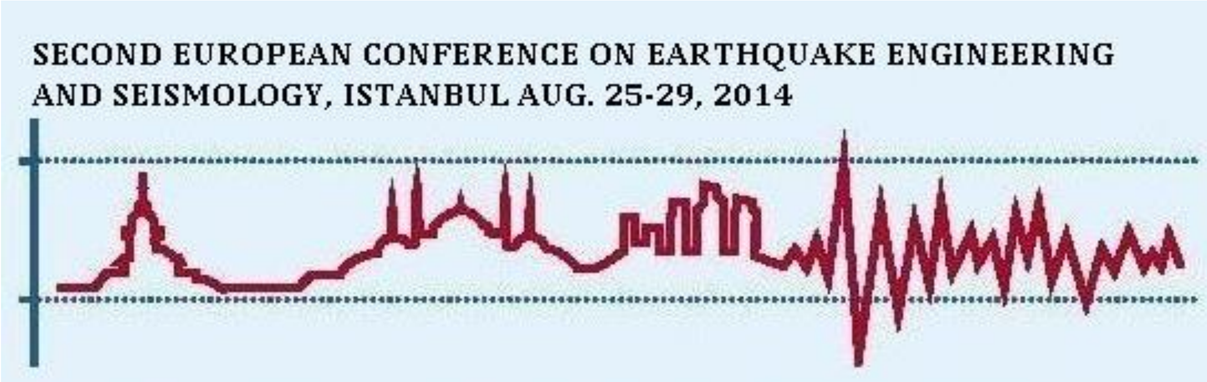

DETERMINATION OF DEEP SHEAR WAVES VELOCITY STRUCTURE USING COMBINATION OF DISPERSION CURVES FROM CROSS-CORRELATION AND ARRAY OF AMBIENT NOISES RECORDING

Mohsen FAZLAVI[1], Mojgan NOROOZI[2], Ebrahim HAGHSHENAS[3] Hossein PEZESHKI[4]

ABSTRACT

In this paper two approaches applied (cross-correlation and Frequency-wavenumber) to calculate velocity dispersion curves. Due to distance between stations, group velocity dispersion curve from cross-correlation method is only valid in low frequency range (0.1 to 0.5 Hz) against phase velocity from Frequency in higher frequency (1to 15Hz). 4 month continuous microtremor data for 5 stations were used in Cross-correlation method to extract Green function of alluvium between each pair of stations. Also the group velocities as a function of period are calculated using traditional frequency-time analysis (FTAN). On the other hands, 5 array with 12 broadband seismograph (0.1-50 Hz) arranged near network station. Phase velocity dispesion calculated with frequency-wavenumber analysis. İn this paper we peformed jointed inversion processing with two type dispesion curve for each station. Shear wave velocity near the mountaian indicated shallow contrast (to 3000 m/s, less than 200 meter) contrary the south of Tehran which shows larger depth for contrast (between 800 to 1200 meter), compatible with geological information of each site in Teharn. Jointing two dispersion curve take a reliable shear wave velocity in wide depth range, surface to 3km.

Keyword: cross-correlation, frequency- wavenumber, microtremor, Tehran, shear wave velocity

INTRODUCTION

Tehran, Iran's capital with a population more than 10 million is located in the foothill of Alborz mountains belt. This region is dominated by one of the most seismically active zones with four main faults, Mosha, North of Tehran, Rey and Kahrizak, were pointed surrounding of Tehran (Berberian et al, 1985) and a lot of small length faults were maped especially in north of Tehran (Abbasi et al, 1999)(figure 1). Due to great potential of earthquake, in recent decade, a series of comprehensive studies have been conducted, extracting deep shear wave velocity profile of Tehran Alluviums using event and ambient seismic noises. However earthquake records have a lot of energy to pass along deep layers between both stations of seismic network, but it is a probabilistic phenomenon. On the other hands, recording ambient seismic noises are the stationary data with low amplitude, could be used as an alternative approach, generally produce by different sources.

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According to previous studies (Paul & Campillo 2003, Sabra et al 2004, Shapiro & Campillo 2004, Campillo et al 2005), the Green Function between two stations can be obtained by calculating the cross correlation of noise records. Also, Snieder et al (2010) demonstrate the surface waves are predominant in the resulted Green Function. Shomali et al (2013) and Motaghi et al (2012) using tomography method for 1 year earthquake records and microtremors of TDMMO network in surroundings of Tehran, presented average dispersion curves of each direction between 0.2-0.5 Hz. Frequency-wavenumber method, based on coherency of traveled signals between two stations were applied for frequencies higher than 1 Hz. By arranging array of seismometers with aperture less than 500 to 1000 meters, firstly presented by Capon, 1969. However, dispersion curves between 1 to 15 Hz would be extracted with frequency-wave number analysis, but they are not sufficient for inverting. Actually here, we combined dispersion curves for each station from Frequency-wave number and cross-correlation analysis which illustrates the vast range of dispersion curve (0.2-15 Hz).

