Characteristics of gas hydrate and free gas offshore southwestern Taiwan from a combined MCS/OBS data analysis

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Abstract

In this study, we present the results of the combined analyses of ocean bottom seismometer and multi-channel seismic reflection data collection offshore southwestern Taiwan, with respect to the presence of gas hydrates and free gas within the accretionary wedge sediments. Estimates of the compressional velocities along EW9509-33 seismic reflection profile are obtained by a series of pre-stack depth migrations in a layer stripping streamlined Deregowski loop. Strong BSR is imaged over most of the reflection profile while low velocity zones are imaged below BSR at several locations. Amplitude versus angle analysis that are performed within the pre-stack depth migration processes reveal strong negative P-impedance near the bottom of the hydrate stability zone, commonly underlain by sharp positive P impedance layers associated with negative pseudo-Poisson attribute areas, indicating the presence of free gas below the BSR. Ray tracing of the acoustic arrivals with a model derived from the migration velocities generally fits the vertical and hydrophone records of the four ocean-bottom seismographs (OBS). In order to estimate the Poisson's ratios in the shallow sediments at the vicinity of the OBSs, we analyze the mode-converted arrivals in the wide-angle horizontal component. P-S mode converted reflections are dominant, while upward P-S transmissions are observed at large offsets. We observe significant compressional velocity and Poisson's ratio pull-down in the sediment below the BSR likely to bear free gas. When compared to Poisson's ratio predicted by mechanical models, the values proposed for the OBSs yield rough estimates of gas hydrate saturation in the range of 0–10% in the layers above the BSR and of free gas saturation in the range of 0–2% just below the BSR.

Introduction

Gas hydrates are ice-like non-stoichiometric crystalline solids composed of a hydrogen bonded water lattice entrapping low-molecular weighted gas molecules commonly of methane. Gas hydrate form under specific conditions of relative high pressure and low temperature, when the gas concentration exceeds those which can be held in solution, both in marine and on-land permafrost sediments (Sloan, 1998). Commonly, gas hydrates occupy more than 1% of the bulk sediment volume, resulting in 5–15% saturation of the pore space depending on the porosity (e.g. Paull et al., 1996; Dickens et al., 1997; Guerin et al., 1999). A thick stratum bearing free gas often underlies the hydrated layer.

Gas hydrates have been recovered at numerous locations, from shallow cores as well as on-land (permafrost) and offshore drill sites (Booth et al., 1998). Gas hydrate often occur as segregate bodies in the form of lenses, nodules, pellets, or sheets where the hosts sediments are fine grains (clay and silts), and only display an interstitial or cementing feature in coarser grained lithologies. At the Cascadia margin, ODP Leg 141 Sites 889, 890 and 892 penetrated the gas hydrate stability zone and underlying free gas layer (MacKay et al., 1994). While small pellets and occasionally massive pieces of hydrates were recovered in the shallow silt clays, gas hydrate appeared concentrated at several horizons, often characterized by coarser grained sediments, within the tectonized strata of the accretionary wedge (Kastner et al., 1996;

Tréhu et al., 2004). At the Blake Ridge during ODP Leg 161, small crystals of gas hydrates dispersed in the sediments were sampled. Between 200 mbsf and the base of the gas hydrate stability zone at 520 mbsf, 1.5–6.0% gas hydrate saturation of the pore space was inferred. In zones where cavities in the microfossil shells increase the pore space, gas hydrate concentration reached 10%. The base of the hydrate-saturated zone is generally sharp, but does not coincide with the occurrence and strength of the BSR: On one hand, a strong BSR was observed at 4.25 s TWT at ODP Site 997, where PCS samples taken immediately below the BSR indicate a free gas phase fraction of 12% (Dickens et al., 1997), and underlain by 100 ms of high amplitude reflections. On the other hand, probably no more than 1% gas saturation characterized Sites 995 and 994 where a weak BSR at 460 mbsf and no BSR were respectively, observed. Therefore at the Blake Ridge, some form of gas hydrate coexists with the free gas (over 70 m at Site 995) below the BSR (Guerin et al., 1999). Also, during ODP Leg 170, at Site 1041 offshore Costa-Rica, gas hydrates were encountered between 120 and 280 mbsf, in water depths of 3300 m (Kimura et al., 1998). Massive hydrates were recovered within horizons of fractured claystones and siltstones, and several ash layers strongly cemented by hydrates were observed.

