IWT SBO PROJECT 120003 "SEARCH"

Archaeological heritage in the North Sea

Development of an efficient assessment methodology and approach towards a sustainable management policy and legal framework in Belgium.

Archeologisch erfgoed in de Noordzee

Ontwikkeling van een efficiënte evaluatiemethodologie en voorstellen tot een duurzaam beheer in België.

NON-CONVENTIONAL SURVEY TECHNIQUES FOR MARINE ARCHAEOLOGICAL INVESTIGATIONS

WP 1.1.3

Responsible partners: Deltares, UG-RCMG

Authors: P. Kruiver, O. Zurita Hurtado, G. Diaferia, T. Missiaen

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1 Introduction

For the last 70 years, most seismic survey studies have been limited to extracting information from compressional (P) waves. However, there are cases when the physical properties of earth materials constrain the effectiveness of compressional waves. Other types of seismic waves, like shear (S) waves and surface waves, depend on different elastic properties than P-Waves. Hence, they respond differently to certain materials, yielding supplementary information about the substrate. Additionally, since each wave travels at its particular speed and responds to changes in elastic moduli and density differently, achievable resolution and observed reflections may be different.

Pore materials filled with gas, even in small quantities, represent one of the most important limitations of conventional seismic survey techniques, because it disrupts P-wave transmission and thus obscures underlying strata. Shear waves and Surfaces waves are less affected by gas. Therefore, they represent a valid alternative to conventional acoustic techniques in gas-rich areas, like the Belgian sector of the North Sea.

In addition to acoustic techniques, other geophysical techniques might be useful for imaging the subsurface for the purpose of underwater archaeology. These techniques are based on the difference in e.g. electrical conductivity, magnetic susceptibility or dielectrical properties of the subsurface.

In this report, we describe the basic principles governing shear and surface waves propagation and non-acoustic alternatives. We list advantages and limitations for each method. The use of each method to investigate the substrate in shallow marine environments is described, most specifically in marine archaeological investigations. The scope of the report is restricted to underwater methods only. Geophysical techniques suitable for the transition zone from sea to land and land techniques are not specifically described in this report.

2 Shear waves

Shear or secondary waves (S-waves) propagate through the body of a medium making particles to displace perpendicular to the direction of wave propagation (Figure 1). They travel more slowly than P-waves and as a consequence they are the second disturbance to be recorded on an earthquake register (hence the name).

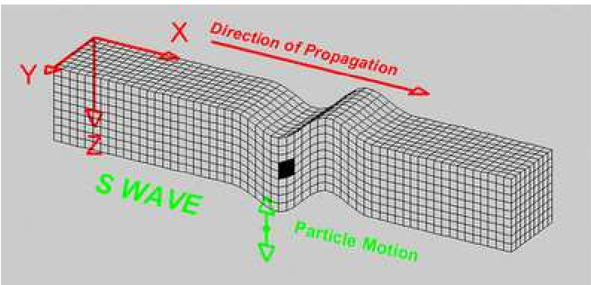

Figure 1. Shear wave particle motion.
From Purdue University - Department of Earth and Atmospheric Sciences at website (http://web.ics.purdue.edu/~braile/edumod/waves/Swave.htm).

As opposed to P-waves, which produce particle displacements along the radial direction, the general S-wave motion within the plane of the wavefront presents two degrees of freedom and can be resolved into components polarized in the horizontal plane and those polarized in the vertical plane. These components are known as SH (for horizontal) and SV (for Vertical) waves. Because of this polarization, S-wave amplitudes may vary with direction.

2.1 Properties of Shear waves

Velocity

The S-wave propagation velocity is determined by the rigidity or shear modulus (μ) of the propagating media and its density (ρ).

$$V_s = \sqrt{\frac{\mu}{\rho}}$$

Since fluids and gases cannot support shear stress, S-waves are incapable to propagate through them. In the case of pore materials, saturation modifies the bulk modulus or incompressibility of a medium, changing P-wave velocities, but does not affect the shear modulus, leaving S-wave velocities unaffected. This insensitivity of shear waves to pore materials is an especially beneficial property when trying to image beneath shallow gas or to distinguish layers in saturated material.

Type of formation	P wave velocity (m/s)	S wave velocity (m/s)	Density (g/cm^3)
Scree, vegetal soil	300-700	100-300	1.7-2.4
Dry sands	400-1200	100-500	1.5-1.7
Wet sands	1500-2000	400-600	1.9-2.1
Saturated shales and clays	1100-2500	200-800	2.0-2.4
Marls	2000-3000	750-1500	2.1-2.6
Saturated shale and sand sections	1500-2200	500-750	2.1-2.4
Porous and saturated sandstones	2000-3500	800-1800	2.1-2.4
Limestones	3500-6000	2000-3300	2.4-2.7
Chalk	2300-2600	1100-1300	1.8-3.1
Salt	4500-5500	2500-3100	2.1-2.3
Anhydrite	4000-5500	2200-3100	2.9-3.0
Dolomite	3500-6500	1900-3600	2.5-2.9
Granite	4500-6000	2500-3300	2.5-2.7
Basalt	5000-6000	2800-3400	2.7-3.1
Gneiss	4400-5200	2700-3200	2.5-2.7
Coal	2200-2700	1000-1400	1.3-1.8
Water	1450-1500	-	1.0
Ice	3400-3800	1700-1900	0.9
Oil	1200-1250	-	0.6-0.9

Table 1. P- and S-wave velocities [Bourbie, 1987]

Since S-wave velocities (V_s) depend on the shear strength of the material, knowing its propagation velocity, in combination with the velocity of P-waves (V_p), allows us to determine the elastic constants of the substrate. These elastic moduli provide key geotechnical information required in e.g. foundation studies. These elastic constants include the following:

i. Young's Modulus (E): The ratio of the applied stress to the fractional extension (or shortening) of the sample length parallel to the tension (or compression). It predicts how much a material sample extends under tension or shortens under compression.
ii. Bulk Modulus (K): The ratio of the confining pressure to the fractional reduction of volume in response to the applied hydrostatic pressure. Hence it is a measure of incompressibility.
iii. Poisson's ratio (σ): The ratio of lateral strain (perpendicular to an applied stress) to the longitudinal strain (parallel to applied stress).

Resolution

Due to their slower velocity, S-waves show shorter wavelengths than P-waves of the same frequency. This property should theoretically suggest an increase in resolution compared to P-waves. However, this is only valid for soft rocks, where S-waves are three to four times slower than P-waves of comparable frequency content, implying S-waves could substantially increase resolution. On the other hand, on hard rocks shear wave's velocity is usually about half of the P-wave velocity, but the predominant frequency is also about half, indicating the S-waves will not increase resolution. Since our study involves shallow and soft sediments, we should be able to take advantage of this property of shear waves.

P-S conversion

Another important property of Shear Waves is that when a wavefront from a compressional seismic source strikes an interface at an angle other than 90 degrees, the reflected and transmitted energy is partitioned into P and SV (vertical shear) wavefields (Figure). The SV wavefield produced by such a source is often more robust than its companion P wavefield [*Hardage*, 2011]. A minor amount of SH – horizontal shear – energy also radiates away from the application point of a vertical impact, but this S-wave mode is weak and can be considered negligible. It is this P-S conversion that is useful for underwater seismic surveys.

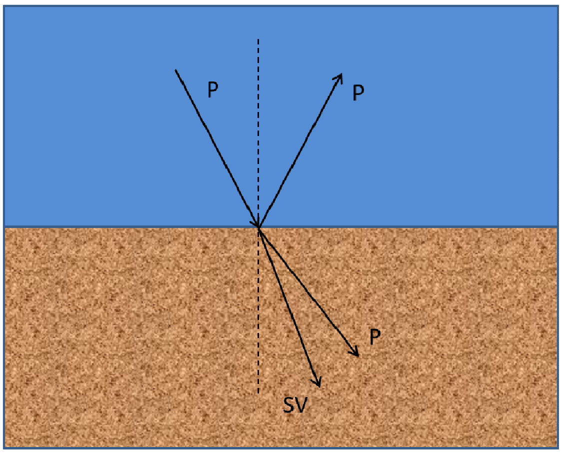

Figure 2. Waves generated at a fluid-solid interface due to an incident fluid wave. Modified from PhD Thesis [El Allouche, 2011].

A P-wave generated by e.g. an airgun will thus give rise to a transmitted S-wave when hitting the seafloor. Likewise, an up-going SV-wave reconverts to P-wave on return to the solid-liquid interface and so can be detected with conventional pressure detectors.

2.2 Shear wave imaging

In theory, shear waves can be created and measured as in conventional P-wave reflection. An energy source generates elastic waves in the ground, and these elastic waves are detected at multiple locations by vibration sensors. However, because of their lower velocity, their arrivals are imbedded somewhere in the seismic record after the P-wave arrivals. The solution would be to avoid the generation of P-waves and have S-waves only. This can be achieved by using seismic energy sources that generate pure shear waves, and multicomponent receivers capable of detecting shear waves. On land, this is relatively easy to achieve: a special oriented source is used along with orthogonally oriented receivers firmly planted in the ground. Under water, pure shear waves are not possible, because they cannot be transmitted through fluids. Therefore pure S-waves cannot be generated by marine sources, nor can hydrophones detect them.

As a consequence, most developments on marine shear wave survey require placing the source and/or the receivers in contact with the water bottom. A recent example is the Norwegian Geotechnical Institute's prototype seabed-coupled shear wave vibrator [*Vanneste et al.*, 2011]. Along with multicomponent ocean-bottom cable, it is capable to acquire all components of the seismic wave field (Figure).

However, several studies have shown evidence that contact with the seafloor is not necessary to produce shear wave in a marine environment [*Drijkoningen et al.*, 2012]. While generating direct shear waves requires special sources, generating converted shear waves from the reflection of P-waves on an acoustic interface does not require any particular equipment (as explained in section 2.1).

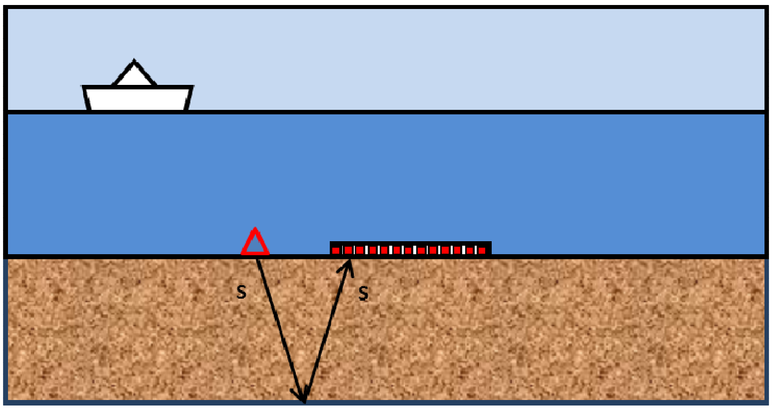

Figure 3. Marine shear wave reflection using both source and receivers on the seafloor.

There are several methods to record S-waves in marine environments. The first requires placing individual receivers on the seafloor with a remotely operated vehicle, which can be difficult, slow and therefore expensive. The second option involves placing instrumented cables packed with sensors directly on the seafloor (Ocean Bottom Cable or OBC) (Figure 4). In order to record shear waves, multicomponent receivers, composed of one hydrophone and one or three orthogonal accelerometers, are used. This receiver array is capable of recording the reflected SV component (using one vertically oriented accelerometer) or the full particle motion vector (using three orthogonal accelerometers) along with the conventional pressure wavefront (with the conventional hydrophone). The cable is positioned on the seafloor either by releasing the cable into place from the sea surface or by dragging it from one bottom location to the next. In this option, however, it is difficult to precisely control the exact position of the individual sensors.

