 3.11. Vertical cross-section of boreholes in center and east of Adapazarı Basin (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

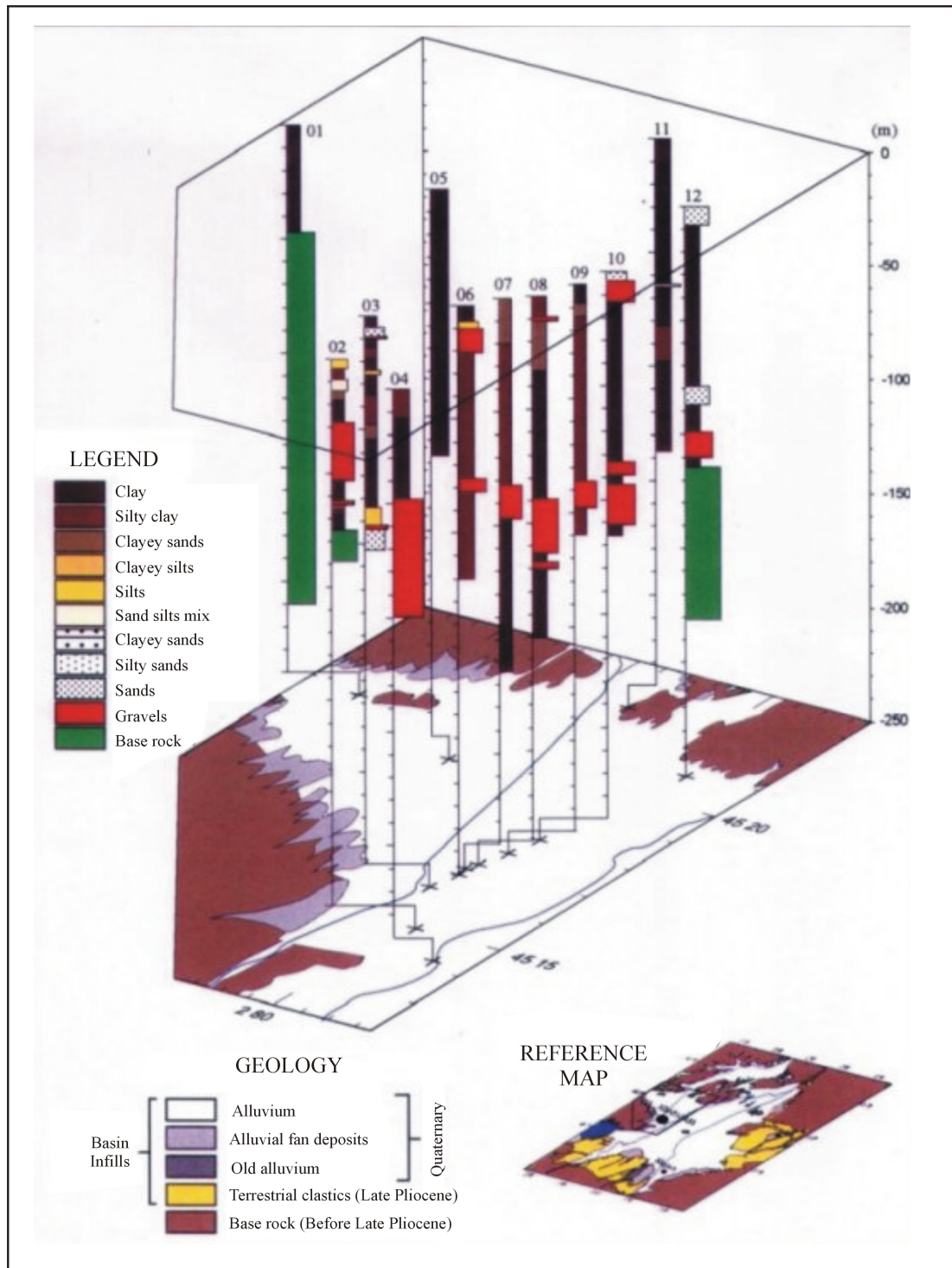


Figure 3.12. Vertical cross-section of boreholes in northwest of Adapazarı Basin (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

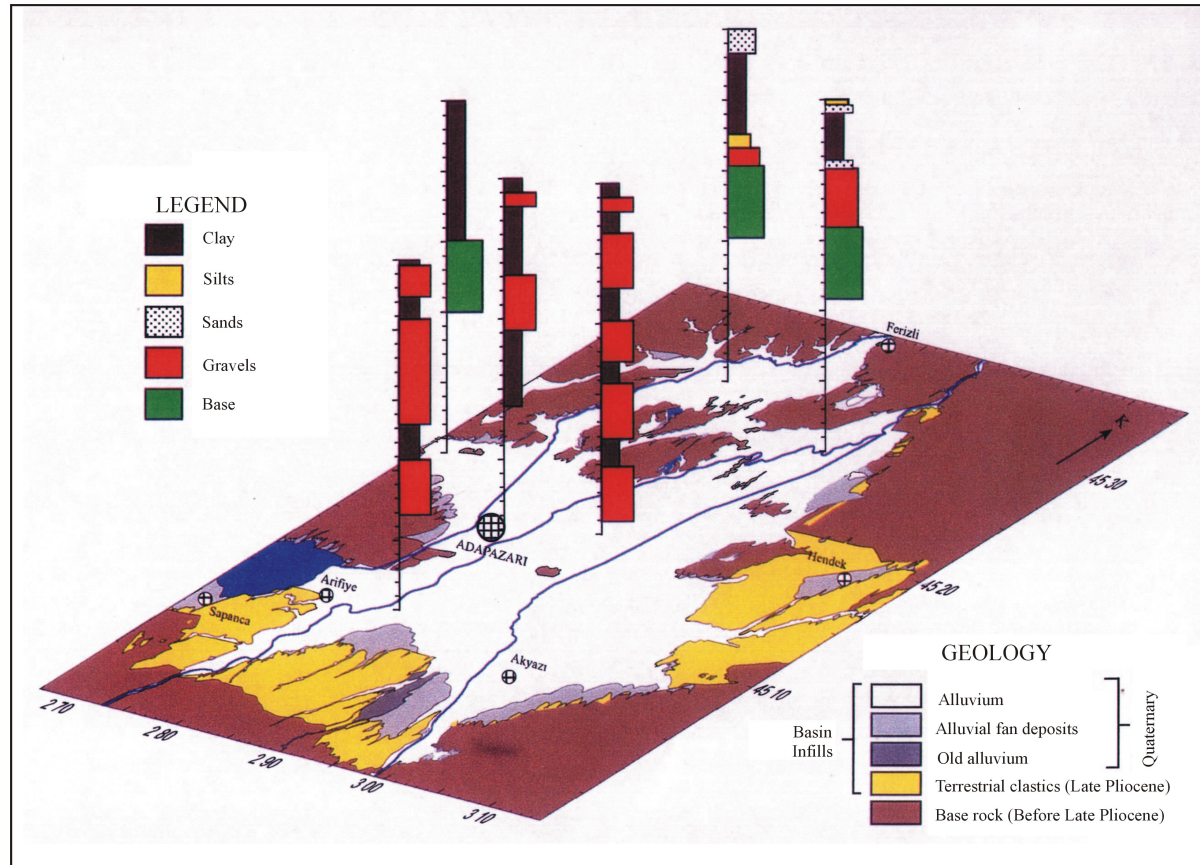


Figure 3.13. Vertical cross-section of boreholes in north, south, west and the centre of the Adapazarı Basin (TÜBİTAK, ODTÜ, MTA report for Adapazarı settlement, 1999).

3.2.Geotechnical characteristics of Adapazarı basin

Adapazarı-Arifiye area is the most perplexing and peculiar from the point of view of ground conditions. This city, with a history going back over 1000 years, is located in the vicinity of the River Sakarya. Its name translates as “island market”, indicating a long past of annual flooding and river meandering, which frequently necessitated the use of boats to reach the central market here. Flooding by the River Sakarya was last recorded on the Adapazarı plain in 1963, after which dams constructed upstream have checked the excessive flows of the past. The other stream, Çark, currently flows along the western side of the city, draining the nearby Lake Sapanca northwards. No record of flooding caused by this stream is available, but evidence exists to suggest that it has periodically altered course towards the city centre, leaving behind marshes. The altitudes on the axis from Arifiye to the northern limits of the city of Adapazarı vary between 28 and 32 metres and the area of study can, for most practical purposes, be assumed to be a flat site. The hills to the west of Adapazarı rise up to 250 m.

The Upper Cretaceous-Eocene flysch bedrock makes a notch shaped valley along a north- south axis under the city centre which has been filled with fluvial and lacustrine sediments for the past 7000 years. The current depth of the basin is estimated to be over 1000 m with the result that earthquakes that hit at intervals of a decade have had devastating effects, possibly due to ground amplification effects and liquefaction (DRM Report, 2003).

The sediments are soft and loose with the ground water table rising to the surface in spring months in many locations. Carbon dating on fragments of a reed found at a depth of 7.5 m revealed that it could not be older than 700 years (DRM Report, 2003). This corresponds to an excessive rate of sedimentation in the Holocene. Overwhelmingly green and occasionally brown non-plastic silts are liquefiable and are found in the top 10 m. The colour indicates the genetic property, which are metamorphic and ophiolitic rocks along the Geyve Valley through which River Sakarya flows. The brown clays are found below 6m and are deposits in shallow lakes. Grey clays of high plasticity are sediments likely to be deposits of temporary swamps of the past. A majority of soils in Arifiye appear to be predominantly clays

of high plasticity and organic content, suggesting the presence of large marshes in the past.

A total of about 600 high quality boreholes and soundings have been implemented since the 1999 earthquake by Sakarya University. Most of the boreholes were 15 m deep with several selected spots reaching 30 m additionally, a pilot borehole of 200 m was drilled in the middle of the city. A total of 3850 samples was taken from these boreholes and these samples were classified according to the Turkish Standard TS1500/2000. All the data obtained from this work points out to the fact that Adapazarı-Arifiye region is a fine soil zone. On the other hand, the information collected from the different surveys that were made during the past three years indicates that the soil conditions in the city can be summarised as follows.

The soil layering is exceptional: The velocity and the duration of flooding formed layers as thin as a few centimetres and the thickness of most layers rarely exceed a few meters. The top 5 m is dominated by silts and their age is between 100 and 1000 years .

Clays and sands appear as bands, high plasticity clays being deposited in former marshes or possibly shallow lakes, and the sands and silty sands along former and buried streams. Gravel is rarely encountered, and its presence is a clear indication of the former beds of Sakarya and Çark streams.

3.3. Tectonic setting of the region

Western part of the North Anatolian Fault (NAF) controls the tectonic regime in the Marmara region. West of Bolu, toward the Marmara region, the NAF begins to loose its single fault line character and splays into a complex fault system.

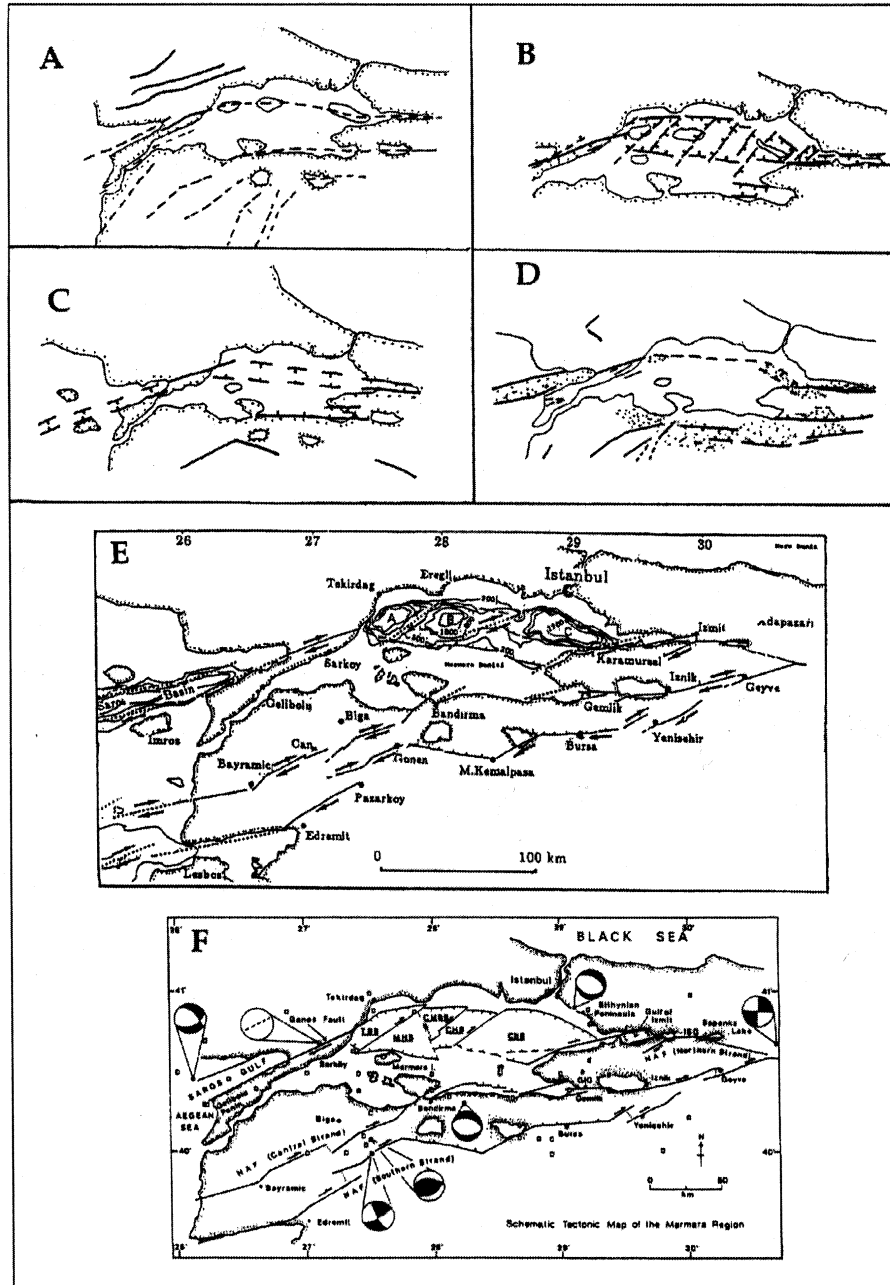


Figure 3.14. Compression of the structural models suggested for the Marmara region ([A] Pinar,1943 [B] Pfannenstiel et al.,1944 [C] Crampin et al.,1986 [D] Şengör,1987 [E] Barka et al.,1988 [F] Wong et al.,1995 Ergun et al.,1995).