In order to detect and quantify the amount of free gas and gas hydrate present in sediments using seismic techniques, a predictive model on the elastic properties of the sediments is required, where porosity, water saturation, and gas hydrate or free gas saturation are accounted for. Methane clathrate hydrate (structure I) has been found to be very strong, based on triaxial deformation experiments (Durham et al., 2003). However, the elastic moduli for gas hydrate saturated sediments depend largely on the microscopic distribution of gas hydrates in the host sediments and sediment pore water. Three principle models involving the cementation between grains, replacement of the matrix, and dissemination (“floating”) in the pore space have been proposed (e.g. Helgerud et al., 1999), resulting in major differences in the predicted seismic velocities. The cementation model is appropriate only for low porosity (36–40%) granular sediments (i.e. sands), while load-bearing sediment frame components or pore-fluid component are appro-

priate for both sands and clay-rich ocean bottom sediments (Helgerud, 2001). Dispersed gas hydrates at typical concentrations of < 5% probably will have small rheological impact, but higher concentration should show noticeably higher strength (Durham et al., 2003).

Helgerud (2001) reports measurements of compressional and shear-wave speed in laboratory synthesized methane hydrates as a function of pressure and temperature. This author also analyzes data from hydrate bearing onshore sands in the Arctic (Northwest Eileen Well#2) and hydrate-bearing, high porosity, clay-rich ocean bottom sediments from offshore of the southeastern United States (ODP Site 995). The modeling results show that the methane hydrate does not act as grain contact cement at Northwest Eileen Sate Well#2, but founds not enough independent data to choose between a pore fluid and sediment frame component behavior. At ODP Site 995, methane hydrate acts as a sediment frame component, about 2–4% of the sediments (by volume) from 200 to 450 mbsf, with peak concentration of 8–9%. Samples recovered from the Mallik 2L-38 gas hydrate research well reveal that the dominant effect of the gas hydrate-bearing sediments is a pore-filling constituent in northwestern Canada (Lee and Collet, 2001). At this site, most recovered gas hydrate samples are very small, occurring mostly in intergranular pores of sandstone units; occasionally, larger pieces of 2 cm in diameters occur as clasts and fill intergranular porosity, while some visible hydrate occurs as pore filling or as coatings on granules (Uchida et al., 1998). Thus, in the vast majority of gas hydrate bearing samples that have been recovered, the observed and predicted P and S wave velocities increase gradually, not dramatically, as the hydrates concentration increases.

In the last decade, vast areas of sediments bearing gas hydrates and free gas have been discovered offshore southwestern Taiwan. Reed et al. (1992) first recognized the occurrence of a BSR in the vicinity of Taiwan. Numerous studies have since been dedicated to the distribution of the BSR in this region (e.g. Chi et al., 1998; Liu et al., 1999; Schnurle et al., 1999; Chen and Liu, 2002), concluding that BSRs are observed in both the Manila accretionary prism and the passive margin of the South China Sea continental slope (Figure 1). At least 20,000 km² of the Manila accretionary prism,

Figure 1. Bathymetric map of the surveyed area. Ship-track of EW9509 Line 33 is annotated with shot numbers. OBS locations are marked with circles. The distribution of BSR is shown as inset (after Liu et al., 2003).

in water depths ranging from 500 to 3500 m, is covered by BSRs (Liu et al., 2003). The widest aerial distribution of BSR is located in the northwestern part of the Manila Trench, where rapid deposited terrigenous sediments with relatively high amounts of organic carbon combined with strong dewatering (upward fluid flow) may pro-

vide a source for the methane and motor for gas hydrate accumulation. Until samples of these hydrates can be recovered, much of our ability to propose estimates the amounts of gas hydrate and free gas contents offshore Southwestern Taiwan relies on the analysis of seismic data (Schnurle et al., 1999; 2002).

In the summer of 1995, an US–Taiwan collaborative geophysical study of the Taiwan arc–continent collision zone was conducted. (Liu et al., 1996, 1997). Along the 22°12'N parallel six ocean-bottom seismographs (OBS) stations were deployed, and an east west combined seismic reflection–refraction profile, 180 km in length, was acquired across the northern part of the Manila Trench. Analyses of these OBS vertical components reveal the complex margin wedge structures at depth immediate west and underneath the Hengchun Peninsula (Nakamura et al., 1998). Schnurle et al. (2002) have presented the initial results for the processing of the reflection data, and the analysis of the OBS records. In this study, we present the results of the seismic reflection and 4-component refraction data analysis at the four ocean-ward OBS stations, and focus on the identification of gas hydrates and free gas within the accretionary wedge sediments.