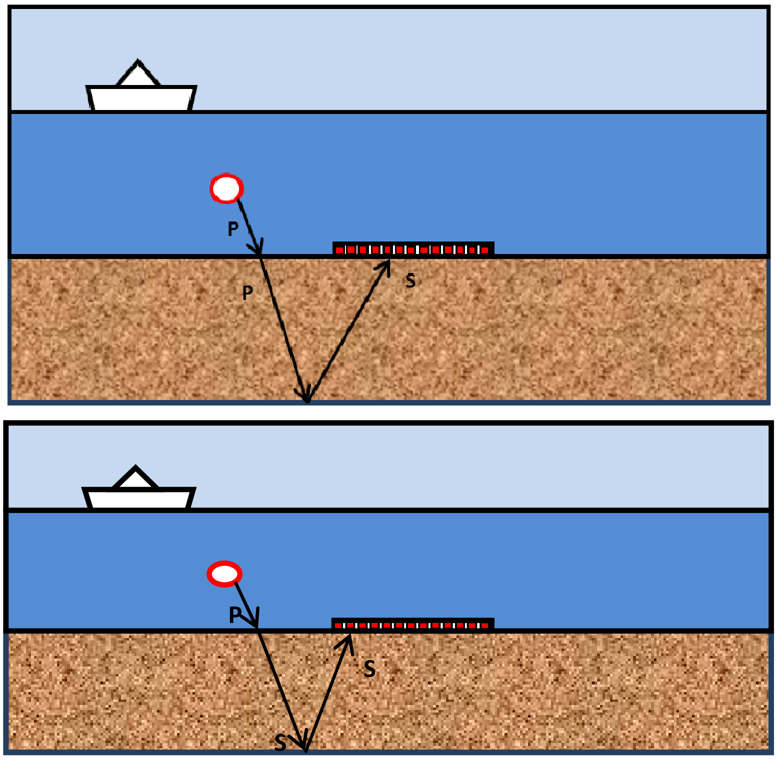

Figure 4. Marine shear wave reflection using converted waves. Detection using OBC on the seafloor.

A third method involves hydrophones. Based on the same principle used to generate shear waves with conventional (P-wave) sources, we can use conventional streamers towed through the water column to record the transmitted up-going shear wave field. When converted up-going shear waves reach the seafloor part of the energy is transmitted into the water as P-wave and can be recorded by the hydrophones (Figure). As mentioned in section 2.1, however, the distance between the water bottom and the streamer must be within a quarter of the dominant wavelet for the P-wave. The main problem with this approach is how to identify shear waves on the seismograph given that the converted waves will have much lower amplitudes than the primary P to S reflection.

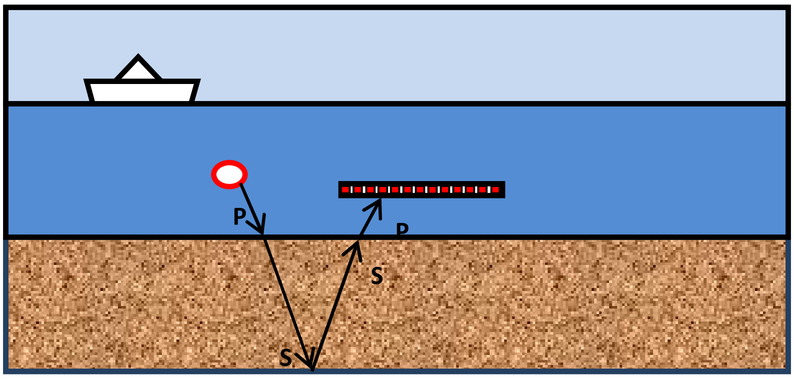

Figure 5. Marine shear wave reflection using converted waves. Detection by streamer.

Once data has been acquired, data processing is in principle relatively similar to that of conventional P-wave reflection surveys. However, this is true only when both devices are in contact with the seafloor (Figure) or when using conventional sources and receivers towed through the water column and down-going transmitted and up-going reflected S-waves are used (Figure). However, if the source or the receivers are not in contact with the sea bottom (Figure 4), then reflection is asymmetric and source-receiver reciprocity will not be fulfilled. Asymmetric reflection (
Figure *2*) is due to the fact that the reflection point is not halfway between the source and the receiver because converted or transmitted shear waves have slower velocities than compressional waves.

The source-receiver reciprocity consists that a trace recorded for a given source-receiver pair is the same as one for which the source and receiver have exchanged positions. Many basic processing steps like normal move out (NMO) corrections, trace interpolation and migration rely on these two assumptions. Finally, if the receivers are not oriented perfectly parallel and perpendicular to the radial and transverse coordinate system, S-waves components will be recorded on all orthogonal accelerometers. Polarization filters can be used to rotate the sources and receivers.

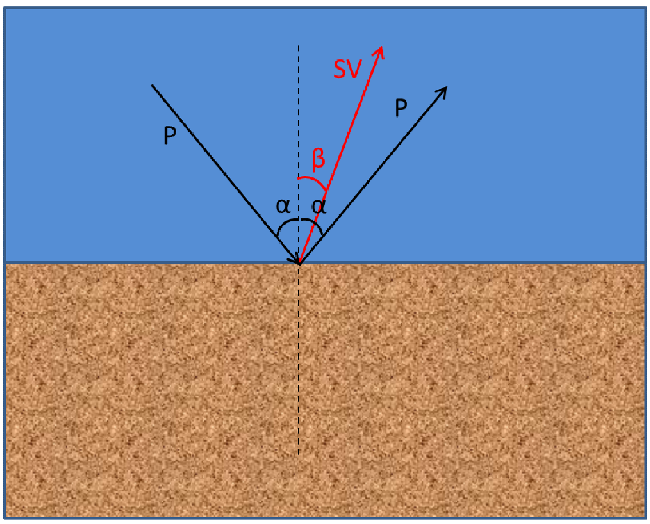

Figure 2. Asymmetric reflections

2.3 Advantages and limitations of shear waves

The main advantage of using shear waves to study the sub-seafloor is their capability to image through and below gas accumulations in the sediments (Figure 3). Moreover, their insensitivity to rock's fluids content may allow identification of layers that are undetected by compressional waves in saturated materials.

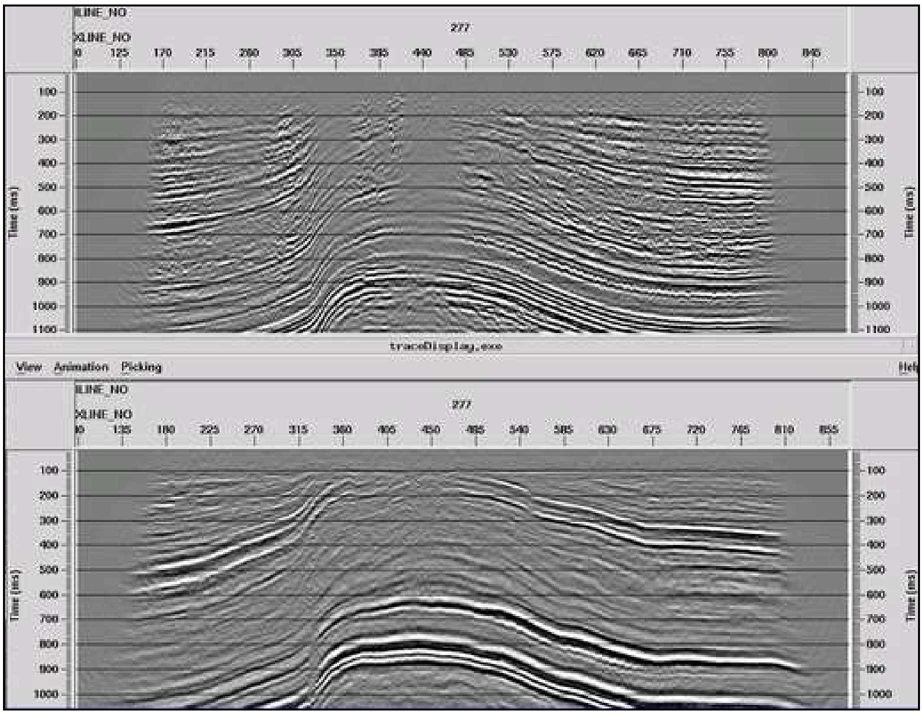

Figure 3. Seismic profile on gas area. PP reflection (top image) and PS (bottom image). From RXT website (http://www.rxt.com/imgaing-through-gas-clouds/category181.html)

Considering the V_p/V_s ratios (> 1), for a given frequency, the resolution should theoretically improve on a shear wave stacked section from unconsolidated soils in comparison to an equivalent compressional wave survey [*Miller et al.*, 2001].

Finally, since shear velocities are directly related to the stiffness of the rock, important engineering rock properties can be reasonably estimated by collecting both shear and compressional waves along coincident profiles.

On the other hand, shear waves are not as easy to generate and register as compressional waves, in particular in marine environments. A dedicated layout is needed in order to perform an optimal shear wave survey. Additionally, processing shear waves requires extra effort and in some cases different software than processing conventional P-waves. This makes the technique more expensive and as a consequence economically unsuitable for many near surface investigations.

Due to polarization of shear wave components, velocities are rarely the same in the vertical (SV) and in the horizontal (SH) plane. SH velocities may also vary in different azimuths, a situation known as horizontal anisotropy. In this case, the ground spectral response will be different, depending on which way the source and receivers are oriented. This does not influence the processing in itself, but has consequences for interpretation. Therefore, during acquisition, the orientation of the source and the receivers has to be noted. Differences at crossings in a grid might be explained by anisotropy (apart from the influence of noise).

One of the most common limitations of shallow shear wave reflection surveys is that Love wave arrivals can be mistaken as coherent events and might be stacked after NMO corrections, masking true reflections that might be present. Love waves are horizontally polarized shear waves (SH waves) that are guided by an elastic layer (see also chapter 3). These waves are observed only when there is a low velocity layer overlying a high velocity layer, so appears in a layered earth. The Love wave travel path can produce surface wave arrivals to show apparent hyperbolic curvature on the seismogram. Although surface wave arrivals in theory are linear and should possess no curvature, the apparent hyperbolic curvature often forms at near offset traces due to the near-field effect of surface waves, the wave interference effect, or a combination of both. Consequently, if Love waves are not efficiently removed from the record during processing, misrepresentations of the subsurface will be common [*Miller et al.*, 2001].

2.4 Recommendations on shear waves

Shear wave imaging can be enormously beneficial to investigate shallow sediments in areas like the Belgian North Sea where shallow gas accumulation are common and no information can be obtained from conventional P-wave reflection surveys. Their theoretical increase in resolution could be potentially valuable for our objectives but as many studies have proved this only occurs in very limited geological settings.

Processing P-SV converted shear waves requires special attention and must take into account the inherent characteristics of converted shear waves, like asymmetric reflection, source-receiver differences and multicomponent polarization into the processing flow.

Shear waves imaging should not be conducted as a stand-alone technique, it must be performed in combination with compressional wave techniques. Applying both methods has the potential to reveal more information about the subsurface than either wave type alone.

To our knowledge, shear wave imaging has never been used for archaeological investigation purposes.

3 Surface waves

Surface waves are waves that exist only near a boundary or interface between two media. Surface waves generated at the air-soil interface are called *Rayleigh waves*. Those generated in a marine setting at the water-soil interface are named *Scholte waves* [*Scholte*, 1947] and when generated at any solid-solid interface they are called *Stoneley waves*. *Love waves* are horizontally polarized shear waves (SH waves) guided by an elastic layer in between an elastic half space on one side and a vacuum (or air) on the other side. They can only exist if the elastic layer has a lower shear wave velocity than the underlying halfspace. Particle motions in the Rayleigh or Scholte waves and in Love waves are shown in Figure .