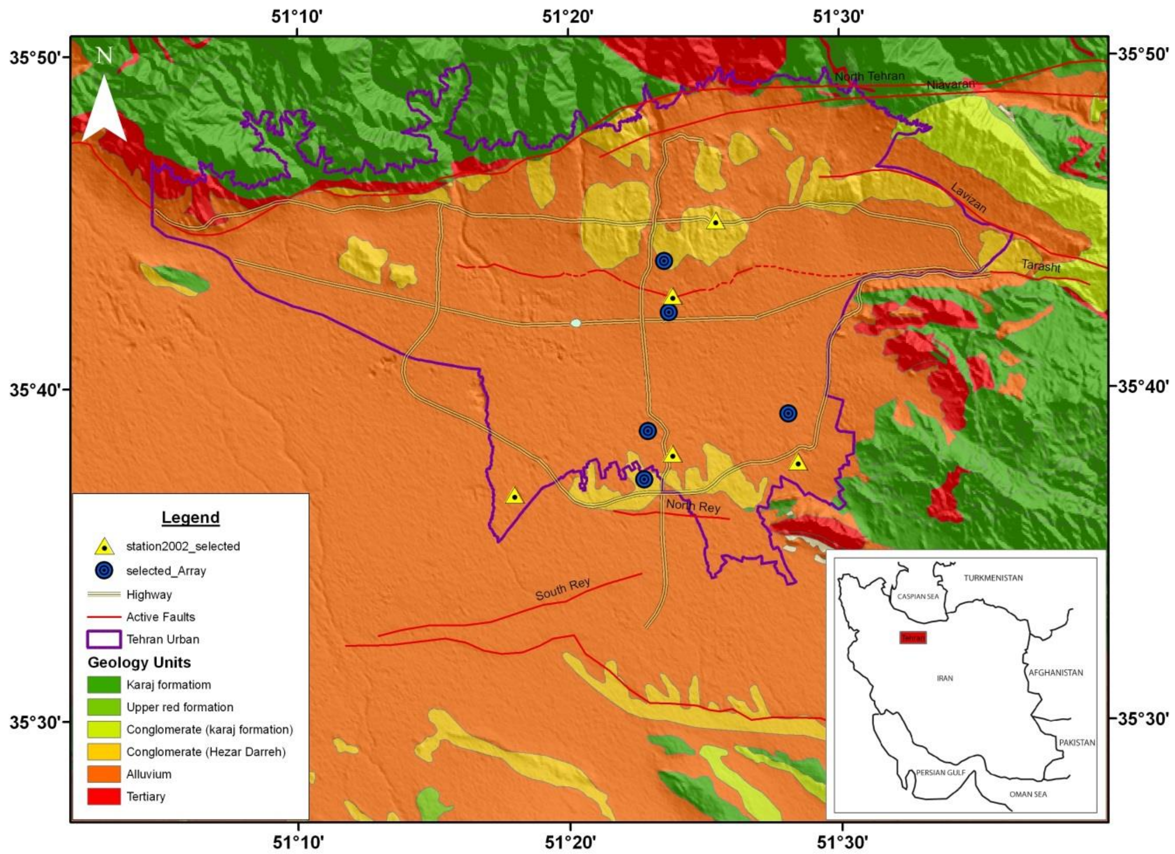

Figure1. Geological map of Tehran. Yellow triangular indicated 4 month temporary network stations and blue circles shown 5 array stations in Tehran.

In this study we try to perform joint inversion of two different dispersion curves extracted from cross-correlation and frequency-wavenumber methods. The goal of the present study was the assessment of the ability of this technique in determination of deep shear wave velocity of alluvial deposits.

DATA ACQUISITION

To analyze cross-correlation method, the continuous data recorded in 5 seismological stations network installed in Tehran during 4 months with distance between 4-9 Km were used. The recorded data of vertical components were used to calculate the cross correlation. By calculating the cross-correlation

on vertical components, Rayleigh wave propagation between two stations could be reconstructed (figure 2). The raw data used for cross-correlation technique had recorded for 4 months (from March to August 2002) by Guralp devices (CMG40) with 100 samples per second.
On the other hands, 5 array of seismograph with aperture between 50 to 1000 meters, were be arranged near network stations (figure 2). These array measurements performed in 2011 using 12 CMG6TD Guralp seismometers. All broadband sensors had been calibrated and are reliable for frequencies from 0.1 to 50 Hz. Ambient seismic noise was recorded at 100 Hz sampling rate. Some of these arrays consist of 2 to 3 concentric circles installed simultaneously or arranged separately with different apertures vary from 50 to 1000 meters. Given the number of available stations, i.e. the range of inter-station distances, this configuration is a good compromise between sufficient resolutions at measurement depths of investigation. The absolute times of recordings were provided by GPS antenna connected to each station to synchronize all recorders to GMT. If the locks to the time signal received by the GPS antenna, then common time for all the records is guaranteed. CMG40 and CMG6TD recorded seismic data by three components but in this paper we use only correlation of vertical component.