Seismic reflection–refraction data analysis

The data reported in this study were acquired using six OBS of the Institute for Geophysics, University of Texas (UTIG), and National Taiwan Ocean University (NTOU). The instruments were deployed along the seismic line from R/V Ocean Researcher I ahead of air-gun shooting from the R/V Maurice Ewing, and were retrieved after shooting. Thus, six OBS (3-component plus hydrophone, except no hydrophone record for OBS45), with a 25-km average spacing, recorded the shots of EW9509-33 multi-channel seismic profile offshore southwestern Taiwan (Figure 1).

The seismic source used was a 20-airgun array with a total volume of 8420 in.³ fired at 20 s intervals. For the seismic reflection data, 16 s were recorded by a 160-channel digital streamer 4000 m in length. Navigation was provided by differential GPS, using data received simultaneous onboard the R/V Maurice Ewing and at a base station at the National Taiwan University, respectively. The reflection profile was processed with the Omega processing package. An acquisition gap in the vicinity of the Kaoping Canyon occurred, and the seismic profile was processed in two separate sections. Seismic geometry was assigned based on a 12.5-m common-reflection-point (CRP) interval. Instrument signature and

spiking predictive deconvolution have been applied. Stacking velocities were interactively analyzed at every 25th CRP based on maximum semblance of a 5-CRP super-gather. Horizon velocity analyses were performed in order to optimize the stacking velocity field above, at and below the BSR. Based on the stacking velocity field, true amplitude recovery was applied to correct for spherical divergence attenuation. Frequency analysis revealed a frequency content centered around 35 Hz with –6 dB offset below 15 Hz and –18 dB offset above 90 Hz. Since the BSR is located above the sea floor multiple, no multiple attenuation was necessary in this study.

Reflection data analysis

Special efforts have been made to derive the best velocity information from the seismic reflection data. A series of Kirchhoff pre-stack depth migrations of the common-offset panels was implemented in a layer stripping streamlined Deregowski (1990) loop. In order to account for the 2D complexity of the surveyed area, about 15 iterations were necessary. Figure 2 presents the depth section of EW9509-33 and the migration interval velocity field.

During each iteration, finite difference ray-tracing of every four shots (200 m spacing) was performed; the travel-times were then used to pre-stack depth migrate the reflection records. Residual move-out in the image gathers was analyzed and the migration velocity field was updated from the resulting depth/depth-error. A velocity pull down (initial over-estimated velocity) is characterized by a positive residual-parameter (downward pointing move-out), while a negative residual-parameter results in a positive velocity update. During the early iterations, CRP were computed at 50 m (the average shot spacing) in order to keep a full-fold for residual move-out analysis; Manual picking of the residual move-out parameter (r-p) at 1000 m spacing was necessary, in order to constrain the velocity updates by integrating a-priori knowledge from our structural interpretation, and stabilize the inversion process by keeping a smooth migration velocity field. During the later iterations, the CRP spacing was reduced to 25 m (80 fold), then 12.5 m (40 fold) in order to lower spatial aliasing and refine lateral velocity variations. The migration velocity field above the

Figure 2. (a) Pre-stack depth migrated section of EW 9509-33 and migration interval velocity field from the deformation front to the upper slope of the accretionary complex.

Figure 2.(b) Pre-stack depth migrated section of EW9509-33 and migration interval velocity field in the southern Taiwan shelf.

BSR converged rapidly. However, the flattening of the image gathers for reflection from within low velocity zones appeared more tedious, since their width and the amount of velocity pull-down have to be adjusted together. In the later stages, a spatially continuous automatic picking has proven efficient for the refinement of the velocity model: performing cross-correlation of each CRP generated r-p move-out and semblance maps versus depth. Automatic picks were then discriminated into horizons at depths when semblance was significant and the move-out values were similar within five adjacent CRP. The migration velocity field was subsequently updated. During the last few iterations, depth-errors were generally small down to depths of 750–1000 m below the sea-floor, particularly in the areas where the sediments are less severely deformed. However, due to the 3D roughness of the sea-floor at the transition between the lower and middle slope (between -88 and -84 km), where a narrow canyon cuts the accreted units (Figure 2a), the seismic imaging is relatively poor and the migration velocity field

weakly constrained. Furthermore, while strata of the accreted units with dips up to 40° are well imaged, steeper events (greater than 60°) are locally observed, as for instance across the elevated ridge (Figure 2a) that marks the transition between the middle and upper slope (between -70 and -64 Km). At these locations, the complexity of the structures greatly compromises the velocity inversion. Figure 3 shows the image gather, residual move-out, and reflection angle of four CRP that correspond to the four OBS stations (OBS42, 43, 44, and 45) analyzed in this paper.