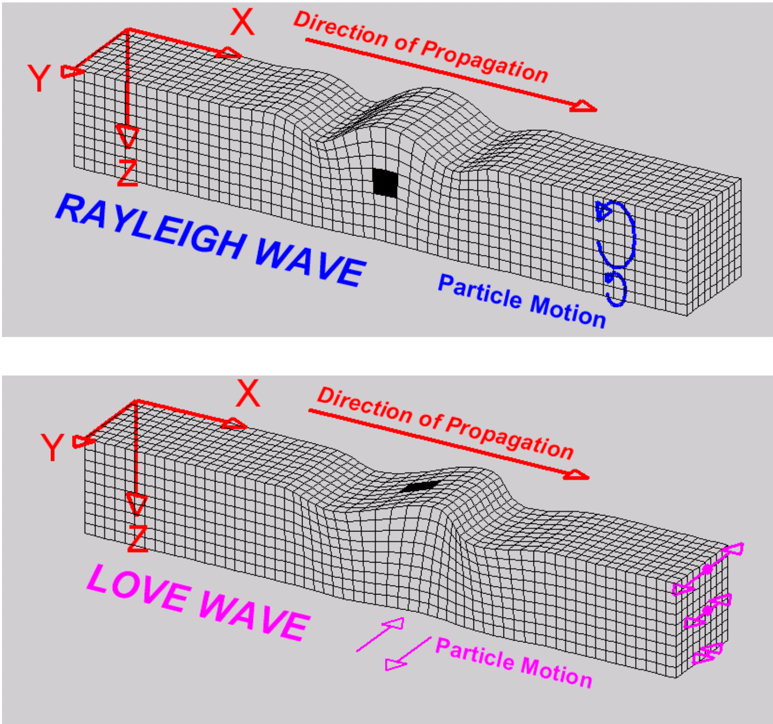

Figure 8. Particle motion in Rayleigh and Scholte waves (top) and Love waves (bottom). From Purdue University - Department of Earth and Atmospheric Sciences at website:
http://web.ics.purdue.edu/~braile/edumod/waves/Rwave.htm
http://web.ics.purdue.edu/~braile/edumod/waves/Lwave.htm

Surface waves are highly energetic: up to 60-70% of the seismic source energy is transferred to surface waves. Surface waves are well known for their destructive potential during earthquakes. These high amplitudes are mainly concentrated at the surface of the Earth and they decay exponentially with depth. Due to the fact that surface waves are 2 dimensional, rather than 3 dimensional, their geometrical attenuation with distance is lower compared to body waves.

Propagation velocity of surface waves is slightly lower than that of shear waves. On land, the Rayleigh wave velocity (V_R) is about 95% of the shear wave velocity (V_S) [*Telford et al.*, 1990]. Under water, the ratio between Scholte wave velocity and shear wave velocity is related to the ratio of wavelength to water depth. [*C.B. Park et al.*, 2005] show that Scholte wave velocity is slightly lower (~ 5%) for deep water conditions (see Figure) where the wavelength is shorter than several times the water depth. However, the correction needed for that usually falls within the uncertainty level of the measurement. Because of this close resemblance in values between V_S, V_R and V_{Sch}, surface wave inversion has become so important in the last twenty years.

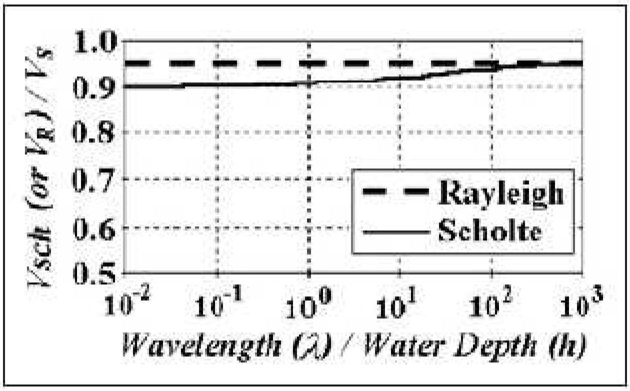

Figure 9. Approximate relationship between velocities of Scholte waves (V_{Sch}), Rayleigh waves (V_R) and shear waves (V_S). Taken from [C.B. Park et al., 2005]

Since we are concerned with marine environments, we will focus on Scholte Waves.

3.1 Properties of Scholte Waves

For the Scholte wave to exist, an interface system of an elastic solid half-space coupled to a liquid half-space is required. The wave will propagate along the path described by the interface, as long as any curvature effects of the interface are large compared to the wavelength of the wave.

Particle motion in Scholte Waves is the same as in Rayleigh waves and consists of elliptical motions (generally retrograde elliptical as shown in Figure) in the vertical plane and parallel to the direction of propagation.

The amplitude of the particle motion of a surface wave decreases exponentially with depth. Therefore, the large majority of the wave energy is contained within one wavelength and the velocity at which the surface wave propagates is influenced by the properties of the ground down to about one wavelength. Considering the case of a layer of thickness *h* overlaying a thicker stratum: a surface wave of wavelength shorter than *h* will propagate quasi entirely within the upper layer and will travel at a velocity dependent of the properties of this soil layer. Conversely, a surface wave of wavelength significantly longer than *h* will be principally affected by the lower stratum. Surface waves of intermediate wave lengths will be influenced by both layers.

In a uniform isotropic half-space all surface waves would travel at the same velocity. However, in a multi-layered soil or soil with stiffness properties varying with depth, the velocity of a surface wave depends on its wavelength (or frequency, see Figure). The dependence of phase velocity on the frequency of a propagating wave is known as *dispersion*. Theoretically, the dispersion property of surface waves is determined by several elastic properties; however, the depth-variation of the shear wave velocity (V_s) is the most influencing factor. This fundamental property is often used to deduce V_s properties of near-surface earth materials.

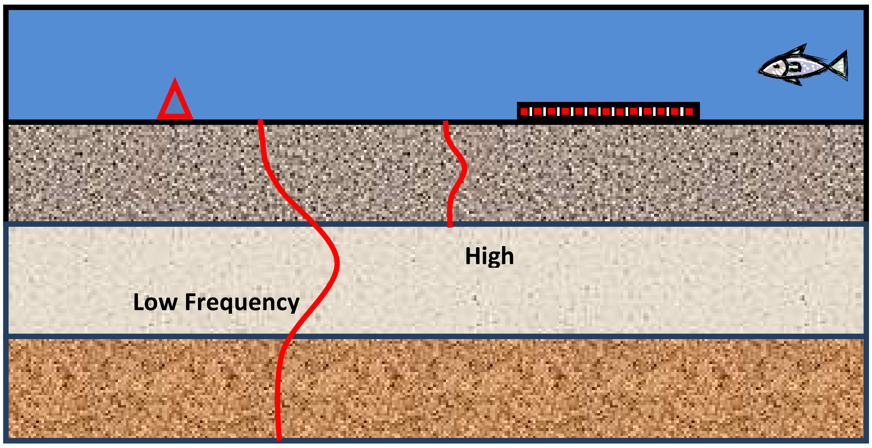

Figure 10. Amplitude of vertical displacement of Scholte waves as a function of wavelength (freq.)

The propagation velocity of Scholte waves is slightly different than that of Rayleigh waves for small wavelengths and gradually approaches the Rayleigh wave velocity for long wavelengths [*C.B. Park et al.*, 2005]. So in the long-wavelength limit, the influence of the water-layer is negligible and the Scholte wave velocity equals the Rayleigh wave. For smaller wavelengths, it is a modified version of the Rayleigh wave that is trapped near the fluid-solid interface [*Kugler et al.*, 2007].

Scholte waves are low frequency waves, and their frequency range is very narrow. Whereas Rayleigh waves contain frequencies of 5-10 to 50-60 Hz, Scholte wave frequencies range typically from 2 to 20 Hz. For shallow water conditions (water depth < 10 m), the majority of Scholte waves comply to the long-wavelength limit.

3.2 Scholte Wave Surveys

The procedure for Scholte wave surveys is commonly known as Underwater Multichannel Analysis of Surface Waves (U-MASW). It consists of:

1) Measuring seismic surface waves generated from the seismic source (acquisition)
2) Analysing the propagation velocities of those surface waves (Dispersion Analysis)
3) Inversion and -calculation of shear-wave velocity (V_s) profiles.

This method was originally developed as a land survey method in the nineties by the Kansas Geological Survey in order to estimate near-surface Shear wave velocities [*C. B. Park et al.*, 1999]. Owing to the similarities between Rayleigh and Scholte waves, MASW was later adapted to marine environments to characterize stiffness distribution of waterbottom sediments (e.g. [*Bohlen et al.*, 2004; *Kruiver et al.*, 2010; *Kugler et al.*, 2005; *Kugler et al.*, 2007; *Nguyen et al.*, 2008; *C.B. Park et al.*, 2005; *Shtivelman*, 1999] .

3.2.1 Data Acquisition

The goal is to record surface waves with the highest possible signal to noise ratio (S/N) over a wide frequency range, in order to identify dispersion of modal curves. High signal to noise ratio allows to separate different waves, different modes of surface waves, and allows to estimate uncertainties [*Maraschini*, 2008].

Scholte waves have a rather narrow low-frequency band, between 2-20 Hz. It means that sources with a low frequency signature should be preferred for their generation. This requirement is usually met using powerful sources, that also assure a good S/N ratio. It is generally recommended to use airguns for Scholte wave acquisition because those are powerful and generate adequate low frequency energy [*Diaferia*, 2012].

The distance between the source and the sea-floor is a key parameter for the occurrence of Scholte waves. In fact, the closer the source to the water-sediment interface (0-5 m range), the more effectively evanescent P-waves are converted into propagating surface waves [*Diaferia*, 2012]. Evanescent P-waves are a specific type of P-wave whose energy attenuates quickly. For deep applications, this wave is not relevant. For shallow applications, however, this type of wave is useful, because of the possible conversion to S-waves in the subsurface.

To register Scholte waves, either a multichannel streamer or Ocean Bottom Cable (OBC) can be used. Streamers, containing 24 or more equally spaced channels, are usually towed behind a ship at a constant water depth. When dealing with the acquisition of surface waves, the least contamination by body waves is desired; usually either 8-10Hz or 4-4.5 Hz hydrophones are used to enhance the Scholte wave signal. Experimental data, as well as a forward modelling example, demonstrate that hydrophones should be placed within 4-5 m from the seafloor in order to adequately acquire the entire Scholte wave's spectra. An example of a Scholte wave acquisition set up is shown in Figure .

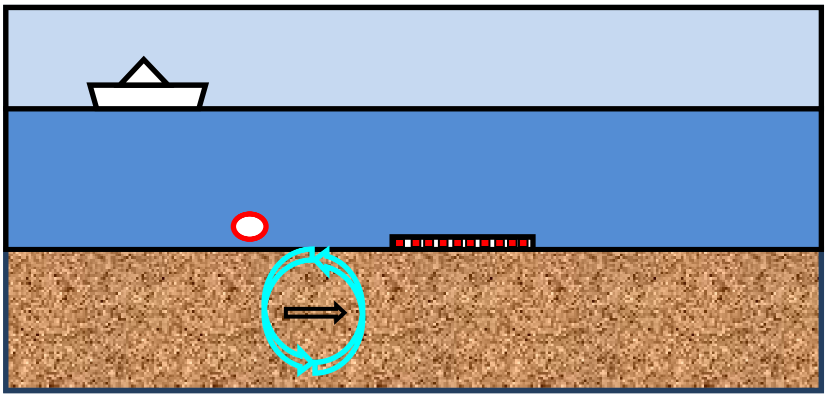

Figure 11. Acquisition set up for Scholte waves

At Deltares, Scholte waves were observed in several marine surveys, but usually by chance. During his MSc thesis research, [*Diaferia*, 2012] performed numerical experiments to obtain the optimal survey set up to detect Scholte waves. The influence of nearest offset, receiver spacing and number of receivers were investigated. The length of the streamer influences modal separation, dependence on lateral variation and can produce aliasing. For a fixed nr. of receivers, longer arrays improve the modal separation and reduce data uncertainties. Shorter arrays, however, are less sensitive to lateral variation in the subsurface.