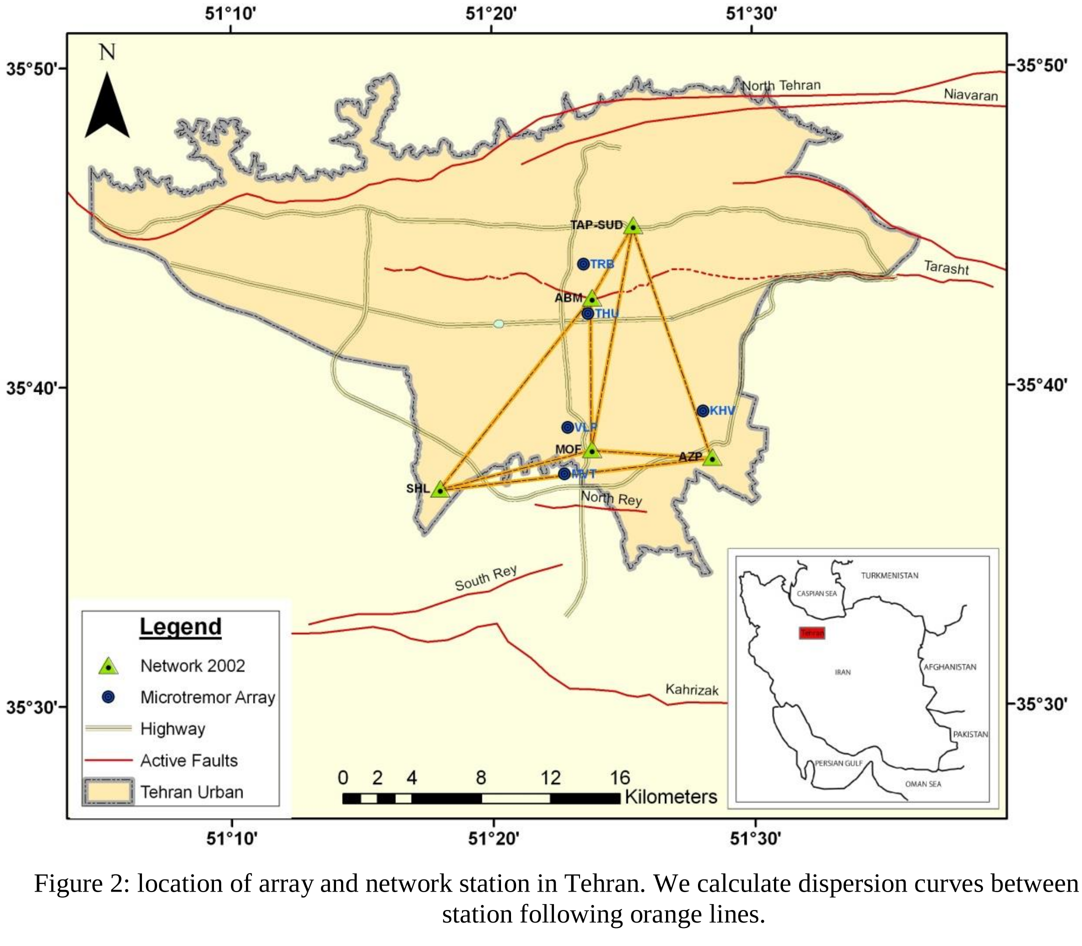

Figure 2: location of array and network station in Tehran. We calculate dispersion curves between network station following orange lines.

THEORY OF CROSS-CORRELATION & FREQUENCY-WAVENUMBER METHOD

Cross-correlation technique

Ambient seismic noise is composed of seismic waves with random amplitude and phase that can propagate in all directions, independent of source. Thus these waves have valuable information about the propagation path which is not easily obtained in the other ways (Shapiro & Campillo 2004). This is the reason of the studies growth on ambient seismic noise topic in the recent years.
In wave physics, it is important that Green function of the sublayers is investigated. The Green function between two points A and B is determined by a direct pulse measurement and gives information of the propagation path and consists of both body and surface waves (Sabra et al. 2005).

