The Influence of Sediment Layering and Geoacoustics on the Propagation of Scholte Interface Waves

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Abstract — Results of recent Scholte wave measurements at two diverse test sites showed considerable differences in the dispersive behavior of these waves. For the southern California site, the Scholte waves show strong, normal dispersion with group velocities ranging from approximately 30–75 m/s. On the other hand, at the Oregon Margin site the dispersion was less clearly defined, with group velocities ranging from approximately 30–205 m/s. Using a full-wave numerical model (SAFARI/OASES) and a seismic normal mode method, forward modelling (iterative inversion) was successful in matching the dispersion curves obtained from the measured Gabor matrix. A thin soft sediment overlying a harder subbottom at the Oregon site is shown to be the crucial factor in the differences; moreover, the precise thickness of the sediment is significant. For the thick sediment site off southern California, the higher frequency fundamental Scholte mode is controlled largely by the upper few (2–10) meters of sediment, whereas the lower frequency dispersion is dominated by deeper layers, which also strongly influence the higher-order Scholte modes. The conclusions are believed to be of more general applicability than just the two sites examined.

I. INTRODUCTION

Knowledge of elastic shear wave velocities and attenuations in marine sediments is important in a number of disciplines, ranging from commercial oil prospecting to acoustic surveillance. However, direct methods of acquiring a detailed knowledge of these parameters are time consuming, expensive, and not always feasible, particularly in the largely inaccessible deep-sea areas. A far more practical approach involves the use of elastic surface waves, particularly of the Scholte/Stoneley type, to remotely sense the desired parameters. The possibility of deducing the properties of shear body waves from Scholte/Stoneley interface waves derives from the strong dependence of the latter on the geoacoustics of the seabed/subbottom, particularly the shear velocities and attenuations. As a result, the behavior of interface waves, notably their dispersive nature, can be "inverted" to obtain the desired shear properties of the sediments.

The dispersion of explosion-generated seismic interface waves is a manifestation of the frequency-dependent penetration of the energy into the layered sediments of the sea bottom. The longer-wavelength, low-frequency energy penetrates more deeply into the sediments, thereby sensing the higher-speed sediment layers (in the typical case, shear speeds increase with depth in the sediment). At higher frequencies, much of the energy is confined to the surficial, slower-speed layers. Consequently, a broadband wave train propagating in a layered environment will be dispersed, the degree of the dispersion being determined by the spatial rate of change of the layers—i.e., by the depth gradient of shear velocity. Not surprisingly, propagation in a uniform medium (e.g., a half-space) is non-dispersive. The frequency selective behavior of the sediments forms the basis for exploiting the Scholte wave dispersion, as will be clear subsequently.

In this paper, we present preliminary results of the examination of the influence of sediment layering on the dispersion of Scholte waves. This investigation was prompted largely by the results of recent experimental observations of markedly different characteristics of Scholte waves measured at two diverse test sites [1]. Although the pertinent experimental results are presented, the primary emphasis here is on the numerical modelling procedure and results. The goal of the modelling study was two-fold: (1) an immediate one, viz., to quantitatively explain the differences in the observed behavior in terms of the contrasts in the sediment geoacoustics of the test sites and (2) more generally, to examine the effect of a thin soft sediment layer overlying a hard subbottom. The second objective was prompted by our earlier tentative conclusion that fine-scale stratigraphy at one of the sites was a significant factor in the observed differences.