Amplitude versus angle analysis

Travel times analysis of the reflection seismic data is efficient in determining the low wave-length of the velocity field. Residual move-out after pre-stack Kirchhoff depth migration has proven appropriate to determine the central wave-length of the velocity field when the stationary phase assumption is valid, as well as

removing the dip bias in RMS velocity analysis (Deregowski, 1990). The high-frequency domain of the subsurface impedance can further be tackled using wavelet analysis (such as attribute or amplitude versus offset analysis). Performing AVO after migration has long been recognized as advantageous, since the migration process better preserves the amplitudes of the wave field and collapses diffractions. Further more, the incidence and emergence angles of the reflection can be approached more accurately when using ray-tracing methods in depth domain, rather than ray-path approximations in time domain based on RMS velocities. Hence, in addition to the travel-times, these angles can be computed during the finite-difference ray-tracing, without requiring the knowledge of an accurate reflector position (Bleistein et al., 1987). Therefore amplitude versus angle analysis (AVA) can be performed within the process of pre-stack depth migration with little additional computing time. Since the pre-stack depth migration is stacking the input data along the diffraction hyperbola associated with any given event, we can compute the associated reflection angle as a running average of the reflection angles of the input data within this stack. This running average can further be weighted by the data amplitude or energy accordingly (Schnurle, 2004).

The reflection strength is then related to the reflection angle, the intercept P and gradient G by: $R(\theta) = P + G \sin^2(\theta)$, with the reflection angle $\theta = (\theta_s + \theta_r)$, and θ_s and θ_r are the propagation angles between the ray-path from the source or the receiver and the vertical, respectively. Following Aki and Richard (1980) and Shuey (1985), the intercept P and gradient G of the AVA analysis can be related to the local acoustic, shear-wave velocity and density impedance as:

$$2P = [\Delta V_p/V_p + \Delta \rho/\rho]$$

and

$$2G = [\Delta V_p/V_p - 2\Delta V_s/V_s - \Delta \rho/\rho],$$

provided that the variations in velocities and densities ΔV_p , ΔV_s , and $\Delta \rho$ are small when compared to V_p , V_s , and ρ , respectively, and $V_p/V_s \approx 2$.

The intercept P is representative of the true normal-incidence P-wave reflectivity, while G is

related to both acoustic and shear-wave impedances. Thus, the Poisson's ratio's impedance A_σ is given by:

$$\begin{aligned} A_\sigma = \Delta \sigma / \sigma &= 4[\Delta V_p/V_p - \Delta V_s/V_s]/3 \\ &= 4[P + G]/3. \end{aligned}$$

Finally, the signal to noise ratio of AVA attribute profile can be increased by weighting the attribute with the average amplitude of the reflection within the CRP gather, or with the regression coefficient of the least-square estimate of the linear fit of the CRP to P and G .

Figure 4 presents the AVA intercept P and pseudo-Poisson A_σ , weighted with the regression coefficient for the main section of EW9509-33 where the BSR is most clearly observed. Below the sea-floor, the P impedance decreases on most events as the acoustic velocity and density increases with depth. The BSR is characterized by a sharp reversal of the P impedance at the bottom of the hydrate stability zone or top of a free gas layer. Thus the BSR is marked by a strong increase of the pseudo-Poisson attribute, commonly underlain by a negative reflector in the middle and upper slope of EW9509-33, and sometimes overlain by a broad negative region in the upper slope of EW9509-33. The negative pseudo-Poisson attribute below the BSR is likely to mark the bottom of a free gas zone (see discussion). Finally in east-dipping strata at the rear of each accreted units, small clusters of positive intercept P , accompanied by pseudo-Poisson events are observed further deep.

OBS acoustic data analysis

The OBS pre-processing sequence consisted of: compute precise location and orientation of each instrument on the ocean floor using water-wave direct arrivals, merging with navigation data, and rotating horizontal axes to radial and tangential directions. Then, source signature and predictive deconvolution, separately designed from each components of individual OBS, were computed. Finally, a 1.4 time-power and 0.5 offset-power amplitude scaling, and band-pass filtering were applied. The OBS records presented in this study have not received further gain correction in order to preserve relative amplitude.