Recent theoretical and laboratory studies have demonstrated that the cross-correlation function of the random wavefields computed between a pair of receivers is included the Green function between them (Weaver & Lobkis 2001; Lobkis & Weaver 2001; Derode et al. 2003; Snider 2004; Larose et al. 2004; Sanchez-Sesma et al.2006a). In seismology, Seismic observations based on cross-correlation between a pair of station have confirmed for surface waves using coda waves (Campillo & Paul 2003; Paul et al. 2005), long ambient noise (Shapiro & Campillo 2004, Sabra et al. 2005a), for crustal body waves (Roux et al. 2005) and for oceanic applications (Lin et al. 2006). It is commonly thought that Green function estimates obtained from cross-correlation calculation accurately represent the full Green function, but this is not always available for all cases. Forghani & Snider (2010) reviewed that the estimated Green function is predominantly including the surface waves. Constantly the analysis of surface waves, dispersion measurements is performed and usually is used for tomography or velocity structure of the ground.

Derode et al (2003) proved that the Green function between A and B points is extracted from cross-correlation calculation; if C was a source,φ_A and φ_B are the wave fields recorded in A and B:

$$C_{AB}(t) = \int \varphi_A(t+\tau)\,\varphi_B(\tau)d\tau = h_{AC}(-t) * h_{BC}(t) * f(t) \tag{1}$$

Where h introduces the Green function and f(t) refers to the function of the excitation source and✳ representing convolution. The ambient noise processing is divided into three phases:

Signal station data preparing

The first step of data processing consists of preparing waveform of data from each station. Firstly, we try to remove instrumental irregularities and random earthquake signals. It's important for the cross-correlation technique to use 1-day data, but our experience shows that averagely 80 per cent on time, data is receivable. All of data were selected from 2 AM to about 19 PM so we have synchronized signals. Decimating data to decrease the time of the future processing and, de-training, removal of the instrumental response, Band Pass filtering (the Butterworth Band-pass filter between 0.1 and 0.5 Hz) the seismogram and time domain normalization (for reducing the effect of the earthquakes, instrumental irregularities and non-stationary noise sources near to stations) were performed to all of data. In different methods, for time domain normalization it has been considered by Bensen et al (2007): one-bit normalization, clipping threshold, automated event detection, Running-absolute-mean normalization and Water-level normalization. We applied one-bit normalization because it gave an utterly coherent noise. It retains only the sign of the raw signal by replacing all positives amplitudes with 1 and all negative amplitudes with -1. So the sharp amplitudes of probably earthquakes on the seismogram are removed.

Cross-correlation calculation

The second step of data processing is consisted of calculation of cross-correlation and stacking between each pair of stations. Cross-correlation was performed in the frequency domain between 0.1-0.5 Hz and then was returned to time domain. The stacking of daily cross-correlation functions on 4 month made an accurate signal with the highest possible signal to noise ratio. Figure 3 shows these functions in some of the chosen paths.

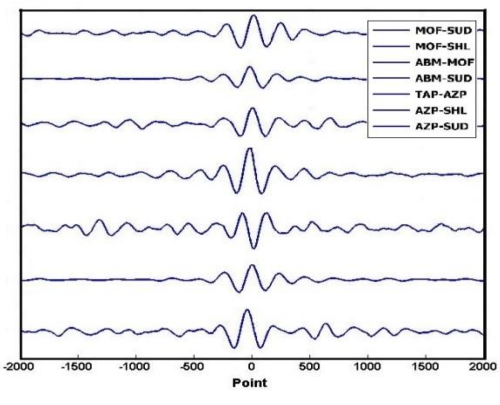

Fig 3: extracted signal from 4 month noise processing for each direction

Cross-correlation is a two-sided function that has organized from positive and negative lag parts that is called causal and acausal signals. The arrival time of wave is recognized in two sides if the noise sources surrounding the stations. If the sources are located in one side, the wave is seen in one side of cross-correlation function. We compressed the two-sided signal into a one-sided signal by averaging the causal and acausal parts to make a signal similar to the Green function (Stehly et al. 2009). It is significant that arrival times are the same in two sides, so averaging did not cause any mistakes. Since only vertical components used for processing, Rayleigh wave is reconstructed. This also is appearing from wave particle motion (Roux et al. 2005). All the functions were filtered again to reduce the effects of noise that has created on the Green functions. Finally, the Green functions with the signal to noise ratio more than 2 were noted to choose an accurate Green function in each path.

Dispersion measurements

After estimation of the Green function between each pair of stations, the group velocities as a function of period can be measured by using traditional frequency-time analysis (FTAN) (Levshin et al. 1972, 1989; Ritzwoller & Levshin 1998; Levshin & Ritzwoller 2001). The method that we promote is based on the version of FTAN described by Levshin et al (1989), which obtained measurements on the analytic signal. Group velocity is equal to the velocity of envelope function that is achieved from the analytic signal. So the analytic signal should be considered.