Another important parameter is the receiver spacing (Dx). It has a strong impact on the minimum wavelength (and highest frequency) that can be recorded without aliasing. Since unconsolidated sediment in marine environments can have S-wave velocities as low as 50-100 m/s, even a 0.5-1 m spacing might be necessary [*Diaferia*, 2012]. Finally, the distance between the source and the first receiver must be a compromise between the signal to noise ratio, which decreases with distance, and the necessity of neglecting near field effects, which is improved with long distances. The best survey set-up according to [*Diaferia*, 2012] is to use a receiver spacing of 1 to 2 m, with nearest offset of 10 m and at least 48 channels.

Additionally, the acquisition time length must be long enough to record all the surface wave energy, and the sampling rate must take the Nyquist frequency into account. However, with the low frequency content of Scholte waves, aliasing in the time domain is never an issue.

3.2.2 Data Processing

The goal of the processing is to remove data related to body waves and to attenuate noise in order to be able to extract the dispersion curves from the recorded Scholte energy. Surface waves are defined as dispersive, because each phase travels with a different velocity depending on its frequency. Lower frequencies have higher penetration and are more sensitive to deeper (often harder) layers. As a consequence they travel faster. In contrast, high frequencies have scarce penetration and therefore travel with the (generally lower) velocity of the shallow part of the subsurface. This frequency dependence is difficult to observe in normal seismograms (distance - time domain, x-t), but can become obvious if data is transformed into other domains like the phase slowness – frequency (p-f) or velocity – frequency domain (v-f). An example of dispersive Scholte waves in x-t and v-f domain is shown in Figure .

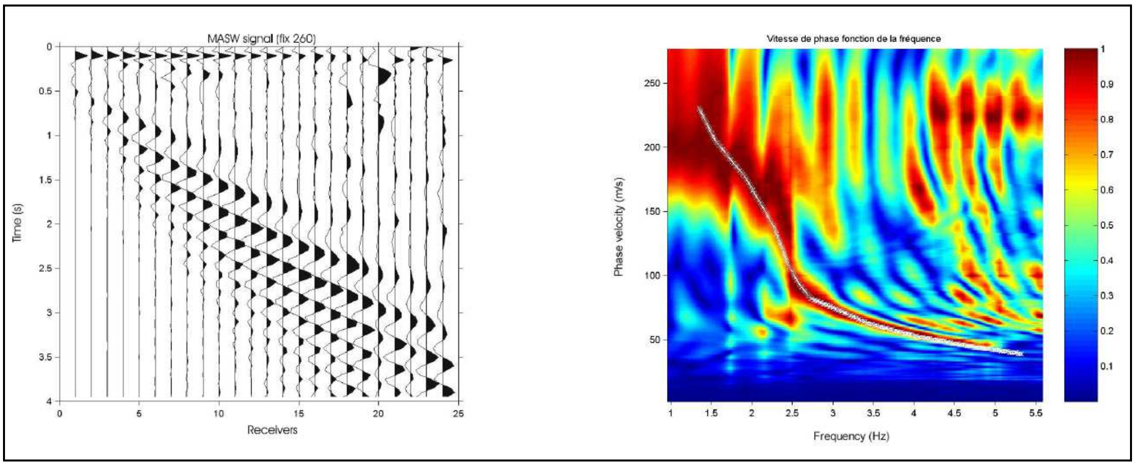

Figure 12. Field data and related dispersion curve. From [Paoletti et al., 2006].

3.2.3 Inversion

From the dispersion plots, information about the subsurface is to be inferred. For that, the appropriate subsurface model has to be found, representing the sediment structure which gave rise to the observed seismic wave propagation and hence the dispersive nature of the recorded wavefield. This has to be achieved by an inversion algorithm which finds a shear wave velocity profile whose dispersion curve, calculated by means of the forward model, is as close as possible to the real one (example see Figure). This process is repeated until the model and picked dispersion curve show a good match.

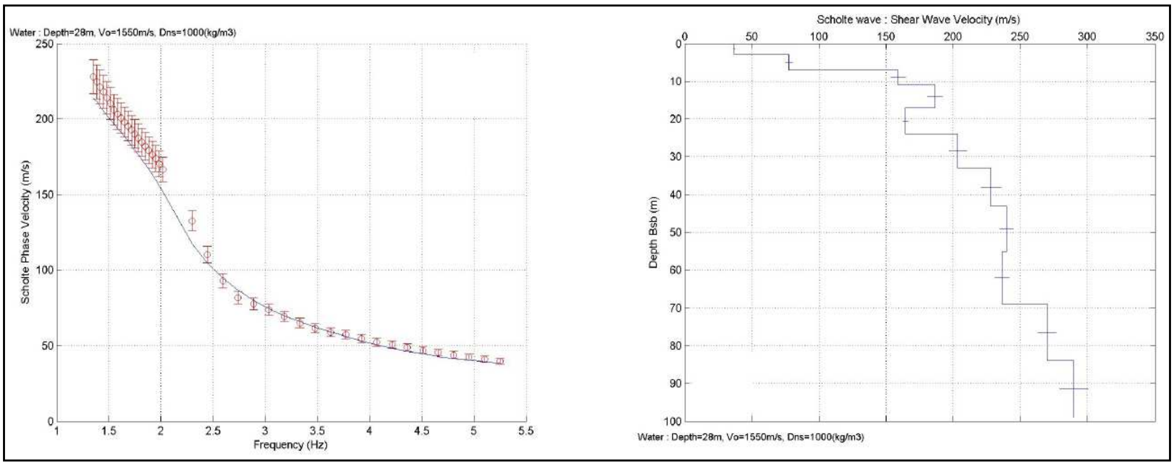

Figure 13. Dispersion curve picking and shear wave velocity profile. From [Paoletti et al., 2006]

There are several ways to search the model space for the model that "best" fits the data. Common algorithms are the nearest neighbourhood [*Wathelet*, 2008] or Monte Carlo e.g. [*Socco and Boiero*, 2008].

Inversion of the Scholte wave dispersion curve to a V_S profile requires a proper modelling scheme that accounts for the existence of the water layer above sediments. However, considering that the maximum deviation of V_{Sch} from V_R is usually less than 5 %,that correction usually falls below the uncertainty level of the measurement. Treating Scholte waves as identical to Rayleigh waves during the inversion analysis does not significantly degrade the confidence level of the calculated V_S profile for the soft substrate case (Park, 2005).

Once all locations have been carefully processed and its corresponding shear wave velocity profile obtained, a section showing shear save velocity variations along the profile is produced.

Figure shows an example of a shear wave profile derived from airgun and OBC data in the Danube River in Hungary. The top layer has velocities ranging from 400 to 540 m/s and is interpreted as a muddy layer. From south to north the thickness of the top layer consistently increases. The second layer is a consolidated clay with velocities from 400 to 850 m/s. At a depth of 15 to 20 m the sharp contrast in shear wave velocities coincides with the transition from clay to sand observed in a borehole at the survey site.

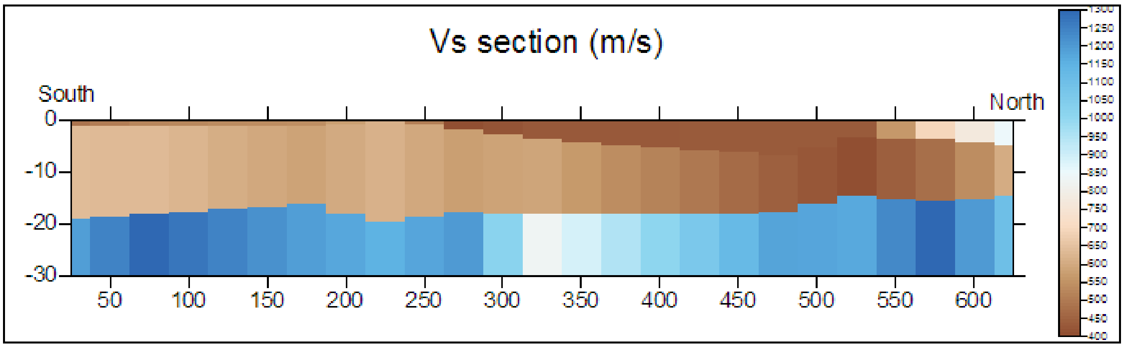

Figure 14. Profile showing Shear Wave Velocities along the acquisition line, example from Danube river, Hungary [Kruiver et al., 2010]

Other examples come from the research group at Kiel University (Germany), e.g. [*Bohlen et al.*, 2004; *Klein et al.*, 2005; *Kugler et al.*, 2005; *Kugler et al.*, 2007]. In [*Bohlen et al.*, 2004], they present shear wave velocity profiles for a site in the Baltic Sea . The data were acquired using airgun and Ocean Bottom Seismometers (OBS). The combination of 1D inversions gives information on the lateral variations (2D), the result is called 1.5 D. Generally, they observe a 3 layer model with V_s~ 250 m/s in the top layer, 300 m/s in the second layer and 300-350 m/s in the bottom layer, up to approximately 35 m below the seafloor (Figure).

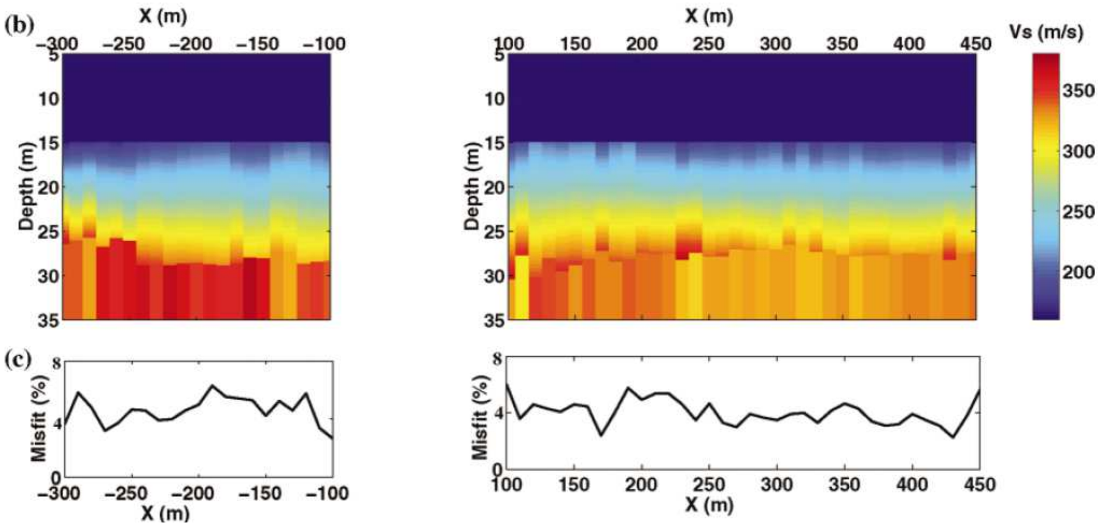

Figure 15. Example of V_s profiles from a location in the Baltic Sea. b) Inverted 1D VS models for the upper 35 m. c) Misfit between measured and modelled dispersion curves according to equation 6 in [Bohlen et al., 2004]

This approach is extended to 3D in [*Kugler et al.*, 2007]. For a site in the Baltic Sea, the Scholte waves were excited by air-gun shots in the water column and recorded at the seafloor by ocean-bottom seismometers as well as buried geophones. To acquire a 3D image of in situ shear-wave velocities, reference phase slowness-maps and residuals at different frequencies were calculated. From that, a model of the depth-dependency of shear-wave velocities for every location is obtained. A slice of the 3D result is show in Figure 6. In the top four meters, where fine muddy sands can be observed, V_s is very low (60–80 m/s). Below that, V_s exceeding 170 m/s is found for the silts and sands. The upper edge of glacial till is situated approximately 20 m below the seafloor, showing V_s of 300–400 m/s. A sensitivity analysis reveals a maximum penetration depth of about 40 m below the sea bottom and that density may be an important parameter, best resolvable with multimode inversion.

