Estimation of shear wave properties in the upper sea-bed using seismo-acoustical interface waves

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Abstract

Acoustic interface waves have been recorded with a bottom deployed acoustic array at several locations. High resolution array processing technique (Prony) is used to estimate velocity versus frequency for the interface waves. An inversion scheme based general linearization is used to obtain estimates of shear velocity as function of depth. The data collection with subsequent processing and interpretation is intended for operational use in conjunction with normal refraction surveying using the same equipment. This paper discusses survey design in terms of measurable shear velocities, depth of investigation and resolution as function of frequency and array configuration. Examples of processing and interpretation of field data are presented.

1. Introduction

Marine refraction seismic surveys are usually done with a linear array of hydrophones and an "explosives" type source for generation of the refracted waves in the sea floor. The refracted arrivals are then picked as the first break, and estimation of compressional wave velocity(v_p) and thickness together with dip is determined for the different layers within the sediment [1].

The refraction technique gives however no information of the shear wave velocity structure in the bottom and for many applications shear information is very useful. The shear modulus of the bottom is very important for underwater acoustic propagation. For geotechnical use it is important that acoustic shear modulus of the bottom material is related to the strength, although the direct relation is generally not known. Another attractive feature is that shear wave velocity is often a better sediment identifier than the compressional wave velocity. Several different sediment types may have quite similar compressional wave velocity (e.g. clay/silt/sand) but very different values for the shear velocity.

A well known technique for measuring the shear wave velocity structure of the sea floor is based on the measurement and analysis of the seismo-acoustic interface waves. In the current project we have modified the data gathering and analysis of the refraction surveys to allow for the recording and analysis of acoustic interface waves without interference with the refraction survey. The main modification is extending the recording time, which allows recording of the late arrivals with in wave field. Modification of gain and filter settings avoids saturation and provides the recording of the interface waves. In this paper the design of the array and the processing to establish shear wave parameters are described together with results from three field surveys.

2. Theoretical aspects

The seismo-acoustical interface waves belong to a class of slowly propagating waves at the interface between two media, they are therefore often called boundary waves or just interface waves. The Scholte wave belongs to this class and propagates at the boundary between a solid and liquid (e.g. water and sediments). The properties of the Scholte wave can be summarized as ([2],[3] and [4]):

- It has a rotational particle movement in the sagital plane
- If the shear wave velocity is varying with depth, the Scholte wave becomes dispersive
- It is generated by a source close to, or at the sea floor
- The radiated pressure can be recorded by hydrophones close to or at the sea floor
- The velocity of the Scholte wave is approximately 0.9 times the shear velocity

The frequency dependent velocity (dispersion) can be estimated from the recorded time signals, and there exists algorithms for inverting the dispersion curves to shear wave velocity (v_s) as function of depth [5] and hence converting this to dynamic shear modulus (μ) also as function of depth. Figure 1 shows the flow chart of the working process to establish such estimates in addition to ordinary refraction analysis.

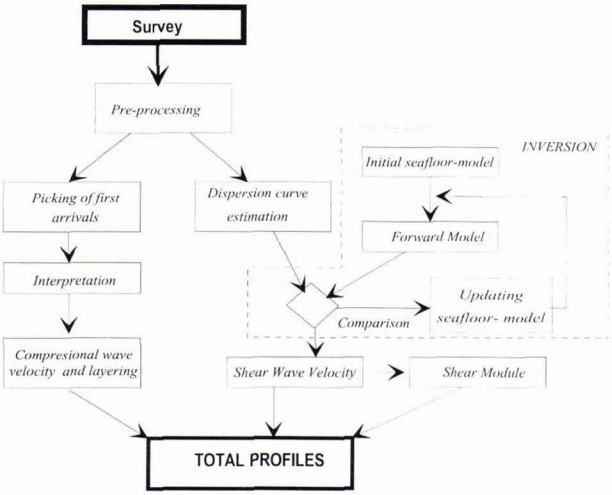


Figure 1: Flowchart of the processing refraction seismic data with extended recording time.