II. PERTINENT EXPERIMENTAL OBSERVATIONS

A. Test Arrangement and Site Geoacoustics

The results to be discussed derive from Scholte wave measurements in two diverse geographical areas: the Oregon Margin off the coast of Washington, and the flat marginal basin off the coast of San Diego. For both experiments the primary goal was to record Scholte waves directly generated by bottom sound sources. At each test site a number of sensors were deployed, including several ocean bottom seismometers (OBS). Each OBS consisted of a set of...
tri-axis geophones and an externally-mounted hydrophone clamped to the side of the OBS. Among the different sound sources used in the experiments, only the bottom explosive shots are of interest here. The bottom shots, which generally consisted of 18 kg (40 lb.) or 23 kg (50 lb.) demolition charges, were dropped over the side of the ship and subsequently detonated by an electric blasting cap. Detonation times were set to allow ample time for the dropped explosives to reach the seafloor, detonation being initiated by a crystal-controlled timer in an expendable pressure case. The bottom shots were deployed in a 2-dimensional horizontal pattern, with the distance from the shots to the receivers ranging from approximately 400 m to 2 km. Ranges were accurately determined using triangulation with at least three sensors.

The seafloor geoacoustic environments at the Oregon main site and the southern California site are generally similar. The Oregon site is located just seaward of the continental slope in water 2600 m deep. The seafloor relief is very low (<10 m), and the bottom consists of 3000 m of flat-lying sedimentary layers which overlie basaltic basement. The southern California site is located just seaward of the Patton Escarpment, which marks the outer limits of the southern California borderland terrain. The 3800-m water depth is somewhat greater than for the Oregon site, but in both cases the depth greatly exceeds the wavelengths of the signals considered here. The seafloor at the southern California site also consists of very flat-lying sediments with low bathymetric relief (<10 m). Deep Sea Drilling Project Site 469 is located within a few kilometers of the site, and shows that the upper sediments are hemipelagic clays and that basaltic basement occurs at a depth of 390 m below the seafloor [2]. In both cases, basement is sufficiently deep that it plays no part in the propagation of interface waves discussed here. Superficially then, the environments appear to be very similar for the propagation of interface waves on the seafloor.

Despite these general similarities, the geoacoustic properties of the upper few meters of sediment are considerably different. The upper 42 m of the sediment at the southern California site are clays with a compressional wave speed of approximately 1510 m/s. Although shallow (<3 m) cores at the Oregon site show hemipelagic muds with sound speeds near 1500 m/s [3], the average sound speed in the upper sediments calculated from the compressional wave arrivals of the OBS data is near 1800 m/s. Since the site is located on the distal edge of the Astoria Fan, the deeper sediments (>3 m) are probably predominantly sand rather than clay, which have higher sound speeds. Examination of the 3.5 kHz seismic records from the site indicates a thin, almost acoustically transparent, sediment layer approximately 3 m thick overlying a strong reflector below which we see no penetration. Our interpretation of these data is that the sedimentary section is composed of a thin layer of hemipelagic muds overlying sand on the Oregon Margin.

B. Selected Results

1. Southern California Test Site

A typical response to a bottom shot is given by Fig. 1, which shows the time series (seismogram) recorded by the vertical OBS component for an approximately 800-m distant explosion during the southern California experiment. In this and subsequent seismograms it is the distribution of energy within each seismogram, not the relative amplitudes of the seismograms, which is of primary interest. As indicated on the figure, the direct waterborne signal, W1, is followed by the first and second water column multiples, W2 and W3, respectively, and then by the arrival of the Scholte wave signal. Of all four OBS components, the vertical component of the geophone showed the clearest presence of Scholte waves. In particular, the Scholte wave shown on this component was larger in magnitude than those of the other components, relative to the waterborne arrivals, and showed strong "normal dispersion" (lowest frequencies arriving first) expected of Scholte waves in typical sediments. The presence of the Scholte wave on the other components, while evident, is not nearly as prominent. Hence, we will examine only the vertical component in this paper.

Based on the arrival times and known source-to-receiver distances in Fig. 1, the Scholte waves group velocities were estimated to range from approximately 30 to 60 m/s. These very low Scholte wave velocities are consistent with the fact that the seafloor at the test site is covered with soft sediments with low shear speeds. It is noted that the Scholte velocities obtained here are in agreement with earlier measurements, by independent researchers, at a nearby test site [4]. Since, as noted, the Scholte wave is normally dispersed, it is reasonable to assume that the early-arriving, low-frequency wave traveling at 60 m/s has penetrated somewhat deeper into the (higher velocity) subbottom than the slower, higher frequency wave. The strong dispersion of the Scholte wave most likely indicates a steep depth gradient in the shear wave velocities.