Figure 4. Amplitude versus angle attributes. (a) Intercept P acoustic impedance; Regression coefficient is shown as inlet. (b) Pseudo Poisson's ratio $4(P+G)/3$; Product $P*G$ is shown as inlet.

Ray tracing of the acoustic as well as the converted wave arrivals was performed with the RayInvr software (Zelt and Smith, 1992), with an input format slightly modified from 10 to 1 m spatial definition and 10 m/s acoustic velocity accuracy. An initial compressional velocity model was constructed from the migration velocity field obtained during the reflection data analysis. Thus in addition to the sea-floor and BSR, we have digitized three layers within the sediments above the BSR, and two layers below the BSR corresponding to reflection from the eventual free gas zone. We then added two additional layers, generally unrelated to the structures but consistent with the interval velocity contours in the deeper part of the model. The BSR in the vicinity of OBS42 in water depth of about 850 m is not clearly observed: only weak reflections with a possible polarity reversal at around 180 mbsf in the center and shallowing seaward could indicate the presence of gas hydrates on the seismic profile, between layers 1 and 2. Velocity nodes at the top and bottom of each layer, with 414 m horizontal regular node spacing, were assigned based

on the interval migration velocities 5 m above and below interfaces, respectively.

For each OBS, we performed tomographic ray tracing of the reflected and refracted acoustic waves. Additionally, we computed arrival times from the head waves arising at the model interfaces. These arrival times were then compared to time picks from the seismic data of the OBS stations. Figures 5–8 present the compressional velocity models and the ray-path of the refracted acoustic waves, reflections and head-waves from the model interfaces for OBS45, 44, 43, and 42, respectively. These travel times from ray tracing are then plotted on top of the vertical component and hydrophone records. RMS travel-time residuals and normalized chi-square values for these forward models range from 0.0012 to 0.0017 and 387–625, respectively, with better fitting for OBS43 (Figure 6) and OBS44 (Figure 7) when compared to OBS45 (Figure 8). A damped least-square to the linearized inverse problem (Zelt and Smith, 1992), using the partial derivatives of the travel times with respect to the model parameters and the travel time residuals could

Figure 5. Acoustic ray-tracing at OBS42. (a) Compressional velocity model. (b) Refracted acoustic wave, reflections and head-waves from the model. (c) Vertical component (d) hydrophone seismic data recorded at OBS42 (with a 2 km/s reduction velocity); Travel times from ray tracing are overlain.

Figure 6. Acoustic ray-tracing at OBS43. (a) Compressional velocity model. (b) Refracted acoustic wave, reflections and head-waves from the model. (c) Vertical component (d) hydrophone seismic data recorded at OBS43 (with a 2 km/s reduction velocity); Travel times from ray tracing are overlain.

Figure 7. Acoustic ray-tracing at OBS44. (a) Compressional velocity model. (b) Refracted acoustic wave, reflections and head-waves from the model. (c) Vertical component (d) hydrophone seismic data recorded at OBS44 (with a 2 km/s reduction velocity); Travel times from ray tracing are overlain.

Figure 8. Acoustic ray-tracing at OBS45. (a) Compressional velocity model. (b) Refracted acoustic wave, reflections and head-waves from the model. (c) Vertical component recorded at OBS45 (with a 2 km/s reduction velocity); Travel times from ray tracing are overlain.

further constrain the acoustic velocities in the survey area. Unfortunately, the velocity pull-down below the BSR makes such inversion numerically unstable.

As a matter of fact, arrivals are far better identified on the hydrophone data when compared to the vertical component of the OBS stations. Furthermore, while the compression of the seismic wavelet with the source signature is fair, ringing is prominent on the vertical component data (particularly in the near offsets), and source ghosts (180 ms apart) are only partially removed by the predictive deconvolution on the hydrophone data. Also, arrivals from the BSR (in violet) are generally weaker than on the reflection data.