Imagine S(t) is the raw signal. The analytic signal is composed of raw and Hilbert transform of signal:

$$s_a(t) = s(t) + iH(t) = |A(t)|\exp(i\emptyset(t)) \tag{2}$$

Where H(t) is the Hilbert transform of S(t), |A(t)| and $\emptyset(t)$ are the amplitude and phase of analytic function. According to Fourier transform:

$$s(\omega) = \int_{-\infty}^{\infty} s(t)\exp(i\omega t)\,dt \tag{3}$$

The Fourier transform of analytic function is as follow:

$$s_a(\omega) = s(\omega)(1 + sgn(\omega)) \tag{4}$$

To construct a frequency-time function, the analytic signal is subjected to a set of narrow Band-pass Gaussian filters with center frequencies ω_0 on the analytic signal:

$$s_a(\omega, \omega_0) = s(\omega)\big(1 + sgn(\omega)\big)G(\omega - \omega_0) \tag{5}$$

α is a tunable parameter that defines the complementary resolutions in the frequency and time domains and commonly made distance dependent. Inverse transforming to the time domain yields the smooth 2-D envelope function, $|A(t,\omega_0)|$ and phase function, $\emptyset(t,\omega_0)$. Group velocity is measured using $|A(t,\omega_0)|$ and phase velocity using $\emptyset(t,\omega_0)$. The group arrival time, $\tau(\omega_0)$, is determined from the peak of the envelope function so that the group velocity is $U(\omega_0) = r/\tau(\omega_0)$, where r is interstation distance.

After exerting the Gaussian filter in different frequencies, an energy diagram of signal is characterized as the period terms of group velocity. The primary dispersion curve based on maximum amplitude in each frequency is calculated. Fig 4a is an example of primary dispersion curve measurements. Furthermore, matched filter was used to remove the effect of high Rayleigh wave modes. The conclusive dispersion curve is determined by the renewed tracing of maximum amplitude in each frequency (Fig 4b). The synthetic Green function (red wave in Fig 4c) can be calculated based on the conclusive dispersion curve and has compared with the estimated Green function (blue signal in Fig4c). The suitable fitness between the estimated Green function and synthetic Green function represents reliable group velocity measurements.

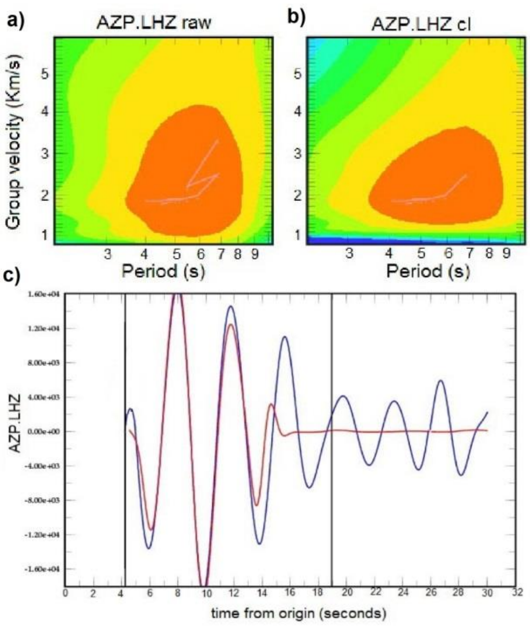

Figure 4: dispersion curve and reconstructed time history for each direction

Considering signal to noise ratio for the Green functions, dispersion curves for some of the path was calculated such as: AZP-SUD, AZP-SHL, MOF-SUD, MOF-SHL, ABM-MOF, ABM-SUD and AZP-TAP.

FREQUENCY-WAVENUMBER METHOD

Frequency-Wave number analysis (F-K) (Lacoss et al. 1969, Capon 1969, Kvaerna and Ringdahl 1986) assumes horizontal plane waves traveling across the array of seismometers. Considering a surface wave frequency, direction of propagation and velocity (or equivalently kx and ky, wavenumbers along X and Y Cartesian axis, respectively) the relative arrival times are calculated at all sensor stations and the phases are shifted according to the time delays (Shabani 2008).

To calculate the frequency–wavenumber power spectral density $\mathrm{h}(\mathrm{k_x}, \mathrm{k_y}, \omega)$ on base of phase velocity, the cross correlation functions, $S(\Delta\mathrm{x}, \Delta\mathrm{y}, \tau)$ should be estimated.

$$S(\Delta x, \Delta \mathrm{y}, \tau) = E\left[\mathrm{X}^*(x_j, y_j, t_j)\mathrm{X}(x_i, y_i, t_i)\right] \tag{6}$$