An effective approach to the analysis of multi-modal dispersion of surface waves is the multiple filter technique of Dziewonski et al. (1969) [5]. This results in a presentation

![Fig. 1. Response of vertical geophone to an 800-m distant explosion (southern California test site).](image)
of narrow-band filtered wave amplitudes as a function of group velocity and frequency, known as the Gabor matrix. In brief, the Gabor matrix is obtained as follows. The Fourier transform of the relevant seismogram (time series) is multiplied by a set of Gaussian functions (windows) centered at a discrete set of frequencies. From the resulting windowed spectrum the quadrature spectrum is obtained (this is needed to recover phase information), and then the results are transformed back into the time domain by inverse Fourier transformation. From the resulting seismograms, instantaneous amplitudes and phases are obtained as a function of time or propagation velocity. Modal group velocity is then determined as the peak amplitudes in the resulting contour plots. Although it is not entirely without its limitation, the multiple frequency method has been successfully used by numerous investigators (e.g., 5–12), since the modal behavior of interface waves can, potentially, provide much insight into the stratification, shear velocities, and attenuation of bottom sediments. The multiple frequency approach used in this paper is equivalent, but somewhat different, from the Dziewonski approach [12].

Fig. 2 is a Gabor matrix obtained by the multiple frequency method from the vertical component seismogram recorded at a source-receiver range of approximately 1 km. The group velocity scale is plotted linear in slowness (reciprocal of velocity) rather than in velocity, since this enables particular features of the Gabor matrix to correspond directly to its original seismogram (shown adjacent). Clearly, the response is dominated by the fundamental Scholte mode, with group velocities ranging from approximately 30–75 m/s. The strong normal dispersion evident in the Gabor matrix, as well as its orientation, suggests a fairly steep positive shear wave velocity gradient. A very weakly dispersed higher order mode, centered about a group velocity of approximately 55 m/s, is also quite evident. The relatively undispersed line at approximately 205 m/s corresponds to the first water-column multiple of the compressional (waterborne) body wave, sound speed approximately 1410 m/s (W2 in Fig. 1). Since the water depth at the site is approximately 3800 m, the first multiple arrives approximately 5 seconds after the direct waterborne arrival (W1). The group velocity of W2 (~205 m/s) is its apparent velocity, based on the fact that horizontal (source-receiver) ranges (here approximately 1 km) are used to compute the group velocities in the Gabor matrix. Also evident in Fig. 2 is the greatly attenuated second water-column multiple (W3) arriving about 10 seconds after W1 (group velocity ~100 m/s). It is noted that all these arrivals are clearly distinguishable in the adjacent seismogram. A feature still under investigation is the somewhat more slowly arriving energy appearing, in Fig. 2, at the bottom right portion of the fundamental mode, and also, in Fig. 1, as the lower energy pulse following the fundamental. As shown later this complicates the modelling efforts. As a final point in Fig. 2, relevant to Gabor matrices in general, it is noted that the Gabor matrix will look different at other ranges, even for otherwise identical conditions, as a result of attenuation effects. In particular, some features will tend to diminish in importance with increasing ranges, particularly those with lower energy and higher frequency content. As a result, it is important to carefully examine dispersion data at both long and short ranges in order to provide a more accurate picture of the true frequency content of interface waves. Dispersion data over a broad frequency range permit better resolution of the surficial sediments and, at the same time, provide information on the deeper layers [7, 13].