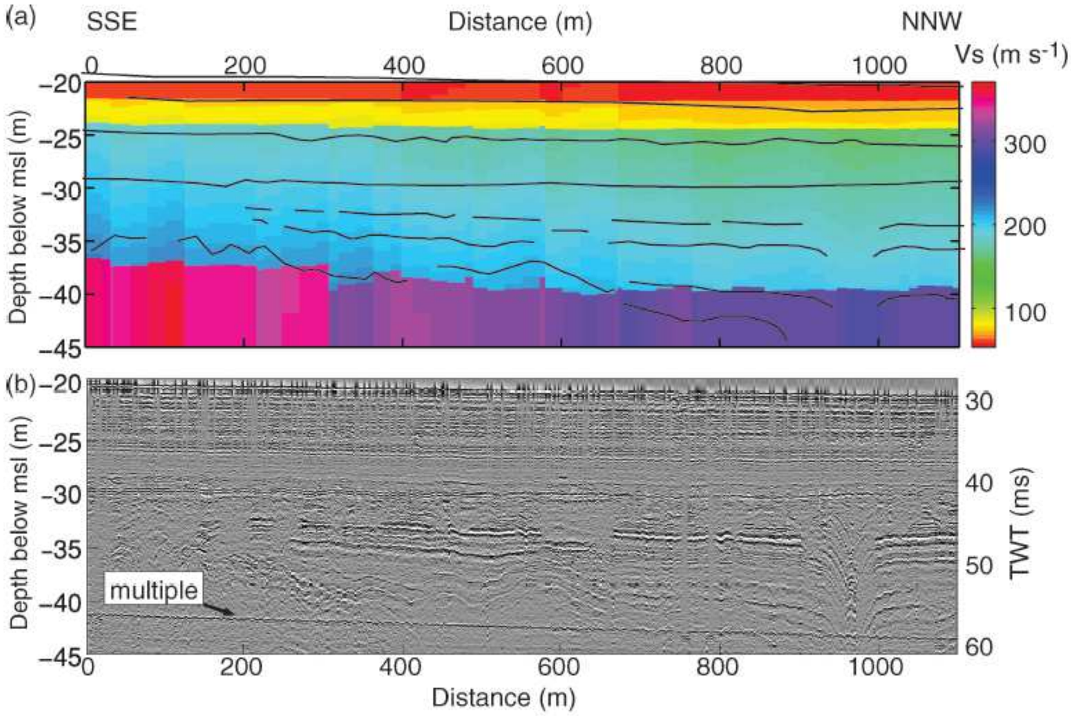

Figure 16. Example of slice through 3D-model of shear wave velocities for a site in the Baltic Sea, Northern Germany, from [Kugler et al., 2007]. a) vertical slice of Vs model, black lines indicate the main reflections of the boomer data shown in b).

3.3 Advantages and limitations of Scholte waves

In comparison to using conventional body-wave methods to achieve similar V_s information the surface-wave method has several advantages:

- Field data acquisition is very simple, because surface waves are very energetic and always represent the strongest energy of the seismic record.
- Data processing procedure is relatively simple.
- No special field equipment is required.
- Because of all above reasons, it represents a cost effective and time efficient methodology to obtain shear wave velocity profiles of the area of investigation.
- Scholte waves are not affected by acoustic masks like shallow gas.
- If combined with information from seismic refraction, sonic profiling or other methods to obtain P-wave velocities, elastic modules can be derived.

On the other hand, a surface-wave survey presents important limitations with respect to conventional body wave imaging techniques:

- Since Scholte waves are mainly low frequency, they can only provide information about shear wave velocities of mayor layers but thin layers are undetected. Hence, vertical resolution is very poor compared to shear wave imaging.

- Long record lengths are required, because of the low V_s velocities. This constrains the offset and speed of the boat. Short record lengths do not permit recording the slower part of surface waves. The error in receiver positions due to ship movement during recording becomes increasingly large for lower boat speeds, because of less control on streamer positions. For high boat speed, binning errors occur (general problem in seismic processing), which is increasingly severe for long records.
- The inversion process must be performed keeping in mind the geological setting. Without this, inversion algorithms can produce velocity profiles that are mathematically accurate but geologically unrealistic. With this, there is a trade-off between resolution in dispersion curves (longer spreads needed) versus lateral resolution (shorter spreads for better resolution).

3.4 Recommendations for Scholte waves

Producing and recording pure shear waves (body waves) require special equipment. Its theoretical resolution benefits in comparison to compressional wave reflection are not easily accomplished, because it is difficult to produce high frequencies shear wave reflection in most near-surface settings. Surface waves may therefore provide an easier (and cheaper) alternative to obtain shear wave velocities.

Surface wave profiling does not produce a high resolution image of the buried layers. It provides an estimate of the velocity of the major layers. However, in conjunction with information obtained from conventional seismic reflection data, we can obtain information about those areas that are “invisible” for compressional waves.

In archaeological investigations, surface waves are expected to be useful in areas where compressional waves suffer from the presence of gas.

4 Resistivity methods

4.1 Introduction

Electrical Resistivity Tomography (ERT) is a well-known geophysical method applied on land. ERT enables the characterization of the subsurface by injecting electrical DC currents and measuring voltage differences at the surface. By measuring the voltages on the surface, the flow pattern of the electrical currents can be derived. From this, the electrical resistivity can be calculated. The electrical resistivity is mainly related to the amount of water/fluid in the pore space, the conductivity of that water/fluid, and the properties of the material surrounding the water. Measurement of resistivity is usually done with four electrodes, two for the current injection and two for the voltage measurements (Figure 17). The distance between the electrodes determines the depth range to which the method is sensitive. Combining multiple electrode combinations/distances enables us to gain enough resolution over the depth range of interest.

Data are then processed (inversion) in order to retrieve the true distribution of resistivity in the subsurface. This method is mainly employed on land especially for environmental, geotechnical and engineering purposes.

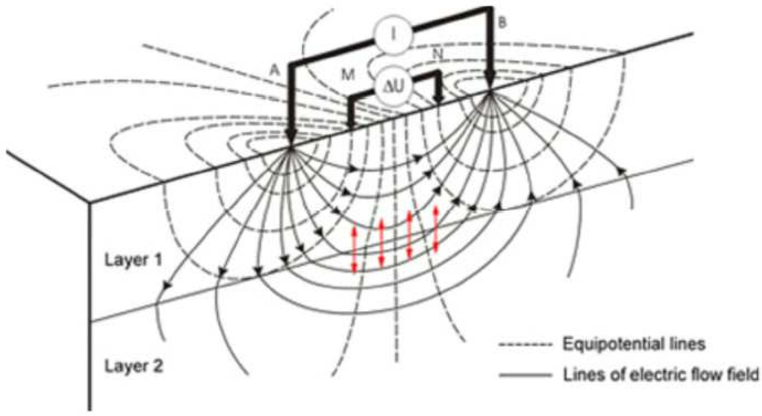

Figure 17. Current flow and equipotential lines generated by two injection electrodes (A and B). Resistivity measurements are obtained from potential difference measurements between the M and N electrodes.

The application of the electrical resistivity method to marine environment is known since the 1980s, but still not common, as confirmed by the lack in literature. Nevertheless, the use of the electrical resistivity method can represent an alternative or complementary procedure for shallow water investigation.

For marine applications, the electrodes are contained within a long cable that is towed behind the survey vessel. Different source-receiver electrode configurations can be applied. Both floating cables and cables towed over the bottom can be used. The advantage of the latter is that it increases the penetration depth and reduces the influence of the (highly conductive) water layer. Through a dense network of 2D lines a 3D resistivity volume may eventually be obtained.

In marine environments, acoustic measurements can be largely affected by the presence of biogenic gas in the sediment matrix. Usually, obtaining an accurate image in areas with gas accumulation is rather difficult. The ERT method can be employed to fill in the gaps in the acoustic record [*Tarits et al.*, 2012].

4.2 ERT for remnants identification

Even if it is not a common practice, the resistivity method can be used in order to identify buried artifacts. Usually, those consist either of metal or wood giving rise to a low and high resistive anomaly, respectively. [*Passaro*, 2010] show a successful application of this method in shallow water for the identification of a buried shipwreck. Resistivity data were acquired with floating electrodes (1 m below sea-level) assuring an easy and fast acquisition with no risk of damaging the cables due to obstacles on the seafloor (Figure). The inversion of the acquired data revealed the presence of a high resistivity anomaly (see Figure) caused by the presence of a military shipwreck, as confirmed by both magnetic and acoustic investigation. [*Passaro*, 2010] suggest that ERT will be successful for large wooden or metal objects, such as shipwrecks. We expect that small objects cannot be detected, the objects should be tens of m in size.

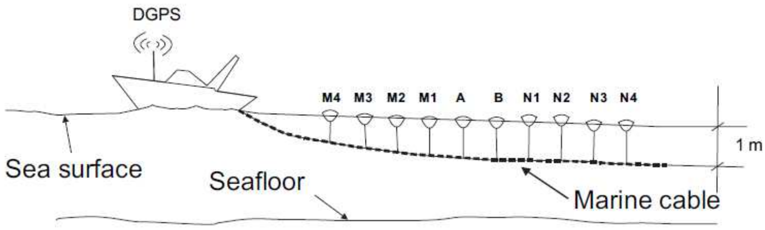

Figure 18. Acquisition set up for an Electrical Resistivity Tomography (ERT) in shallow water. A and B are the injection electrodes while M and N are the potential electrode. From [Passaro, 2010].

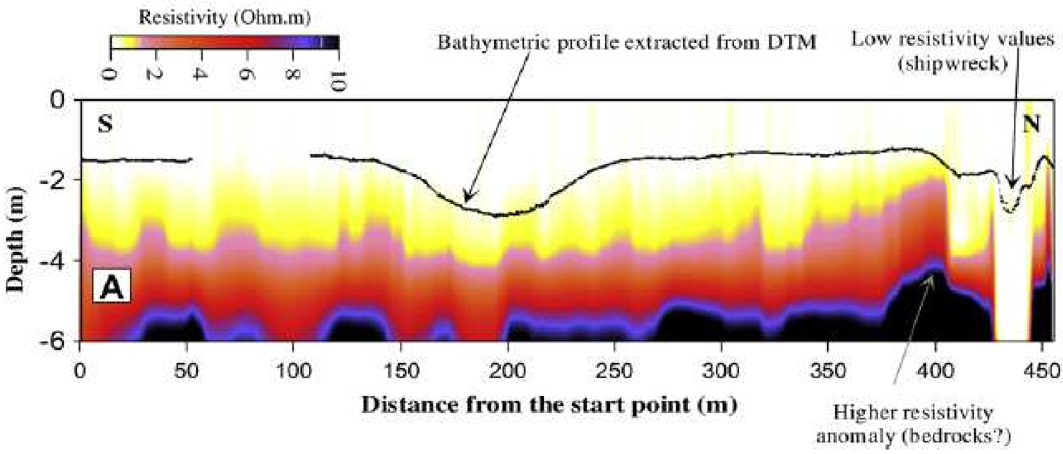

Figure 19. Electrical resistivity profile acquired in shallow water. The very low resistivity anomaly on the right is caused by the presence of a buried military shipwreck. From [Passaro, 2010].

4.3 ERT for geological characterization

In the marine environment, ERT can be also employed to geologically characterize the subsurface. Since different lithologies have different resistivity values, the recognition of sand versus mud is expected to be feasible, especially if combined with other geophysical methods.

Generally, resistivity depends on the porosity of rocks. Sand and clay show lower resistivity values (1-100 Ohm*m) compared to less porous rocks like limestone, granites (> 1000 Ohm*m). Within the same type of rocks, cementation leads to a decrease in porosity and therefore in resistivity. That is why ERT is often used to discriminate different lithologies as well as detecting depth and structure of bedrock. Also for the determination of the groundwater table ERT is commonly applied. Determination of lithologies can be carried out in a marine environment as well, even if this practice is not common.