2.1. Survey design

The recorded wave field can either be processed using single- or multi sensor processing techniques for estimation of the dispersion curves. In the research community the multiple filter analysis [6] has been used extensively. The method gives an estimate of group velocity as a function of frequency. Since the method is a single sensor technique it suffers from problems like lack of resolution (velocity versus frequency) and it produces an average velocity estimate between source and receiver. This problems can be avoided by the use of the multi-sensor, high resolution estimators like the Prony method. The Prony method is based on wave number estimation from the signal recorded within the array, and represent therefor local estimate of the dispersion beneath the array. There are no problem with lack of resolution in velocity estimates. However, to use such a method, the array must be design to ensure a proper sampling of the wave field.

In general the receiver spacing (Δx) determines the lower limit of shear wave velocity which can be estimated. Since this parameter also are the target of the investigations a minimum velocity ($v_{s,min}$) has to be predicted/assumed. From this assumption one can derive the receiver spacing which must be equal or less then half a wavelength at the highest frequency of investigation (f_{max}):

$$\Delta x \leq \frac{\lambda}{2} = \frac{v_{s,min}}{2 \cdot f_{max}} \tag{1}$$

Assuming that $v_{s,min}$ is equal to 100 m/s, and that $f_{max} = 20$ Hz, the receiver spacing $\Delta x \leq 2.5$ meters. The distance from the source to the last receiver (R) is depending upon the attenuation of the waves in the sediments.

Here, it is assumed that the signal must be detected before the transmission loss (TL) is less than 20 dB, and using the absorption factor $\alpha_s = 0.002$ dB/m per. Hz [7]. The distance is determined by:

$$R = \frac{TL}{\alpha_s \cdot f_{max}} = \frac{10000}{f} \tag{2}$$

At $f_{max} = 20$ Hz, $R = 500$ meters.

In many cases a certain distance between the source and the array is required, to separate the waterborne and interface arrivals to avoid aliasing in the ($\omega - \kappa$-domain). Aliasing causes mis interpretation and occur when the surrounding environment is hard. The array length is given by $L = (M-1) \cdot \Delta x$, and it follows from eq. 2 and 1 that the source datum distance (SRD) is:

$$SRD \leq R - (M-1)\Delta x = \frac{1}{2 \cdot f_{max}} \left(2\frac{TL}{\alpha_s} - (M-1) \cdot v_{s,min} \right) \tag{3}$$

Using $M = 48$ receivers and the other parameters as described earlier: $SRD \leq 382$ meters. By use of the earlier assumptions, the minimum recording time becomes:

$$t_{rec} \geq \frac{R}{0.9 \cdot v_{s,min}} \tag{4}$$

Hence, $t_{rec} \geq 5.5$ seconds. In general we have used 8 seconds in our experiments.

2.2. Estimation of dispersion curves and inversion

The Prony method is a so called multi-sensor technique, that is a technique that makes use of an array of receivers to estimate the phase velocity as function of frequency. If N time samples has been recorded with M different receivers (receiver spacing Δx), the Prony method estimates the wave numbers κ. These can further be converted to phase velocity as function of frequency. The Prony method is deterministic, and the signal model for a certain frequency is [8]:

$$S_m = \sum_{p=1}^{P} h_p \cdot e^{\imath \cdot \kappa_p (m\Delta x + x_0)} \tag{5}$$

Here, h_p contains the magnitude and phase for the mode p, while κ_p is the wave number carrying the velocity information for the same mode. The recorded signal $s(t,x)$ in time and space, may be seen as a two dimensional Fourier transform, where $A(\omega, \kappa)$ is the response of the formation. Mathematically this is expressed as:

$$s(t,x) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} A(\omega,\kappa) e^{\imath\omega t - \imath\kappa x} d\omega d\kappa \tag{6}$$

Eq.6 is continuous both in frequency and wave number. In most cases, only a limited number of modes are of interest. In addition, the frequency domain is band limited. Therefore eq. 6 can be rewritten as:

$$\hat{s}(x,t) = \int_{0}^{\omega_0} \Sigma_{p=1}^{D} A(\omega, \kappa_p) e^{\imath\omega t - \imath\kappa_p x} d\omega \tag{7}$$