Where $\Delta x = x_i - x_j = r\cos\theta$, $\Delta \mathrm{y} = \mathrm{y}_i - \mathrm{y}_j = r\sin\theta$ and $\tau = \mathrm{t_i} - \mathrm{t_j}$

$$h\left(k_x, k_y, \omega\right) = \iiint_{-\infty}^{\infty} S(\Delta x, \Delta \mathrm{y}, \tau)\exp(\mathrm{i}(-k_x(\Delta x) - k_y(\Delta \mathrm{y}) + \omega(\tau))\mathrm{d}\Delta x d\Delta \mathrm{y} \mathrm{d}\tau \tag{7}$$

On the other hand, Capon (1969) suggested frequency – wavenumber power spectral density $\mathrm{h}(\mathrm{k_x}, \mathrm{k_y}, \omega)$ calculated using Limited Fourier Transform.

$$h\left(k_x, k_y, \omega\right) = \left[\sum_{i=1}^{N}\sum_{j=1}^{N} \emptyset_{ij}(\omega)\exp(\mathrm{i}(k_x\left(x_i - x_j\right) + k_y\left(y_i - y_j\right))\right]^{-1} \tag{8}$$

Where $\Delta\mathrm{x} = \mathrm{x_i} - \mathrm{x_j}$ and $\Delta\mathrm{y} = \mathrm{y_i} - \mathrm{y_j}$ are both stations distance and $\Phi(\omega)$ is Inverse matrix of Fourier Transform of correlation matrix. The array output (phase velocity dispersion curve) divided by the spectral power which is called the semblance (Lacoss et al. 1969, Asten and Henstridge, 1984). With location of the maximum of semblance in the plane (kx, ky), wavenumber K0 at specified frequency ω_0, we can find an estimate of the phase velocity C0 and azimuth ϕ_0 of the traveling waves across the array.

$$\mathrm{C_0} = \frac{\omega_0}{\mathrm{k_0}} = \frac{2\pi \mathrm{f_0}}{\sqrt{\mathrm{k_{x0}^2} + \mathrm{k_{y0}^2}}} \qquad \phi_0 = Arctan(\frac{k_{x0}}{k_{y0}}) \tag{9}$$

Figure 5 shows the high power zone of phase velocity dispersion using Geopsy software. For each location averaged mean and its standard deviation of dispersion curves were selected. Each curve filtered in the reliable range calculated by Tokimatsu (1997) approach. It is noted that in each location, all dispersion curve resampled with a common interval (0.2-20 Hz, constant phase response of instruments). Also figure 6 shows collective group and phase dispersion curves for 4 location of Tehran which could be applied in inversion process.

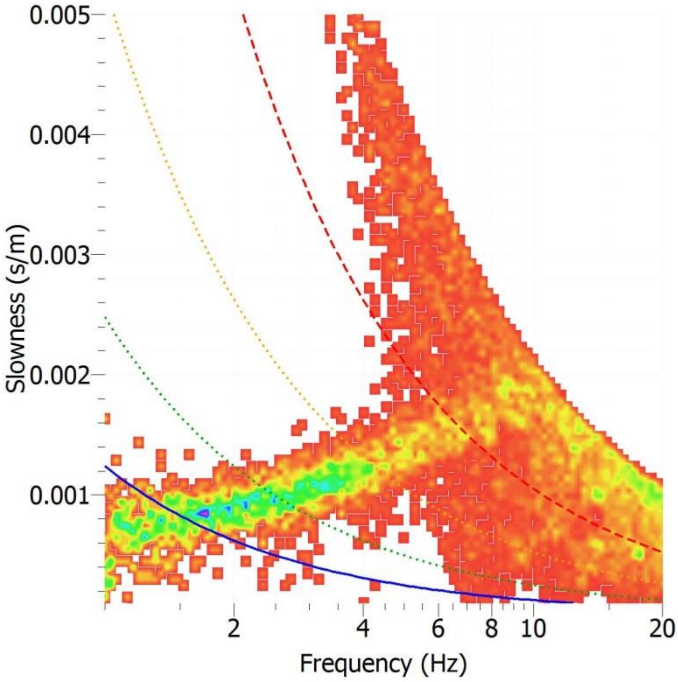

Figure 5: dispersion curve of THU station calculated with Frequency-Wavenumber method.