Due to the presence of highly electrically conductive seawater, either a high current generator and/or highly sensitive electrodes need to be employed. In [*Tarits et al.*, 2012] an example of application of ERT with low power and highly sensitive sensors is reported. The electrodes were deployed along a cable that was towed behind a ship; the two electrodes used for current injection were placed at the ends of the cable. The acquisition set-up was flexible and adjustable to the target depth (longer cable length for deeper penetration). The method was demonstrated to correctly image the highly resistive, calcareous bedrock within a depth of 10-15 m (Figure). Limestone is characterized by a large value in resistivity; it means that it gives rise to a strong electric anomaly, easily detectable and recognizable.

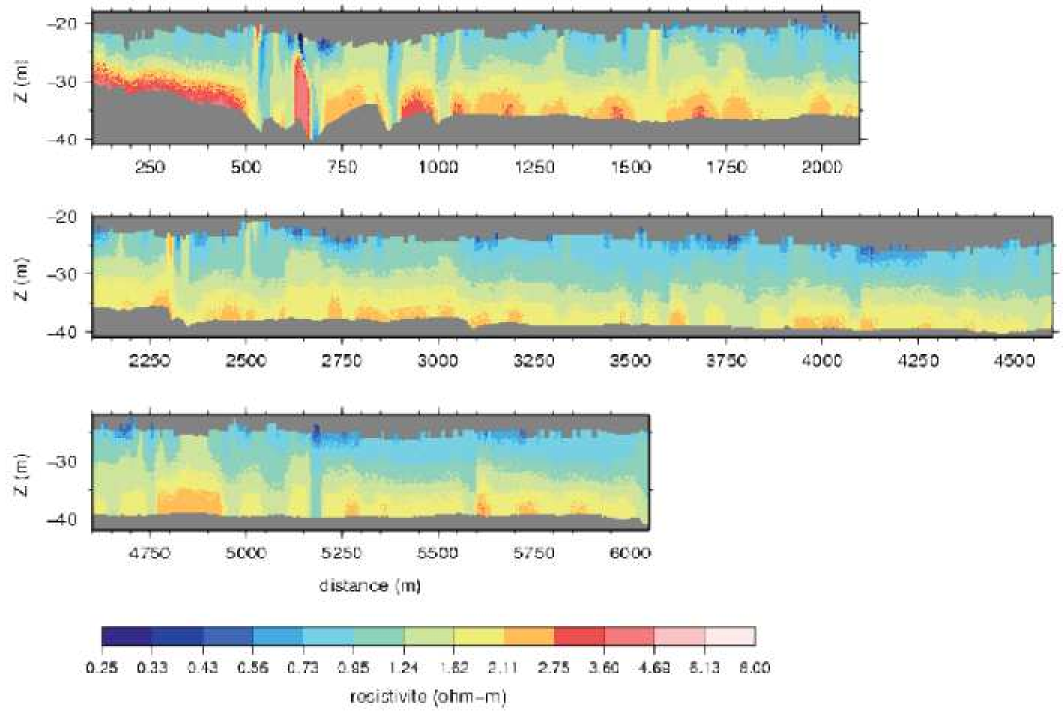

Figure 20. Example of resistivity profile acquired over a marine karst area. The red regions are interpreted as the highly resistive, calcareous limestone overlaid by muddy/sandy sediments (blue). From [Tarits et al., 2012].

For archaeological investigations, the discrimination between the unconsolidated and consolidated sediment is often of major importance. The employability of the ERT method for this scope is not assessed. There is very little mention of the application in literature. However, the depth range of interest is feasible and might contribute to the image of the subsurface in areas with shallow gas. We therefore suggest to try this method in conjunction with other geophysical methods.

4.4 Recommendations on resistivity methods

Underwater ERT can be used for buried (metal) shipwreck identification or for geological characterisation of the subsurface. In this project, the latter seems to be the most promising. ERT covers the depth range for which acoustics can encounter problems with the presence of shallow gas.

For sufficient penetration the electrodes need to be employed close to the seafloor. So far, the method has been employed mostly in shallow water (mostly for practical reasons i.e.

easier deployment of cables and electrodes). The resolution of the method is in general defined by acquisition set-up. As rule of thumb, the maximum resolution is equal to the electrode spacing. In general, this means that maximum resolution is on scale of one or several meters. Resolution decreases with depth, but the lateral resolution is in general good and adequate to identify lateral changes in lithology.

The processing and interpretation of the data is not complicated and comparable to land ERT. For underwater application, the only difference is the presence of the seawater layer, whose influence can be accounted for in the processing phase by standard software.

However, there are still questions whether the method is efficient for the specific case of the southern North Sea (only unconsolidated sediments). The main concern is the capability of this method to distinguish between different lithologies that do not give rise to large resistivity contrasts. For areas with shallow gas, however, ERT might give additional information on the structure of the subsurface that will otherwise be undetectable. In case of employment of the ERT method, it is recommended to use this method only in conjunction with other geophysical methods.

5 EM methods

5.1 Introduction

The use of electromagnetic fields in geophysics is common on land for the detection of conductive items (tanks, buried metal object etc.) as well as for the geological characterization. The method is also based on the measurement of the electrical resistivity, but in a different way from the previous ERT method. In EM surveys, a magnetic field is generated by a coil, which induces a secondary magnetic field. The amplitude and phase variations of this secondary field allow for the calculation of the conductivity/resistivity of the subsoil. Similar to electrical resistivity surveys, also here the resolution will decrease with the target depth and the source-receiver configuration will determine the penetration depth.

For investigation of the deep subsurface, magnetotelluric currents are used as well (see review paper [*Constable*, 2013]). Magnetotelluric currents are caused by EM waves generated by the Earth's magnetic field variation. Those low frequency waves can be also generated by a controlled EM source. However, these magnetotelluric currents are not relevant for the characterisation of the shallow subsurface needed in this project.

Controlled Source ElectroMagnetic (CSEM) surveying is a method used to map commercial-scale hydrocarbon reservoirs from the seabed. It is often used in conjunction with seismics. In this combination, seismics provide structural information, whereas CSEM provides information on pore fluid resistivity.

In CSEM surveying for hydrocarbon purposes, a powerful horizontal electric dipole is towed about 30 m above the seafloor. The dipole source transmits a carefully designed, low-frequency electromagnetic signal into the subsurface. EM energy is rapidly attenuated in conductive sediments. However, it is attenuated less and propagates faster in more resistive layers such as hydrocarbon-filled reservoirs. A typical acquisition set-up for this method is shown in Figure .

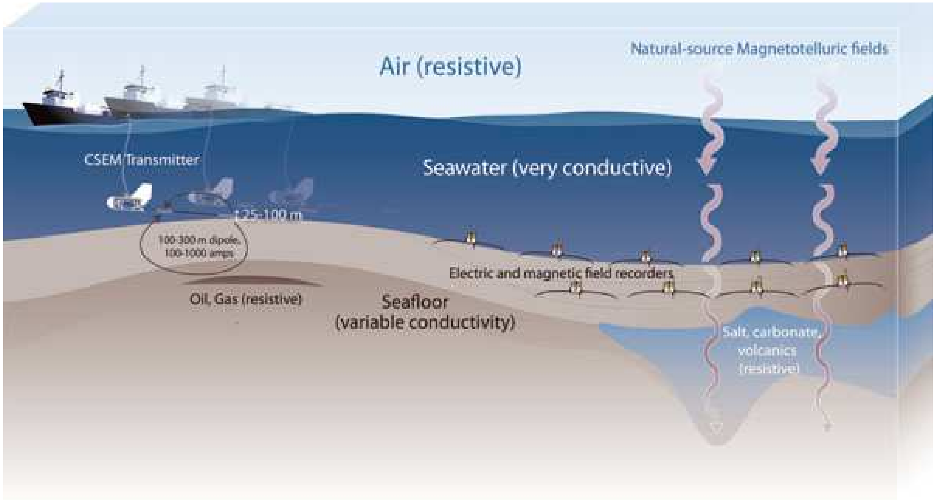

Figure 21. Acquisition set-up for CSEM survey in a marine environment. Usually cables are deployed over several km in order to assure a sufficient penetration for hydrocarbon prospecting. From Constable, 2013.

5.2 CSEM for shallow applications

In [*Evans*, 2007], the CSEM method has been successfully used in a large water depth range (from 10 to 1300 m) for the characterization of the shallow subsurface (first 20 m). The used system consists of a transmitter coil that is towed behind the ship; the coil is in contact with the sea-floor. Three receiver loops are towed at different distances from the transmitter coil, in order to collect data from different depths. The coils need to be in contact. The equipment is tailor-made. Only a limited number of research groups have access to the equipment.

Figure shows an example of porosity values retrieved from resistivity values [*Evans*, 2007]. The resistivity values of the subsurface can be related to the porosity of sediments and therefore to their lithology. The profile in Figure shows a buried low-porosity layer at a depth of about 5 m and with a thickness of about 10 m underlain by a higher-porosity substrate. [*Evans*, 2007] suggests that the method could be particularly promising in the identification of palaeochannels, since their infill is usually characterized by higher porosity values compared to the surrounding sediments (Figure). Since (palaeo)channels are relevant for archaeological studies the EM technique might be helpful here. We need to keep in mind, however, that the lateral resolution is limited. The channel in Figure is approximately 100 m across.

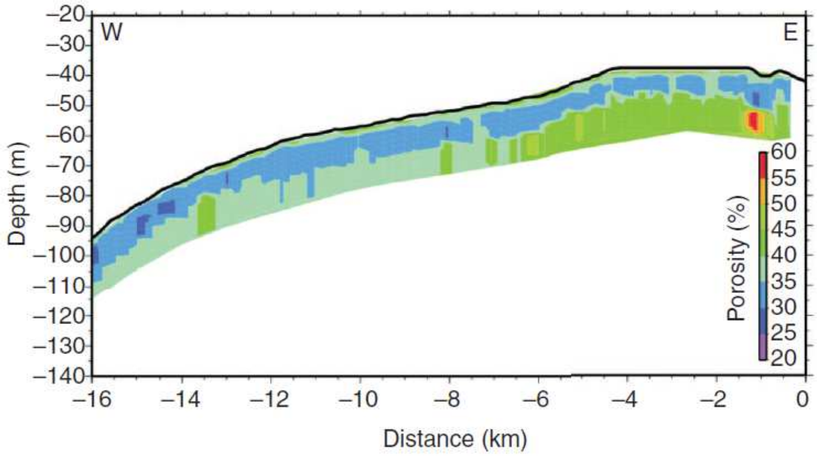

Figure 22. Porosity section derived from resistivity data acquired with CSEM method. From [Evans, 2007].

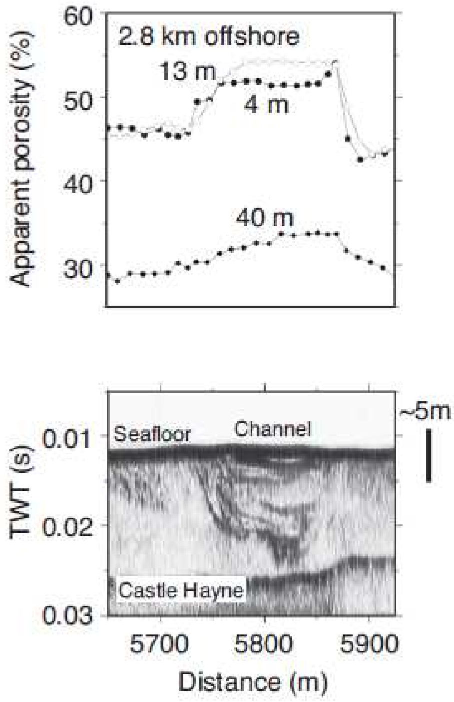

Figure 23. Apparent porosity profiles with 3 different receiver distances (top panel). The increase in porosity is due to the presence of a coarser-grained channel infill, to be seen in seismic section in the bottom panel. From [Evans, 2007].