Here, D represents the discrete and dominating modes. The bandwidth of inspection is set to ω_0. The algorithm for determining the velocity starts with a Fourier transform of each time series corresponding to the M recorded channels. The auto-covariance matrix ($\tilde{B}$), are calculated for each of the frequency-space vectors. After establishing the auto-covariance matrix we have a set of linear equations:

$$\tilde{B} \cdot \hat{g} = \delta \tag{8}$$

where δ is the model error. The solution $\hat{g}$ is further processed to $\hat{\kappa}$ [8]. The phase velocity as function of frequency (e.g. the dispersion curve) is then:

$$\hat{v}_{dispersion} = \frac{2\pi f}{\hat{\kappa}} \tag{9}$$

The inversion process, as described in figure 1, is a typical minimization problem, where the object function is the the difference between the measured and a computed dispersion curves. The algorithm used for the inversion is the one presented by Caiti et. al [5]. The shear wave velocity and the shear modulus is related through:

$$\mu(z) = \rho(z) \cdot \hat{v}_s^2(z) \tag{10}$$

Here, $\rho(z)$ is the density profile. This is not always known, but we have usually some knowledge about the sediments to make an assumption. From the Ranheim experiment the density profile is known from the geotechnical investigations.

Experiment	Nick name	Charge size [g]	Receivers	Δx [m]	SRD [m]	f_s [kHz]
Orkanger	Ork5	200	24	2.5	145	2
Ranheim	Ran7	300	24	2.5	323	2
	Ran16	200	24	2.5	165	2
Sandvik	p437	100	48	2.5	160	2

Table 1: Recording parameters of the different experiments.

3. Result from three field surveys

During the last three years we have performed three field surveys. Two have been carried out in the Trondheim fjord, at Orkanger in August 1994 and at Ranheim in December 1994 [9]. The third experiment was performed at Sandvik on the island Øland in Sweden in October 1996. All the experiments were done in shallow water and close to land. In the Trondheim fjord, one array of 24 receivers was used, while the Sandvik experiment was conducted with an array of 48 receivers. The receiver spacing was 2.5 meters, and "explosive" charges of size 25-300 grams have been used as sources. The parameters of each of the presented shots are given in table 1, and a side view of a typical recording is shown in figure 2.

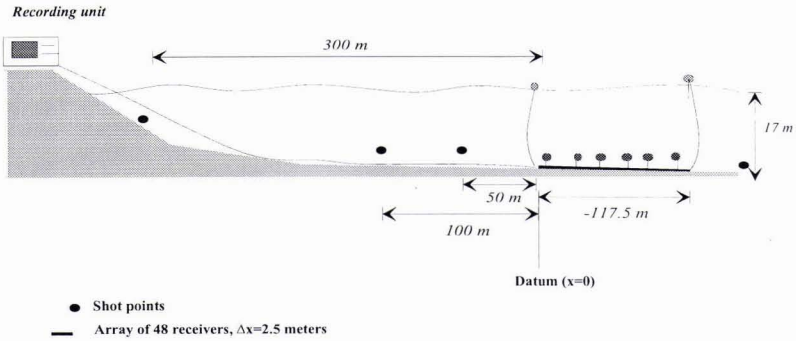


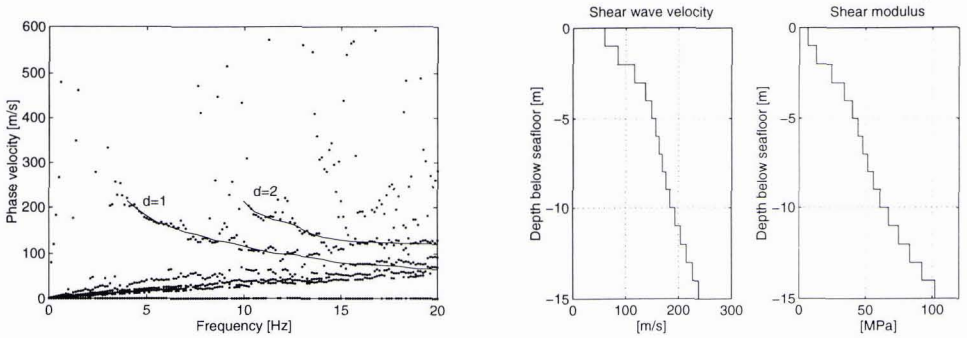
Figure 2: Side view of a typical recording (from the Sandvik experiment 1996).