2. Oregon Margin Test Site

We now briefly consider typical results from the experiment in the Oregon Margin. Fig. 3 shows a response to a 2-km distant explosion. Comparison with Fig. 1 reveals significant differences between the responses shown in the two figures. In Fig. 3, the vertical response again shows the conspicuous presence of the Scholte wave. However, in comparison to the result in Fig. 1, the Scholte wave displays a less clearly defined dispersion pattern. This behavior is consistent with a non-smooth or discontinuous change in the bottom sound speed gradient. Based on the known bottom geoacoustics in this area, cited earlier, this result is quite plausible. Fig. 4, the Gabor matrix at approximately 1 km, shows even more strikingly the relatively complex behavior.
of the dispersion for this environment. The dispersion of the interface wave is very strong: in a narrow frequency band from approximately 0.5 to 1.5 Hz the group velocity decreases from approximately 200 m/s to 30 m/s, suggesting a very steep or discontinuous shear velocity gradient. The higher shear velocities and steep gradient, relative to the sand at the California site, are consistent with a thin soft layer overlying a sandy sediment.

III. NUMERICAL MODELLING

A. Overview of Modelling Procedure

As already noted, the objectives of the modelling effort were to explain quantitatively the major differences between the observed Scholte waves at the two test sites, and to examine the role played by these differences by a thin sediment layer.

The experiment sites were characterized by seafloors with low bathymetric relief and flat-lying sedimentary layers. Consequently, the SAFARI/OASES computer code was used to model the propagation. OASES, which is basically an upgraded version of SAFARI, is a general purpose code for modelling seismo-acoustic propagation in horizontally stratified waveguides using wavenumber integration in combination with the Direct Global Matrix solution technique [14]. SAFARI/OASES was used to compute time series (synthetic seismograms) and frequency spectra for pulse sources, and integrands (Green's functions) and transmission losses at individual continuous wave frequencies. One can also calculate the modal dispersion curves using this numerical model, but it is usually time consuming for a realistic multilayered environment, since a full-wave computation is performed. Hence, modal dispersion curves were computed using a seismic normal mode program [15], which is considerably faster.

The basic modelling approach used here was to attempt to match the data by matching the dispersion curves of the measured Gabor matrices. In particular, beginning with an initial environmental input model, a forward modelling procedure (i.e., trial and error) was performed until agreement was obtained with the observed dispersion curves. The resulting input model was then used in SAFARI/OASES to compute the synthetic seismogram. This iterative modelling of the Scholte wave dispersion required many runs before good agreement was achieved. Clearly, this inversion procedure is unsophisticated, and becomes increasingly unwieldy as the geoaoustic complexity of the environment increases. Apart from its potential tediousness, the method has some more fundamental limitations. To wit: contamination from either lateral inhomogeneity or the effects of refraction arrivals mixing with the interface wave can distort the otherwise fairly smooth modal dispersion curves [16]; local maxima present in the observed dispersion curves (from the measured Gabor matrix) may be attributable to structural effects of the subbottom rather than to the modal structure [17]; group velocity is typically estimated from a single seismogram, and the modal group velocity (dispersion curve) is obtained by correlation of peak amplitudes, a procedure that is often one of subjective interpretation; the method assumes horizontal layering over the entire range between source and receiver, which is generally not the case in realistic environments.

Inversion methods based on the use of phase velocity dispersion also exist, but these, too, are not without their limitations. More sophisticated methods include joint use of group and phase velocity dispersion, iterative semi-automatic schemes based on linearized inversion techniques, etc., [13, 16, 18]. Although some of these approaches are being considered for inversion of surface wave data, they were not used for this investigation. The forward modelling approach used here was, nevertheless, quite successful. Since measured data existed for several source-receiver offsets, and several sensors, we had numerous "redundant" seismograms, thereby lending considerable credibility to our measured Gabor matrices and, hence, confidence in our final modelled results. Further, the iterative modelling approach, albeit somewhat tedious, does have the benefit of allowing considerable physical insight into the effects of sediment layering on the dispersive behavior of interface waves. In particular, the differential role of surficial and deeper sediment layers on broadband interface waves is readily observable using this approach. This, in turn, offers some insight into the method of changing the model to fit the data. It is noted that we are attempting to match modal dispersion curves derived from the Gabor matrices, and not the relative energy distributions in the matrices, as we are concerned with deducing shear speed profiles, not shear attenuation profiles. Iterative inversion of the shear attenuation parameters would enable us, in principle, to match other details of the Gabor matrix, and thereby produce
computed Gabor matrices that resemble the measured ones. We did not take this approach. However, a companion paper [19] does present the results of a detailed investigation into the attenuation of Scholte waves at the two test sites.