A similar application of the CSEM method for shallow subsurface characterization is shown in [*Müller et al.*, 2012]. A specially designed EM profiler has been employed here to record simultaneously conductivity and magnetic susceptibility of the sediments. The uncertainty of data interpretation is largely reduced by the availability of those two independent measurements. The profiler strictly requires a good coupling with the sea-floor (max. 20 cm distance) because of the high damping of the EM field due to the (highly conductive) sea water. The magnetic susceptibility method is mostly sensitive to the upper 1 meter below the sea-floor, the EM method to a depth range related to the coil spacing. The data in this example, require measurements of conductivity and susceptibility on cores as reference in order to be processed. The measurements of resistivity were related to sediment porosity. Figure shows the combined results: the sediment porosity shows a clear distinction between sandy and silty/clayey deposits (in the first 10 m). This is an example of shallow CSEM for a large scale structure.

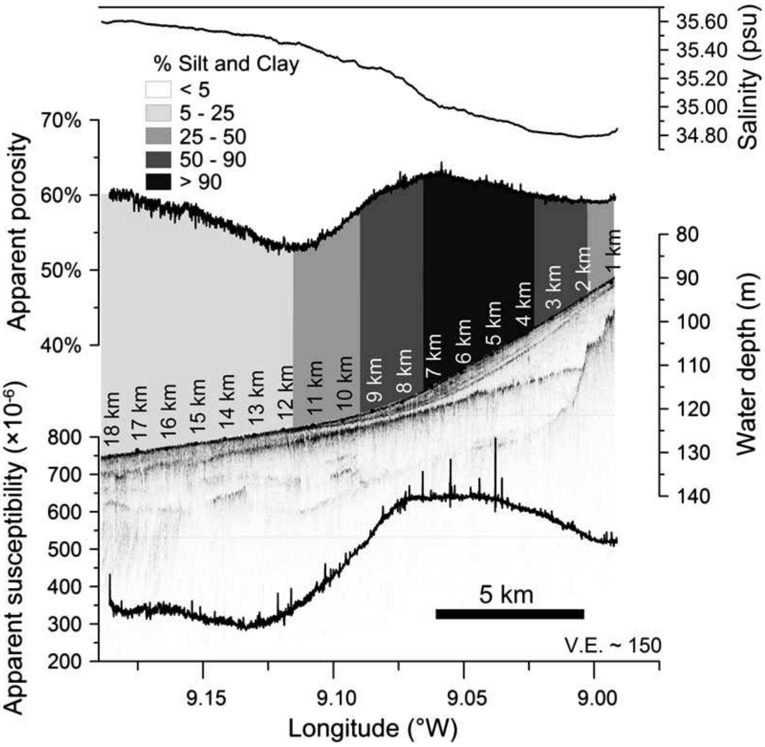

Figure 24. Diagram showing susceptibility, porosity (derived from resistivity data) and the corresponding boomer seismic profile. Porosity and susceptibility allowed deriving silt and clay content along the profile. From [Müller et al., 2012].

As already pointed out in [*Evans*, 2007], the CSEM method for shallow subsurface characterization requires receivers and transmitters loops to be placed as close as possible to the sea-floor. [*Müller et al.*, 2012] suggest to use a strict value of 20 cm as maximum distance from the seafloor. It is clear that this requirement represents a major issue while surveying due to the presence of obstacles and irregularities on in the seafloor.

5.3 Recommendations on EM methods

The EM method has been applied successfully several times in shallow water environments to detect shallow structures of varying resistivity/ porosity. The applicability of the method seems to be most appropriate for the identification of large-scale (km-size) buried palaeochannels.

The receiver loops should be placed close to the seafloor, representing a practical issue. An example of a specifically designed EM profiler (not commercially available) is given in the literature. It seems to be a fast method for data collection over long distances. It is mostly sensitive to the first meter of sediment from the sea-floor. It strictly requires a good coupling (max 20 cm distance) with the sea-floor, requirements that cannot be always met due to sea-floor condition. According to the literature, processing is relatively easy. The most important drawback of this method is that equipment is not off-the-shelf. If we want to use the EM method, we have to arrange cooperation with relevant research groups. Then the EM method should be used in conjunction with other geophysical methods.

6 Magnetic methods

6.1 Introduction

Measures of anomalies in the magnetic field are broadly employed for the identification of ferromagnetic object in the subsurface. This method can be successfully employed in marine environment as well. In this set-up, a magnetometer/gradiometer is towed behind a vessel allowing continuous recording of the magnetic field. After relatively easy processing, the resulting magnetic anomalies provide information on the location (and sometimes also depth) of ferromagnetic objects.

Magnetic surveys at sea often involve the use of two (sometimes more) spatially separated sensors to measure the gradient of the magnetic field (the difference between the sensors). This so-called gradiometric method provides a better resolution and is able to detect smaller phenomena than single-sensor methods. Magnetometers may also use a variety of different sensor types. Proton magnetometers have largely been superseded by faster and more sensitive fluxgate and cesium instruments. Marine magnetometers can either be towed at the surface or near to the bottom. In both cases, a sufficient distance away from the ship to avoid pollution from the ship's magnetic properties. The resolution of the magnetic/ gradiometric data will mainly depend on the type of magnetometer sensor, the number of sensors, and the tow depth (closer to the bottom = higher resolution).

6.2 Examples of magnetic methods in underwater archaeology

The main application of magnetic/gradiometric methods in underwater archaeology is related to (buried, partially buried or exposed) shipwrecks (e.g. [*Quinn et al.*, 2002]). In general, the magnetic investigation is part of a suite of methods used to study the site. Figure shows the spatial distribution of magnetic anomalies over a survey site, together with a single line measurement of magnetic data. On the side scan sonar image, the shipwreck is clearly visible, as it sticks out of the seafloor. The sub-bottom profile (Chirp) shows that the shipwreck rests on a hard surface and is surrounded by mud.

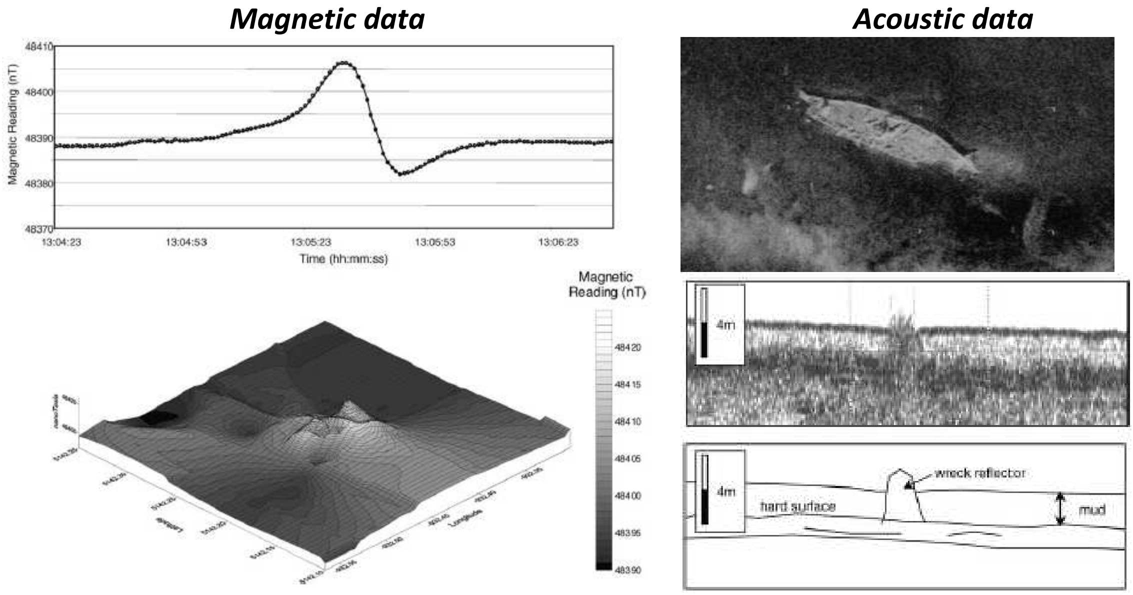

Figure 25. Example of magnetic data on a shipwreck (left), with SSS image (right, top) and Chirp data (right, bottom). From [Quinn et al., 2002].

Another successful example of application of the magnetic method for archaeological purposes is found in [*Weiss et al.*, 2007]. A high resolution, magnetic survey was carried out for the identification of a buried aircraft. A magnetometer/gradiometer was towed behind a catamaran whose position was accurately recorded by DGPS. High spatial resolution was achieved by applying a high sampling rate, the low noise level of the instrument and the accurate positioning. In an area of 3.5 km^2 the method allowed to find several magnetic anomalies (Figure). The aircraft was successfully detected as well as several other buried or partially buried ferromagnetic object (metal parts, a 5th century Byzantine iron anchor, etc., see Figure), as confirmed by inspection by divers and direct probing.

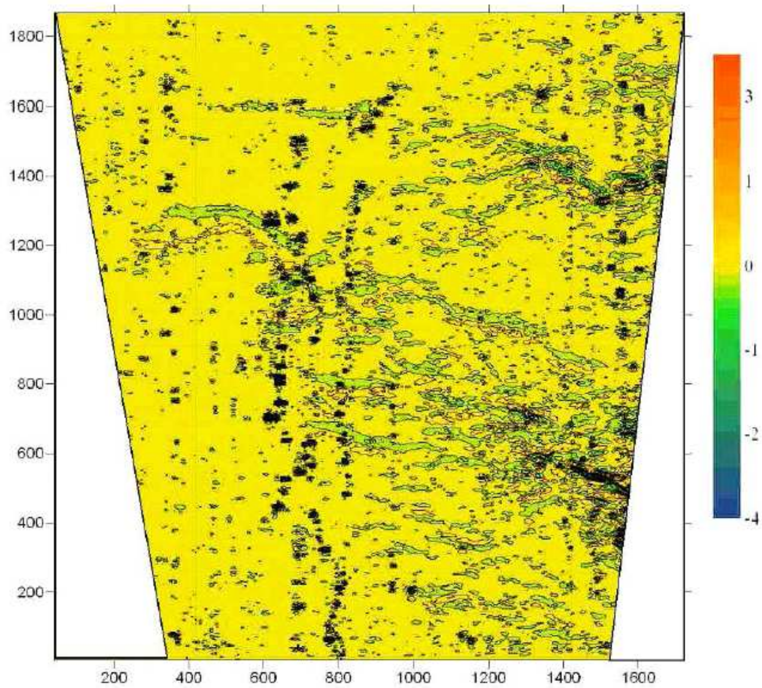

Figure 26. Map of magnetic anomalies in a shallow water environment. Scale in m on both X and Y axes. Anomalies are shown in green. From [Weiss et al., 2007].

Figure 27. Left: 5th century Byzantine iron anchor detected as one of the magnetic anomalies. Right: Example of an aircraft's dud of an found under 1 m of sand. From [Weiss et al., 2007].

6.2 Recommendations on magnetic methods

The main benefit of the magnetic/gradiometric method lies in the combination with other geophysical methods (i.e. seismic and side scan sonar) in order to reduce uncertainty in the object identification. If ferromagnetic objects such as shipwrecks or remnants of shipwrecks (e.g. anchors) are expected to be present at the site, then the magnetic method is able to contribute to the full description of the site. In our project, however, we are mainly focusing on buried features. In that case, the main benefit from magnetics is in gas-rich areas (where we lack sub-bottom profiling data) and/or in iron-containing artefacts that may be missed out by acoustics. In view of the scope of the project (sub-seafloor archaeology) the magnetic/gradiometric method is only appropriate when expecting buried iron objects. On the other hand, magnetometer data are easily acquired simultaneously with seismic data.