3.1. The Orkanger experiment

The survey was performed in a natural river delta at Råbygda. Geotechnical investigations made close to the actual site uncovered sediments containing sand and gravel. Deposits of silt and fine sands was found both on top of and as thin layers buried within the sea bed [9]. Figure 3(a) shows the dispersion curve obtained with the Prony method using $p = 6$. The first two modes are interpreted ($d = 1$ and $d = 2$). The fundamental mode ($d = 1$) has been used for the inversion. The result of the inversion is shown in figure 3(b), as shear wave velocity as function of depth (left) and as shear modulus (right). The shear modulus is obtained using a iso-density profile of $1800 kg/m^3$. The shear profile shows a $3-4$ meter thick loose layer on top of harder sediments. This is consistent with a layer of silt above sand and gravel.

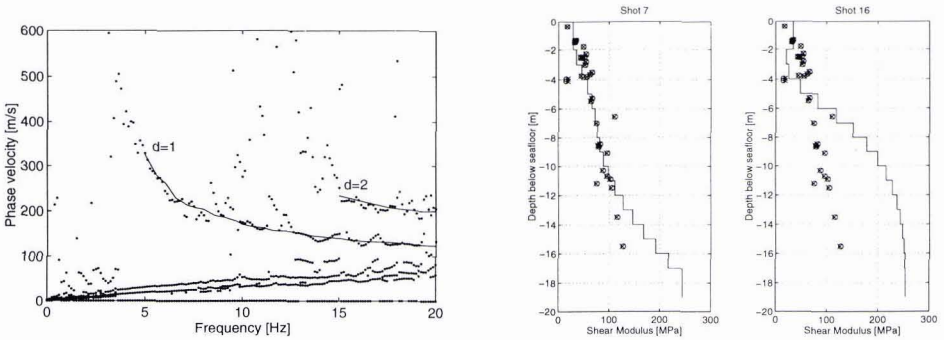
3.2. The Ranheim experiment

From the Ranheim experiment, results from two shots are shown (Ran 7 and Ran 16). The dispersion curve from shot Ran 7 is shown in figure 4(a). Two modes has been interpreted, only the fundamental mode (d=1) have been used in the inversion. (The same has been done for shot Ran 16). The obtained shear modulus profiles are shown in figure 4(b), together with the geotechnical measured shear modulus (G_{max}) derived using the relations by Hardin and Black (o) and Hardin (x) [9]. The agreement between the two independent measurement techniques are quite good for Ran 7, while it has a deviating fit below 6 meters in case of Ran 16. Ran 16 is performed across the direction of deposition. The shot was made with an angle between the shot point and the array. Even after compensation for this angle, it is not possible to correct for the fact that the hard sediments are shallower around



(a) (b)

Figure 3: a) Interpreted dispersion curve using Prony. b) Left: Inverted shear wave velocity profile. Right: Shear modulus profile



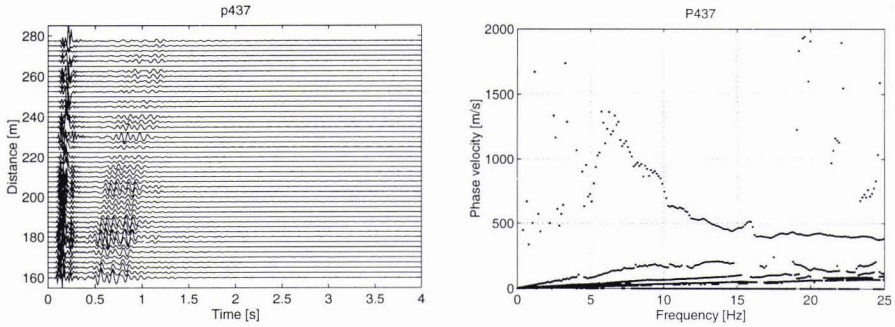
(a) (b)

Figure 4: Result form the Ranheim experiment;a) Interpreted dispersion curve using Prony on Ran7. b) Inverted-shear modulus profiles compared with geotechnical data Left: Ran 7, Right: Ran 16

the shot point than it is around the deployed array. This explains the disagreement between the modulus of Ran 7 and Ran 16, and hence the difference between Ran 16 and the G_{max} measurements.