Based on low resolution bathymetric data and earthquake occurrences, several researchers have developed different tectonic models for Marmara region. Pınar (1943), Pfannenstiel et al. (1944), Crampin et al. (1986), Şengör (1987), Barka et al. (1988), Wong et al. (1995), Ergun et al. (1995) can be cited among them (Fig. 3.14).

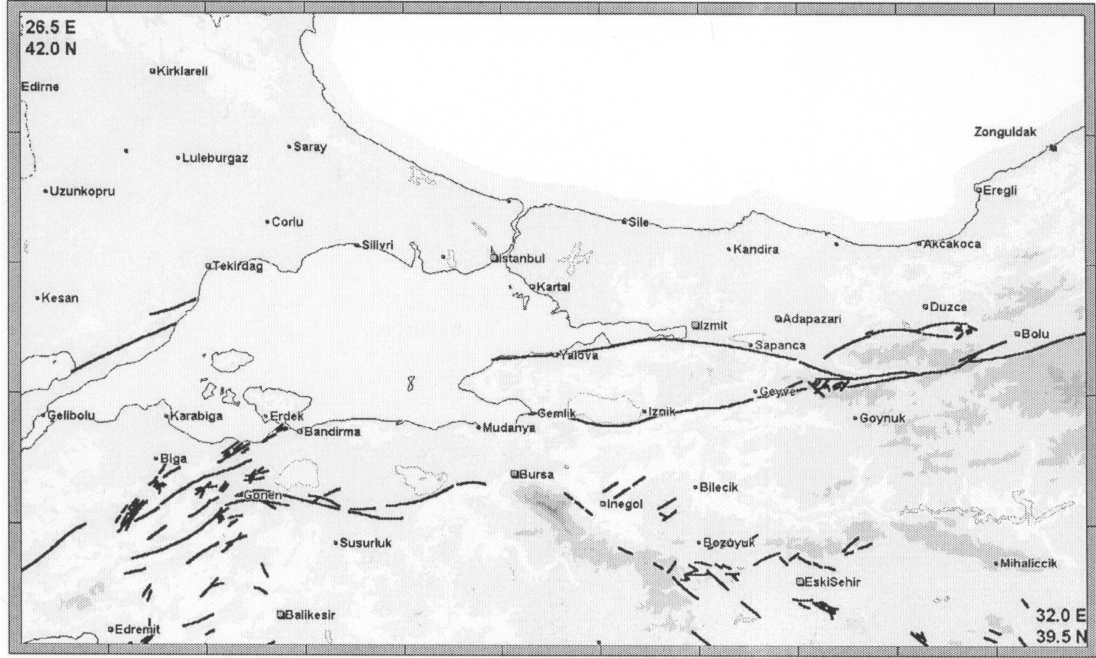


Figure 3.15. Active fault map of the region. (Şaroğlu et al., 1992).

The active tectonic map of the region prepared by the General Directorate of Mineral Research and Exploration (MTA) Turkey is given in Fig. 3.15. Le Pichon et al. (2001) developed a fault model based on the data collected in 1997 by the ship 'MTA Sismik-1'. Data obtained during the recent high resolution bathymetric survey of the Ifremer RV Le Suroit vessel indicates that a single, thoroughgoing strike slip fault system (Main Marmara Fault) cuts the Marmara sea from east to west joining the 17.8.1999 Kocaeli earthquake fault with 9.8.1912 Şarköy-Mürefte earthquake fault (Fig. 3.16).

Building upon and interpreting the extensive studies conducted defines the tectonic evolution of the Marmara sea region as the superposition of two different aged fault system as illustrated in Fig. 3.17. These are the early Miocene-early Pliocene Thrace-Eskişehir fault zone and its branches and the late Pliocene-recent NAF and its branches. The northwest-southeast trending Thrace-Eskişehir fault is a major dextral strike slip system, which was active during the early Miocene-early Pliocene. It has divided into four parts by the NAF at the end of the late-Pliocene. This event marked

the initiation of the late neotectonic period. During that period the NAF extended westward as a number of splays by joining the Ganos, Bandırma-Bahramkale and Manyas- Edremit fault zones (Yaltırak, 2002).

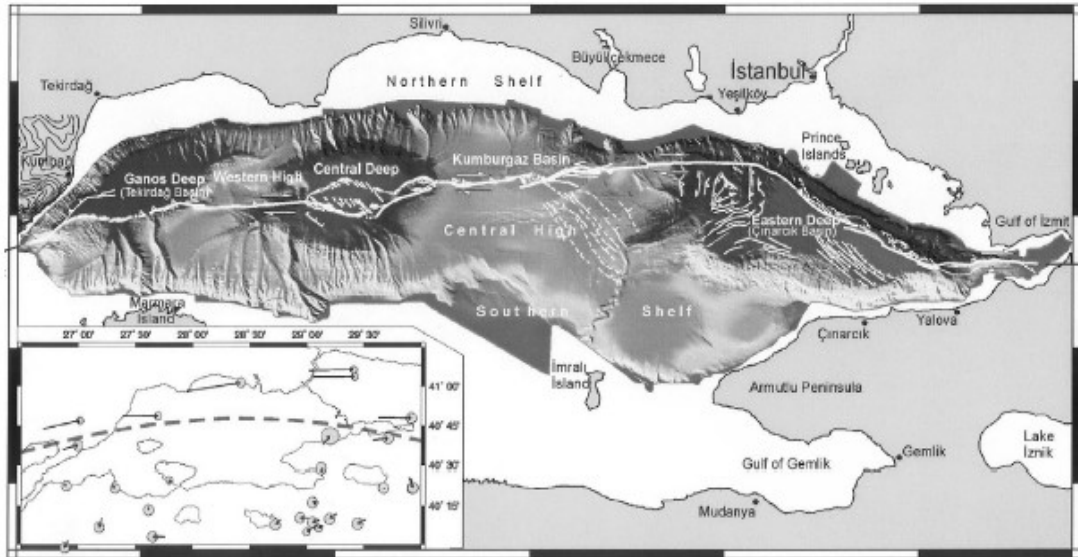


Figure 3.16. The recent high resolution bathymetric map obtained from the survey of the Ifremer RV Le Suroit vessel that indicates a single, thoroughgoing strike slip fault system (Le Pichon et al., 2001).

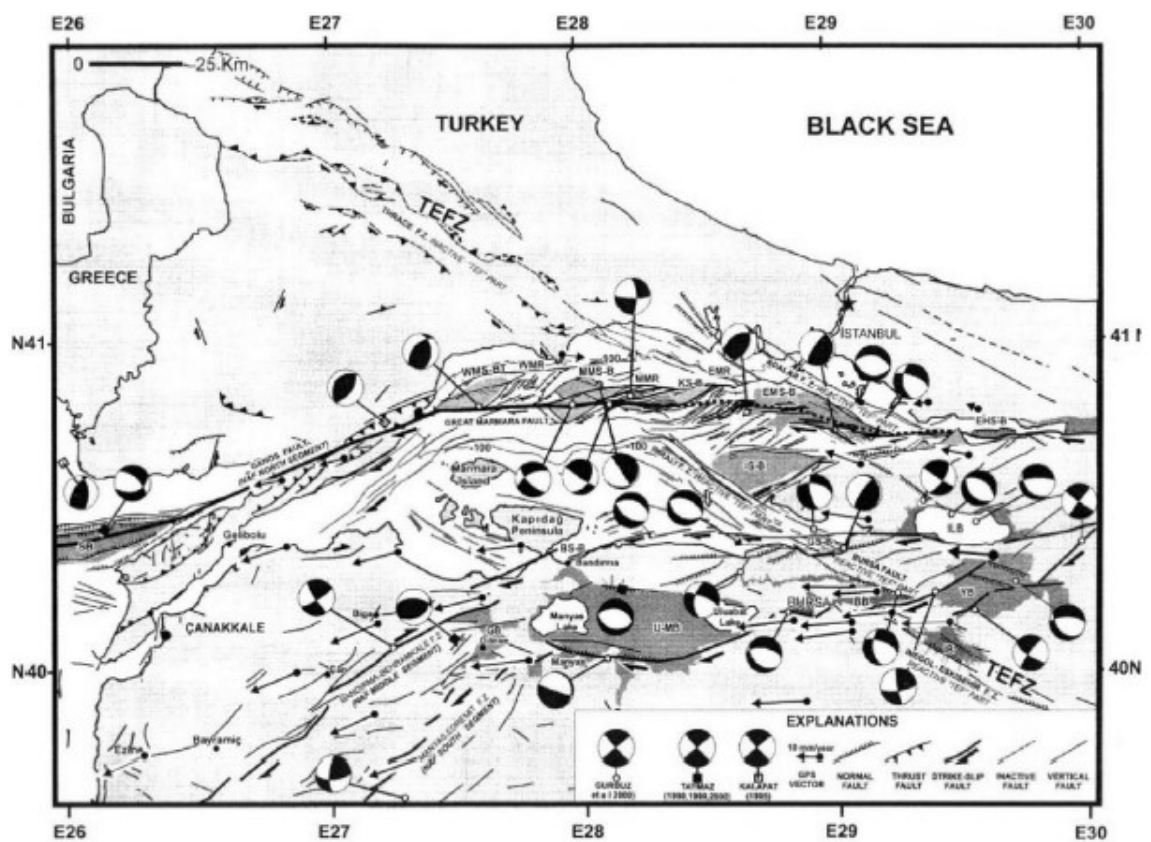


Figure 3.17. Tectonic map of Marmara region compiled from various studies (Yaltırak, 2002)

The connection of the northern branch of the NAF to the Ganos fault zone in the west caused the development of a single buried fault in the Marmara sea and the formation of the troughs and ridges, superimposed onto the negative flower structure formed by the Ganos fault in the early neotectonic period. The middle strands extends East-West from İznik lake to Bandırma and connects to the N60°E-trending Bandırma-Bahramkale zone and turns southward near Bandırma.

The southern branch of the NAF connects to the Manyas-Edremit fault zone, forming three pull apart basins along Yenişehir, Bursa and Manyas segments. The branches of the NAF cut the Thrace Eskişehir fault at three places: The eastern Marmara Sea region, in Gemlik Bay and to the east of Bursa. The lateral offsets at those locations which amount to 58-59, 7-8 and 10-11 km respectively give a clear idea about the relative displacements and slip rates along each of the three branches of the NAF in the Marmara region (Erdik et al., 2004).

4. DATA ANALYSIS AND DETERMINATION OF V_s PROFILES

4.1. Data acquisition and analysis procedure for SPAC method

When there are several dominant modes of surface waves in the microtremors, or when it contains strong body waves, SPAC method has not the potential to separate these waves. The SPAC method can not separate the fundamental-mode surface wave. Such a case is rather rare, but where such a condition exists, the other array analysis methods are available. Additionally, by SPAC method the direction of the source of the wave can not be estimated (Okada, 2003).

However the SPAC method is simpler in field data acquisition and data analysis than the other array analysis methods. The other methods require a relatively large number of observation stations. Therefore their field effort is more intensive than that of the SPAC method. At the same time, as stated in chapter two, the spatial extent of the array becomes larger than that of the SPAC method, resulting in the violation of the assumption of “horizontal layers” (Okada, 2003).

The observation array to collect data for analysis by the SPAC method requires, as seen in Figure 4.1, at least four seismometers, three of them arranged on a circle of radius r and one at the centre. It is expected that a larger number of seismometers will give a better result.

If the three stations on the circle form a equilateral triangle, the directional calculation becomes simpler, and the distance r in (2.30) can take two values, *i.e.* the radius, r , and the length of the side of the equilateral triangle, $\sqrt{3}r$. If two or more arrays with different r are spread at the same location, a wider range of phase velocity can be estimated.

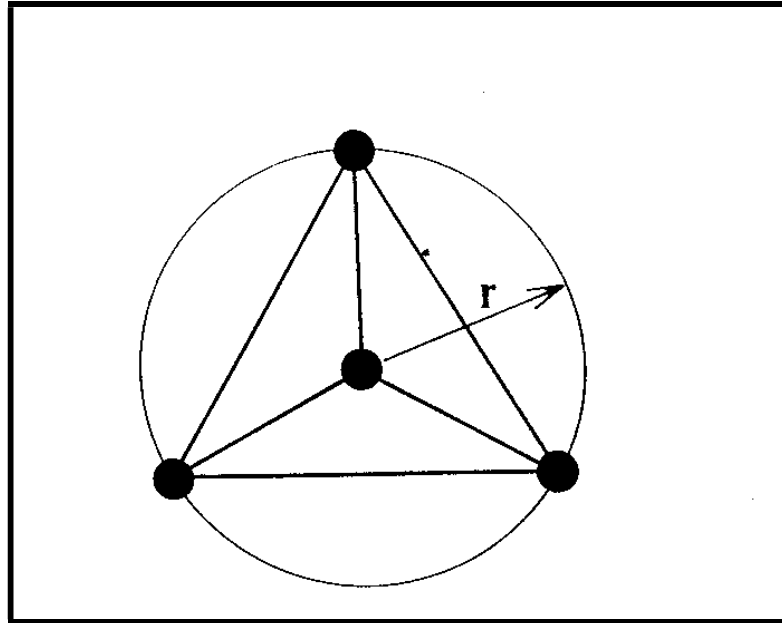


Figure 4.1. Simple array geometry for SPAC that is composed of four geophone (Kudo et al., 2002).

Generally, there are two types of data acquisition system; one with all the stations of an array connected and recorded with a multi-channel recorder, and the other with each station recorded independently.

The former system has the advantage that no time correction is necessary, as all the stations are recorded against the same clock. This system is well-suited to a small array, for estimation of shallow structure. However, there are operational disadvantages for large arrays; the difficulty of wiring restricts selection of station sites, and the requirement for maintenance of the wire throughout the observation.

The second system requires time calibration amongst the stations, for example by using a Global Positioning System (GPS) clock installed in each recorder to synchronise the records. However it imposes little restriction in selection of the station sites. Considering all the advantages and disadvantages, the independent system seems more suited to the recording of microtremors.

4.1.1. Data analysis by Spatial Autocorrelation (SPAC) Method

The work flow for data analysis by the spatial autocorrelation method is shown in Figure 4.2. The theory of the analysis was explained in Chapter 2, and several problems in practical data analysis are presented in this section.