7 Ground Penetrating Radar (GPR)

7.1 Introduction

The use of electromagnetic waves in the frequency band of 50-500 MHz is common and well established in geosciences for applications on land. Usually this method, called Georadar or Ground Penetrating Radar (GPR), is employed for hydrological purposes, ice thickness evaluation as well as several environmental and engineering purposes. In recent years this method is increasingly employed for archaeological investigations on land as testified by the large literature available on the topic (e.g. [*Nuzzo et al.*, 2009] [*Novo et al.*, 2008] [*Grasmueck et al.*, 2005]).

The principle of the GPR method is fairly simple: short electromagnetic waves from the radio spectrum (UHF and VHF frequencies, tens of MHz up to several GHz) are sent into the subsurface by a transmitting antenna. The signal travels with a velocity which depends on the medium (water content and matrix characteristics). In fact, the principles involved are similar to reflection seismics, except that electromagnetic energy is now involved instead of acoustic energy, and reflections appear at boundaries with different dielectric constants instead of acoustic impedances. The reflected electromagnetic wave is then recorded by a receiver antenna. This working principle is shown in Figure .

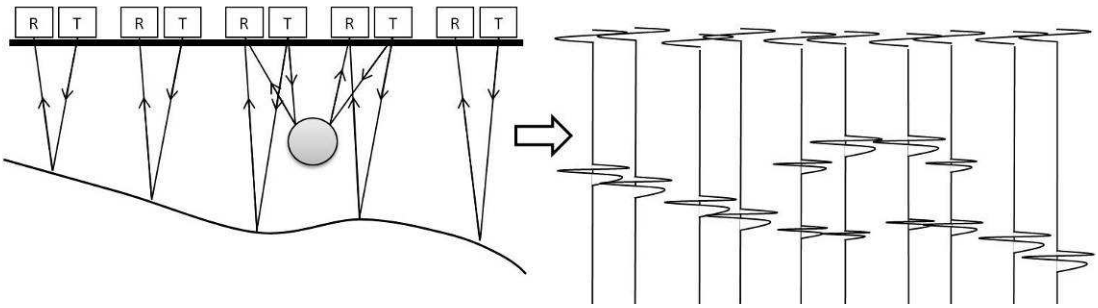

Figure 28. Left: Working principle of the GPR method. An EM wave is generated by a transmitting antenna and sent into the ground. The presence of a layer or object leads to the generation of a reflected wave (recorded by a receiver antenna). Right: example of a radargram.

The depth range of GPR is limited by the electrical conductivity of the ground, the transmitted center frequency and the radiated power. As conductivity increases, the penetration depth decreases. Lower frequencies can reach depths up to tens of meters (e.g. 100 MHz can travel up to 20 meters) with a resolution of tens of centimeters, while higher frequencies can give a resolution of centimeters but up to depths of several meters.

The size and easy use of the instrument as well as the reliability of the retrieved data allow the coverage of large areas in a very cost-effective way, even in a gridded 2D fashion. Moreover, the method is largely non-destructive and particularly suitable for archaeological purposes. In Figure an example of radargram for the detection of a buried grave is given.

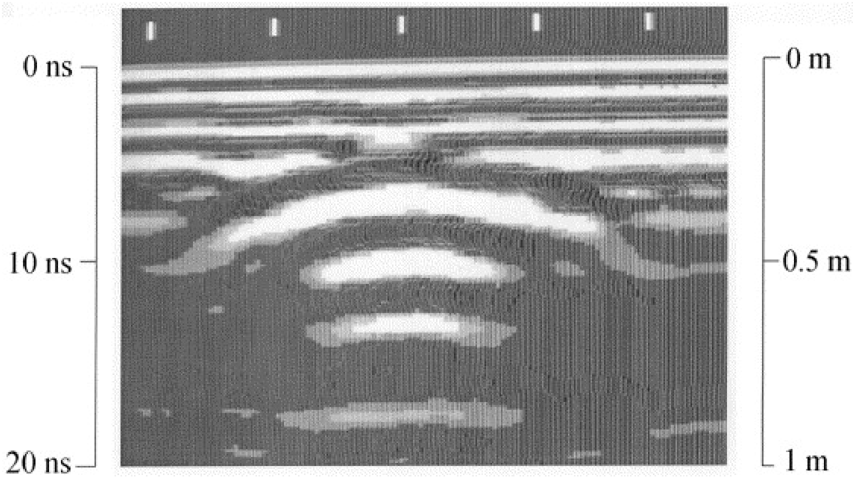

Figure 29. Portion of a radargram acquired in Valencian Cathedral (Spain). A main, clear reflection is identified at 0.25 m and caused by a grave under the church floor. From [Pérez Gracia et al., 2000].

7.2 Underwater GPR

The application of the GPR method underwater has had only limited success in the past. The generated EM wave is strongly damped by the presence of water. As a result, the amplitude of the transmitted signal as well as the reflected signal is very low.

GPR has been used for object detection on a freshwater river in the Netherlands (personal communication Pauline Kruiver). A land GPR (500 MHz) was encased in a waterproof box and attached to a pole at the side of a vessel. To avoid attenuation of the signal, the GPR was kept at a maximum of 1.5 m above the water bottom. The resulting data allowed to detect transects of telecom cables, pipes and gas tubes to a depth of up to 2 m below the water bottom.

Another example of underwater GPR is in the IJssel project of Deltares (the Netherlands). In that case, a GPR was towed behind the survey vessel in a rubber boat. No GPR results could be retrieved, because of the attenuation of the signal; the water depth was too large. A successful application of underwater GPR comes from Gent University (Belgium) for a location in the Danube River, Hungary (personal communication Tine Missiaen). This data is not published. An example of an underwater radargram is given in Figure .

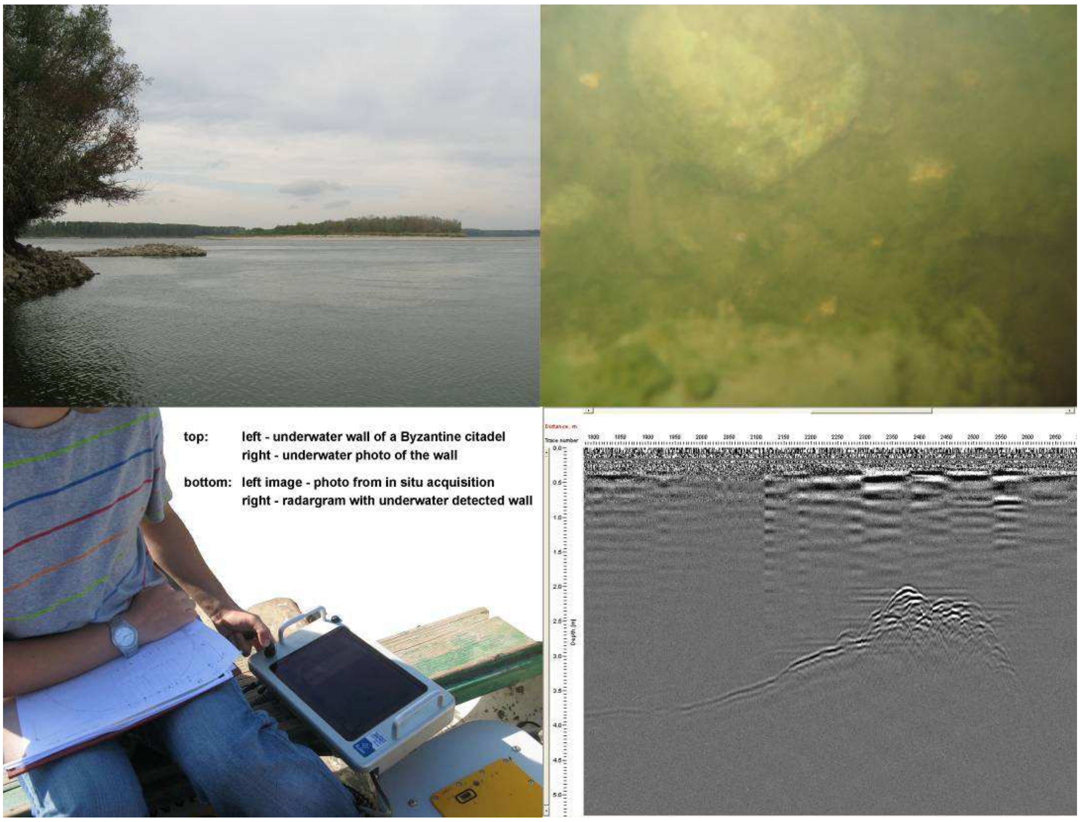

Figure 30.GPR on an underwater site of the Danube river, Hungary.

The application of GPR in salt water environments is even more problematic than in fresh water. Due to the high conductivity of salt water, the damping is so severe that no EM wave can penetrate the subsurface. In the 1990s, the company Groundtracer developed a prototype of a saltwater GPR. Although the results of this newly developed instrument were promising, it never reached the GPR market.

7.3 Recommendations on underwater GPR

In view of the above findings it is recommended not to include GPR in the techniques used in this project.

8 General conclusions and recommendations

Several geophysical, non-conventional techniques for underwater application have been presented assessing the advantages, limitations and the possible use in the project. The techniques are non-conventional either because of a new application of acoustics (shear waves, Scholte waves) or because of the limited experience of land based techniques for marine applications (EM, resistivity, magnetic, GPR). The techniques are summarised in Table .

Method	Application	Remark
Shear wave reflection	Detect subsurface structures, e.g. geological layers, similar to SBP but with higher resolution	Promising technique, because of resolution and insensitivity to gas. Acquisition by airgun and streamer, close to the seafloor. Sophisticated processing needed.
Scholte waves	Detect subsurface structures, e.g. geological layers	Low spatial resolution. Suitable for areas with gas. Acquisition by airgun and streamer, close to the seafloor. Easy processing of data.
Resistivity method	Detection of subsurface structures and detection of (large metal) objects	Promising technique, but application for unconsolidated sediments is not proven. Electrodes need to be close to seafloor. May provide extra information in areas with gas.
EM method	Detect subsurface structures of varying porosity, e.g. channels	Promising technique but application for unconsolidated sediments is not proven. Coils need to be close to seafloor. May provide extra information in areas with gas.
Magnetic method	Detection of ferromagnetic objects on seafloor and buried	Only for ferromagnetic objects and therefore limited application.
Ground penetrating radar	Detection of subsurface structures, e.g. geological layers and detection of objects	Not available for salt water environments.

Table 2. Summary of non-conventional geophysical techniques for underwater archaeology

Shear wave imaging can be enormously beneficial to investigate shallow sediments in areas like the Belgian North Sea where shallow gas accumulation are common and no information can be obtained from conventional P-wave reflection surveys. Shear wave imaging should be conducted in conjunction with classical P-wave reflection techniques. The processing of shear wave data, however, will be challenging.

Surface wave profiling using Scholte waves does not produce a high resolution image of the buried layers. It provides an estimate of the velocity of the major layers. However, in conjunction with information obtained from conventional seismic reflection data, we can obtain information about those areas that are "invisible" for compressional waves. In archaeological investigations, surface waves are expected to be useful in areas where compressional waves suffer from the presence of gas. Additionally, processing of Scholte wave data is relatively easy, compared to shear wave processing.

The resistivity method, usually employed on land, has a potential for a marine application. Equipment is available and processing is relatively easy. For our project, we expect that the application is for geological characterization of the subsurface rather than for the identification of buried artefacts.

The EM method has been applied successfully several times in shallow water environments to detect structures of varying resistivity/ porosity. For the identification of e.g. channels, this method can be used in this project. The equipment used for shallow underwater EM, however, is not readily available.

For both resistivity and EM methods there are practical limitations during acquisition. The electrodes or loops should be placed close to the seafloor.

The magnetic method is useful for the detection of ferromagnetic objects only. It depends of the pilot sites whether these objects are to be expected or not. On the other hand, acquisition is fast and cheap, facilitating a reconnaissance study with magnetometer.

The underwater GPR method is not available for the project at the current status of the prototype instrument.

The non-acoustic methods should not be used stand-alone, but in conjunction with each other and the acoustic techniques. They can provide extra information in areas with gas in the subsurface.

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