3.3. The Sandvik experiment

The Sandvik experiment was conducted to record both compressional and interface waves. Figure 5(a) shows a typical time recording after filtering with a 30 Hz FIR (low pass) filter. The interface waves are found between $0.4 - 1.5$ seconds. The corresponding dispersion curve is shown in figure 5(b). The dispersion curve has been inverted from $10 - 20$ Hz, and the resulting shear wave velocity and shear modulus profile is found in figure 6. The shear modulus profile is generated using a iso-density profile of $2000kg/m^3$. The dispersion below 10 Hz was not taken into account since this part might be influenced by aliasing due to interaction between waterborne and interface waves. The result from the standard refraction seismic analysis is plotted in figure 6. The comparison between shear wave and the compressional wave velocity shows that there is better resolution in the shear wave



(a) (b)

Figure 5: Result form the Sandvik experiment; a) Time series of profile p437. b) Dispersion curve using Prony.

velocity in the upper sediments.

To increase the information about the area data acquired with a bottom penetrating parametric sonar [10], *Topas*, (Simrad Norge) was available. The data were recorded using a Rickert wavelet as source pulse with a center frequency of 5 kHz. The data were taken several hundred meters south-west of profile p437, but close to the other three profiles. A stacked, amplified and processed signal from this system is shown in figure 7(a). Several reflectors have been interpreted, shown in 7(b). The reflector B defines the sea bottom. The water depth is estimated to be in the range of $12.9 - 11.8$ meters (using $v_p = 1450$ m/s). The bottom is quite hard in this area and only small amounts of energy has penetrated into the sediments. The reflector $R1$ determines the depth of the first layer , the depth is less than a meter using $v_p = 1700$ m/s, while $R2$ is the lower boundary of layer two, using $v_p = 2200$ m/s, this layer is 3 meter thick. There are just small variations in layer thickness for both these layers. The MB reflector is just a multiple reflection of the sea bottom. The *Topas* has here been used as a bottom profiler, the result gains no further knowledge about the area than already achieved by the refraction measurements. If the *Topas* had been used in a sweep modus, covering the area with angle dependent reflections/backscatter, such data could have been used for estimation of compressional wave velocity and layer thickness, but in this experiment this could be difficult due to the hard environment.

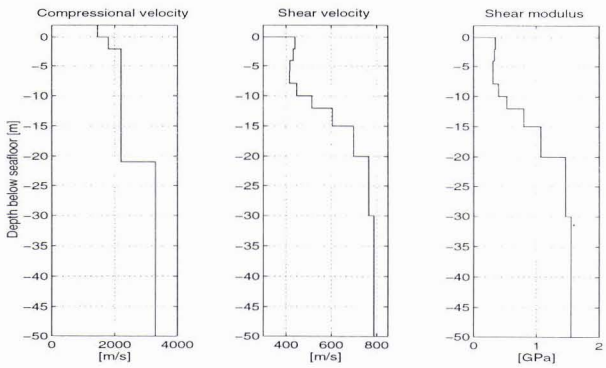
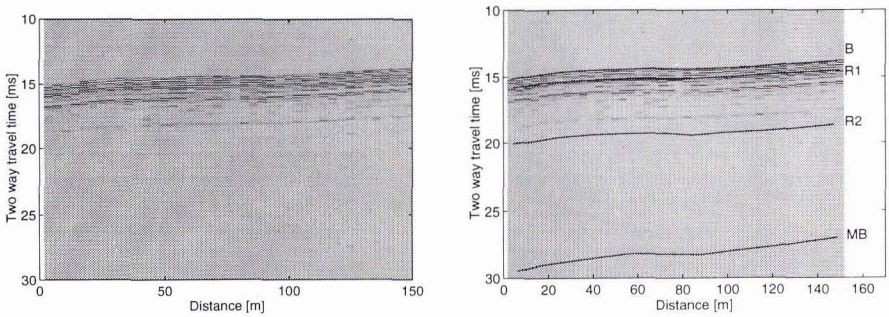


Figure 6: Result from the Sandvik experiment. Left: Shear wave velocity. Center: Shear modulus Right: Compressional wave velocity. All as function of depth.