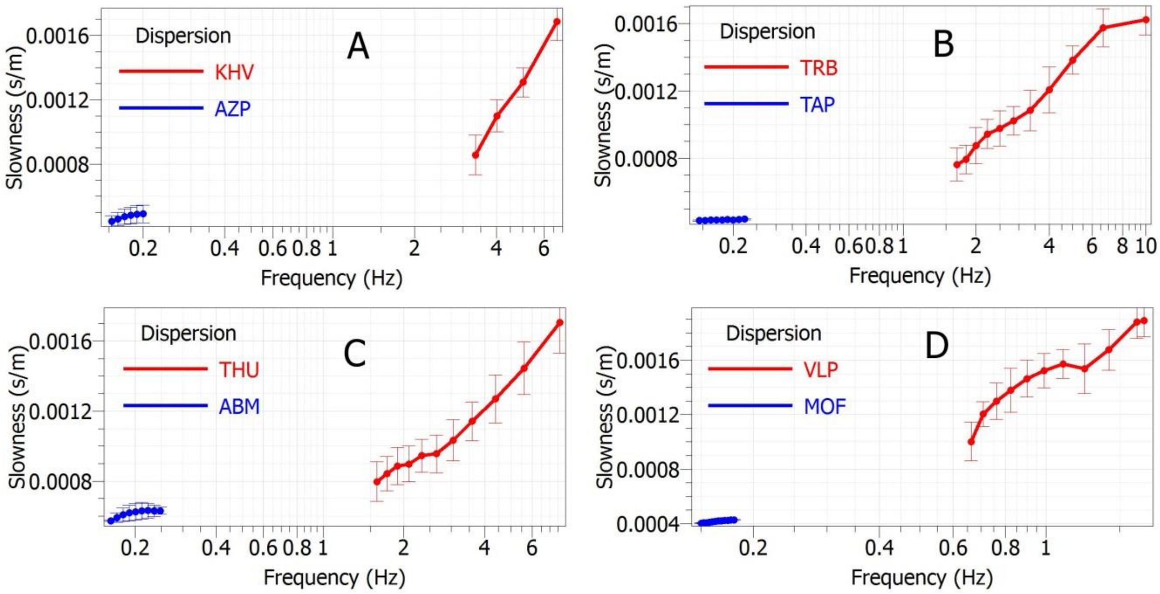

Figure 6: collected dispersion curves of two approach, cross-correlation (group velocity) and Frequency-Wavenumber (phase velocity)

INVERSION PROCESSING

Inversion techniques have spread in science to convert observational measurements into information about the physical characteristic of the environment in which usually are not directly measurable. Inversion problems are mainly defined as a linear or nonlinear problem and due to the complexity of most systems; these problems are expressed as nonlinear. Inversion can be done in several ways such as Mont-Carlo or Genetic Algorithm. Recently, Semberage (1999) presented the Neighborhood Algorithm as the direct search inversion techniques and Wathelet (2008) developed Conditional Neighborhood Algorithm in Dinver package. In this method, four parameters (Vs: shear wave velocity, H: thickness, υ: poisson ratio and γ: density) defined to explaining an elastic properties of alluvium layer. Poisson's ratio usually used to generate compression wave velocity (V_p) in relationship with V_s that generally varies from 0.2-0.5. For initial parameterization, consider to one fourth of the longest wavelength of group velocity ($\lambda/4$), we designed 10 layers with different possibility depth and shear velocity as detail in Table 1. We inverted phase and group velocity curves (figure 6) simultaneously to attain a shear wave velocity model of sedimentary Alluvium (figure7).

Table1. The preliminary parameters applied for inversion. H considered as bottom of depth, Poisson's ratio varied between 0.2-0.4 for each layer and density is 2000 Kg/cm^3.

Layer	1	2	3	4	5	6	7	8	9	10
Depth (m)	1-100	1-300	1-500	1-1000	1-1500	1-2000	1-3000	1-4000	1-5000	inf
V_p (m/s)	200-5000	200-5000	200-6000	200-8000	200-8000	200-8000	200-10000	200-10000	200-10000	200-10000
V_s (m/s)	150-3500	150-3500	150-4500	150-4500	150-5000	150-5000	150-5500	150-5500	150-5500	150-5800

CONCLUSIONS

In this study, we try to join two type of dispersion curve which were calculated from two methods. First, using long period temporary network (about 5 months), dispersion curves of group velocity were calculated by cross correlation method. In next step, we applied frequency-wavenumber method on array measurement data which was arranged near network stations. In this method, phase velocity dispersion curves could be extracted for each station.

Due to large distance of stations in cross-correlation (greater than 5 Km), obtained dispersion curves are valid for low frequency between 0.1 to 0.5 Hz. Also aperture of arrays were less than 1000 meter has caused dispersion curves of phase velocity only appropriate for high frequency more than 1 Hz.

Actually, by jointing inversion of two dispersion curves, we could extract deep shear wave velocity profiles on 4 stations in east, north, middle and south of Tehran (north : TRB- TAP, east: KHV-AZP , middle THU-ABM & south: VLP-MOF).