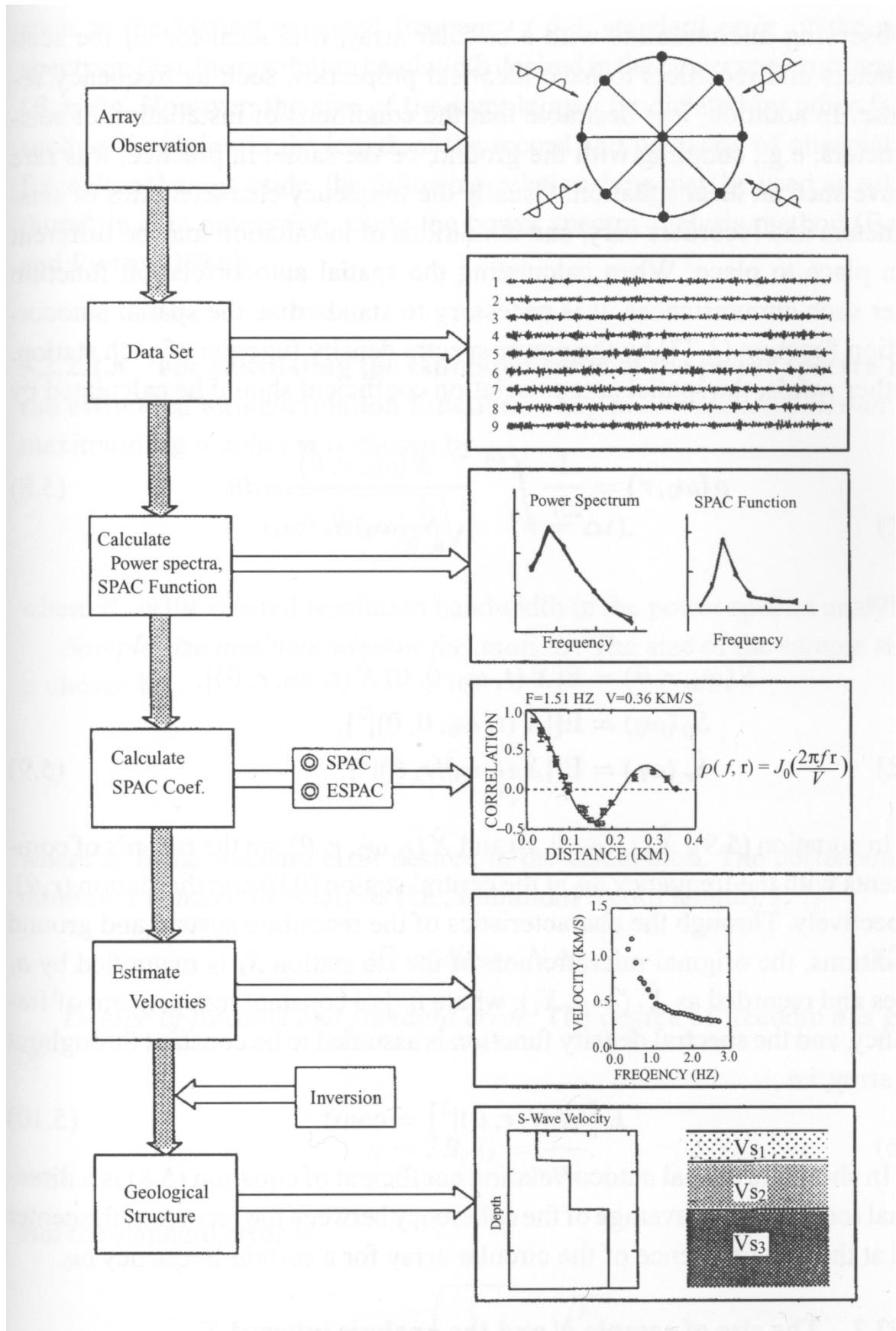


Figure 4.2. Flow of data analysis by the SPAC method (Okada, 2003).

4.1.1.1. Standardization of the spatial autocorrelation function

In observing microtremors with a circular array, it is ideal for all the seismometers and recorders to have all identical properties, such as frequency response. In addition, it is desirable that the conditions of installation of seismometers, e.g. coupling with the ground, be the same. In practice, it is rare to have such an ideal situation. Usually the frequency characteristics of seismometers and recorders vary, and conditions of installation may be different from place to place. When calculating the spatial autocorrelation function under such circumstances, it is necessary to standardize the spatial autocorrelation function (2.27) by the power spectra density function of each station. In other words, the spatial autocorrelation coefficient should be calculated by:

$$\rho(\omega_0, r) = \frac{1}{2\pi} \int_0^{2\pi} \frac{\hat{S}(\omega_0, r, \theta)}{\sqrt{\hat{S}_0(\omega_0) \hat{S}_r(\omega_0)}} d\theta \quad (4.1)$$

where:

$$\begin{aligned} \bar{S}(\omega_0, r, \theta) &= E[\bar{X}(t, \omega_0, 0, 0) \bar{X}(t, \omega_0, r, \theta)] \\ \bar{S}_0(\omega_0) &= E[|\bar{X}(t, \omega_0, 0, 0)|^2] \end{aligned} \quad (4.2)$$

$$\bar{S}_r(\omega_0) = E[|\bar{X}(t, \omega_0, r, \theta)|^2].$$

In (4.2), $\bar{X}(t, \omega_0, 0, 0)$ and $\bar{X}(t, \omega_0, r, \theta)$ are the records of components with the frequency ω_0 at the central station (0,0) and the station (r, θ) , respectively. Through the characteristics of the recording system and ground conditions, the original microtremors at the I -th station X_i is magnified by a_i times and recorded as $\hat{X}_i (= a_i X_i)$, where a_i is a constant independent of frequency, and the spectral density function is assumed to be constant throughout the array, *i.e.*:

$$E[|X(t, r, \theta)|^2] = \text{const.} \quad (4.3)$$

In short, the spatial autocorrelation coefficient of equation (4.1) is a directional (or azimuthal) average of the coherency between the records at the centre and at the circumference of the circular array, for a certain frequency ω_0 .

4.1.1.2. The size of sample N and the analysis interval T_r

There is no standard in the size of sample, *i.e.* the length of the data segment for analysis. It is chosen based on the required precision for the estimated value. It may be practical to decide these values by considering parameters such as the highest expected frequency (f_N), standard error of the power spectrum (ε_r), the resolution bandwidth desired in the power spectrum analysis (B_e), etc. However, the size of the sample may be dictated by other factors such as limitation in the length of the record and precision of observations. Exceptional cases aside, the following relationships may be used as rules of thumb in data processing, using the power spectra analysis method (Bendat and Piersol, 1986):

For calculating the estimated value of the power spectra from the estimated autocorrelation function;

Lag in correlation function: The maximum lag number m is chosen by:

$$m = \frac{1}{B_e h} \quad (h = \Delta t) \quad (4.4)$$

where B_e is the desired resolution bandwidth in the power spectra analysis.

Sample size and time window for analysis: The size of the sample size N is chosen by:

$$N = \frac{m}{\varepsilon_r^2} \quad (4.5)$$

where ε_r is the standard error desired in the calculation. The corresponding minimum window for analysis (*i.e.* minimum record length) T_r is:

$$T_r = Nh = N\Delta t \quad (4.6)$$

Degree of freedom and standard error: the degree of freedom n is given by:

$$n = 2B_e T_r = \frac{2N}{m} \quad (4.7)$$

and the standard error is:

$$\varepsilon_r = \sqrt{\frac{1}{B_e T_r}} = \sqrt{\frac{m}{N}} \quad (4.8)$$

Therefore, when N is constant, a smaller m gives a smaller ε_r .

For calculating the estimated power spectra by the Fourier transform

Here, the commonly-used smoothing in the frequency domain is assumed. When the spectra are white noise within a limited range, the estimated spectra at the frequency interval of $1/T_r$ essentially have no correlation. Therefore, when l frequency components near the estimated values of the raw spectra are averaged, the smoothed estimate value of the spectra $\bar{G}_k$ results in:

$$\hat{G}_k = \frac{1}{l} [\bar{G}_k + \bar{G}_{k+1} + \cdots + \bar{G}_{k+l-1}]. \quad (4.9)$$

This $\bar{G}_k$ is a χ^2 variable with degree of freedom approximately $n=2l$ by the addition theorem of χ^2 of independent variables. The resultant effective resolution bandwidth is approximately $B'_e = lB_e$, where $B_e = 1/T_r$. Therefore this averaging in the frequency domain gives:

$$B'_e = lB_e = l/T_r = l\Delta f \quad (4.10)$$

$$N = 2B'_e T_r = 2l. \quad (4.11)$$

The standard error is:

$$\varepsilon_r = \sqrt{\frac{1}{B'_e T_r}} = \sqrt{\frac{1}{l}} \quad (4.12)$$

The estimated value $\bar{G}_k$ is considered to represent the mid-point of the frequency range from f_k to f_{k+l-1} . Altogether N/l such estimated values are calculated.

4.1.1.2. Calculation of the phase velocity

When the spatial autocorrelation function method is used, the phase velocity at the array centre is obtained by equation (2.30). From $x_0 = 2\pi f_0 r_0 / c(f_0)$, the phase velocity for the frequency f_0 is:

$$c(f_0) = \frac{2\pi f_0 r_0}{x_0} \quad (4.13)$$

where x_0 is the variable of the Bessel function of the first kind of zero order. x_0 satisfies the value of the spatial autocorrelation coefficient ρ_0 for a certain frequency obtained from a dataset observed by a circular array of radius r_0 .

In this case, there may be a frequency range in which the spatial autocorrelation function does not satisfy the Bessel function due to errors in data acquisition and processing and insufficient power in the microtremors. Estimation is not possible in such cases. To fill the gap in the phase velocity, it may be necessary to reprocess using the data from a different time window, or re-acquire data with a different array radius.

4.2. Field study

Array measurements of microtremors was performed, between 10th and 20th of the September 2003, to estimate the V_s velocity of the soil formations in and around of Adapazarı. The research group was composed of Turkish and Japanese engineers and scientist. Hiroshi Okada from Hokkaido University and Kazuyoshi Kudo from Tokyo University were the leader of the group.

Array measurements of microtremors were carried out near the permanent strong motion observation sites, the damaged areas after 17 August and 12 November 1999 earthquakes, and all over the city of Adapazarı. In this thesis, array measurement of microtremor that was collected Toyota automobile factory (TYT), was used. The major objective of this thesis is to determine the S wave velocity structure using by SPAC method in this factory. In Figure 4.3, some photos were given from the field study.



Figure 4.3. Some photos from the field work. (a) Arrangement of the stations of smallest array (2 m). (b) Arrangement of the signal conditioner and digitizer unit for the smallest array measurement.

4.2.1. Instruments

The portable seismograph used in this study was originally developed for temporal observation of strong and weak earthquake motions (Kudo et al.,1998). It is composed of a triaxial accelerometer (Akashi Co. Ltd) of the highly damped ($h \sim 26$) moving coil type (natural frequency of 3 Hz), a signal conditioner (amplifier and filter), and a data logger (24 bits digitizer, 20 megabytes flash memory and global positioning system time synchronization; Hakusan-Kogyo Co.Ltd.). A flat response (-3 dB) of ground acceleration is attained from 0.1 Hz to an aliasing frequency. The sensitivity of the sensor is 1 V/g and optionally 5 V/g; g denotes the gravity. The clipping level of a sensor is 150 cm/sec; the maximum observable acceleration is 1g at 1 Hz or 10g at 10 Hz. The allowable input level of the data logger is selectable, 1 or 5 V. Total weight including an inner battery (2 kg) is 7 kg. The low-pass filter for cut of frequencies at 2, 5 and 30 Hz is provided. We used the filter at a cutoff frequency of 2 Hz for large arrays and one of 5 Hz for small ones, throughout microtremor observations.

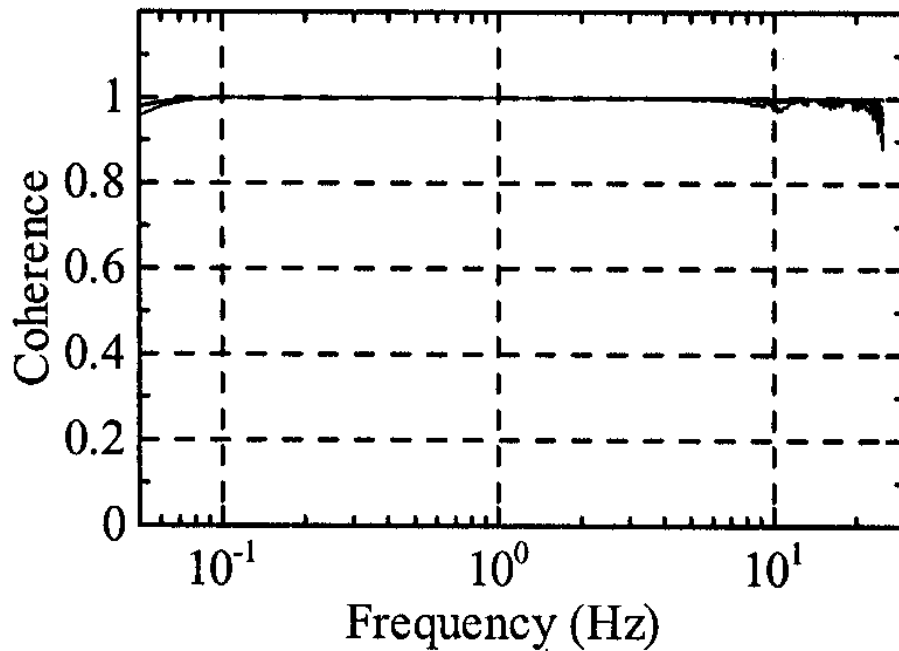


Figure 4.4. The coherences of instrumental responses used in this study.

Figure 4.4 shows the coherences of instrumental responses obtained using simultaneous recordings of microtremors at the same foundation. It is apparent that there is no use for instrumental correction in the frequency range between 0.1 and 10 Hz. In figure 4.5 field instrument photos that was used for this study was given.