(a) (b)

Figure 7: Processed Topas data. a) un-interpreted b) interpreted (in color on PC-screen).

4. Discussion

The high resolution spectral estimators like the Prony method reduces the problem of averaging velocity estimates between source and receiver. The of resolution is increased, but with the result of increased computational cost. If one uses statistical methods the energy content can be estimated for the different modes, and the attenuation factors can be estimated as function of frequency. With an inversion technique for this parameter one could gain further knowledge about the upper sediments.

In the Ranheim experiment we have shown that the described technique fits quite well with geotechnical data. The shot must be set off in the in-line direction of the array. Further more, if the environment is laterally changing the size of the array must be small compared to these changes, since the forward model is based on plane layering. Therefor, if the environment is not piece-wise lateral plane layered, erroneous estimates will appear as results. To ensure good quality information the method should be combined with coring, but the number of cores can be reduced. The result will be a continuous profile of shear wave velocity and compressional wave velocity that can be correlated with the coring, resulting in a high quality identification of the different sediment layers.

As seen from the Sandvik experiment the method gives higher resolution in the upper sediments than what could have been achieved with only ordinary refraction seismic methods. Therefor this method can be used for better classification of the sea floor content in the top 10-20 meters. If the sediments are hard (like at Sandvik) the distance between the source and the array must be increased to separate the different arrivals in time to avoid aliasing effects. However, this problem have only appeared in the Sandvik experiment, it is no problems in the loose areas in the Trondheim fjord. The problem has trigged the idea of building an equipment with adjustable receiver spacing and source-array spacing. Such an array could be optimized to the actual environment, and combined with a source who generates less waterborne arrivals, like a vibrator. The source energy will then be coupled into the sediments and reduce the waterborne part of the wave field.

The Prony method is also able to reveal more than the fundamental interface wave mode. In figure 3 (a) and 4 (a) two modes are interpreted. Attempts to include such modes in the inversion [11], shows that the estimates for the deeper sediments becomes more reliable. The resolution of the upper sediments depends on the highest frequency in the analysis and thereby the receiver spacing (eq.1). By designing a tool with the appropriate settings, a new flexible equipment would provide velocity estimates lower than $100m/s$ (eq.1).

In future we will continue the work on development of processing techniques for analysis of hydrophone data as well as data recorded on three component geophones. A study of how the method functions in laterally changing environment is also necessary. This can be done with scaled models in laboratory or by numerical modeling. We have also started work which will investigated the the performance of the method in transversally anisotropic materials.

The method is well suited for investigations of cables and pipe-line routes as well as for evaluation of areas for placing sub-sea installations.

5. Conclusions

In this paper we have presented results from three field surveys. The data has been gathered with a linear hydrophone array deployed on the sea floor and with "explosives" as source. The recording time has been increased compared with standard refraction seismic survey to capture seismo-acoustical interface waves.

The data has been analyzed using standard refraction seismic methods, as well as we have applied more advanced processing algorithm to extract dispersion curves from the interface waves. The dispersion curves has been used as input to general linearization inversion.

The results of the inversion shows that there are more variations in shear wave velocity as function of depth, than it is for compressional waves. This result is important and shear wave velocity should be taken into account in sea floor classification and characterization. In the case of the Ranheim experiment the results has been compared with data from standard geotechnical investigations. The agreement between the different measurement techniques is good. We believe that there are potential in this technique to improve the classification and the characterization of the upper sea floor.

Acknowledgment

The field survey at Orkanger and Ranheim has been performed by O. K. Fjeld at GEOTEAM A/S. GEOTEAM A/S shot the seismic at the two site on their own expense and provided the geotechnical data from Ranheim. The geotechnical data was prepared by S. Kirkebø, Sintef Geotechnical Engineering, and this project was co-sponsored by the Norwegian Research Council under the TEFT-program.

The Sandvik experiment was a joint project between Statoil Gum T & T Kart, Geomap a.s and NTNU. The refraction seismic surveying and interpretation was done by A. Olsen and O. C. Pedersen (Geomap a.s). P. Morèn (FOA) helped us with the survey and J. Dybedal (Simrad Norge) provided the Topas data.

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