B. Modelling the Southern California Test Site

1. Sensitivity Analysis of the Role of Layering

The initial model of the southern California test site was based on results published by Sauter [4], for an experiment conducted in the vicinity of our test site. Sauter reports values of shear velocity and shear attenuation only for the upper 60 m of the sediment, whereas we initially modelled the entire 390 m. Values for the upper 60 m were based on Sauter's; the values for the remaining 330 m were extrapolated from these values. At the termination of the sediment (390 m), a value of shear velocity of 2500 m/s (basalt) was assigned. A similar procedure was followed for the shear attenuation values. Typical values for the compressional velocity and attenuation were chosen to be constant throughout the sediment, since Scholte wave characteristics are known to be largely insensitive to the compressional properties of the sediment. Fig. 5 shows the resulting initial geoacoustic model. This was used as the basis for the iterative inversion process discussed earlier.

Prior to beginning the comparison with the data, a systematic investigation was made of the sensitivity of the Scholte wave to the sediment layering structure. A number of geoacoustic models of the sediment were developed, ranging from the detailed, realistic multi-layered sediment shown in Fig. 5 to a 390-m isovelocity sediment overlying basaltic basement. The progression from the most complex to the simplest was accomplished by making the sediment isovelocity for increasingly larger portions of the subbottom. For example, the first variation of the model in Fig. 5 was to make the sediment below 60 m isovelocity, while not changing the upper 60 m. Then the sediment below 40 m was made isovelocity, and so forth until the original 21-layered sediment was reduced to an isovelocity one. The preceding approach was clearly designed to attempt to ascertain the effects of the deeper layers. To examine the effects of surficial layers, the geoacoustic model below approximately 10 m was retained and the surficial layers were modified in various ways.

For each of the dozen or so models thus derived, the Scholte wave responses, as revealed in the SAFARI integrands and transmission losses (TL), were computed at several frequencies, ranging from approximately 0.5 to 5 Hz. Examination of these results provides information on the modal structure of the propagation (number and type of modes, phase velocities) and attenuation behavior (from widths of modal response and TL curves). In this short paper we cannot show detailed results for the various geoacoustic models, but the main conclusions will be summarized here.

In assessing the effects of the various perturbations of the model, we use the most detailed model (Fig. 5) as the standard against which the others are to be compared. For this model, the response was multi-modal (3–4 modes) at the highest frequency (5 Hz). The fundamental mode (zero order) was clearly dominant, the third order mode was relatively very small, and there was a hint of one or two modes of even higher order than the third. With decreasing frequency, the number of modes decreased (but their velocities increased), until between 0.5 and 1 Hz there remained one (high velocity) mode.

It is instructive to consider the results from the model containing the same structure as the "standard" for the upper 10 m, but an isovelocity sediment below 10 m. At 5 Hz, the fundamental and first two higher-order modes were essentially identical (velocities, amplitudes and TL) to the standard. The third and higher order modes were absent in this case. This suggests that only the surficial sediments above 10 m are influential at this frequency—except, possibly, for the very high order modes, which, because of their longer wavelengths, can be influenced by the deeper layers even at the higher frequencies. At 2 Hz, only the fundamental mode was identical to the standard. The first higher-order mode was somewhat different from that of the standard, reflecting the different bottoms. At 1.5 Hz a result similar to that for 2 Hz was obtained—i.e., the fundamental modes were identical but the difference between the first higher-order modes was even greater than at 2 Hz. Skipping the 1 Hz result for a moment, we consider the 0.5 Hz case. Here the results of the two models are quite different. Only one dominant mode exists for each

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Fig. 5. Geoacoustic model of southern California site used to initiate iterative inversion process.
model, but the phase velocities are quite different, reflecting the completely different deeper layers. This indicates that at this frequency the deeper layers are the controlling factor. The 1 Hz case suggested that at this frequency a transition occurs between the regimes in which surficial sediments dominate the behavior and those in which deeper layers are paramount. In other words, at this frequency both the surficial sediments and the deeper layers (to a depth of approximately 60 m) are influential.