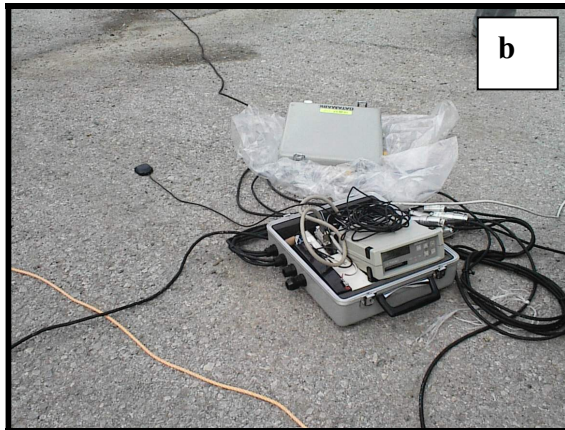
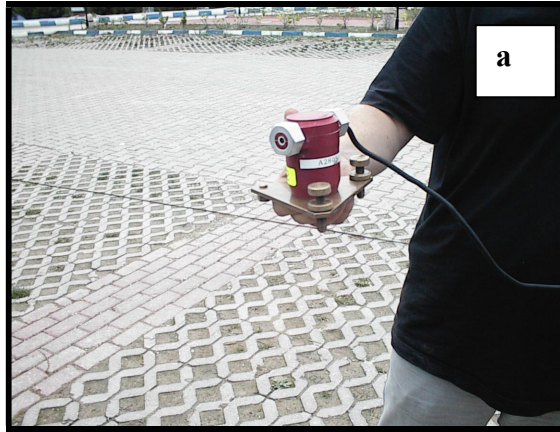


Figure 4.5. Some photos from the field instruments. (a) Triaxial accelerometer (b) Signal conditioner and data logger (c) General view of accelerometer, signal conditioner, data logger, computer and their connection.

4.3. Analysis of microtremors and determination of subsoil structure (V_s profiles)

The microtremor measurements were made on the assumption that the SPAC method is applied to determine the phase velocity dispersion curves of surface waves at the recording site. Only the vertical component data were used to estimate the SPAC coefficients, as we were interested in the Rayleigh wave part of the ambient noise recordings. The original signal trains (Figure 4.6) were divided into multiple time windows and all of them analysed separately to determine best time block that was used to calculate SPAC coefficient. The length of the selected time block varied depending on the wavelength associated with the penetration depth that would be expected. The sampling frequency of the selected time window in Figure 4.7 is 50 Hz and duration of it is 81.92 sec.

The analysis procedure to estimate the SPAC coefficients and to determine the V_s profiles is illustrated in Figures 4.8 - 4.12 for Toyota site. It is located southwest part of the Adapazarı city and compose of mostly recent alluvial deposits. In this section, to explain the analysis procedure of Toyota site, only the largest array ($r = 116$ m) data was processed and presented as an example but the other array data ($r = 58$ m, $r = 20$ m, $r = 5$ m) and the results of them can be found in appendix A.

In the first stage of processing, multiple tests were performed to check the validity of the assumption regarding the stationary nature of the microtremor recordings. If the microtremors could be considered as stationary, then they can be regarded as a stationary random function and the space correlation function between two stations, located one at the center and the other on the periphery of the circular array correspondingly, can be computed as a function of the phase velocity of the contained surface waves in the microtremors signal.

A random function can be considered as stationary when its power spectrum does not vary significantly with time and space. In the case of the present measurements the validity of the assumption in time was investigated through comparisons of the power spectra estimated for successive time window and for each station separately. The validation in space was performed through comparisons of the power spectra calculated for different stations and for the same time window.

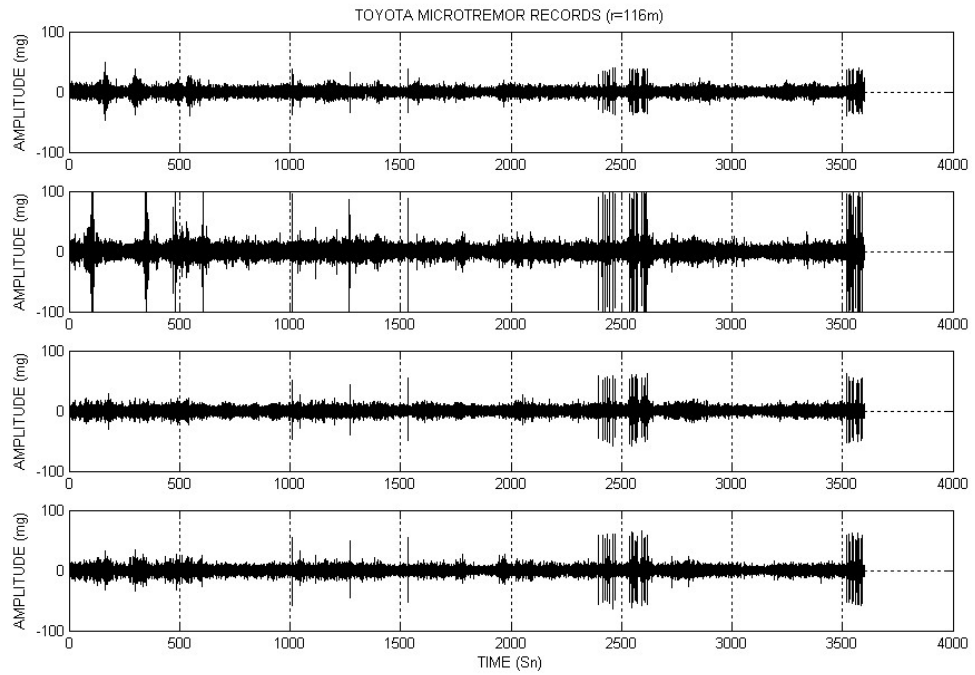


Figure 4.6. Microtremor recordings at site TYT for 60 min total continuous time duration.

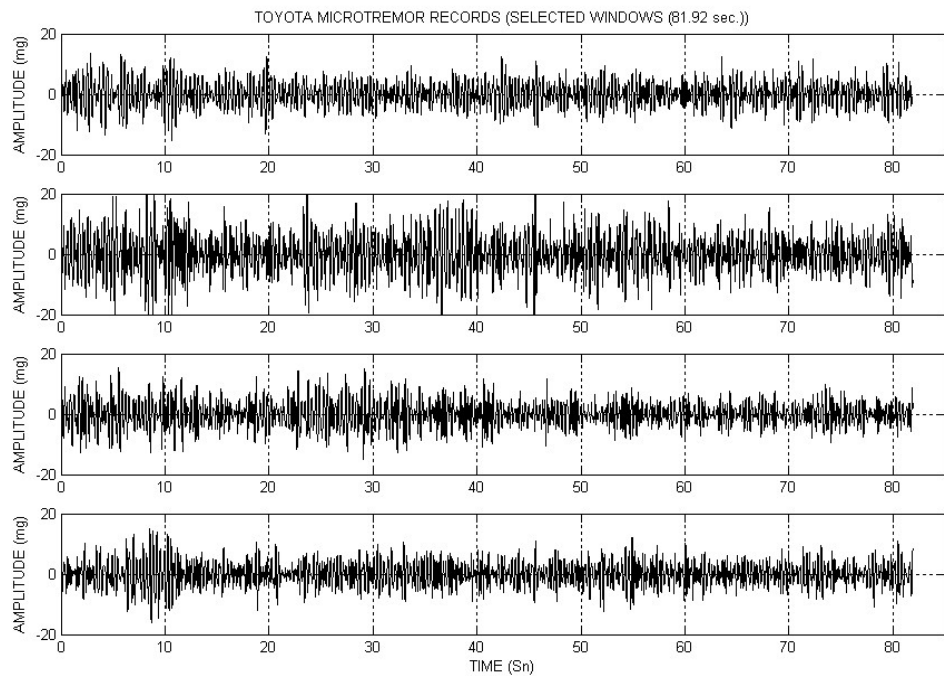


Figure 4.7. Microtremor recordings at site TYT selected time window (81.92 sec).

These comparisons revealed the frequency range in which the microtremor signal appears to be stationary and therefore can be used to estimate the phase velocities. The usable frequency ranges of interest were cross-checked through a comparison of the frequency coherence function estimated for different stations.

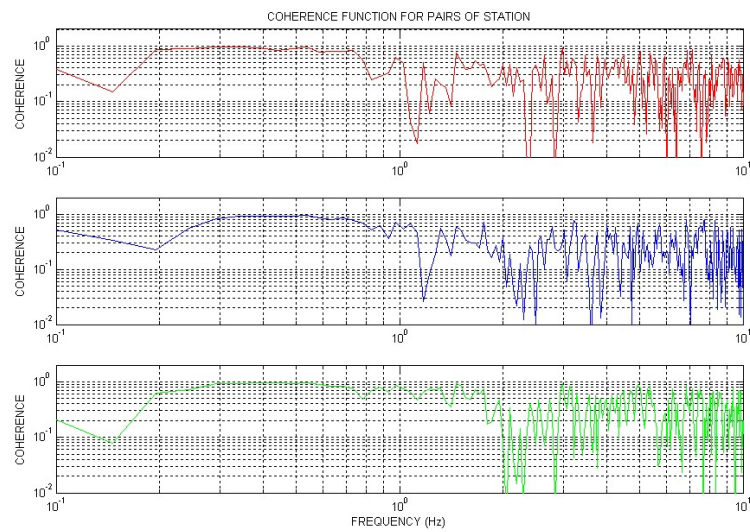
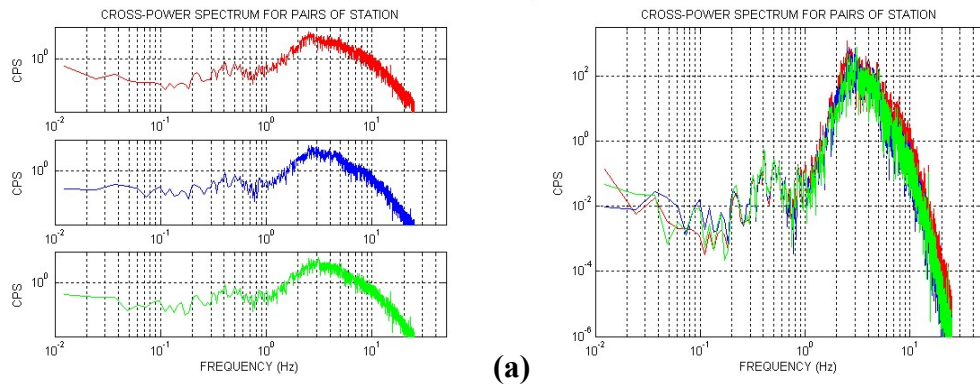
The similarly shaped power spectra calculated for one time window of the four stations (Fig. 4.8a) and their coherency functions (Fig. 4.8b) imply that the recorded ambient noise is stable in the frequency range from 0.4 to 1 Hz, which in turn is the frequency range where the SPAC method can be used for this largest array ($r=116$ m). For the calculation of other frequency ranges (above 1 Hz.), other smaller array sizes ($r=58$ m, $r=20$ m, $r=5$ m) was used (Appendix A).

In the second stage, which is the main part of the analysis, the autocorrelation functions were calculated for each pair of stations of equal inter-stations distances, r , included in the circular array. The selected frequency range for each array was subsequently used to estimate the autocorrelation functions for the corresponding radius r of the array. The autocorrelation functions were finally used to derive the SPAC coefficient for each frequency, which are related to the phase velocity.

In Figure 4.9a, and b is shown autocorrelation functions of every pair of stations and average autocorrelation functions with respect to the frequency of the corresponding array at site Toyota. In Figure 4.10 the obtained SPAC coefficients are presented, at finally selected frequencies, not only the largest array size but also the other array sizes which were used to calculate dispersion curve. Open circles correspond to the observations while solid lines are the obtained first kind zero order Bessel functions fitted to the observed SPAC coefficient, using the least squares method.

The inability to estimate SPAC coefficient at frequencies below 0.4 Hz ($T=2.5$ s) and above 5 Hz ($T=0.2$ s) is due to dimensions of the array which could not provide the appropriate wave lengths necessary to describe the particular frequency band.

The inter-distances of the station pairs used in the analysis of the other array at site Toyota automobile factory (TYT), were $r=58$ m, $r=20$ m, $r=5$ m, and the duration of the selected time windows for these array sizes were 40.96 sec and 20.48 sec (for $r=20$ m and $r=5$ m) respectively.



(b)

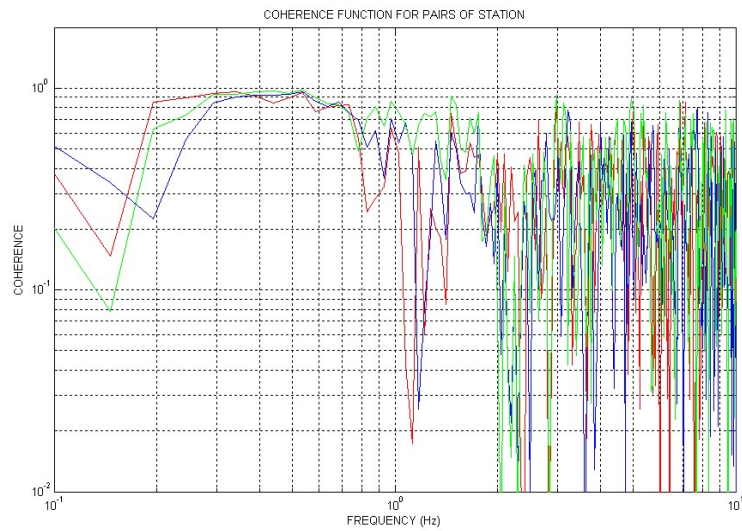
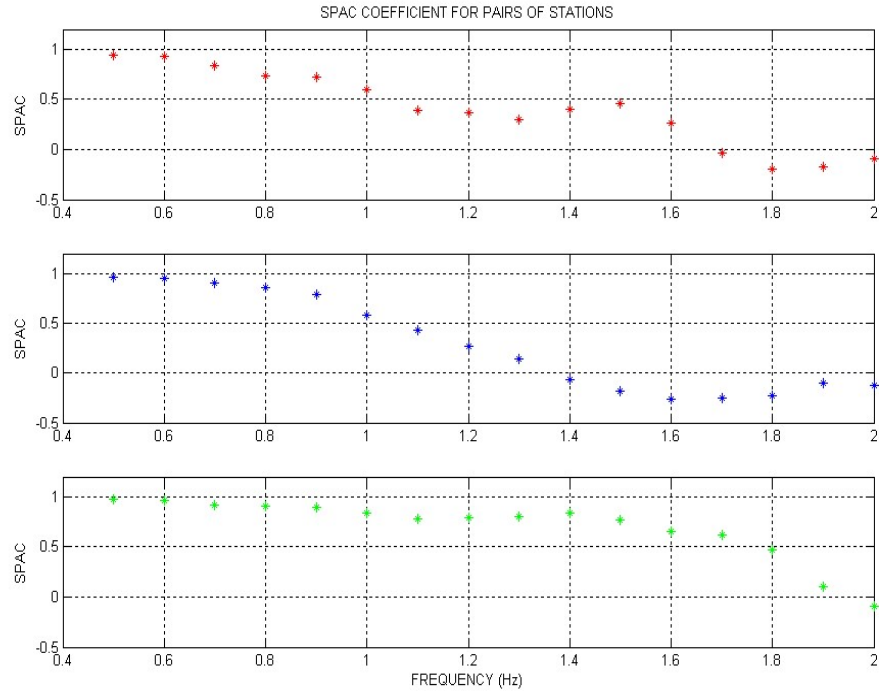
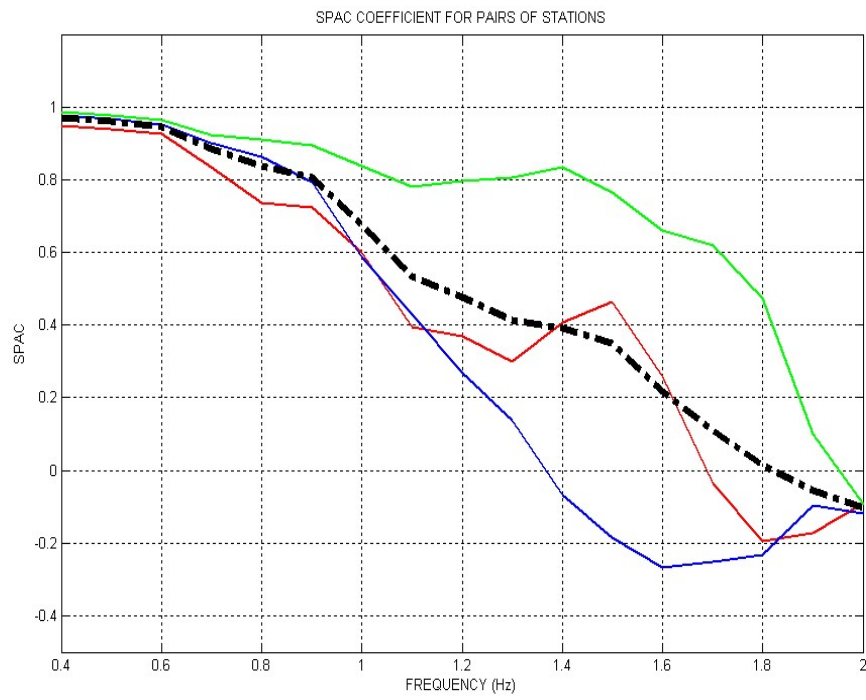