As seen in figure 7 A, shear wave velocity profile in KHV-AZP station has increase gradient to velocity about 3000 m/s for depth less than 200 meter. As well as the same trend repeated for TRB-TAP (figure 7B) where located on north of Tehran. According to geological data, two abovementioned sites (KHV-AZP& TRB-TAP) are located near the northern and eastern mountains (Alborz & Sepaye, respectively) so shallow shear wave velocity contrast depth is acceptable. On the other hands, another stations where arranged in middle and south of Tehran (THU-ABM and VLP-MOF, figure 7 C and D) illustrated deep shear wave velocity contrast between 800 to 1200 meter to geological bedrock with velocity greater than 2500 m/s. these contrast are approved with global geological information of sedimentary of Tehran Alluvium and previous studies. group dispersion curves at low frequency lead to stability of shear wave velocity of bedrock to about 3000 meter depth according to penetration depth of wavelength of signal which travel between two stations in cross-correlation method. As well as phase velocity in higher frequency control the shallow depth velocity above contract.

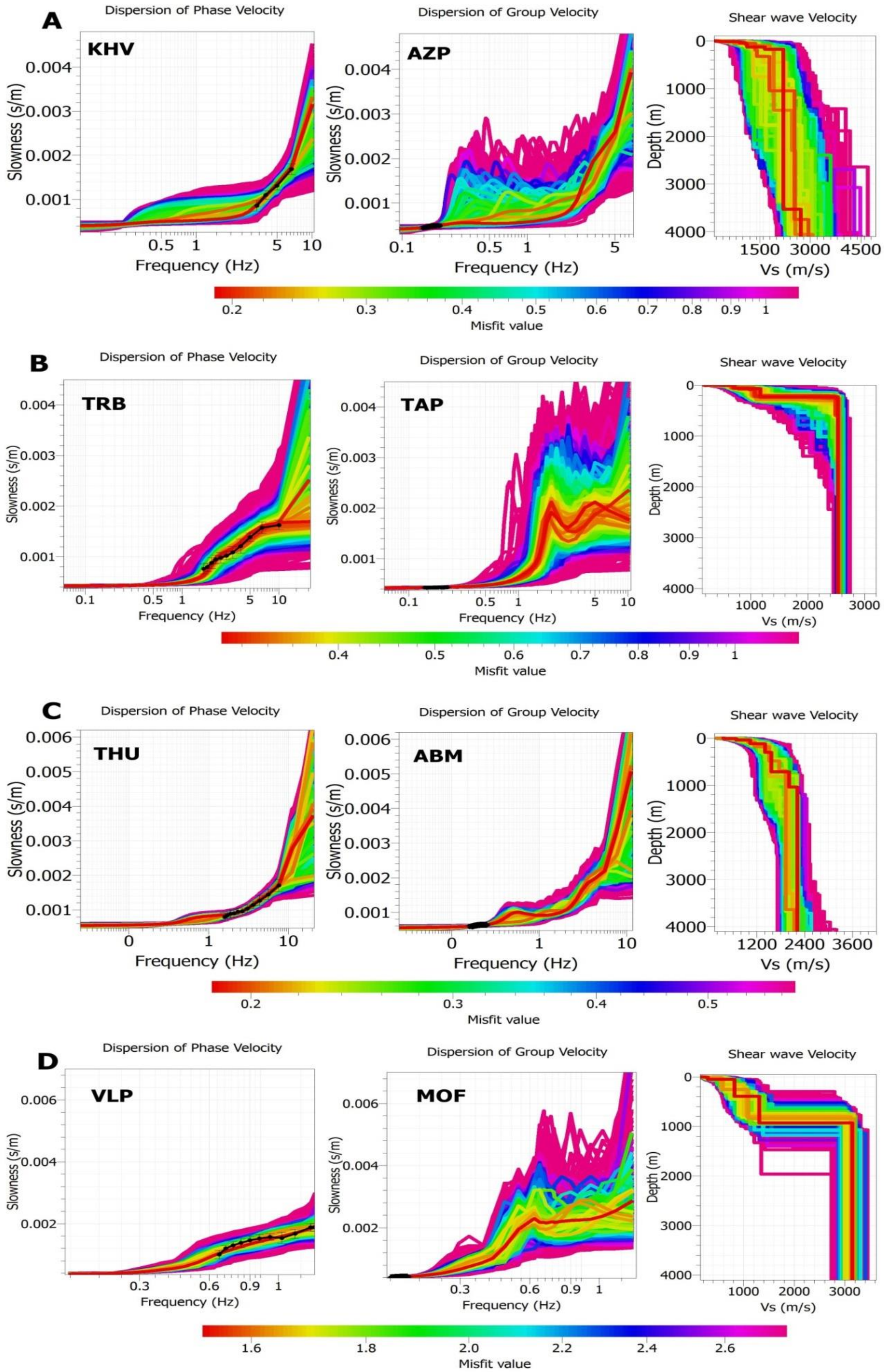

Figure 7: results of inversion processing. Fundamental mode of phase and group velocity was implied to extracting shear wave velocity.

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