Proceeding in the other direction, we considered the results of a model with layering below 10 m identical to the standard, but with a different, simplified upper 10-m structure. In this case we found, as expected, that results at the lowest frequencies gave the best agreement between the two models, and that the agreement got worse with increasing frequencies. The various other models between these two cases generally confirmed the overall conclusions. In particular, for this geoacoustic environment, surficial sediments in the upper 10 m control the higher frequency (above 2 Hz, say) fundamental mode Scholte wave dispersion. In fact, models examining finer details of the surficial layers demonstrated that the top 2 or 3 m may control the higher frequency behavior. At the lowest frequencies (<1 Hz), the fundamental mode is dominated by deeper layer properties. The behavior around 1 Hz seems to be influenced by both surficial and deeper layers. The higher-order modes, owing to their longer wavelengths, are more sensitive than the fundamental mode to deeper layers, even at the higher frequencies. Clearly, the details will vary according to the type of sediments one encounters, but these observations would seem to be of general relevance, since they confirm a priori expectations of the behavior. That is, effective penetration into the sediment of the Scholte wave is of the order of one wavelength, and the effect on the Scholte wave is cumulative, or integrating, over the layers. The results suggest that, in soft sediments at least, the use of the fundamental mode Scholte waves to explore subbottom sediments (deeper than the first few meters) is probably limited to the ULF domain (<1 Hz). However, higher-order Scholte modes may be useful for probing the deeper layers.

2. Dispersion Curves

The geoacoustic model shown in Fig. 5 was used to begin the iterative modelling procedure, already discussed. Fig. 6(a) shows the dispersion curves computed for this geoacoustic model, superimposed on the measured Gabor Matrix. Clearly, the fundamental mode considerably underestimates the higher frequency group velocities. The two higher-order modes also fall short. Based on this result, the next iteration was produced by increasing, in what seemed a rational manner, the shear speed velocities in the pertinent layers. In particular, the surficial sediment velocities were increased by a larger amount than the deeper layers. The resulting dispersion curves (not shown) greatly overestimated the group velocities over the entire frequency range, perhaps reflecting the integrating effect of the Scholte waves. The next iteration resulted in the dispersion curves shown in Fig. 6(b). Agreement is much better, but a disconcerting "hump" from 1 to 2 Hz mars the result. The higher-order modes show much better agreement.

After repeated iterations designed to eliminate the hump, it was necessary to resort to a gradient-smoothing program which, in effect, forces a more gradual change in the shear velocity gradient over the appropriate depth regime. Fig. 7(a) shows the result of this procedure. The agreement is better overall, but the fit for the higher-order modes is actually somewhat worse than before (Fig. 6(b)). There is also some question concerning the more slowly arriving energy associated with the fundamental mode. In particular, it is not clear whether the fit of the dispersion curve through
Fig. 7. Best fit dispersion curves (a) resulting in the inferred shear profiles (b) (southern California test site).

this portion is really appropriate. Moreover, this feature is greatly diminished at greater ranges. As noted earlier (Section III A), other effects can complicate the Gabor matrix. The nature of this feature is presently being investigated in some detail. All this notwithstanding, the computed dispersion curves do represent a good fit to the data. This is noteworthy in view of the difficulty often experienced in matching any of the higher-order modes [10, 17]. The shear velocity profile resulting from this model is shown in Fig. 7(b), along with the corresponding curve for the Oregon site (to be discussed later).