Figure 4.8. (a)Cross power spectrum of the microtremors recorded at the four stations, (b) Frequency coherence functions for pairs of stations corresponding to the distance of 116 m.



(a)



(b)

Figure 4.9. (a) SPAC coefficients of every pair of stations with inter-distance 116m, (b) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 116 m.

In the case of $r=58$ m array the exploitable frequencies vary from 0.8 Hz to 3.0 Hz, while in the case of $r=20$ m and $r=5$ m arrays the corresponding frequencies range from 2.0 to 5.0 and from 3.0 to 5.0 respectively. The results of these three arrays are presented at the same way as for the largest array ($r=116$ m), in Appendix A.

Using by these four arrays, calculated frequency range for the SPAC coefficient is changed between 0.4 Hz to 5 Hz. But, to calculate the experimental phase velocity dispersion curve of Rayleigh waves (Figure 4.11) for Toyota site, we used the frequency range from 0.9 Hz to 4.5 Hz. Because for other frequencies, calculated phase velocities from the observed SPAC coefficients is not in the acceptable error range.

4.4. Inversion procedure and inverted V_s profiles.

Engineering properties of the near surface earth have to be estimated from the experimentally measured dispersion curve. Usually this task is achieved assuming a model of horizontally stratified elastic medium. This process is named inversion. The unknowns for each layer are thickness, density, shear modules and Poisson ratio. Many parametric studies have been devoted to assess the influence of each one of these parameters (Nazarian, 1984, Sanchez, 1987). The general conclusion is that the influence of density and Poisson number is negligible, so that they can be estimated on the basis of experience without significant contribution on the final result.

Many different techniques of inversion have been proposed to obtain the soil stiffness profile starting from the dispersion curve. Recent improvements in computational capabilities have made it possible to use more accurate solutions of the forward problem of Rayleigh wave propagation, which is clearly the basis of the inversion process. Nevertheless it must be recalled that, in general, inversion is a not trivial task since it is not a mathematically well posed problem and the solution is not unique (i.e. more than one stiffness profile can correspond to nearly the same dispersion curve).

A first procedure to invert the dispersion curve is based on a simple trial-error procedure. An initial first tentative profile of the site is set, if it is the case by using a priori information from previous geotechnical tests. The forward problem of wave propagation is then resolved and the computed dispersion curve is compared to the

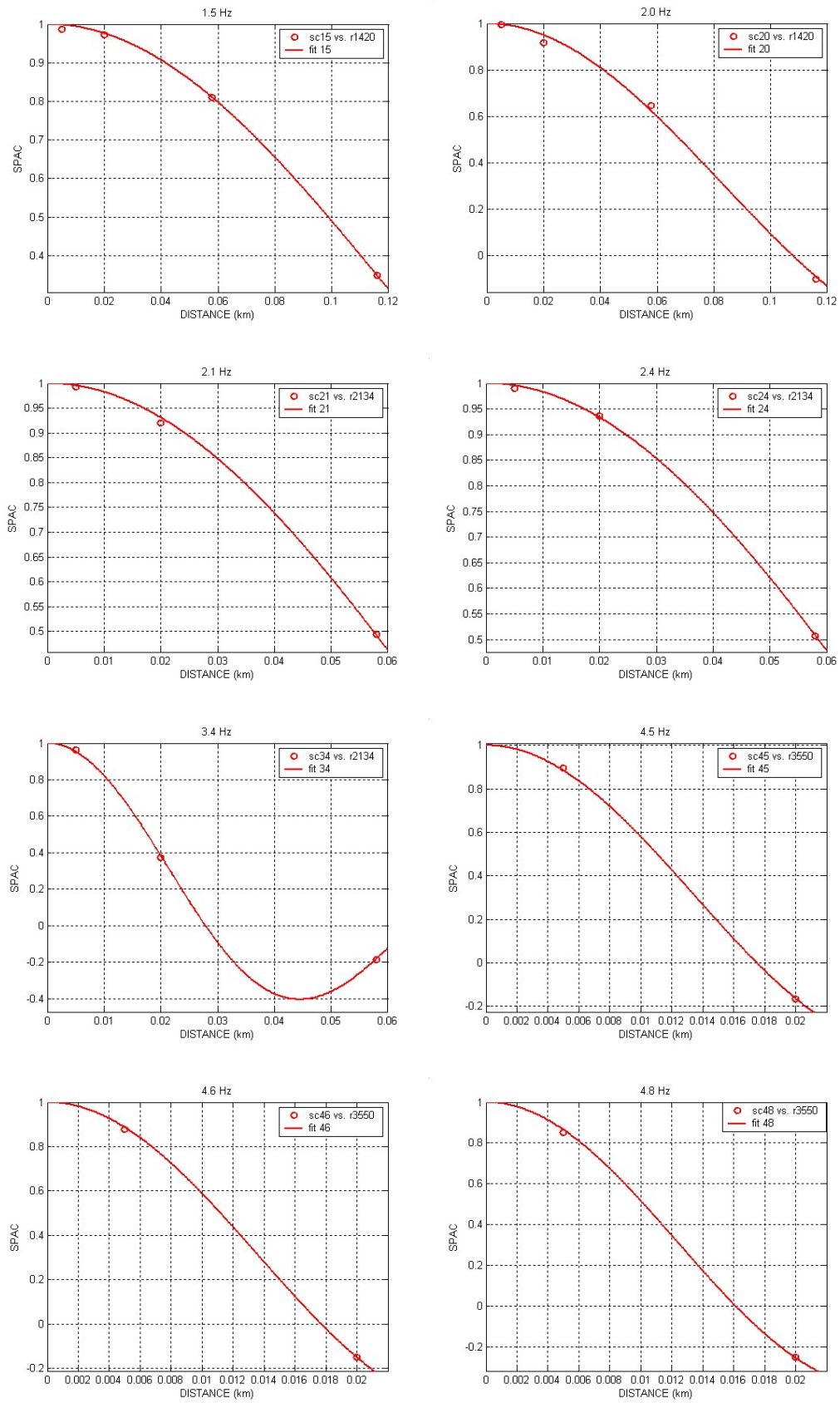


Figure 4.10. Observed SPAC coefficients (circles) and Bessel functions fitted to them (continuous line) for representative frequencies 1.5, 2.0, 2.1, 2.4, 3.4, 4.5, 4.6 and 4.8 Hz.

experimental one. In the successive step some parameters of the initial model are changed in the view of improving the fitting between numerical and experimental data. The judgment about the effectiveness of the fitting is usually done by a least-square criterion and by visual inspection. The process follows an iterative scheme and eventually it converges to an acceptable solution.

In this thesis the experimental dispersion curve (open circles) defined in Figure 4.11 was inverted using an iterative inversion scheme introduced by Hermann and Amon (2002) to reveal the V_s velocity structure below the circular arrays. In Figure 4.11 blue solid line shows the theoretical dispersion curve that was calculated using by SURF96 software package. This employed software has been developed for the surface wave recorded during P-wave refraction measurements, which is being applied recently in earthquake engineering. This inversion procedure has been widely applied and has given reliable results in many studies conducted in several areas of the world (Apostolidis et al, 2004).

For determination of the V_s models a complete initial soil model with thickness, V_p and V_s velocities, and density for each layer to the half-space is needed. When, after using an iterative procedure, the theoretical dispersion curve matches well the experimental one, and then it is considered that the assumed model describes well the soil stratigraphy and the velocity profile.

Figure 4.11 shows the observed dispersion curve (red circles), theoretical dispersion curve (blue solid line) of phase velocity of the Rayleigh wave recorded in the microtremor measurements at the site Toyota and initial dispersion curve for the inversion process (green dash-dot line). The inverted (V_s) soil profile (blue solid line) calculated by the experimental dispersion curve and initial model (green dash-dot line) for iterative inversion process was given in Figure 4.12.

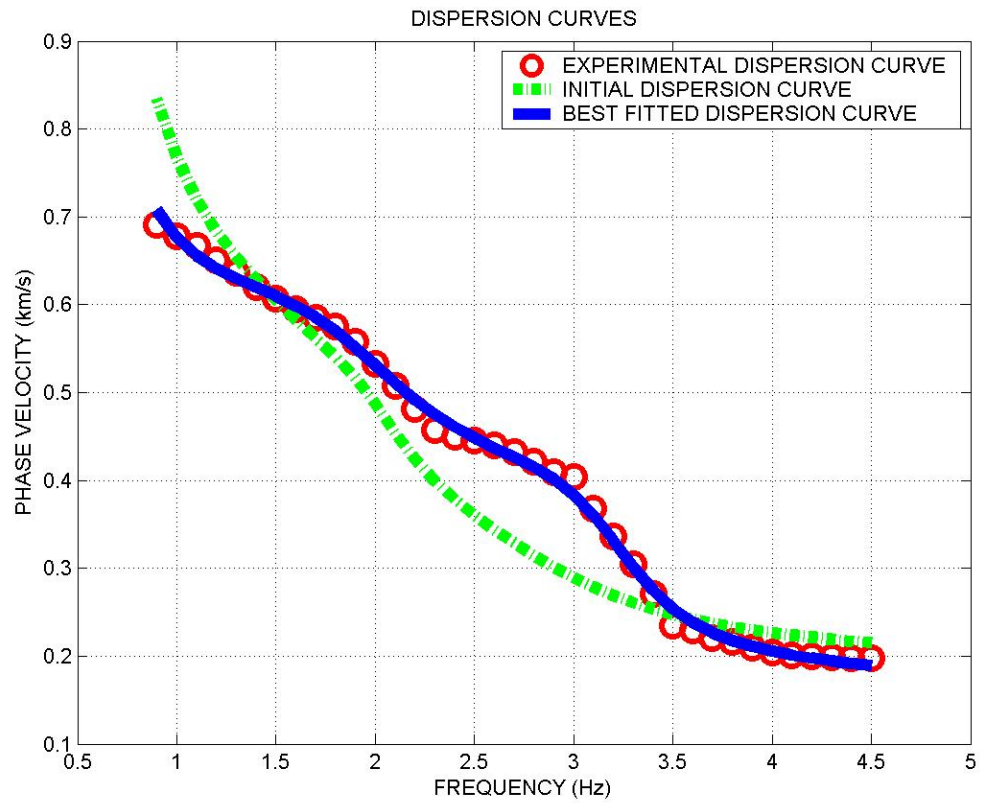


Figure 4.11. Experimental dispersion curve (red circles), initial dispersion curve for inversion process (green dash-dot line) and best fitted dispersion curve (blue solid line) after the inversion process.

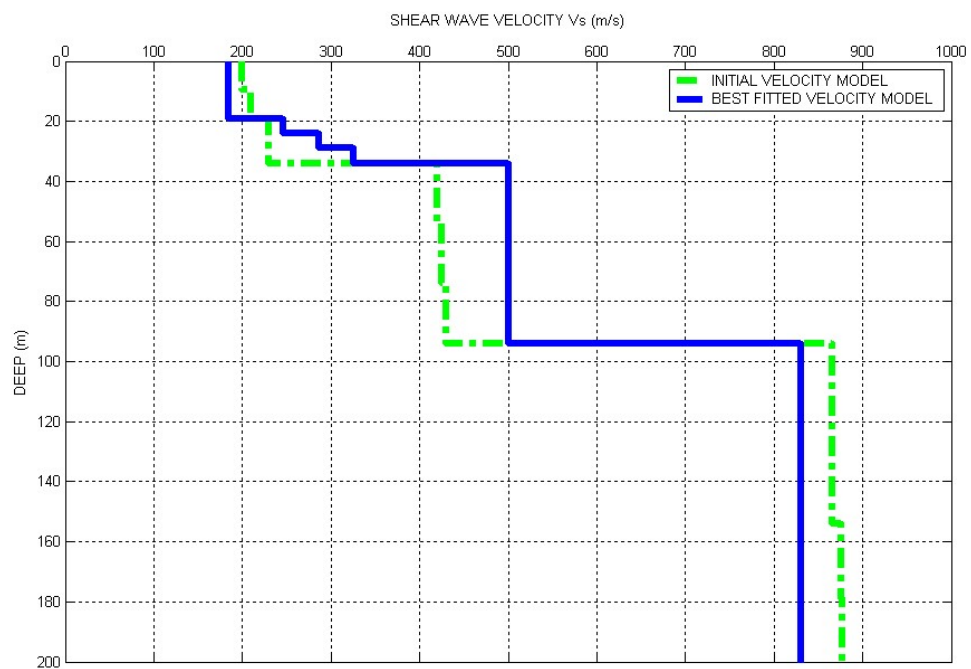


Figure 4.12. Final inverted V_s profile (blue solid line) calculated by the experimental dispersion curve and initial model (green dash-dot line).

5. DISCUSSION OF THE RESULTS AND CONCLUSION

The array observation of microtremors is very promising method for determining S wave velocity structures from shallow to intermediate depths, and it can be conducted with fewer restrictions in urbanized areas. This method relies on two assumption. These are; the property of seismic surface waves that their penetration into the earth is frequency dependent hence the dispersion curve for observed data can be inverted to yield a layered earth model of the subsurface, and the variation of phase velocity with frequency can be measured using array processing methods.

The results presented herein were obtained mainly from array ambient noise (microtremor) measurements at Toyota automobile factory and analyzed using the SPAC method to estimate V_s velocity of soil formations.

Estimated shear wave velocity (V_s) profile is used to assign Uniform Building Code (UBC, 1997) site classification. The Uniform Building Code classification system is based on the average shear wave velocity calculated using the travel time of a shear wave.

The UBC site classification system include hard rock (S_A , V_s greater than 1500 m/s), rock (S_B , V_s between 760 and 1500 m/s), very dense soil and soft rock (S_C , V_s between 360 and 760 m/s), stiff soil (S_D , V_s between 180 and 360 m/s), and soft soil (S_E , V_s less than 180 m/s). A site also can be classified as soft soil if more than 3 m of soft clay is present.

According to this soil classification system and V_s velocity structure that was obtained from the Toyota automobile factory, site classification can be made as follows; the top 19 m was classified soft soil (S_E). From 19 m to 34 m was classified stiff soil (S_D), from 34 m to 94 m was classified very dense soil and soft rock (S_C) and deeper than 94 m was classified rock (S_B). Shear wave velocity profile and soil

classification obtained in this study has a good agreement with other studies that was conducted for similar purposes in the literature.

The analysis of microtremors recordings at Toyota site led to some important conclusions concerning the stationary character and efficiency of the SPAC method. Comparing the power spectra at all different array sizes, it is concluded that within a large city for interstation distances up to 100 m and recording time longer than 30 min, the microtremor can be considered as random stationary function. This fact, which allows to use noise to define V_s profile. It is also important to note that although in some cases the instruments were installed close to areas of intense human activity, the stationary character of recorded noise was not really affected. This means that distances of 50-100 m, which can be easily found even in densely urbanized areas, are not inadequate to provide array ambient noise measurements of good quality.

The most significant advantage of the MSM, which stems from the present study, is that it allows the reliable determination of V_s velocity profiles down to large depths with relatively small apertures of the deployed arrays. This is significant for accurate soil response studies regarding large cities, where open free spaces suitable for the deployment of large arrays are difficult to be found and high energy sources can not be easily accepted.

Compared to the conventional geophysical exploration methods, such as reflection surveys or cross-hole testing, the microtremor array measurements does not require boreholes or any kind of artificial source and therefore has the advantage of exploring large depths with relatively low cost, and less field effort.

The only significant limitation of the method with respect to the array aperture is the condition of horizontal and flat layering below the array. Research in the literature show that at sites where this assumption is probably violated from lateral discontinuities then the possibility to determine V_s values is limited only for surface soil formations. In this case it is difficult to extract accurate information on the total thickness of soil deposits and consequently the depth and velocity of the bedrock. As a result, at sites presenting previously mentioned conditions, microtremor survey method can be used to investigate only the near surface soil formations and obviously a complementary conventional geophysical survey will be necessary, especially for the determination of the depth and the velocity of the underlying rock basement.

To this end, the MSM can be considered as a valuable complementary or alternative tool for determining V_s velocity and soil stratigraphy, which are useful in soil and site characterization, site response analysis, microzonation studies design of infrastructures and urban planning.

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APPENDIX A

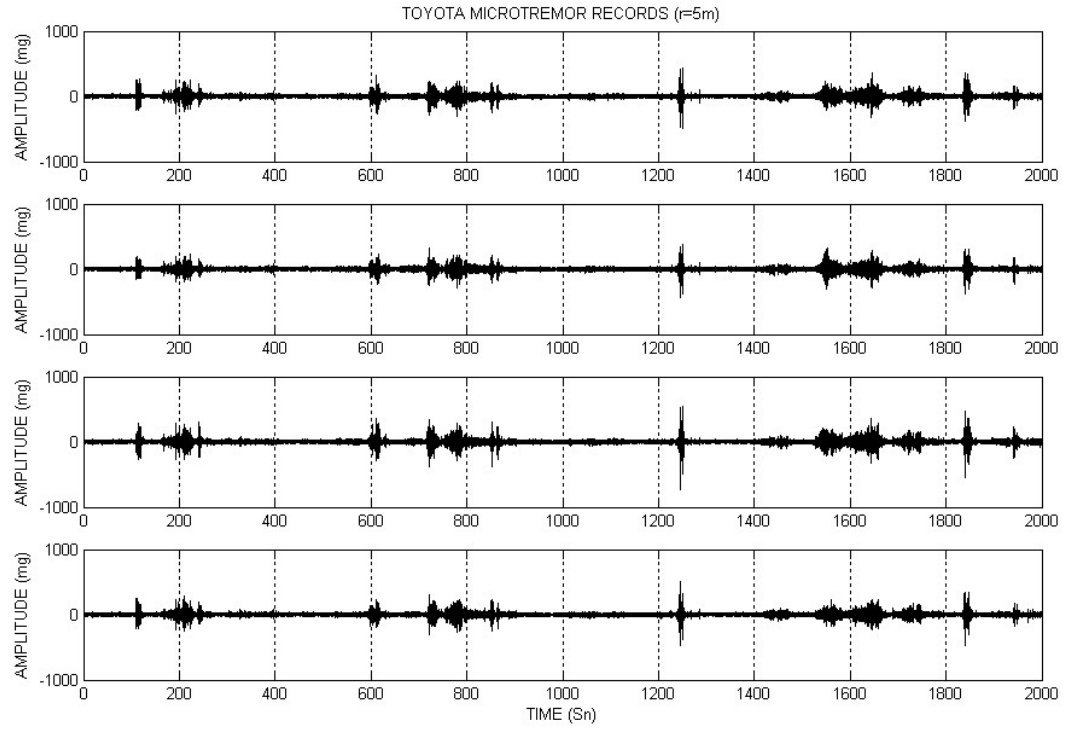


Figure A.1. Microtremor recordings at site TYT for 35 min total continuous time duration ($r = 5$ m).

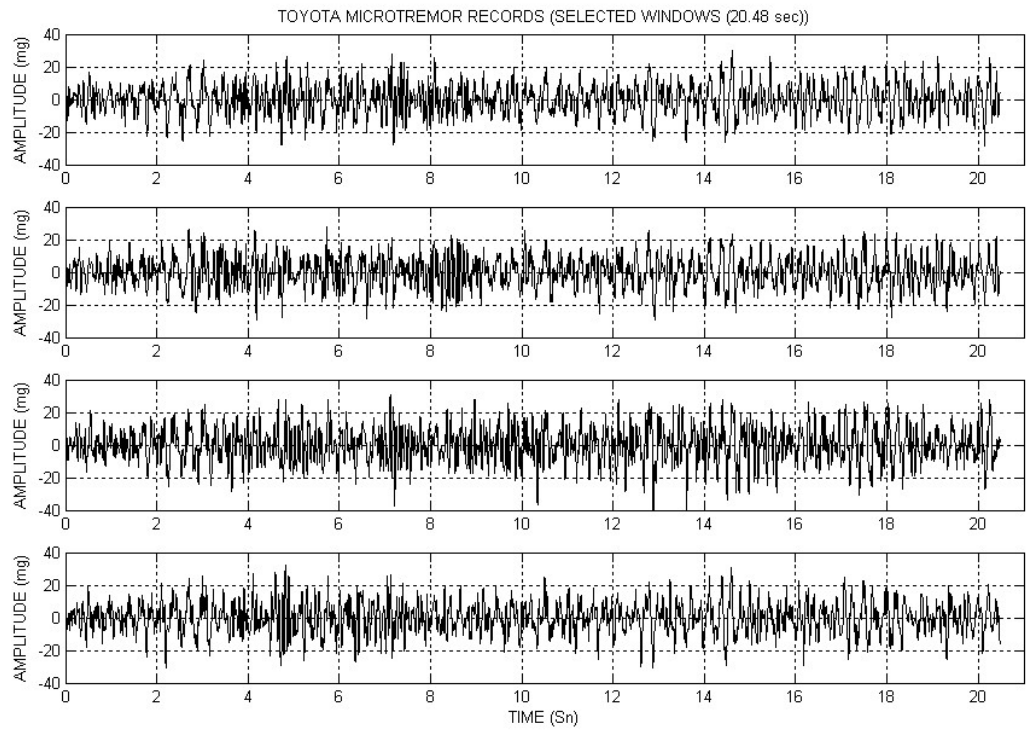


Figure A.2. Microtremor recordings at site TYT selected (20.48 sec) time window ($r = 5$ m).

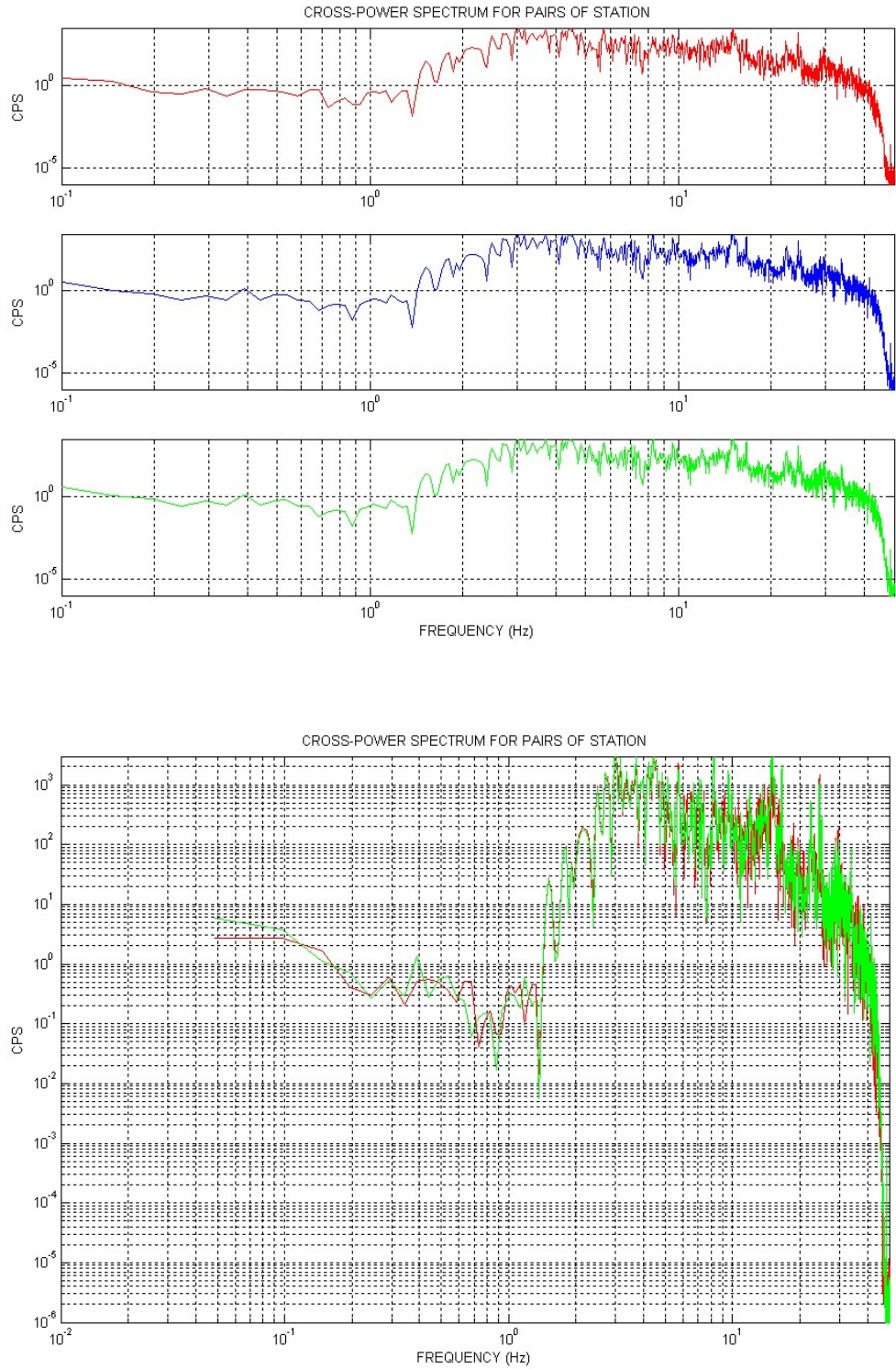
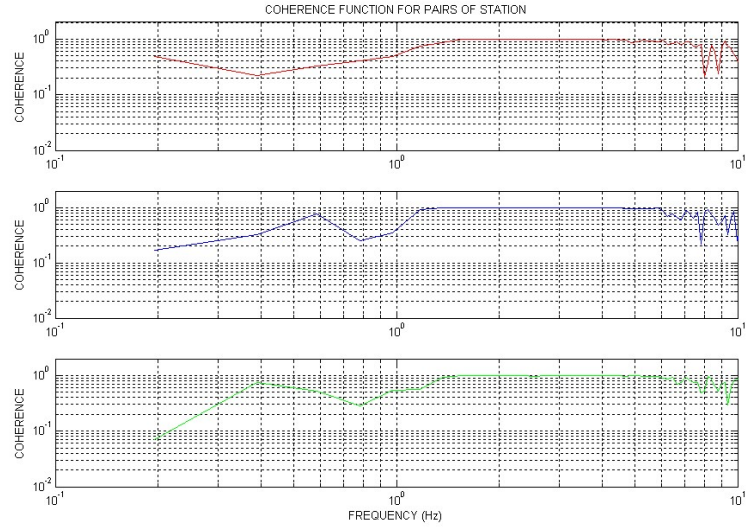
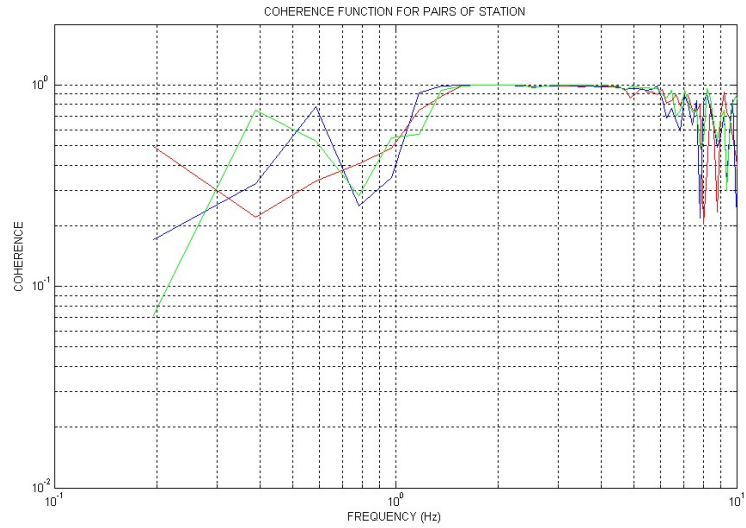