C. Modelling the Oregon Margin Test Site

1. The Effect of a Thin Mud Layer for a Simple Geoaoustic Model

The procedure followed here was similar to the earlier one, but here we focused on the effect of a thin (approximately 3 m) mud sediment layer overlying sand. As before, prior to attempting to match the data, a sensitivity investigation was performed. We began with a simple 3-layered model: 2600 m of water overlying 3 km of sand (shear speed, \( C_s = 600 \text{ m/s} \), overlying basement. The results were examined for a source located on the bottom, for frequencies of 5 Hz, 2 Hz, and 1.5 Hz. As expected, the model resulted in an essentially non-dispersive high velocity (approximately 0.9 \( C_s \)) fundamental Scholte mode, with at least one higher-order mode. Next, a 3-m mud layer (\( C_s = 30 \text{ m/s} \)) was introduced on top of the sand. In this case, the Scholte wave was dispersive: at 5 Hz there existed a slow fundamental mode (~30 m/s) and a much higher velocity first mode. At the lower frequencies, the fundamental mode velocity was approximately the same as the previous case (no mud layer). The transmission losses were also compared, since this is a good measure of what one actually observes. At 5 Hz and a range of 1 km, the TL for the 3-m mud layer case was up to 6-8 dB greater than for the no-mud layer case; at the lower frequencies there was little difference in the transmission losses at the corresponding frequencies (at 1 km). In short, the effect of the 3-m mud layer was to substantially reduce the velocity and increase the TL of the high frequency fundamental mode, while leaving essentially unaltered the behavior of the higher-order mode(s). In a slight perturbation of the model, the 3-m mud layer was divided into three 1-m layers of shear speeds 30 m/s, 45 m/s, and 70 m/s. This affected only the fundamental mode at 5 Hz. The dispersion curves resulting from these two basic models are shown in Fig. 8.

2. Dispersion Curves for the Oregon Margin Geoaoustic Model

The previous simple geoaoustic models clearly indicate the significant effect of the thin mud layer on the high-frequency fundamental Scholte mode. To match the Oregon Margin data, however, required substantially more detailed models. We skip the numerous intermediate stages of the forward modelling (iterative inversion) process and present the final results. Fig. 9(a) shows the best fit to the data, based on the geoaoustic model shown in Fig. 7(b). Fig. 9(b) demonstrates the significant effect of the thin mud layer on the fundamental mode, particularly at the higher frequencies, confirming the results of the simpler models. It is noteworthy that best agreement with the measured data (Gabor Matrix) was obtained with a mud layer precisely 3.5 m thick. Dispersion curves for slightly different mud thicknesses (e.g., 3 m, 4 m), while similar (in steepness) to the correct curve, did not fit the maxima correctly, but, instead, lay too far either to the right or to the left of it.

CONCLUSIONS

Observations of Scholte waves obtained at two diverse sites showed considerable differences in the dispersive
properties. In particular, at the southern California test site the Scholte waves display strong, normal dispersion, whereas the corresponding dispersion results at the Oregon Margin site are less clearly defined. Numerical modelling has produced a good match of the measured data, with the following conclusions particular to the test sites:

- The general features of the Scholte wave dispersion at the southern California site can be explained, quantitatively, by a thick, soft sediment bottom/subbottom with a smooth depth gradient in the shear velocity profile.
- For the Oregon Margin site, the results are consistent with a thin (3.5 m) soft sediment layer overlying a hard subbottom. Moreover, the dispersion is quite sensitive to the precise thickness of the soft layer.

More generally, the following has emerged from the modelling study:

- A thin soft sediment layer overlying a hard subbottom has a significant effect on Scholte wave dispersion, particularly on the fundamental mode at higher frequencies. The effect is to greatly decrease the wave velocities and to increase the transmission loss of the Scholte wave, thereby behaving as a sort of structural low pass frequency filter.
- Generally speaking, in areas characterized by thick, soft sediments, the fundamental Scholte mode at higher frequencies is controlled largely by the first few meters of sediment, whereas the deeper layers control the low frequency behavior of the fundamental and higher-order modes. Nevertheless, there is some interdependence between the two regimes.

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