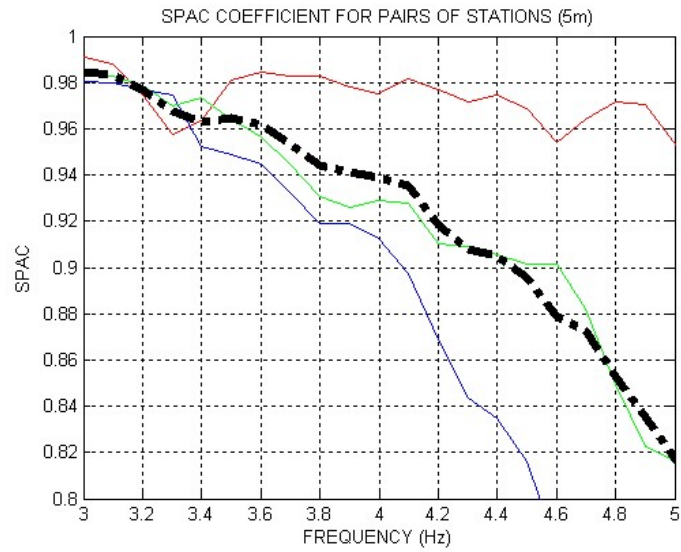
Figure A:3. Cross power spectrum of the microtremors recorded at the four stations ($r = 5$ m).



(a)



(b)



(c)

Figure A.4. (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 5 m (c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 5 m.

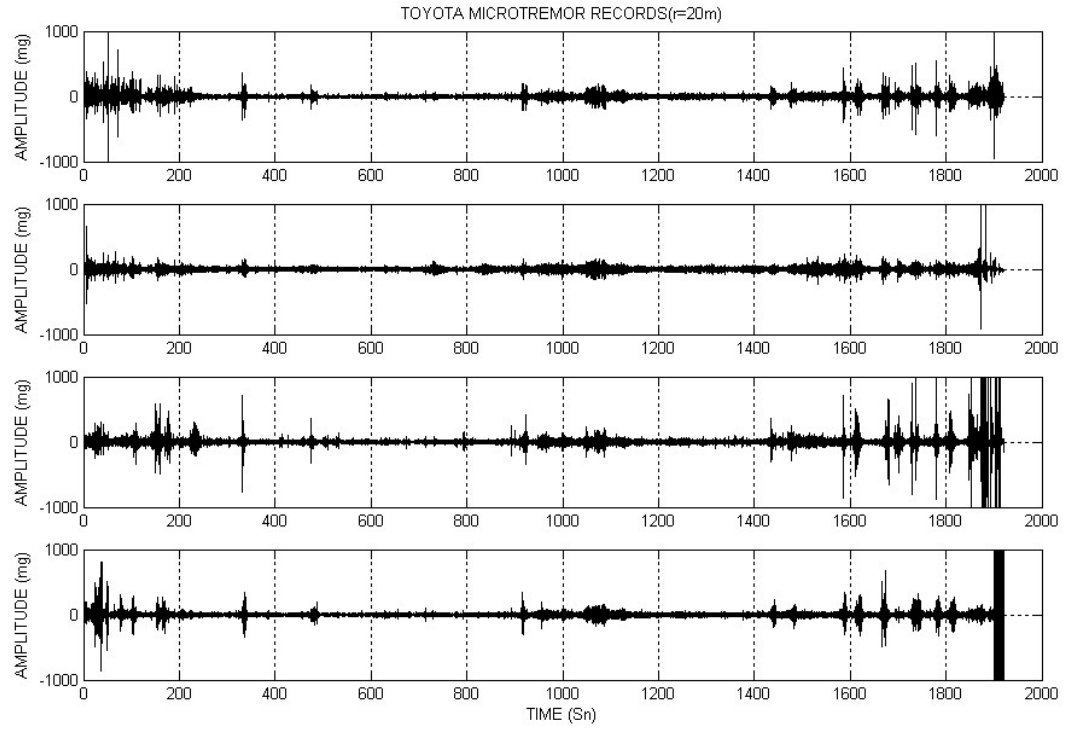


Figure A5. Microtremor recordings at site TYT for 30 min total continuous time duration ($r = 20$ m).

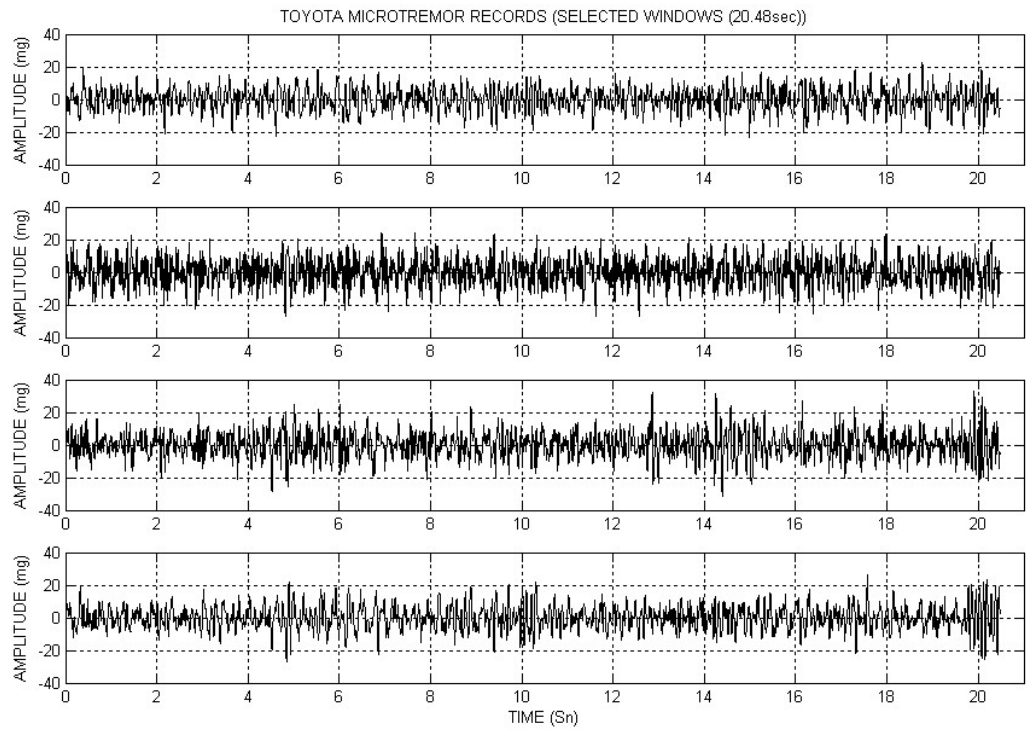


Figure A.6. Microtremor recordings at site TYT selected (20.48 sec) time window ($r = 20$ m).

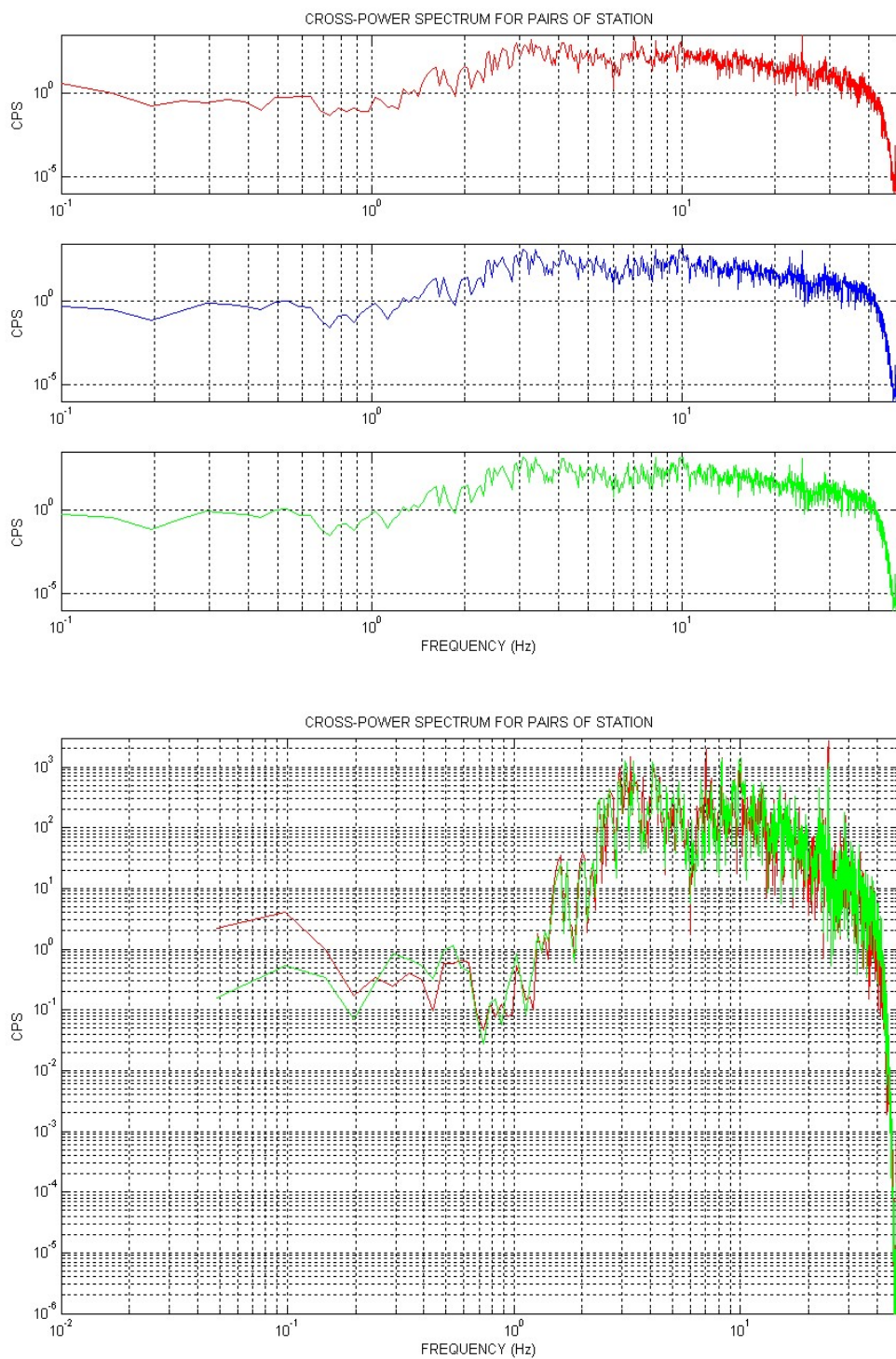
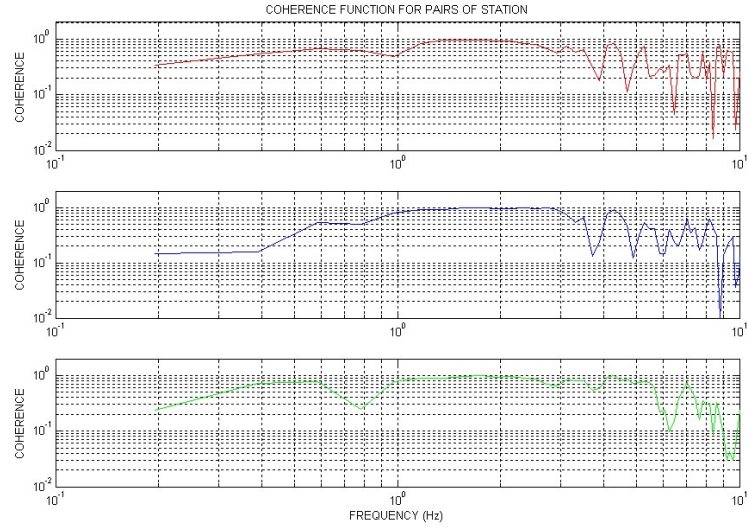
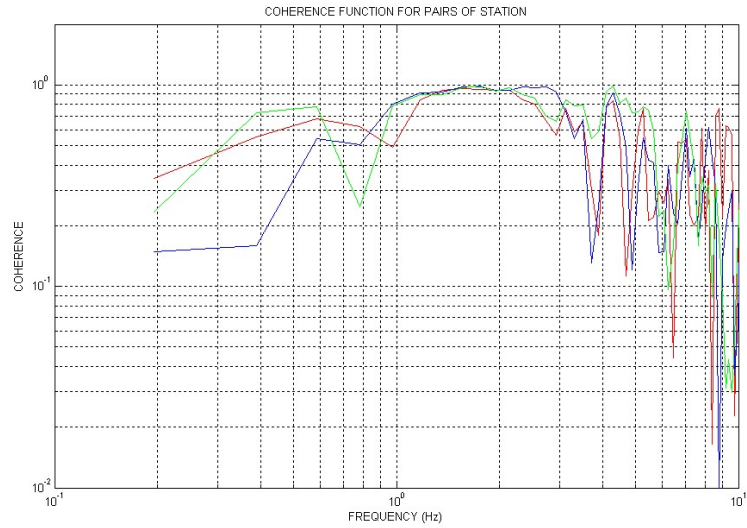


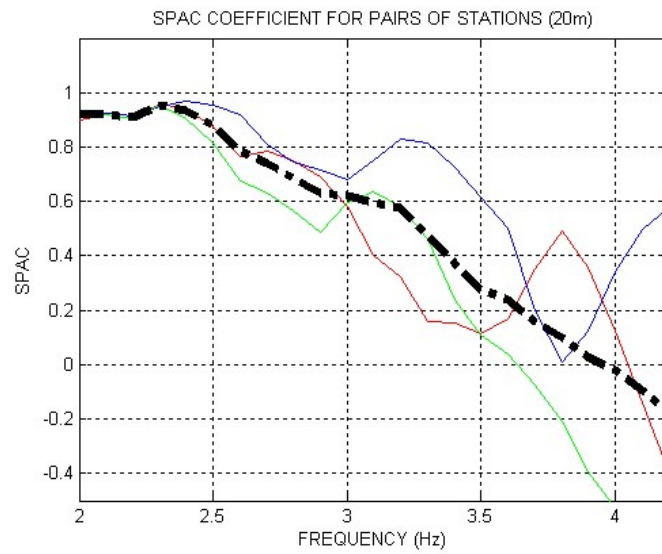
Figure A.7. Cross power spectrum of the microtremors recorded at the four stations ($r = 20$ m).



(a)



(b)



(c)

Figure A.8. (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 20 m (c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 20 m.

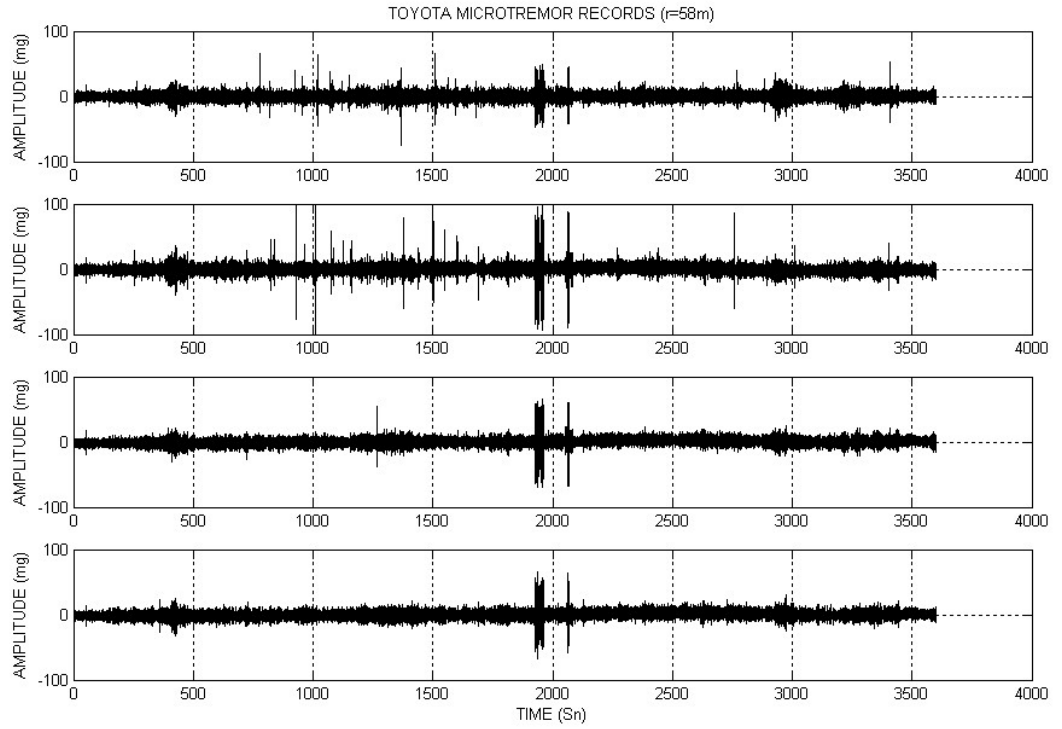


Figure A.9. Microtremor recordings at site TYT for 60 min total continuous time duration ($r = 58$ m).

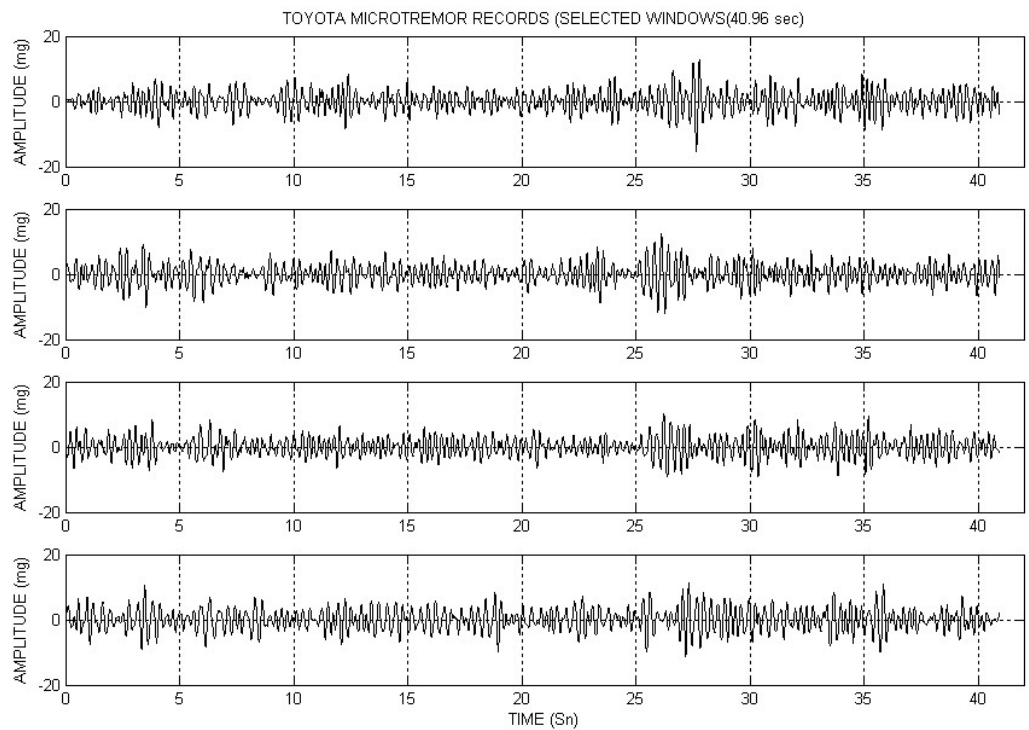


Figure A.10. Microtremor recordings at site TYT selected (40.96 sec) time window ($r = 58$ m).

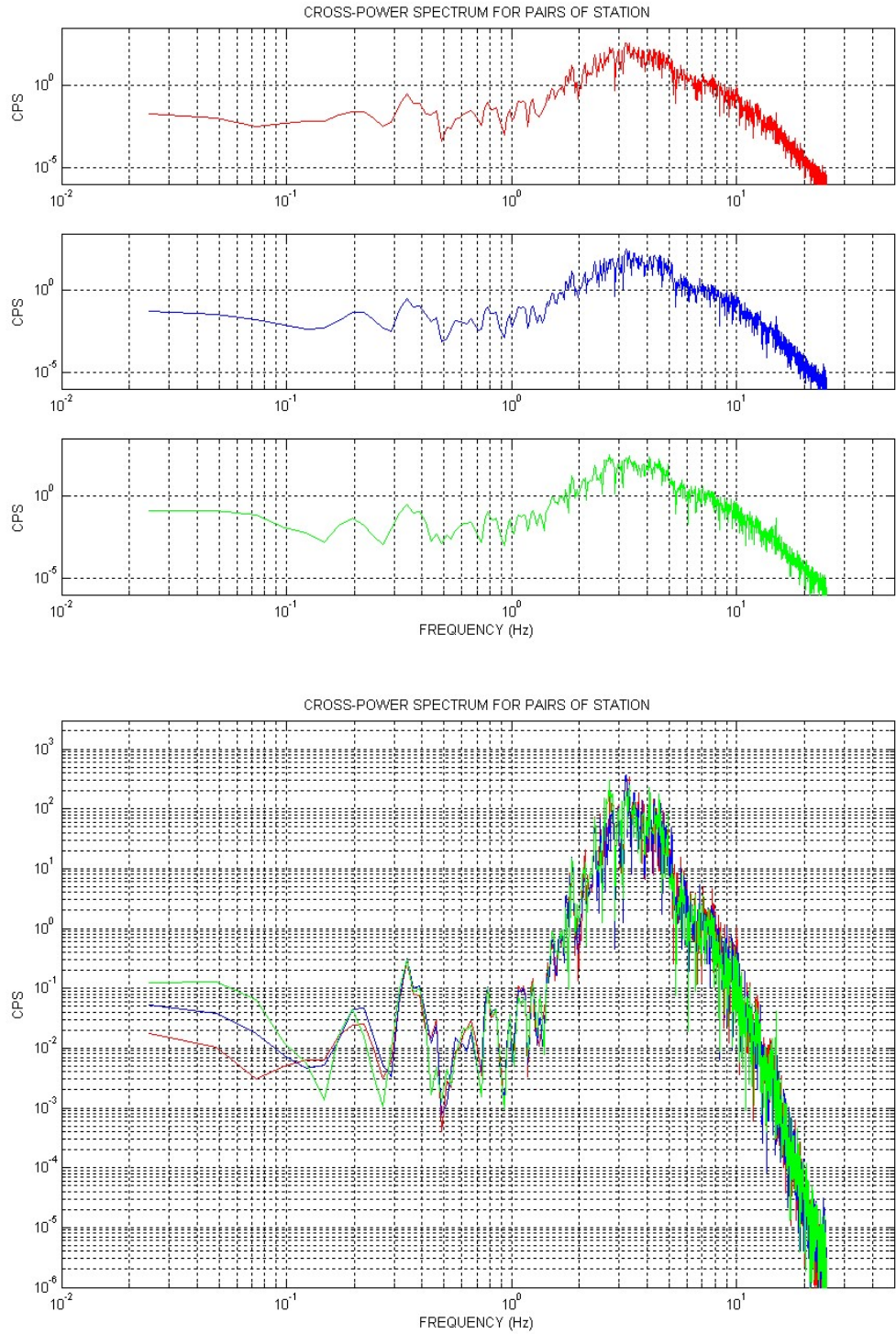
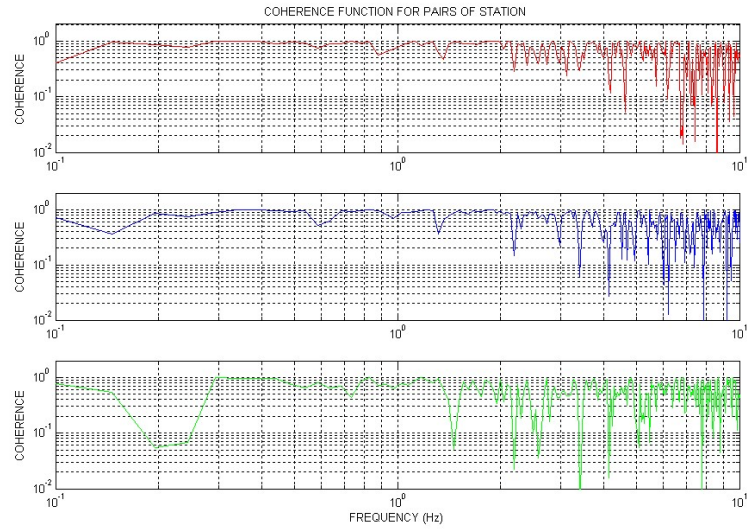
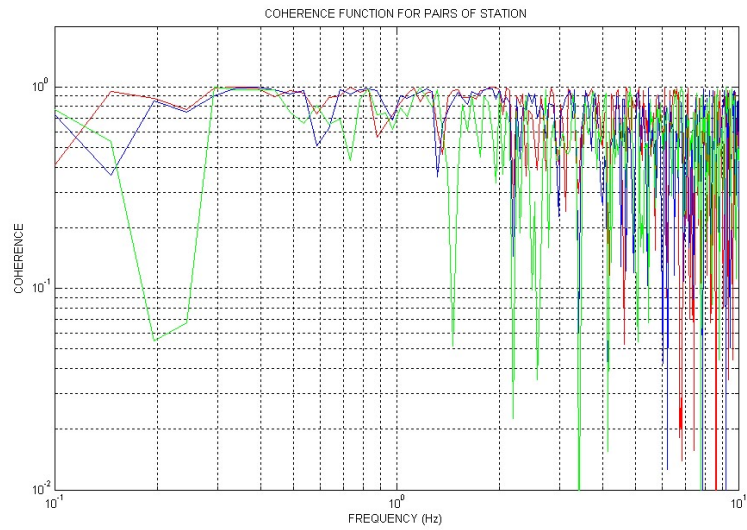


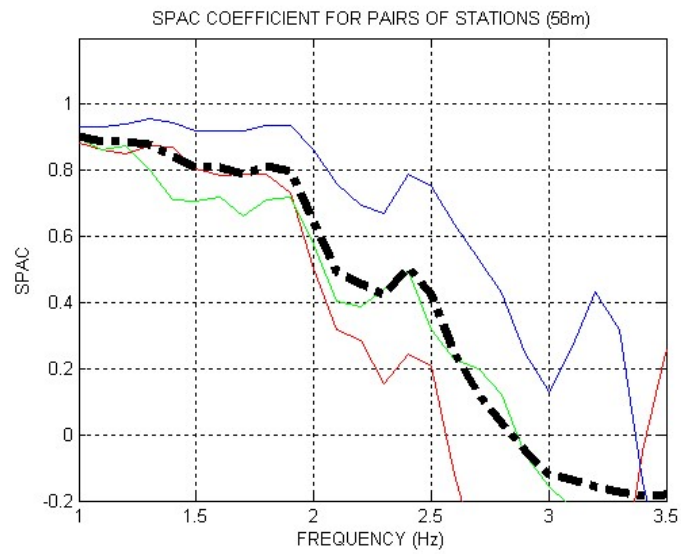
Figure A.11. Cross power spectrum of the microtremors recorded at the four stations ($r = 58$ m).



(a)



(b)



(c)

Figure A.12. (a,b) Frequency coherence functions for pairs of stations corresponding to the distance of 58 m (c) SPAC coefficients (solid lines) and their average (dotted lines) of every pair of stations with inter-distance of 58 m.

RESUME

Aydın Mert was born in İstanbul on January 28, 1973. He finished his high school education at Plevne Lisesi, İstanbul in 1989. After Receiving his Bachelor of Science degree in Geophysical Engineering from the İstanbul University in June 1993, he joined Kandilli Observatory and Earthquake Research Institute of the Boğaziçi University in the same year as a research engineer. At the end of 2001 he applied for MSc programme in Eurasia Institute of Earth Science at İstanbul Technical University. He attended a three months course at University St. Cyril and Methodius in Skopje, Macedonia during his MSc studies. His main research theme has been surface wave propagation and soil characterization. He is married and has a one-year old son.