OBS converted wave data analysis

Our estimates of the compressional velocities are relatively accurate, because our models rely mainly on a-priori knowledge from the velocity analysis and pre-stack depth migration of the seismic reflection data. However, the conversion modes of the recorded horizontal arrivals need to be established first, before estimations of shear-wave velocities from the converted waves travel times and Poisson's ratios can be obtained. As a matter of fact, the conversions from acoustic to shear-waves along specific horizons in the strata may occur in numerous fashions (e.g. Schnurle and Hsuan, 2000). For example in a simple four layered media (water (W), hydrate bearing (H), free gas bearing (G), and normal sediments (N)), we distinguish two primary acoustic reflections, the BSR, i.e. H/G reflection (p_1), and base of the free gas layer, i.e. G/N reflection (p_2), and six mode conversions. Two converted waves are related to the BSR: the sea-floor transmitted P then mode converted reflected S at the BSR, or H/G reflected S (s_1), and the sea-floor transmitted S then reflected at BSR, or W/H transmitted S + H/G reflected S (s_2). In a similar fashion and ordered by increasing zero-offset arrival times, four converted waves are related to reflections at the bottom of the free gas zone: the G/N reflected P, then H/G upward transmitted S (s_3), closely followed by the G/N reflected S (s_4), the H/G downward transmitted S then G/N reflected S (s_5), and finally the W/H downward transmitted S then G/N reflected S (s_6). At each encoun-

tered interface, the strength of P-P, P-S, S-S, and S-P reflections and transmissions depends on the acoustic and shear-wave impedance at the boundary. Based on full elastic synthetic data representative of gas hydrate and free gas bearing sediments, Schnurle and Hsuan (2000) estimate that P-S reflections (s_1 and s_4) are the dominant conversion modes that are likely to be recorded by the horizontal components of our OBS stations.

Figures 8–11 present the horizontal inline and cross-line components of OBS42, 43, 44, and 45, respectively, where the travel times from the shear-wave ray tracing are overlain. The ray-paths corresponding to acoustic waves are marked with lines and shear-waves with dashed lines. Vertical variations of the Poisson's ratio within each layer are not considered in our study. Since the shear-wave ray paths arising from the P-S mode converted reflections hardly extend laterally farther than 100 m away from the OBSs in the shallow part, and 400 m at the deepest layer in our model, lateral variations of the Poisson's ratio within layers can not be constrained by the data analysis. We therefore assume a constant Poisson's ratio within each layer. We have then performed a series of forward modeling of the shear-wave arrivals with a set of Poisson's ratio varying for each layer. Figures 9–12 correspond to our favored model, although we are aware that this model is non-unique.

The seismic signals recorded by the horizontal components are of much better readability when compared to the vertical and hydrophone components. We observed a good fit of the P-S reflected arrivals (type s_1 and s_4) on both inline and cross-line components, up to offset of 3 km. The arrivals recorded at larger offsets thus result from more complex wave propagation. In this study, rather than presenting acoustic reflected rays from deeper layers, later upward P-S converted at the interface of interest (type s_3 , for instance), we present refracted acoustic rays that are upward P-S converted at each interface. The travel-times predicted for these turning-rays extend continuously from the P-S reflections travel times, with a relatively faster move-out, in offsets ranging from about 3 to 6 km (Figures 9–12). Acoustic reflected rays from deeper layers, later upward P-S converted would give a similar move-out velocity trend but with a delay that

Figure 9. Converted wave ray tracing at OBS42. (a) Shear-wave velocity model. (b) P-S converted reflections, S transmitted head-wave and turning rays from the model interfaces. (c) Horizontal inline and (d) cross-line components recorded at OBS 42 (with a 2 km/s reduction velocity). Travel times from ray tracing are overlain.

Figure 10. Converted wave ray tracing at OBS43. (a) Shear-wave velocity model. (b) P-S converted reflections, S transmitted head-wave and turning rays from the model interfaces. (c) Horizontal inline and (d) cross-line components recorded at OBS 43 (with a 2 km/s reduction velocity). Travel times from ray tracing are overlain.

Figure 11. Converted wave ray tracing at OBS44. (a) Shear-wave velocity model. (b) P-S converted reflections, S transmitted head-wave and turning rays from the model interfaces. (c) Horizontal inline and (d) cross-line components recorded at OBS 44 (with a 2 km/s reduction velocity). Travel times from ray tracing are overlain.

Figure 12. Converted wave ray tracing at OBS45. (a) Shear-wave velocity model. (b) P-S converted reflections, S transmitted head-wave and turning rays from the model interfaces. (c) Horizontal inline and (d) cross-line components recorded at OBS 45 (with a 2 km/s reduction velocity). Travel times from ray tracing are overlain.

Table 1. Estimated Poisson's ratio within the 4 layers above the BSR and 3 layers below. For OBS42, a possible BSR is observed within layer2

Station	Layer1	Layer2	Layer3	Layer4	BSR	Layer5	Layer6	Layer7
OBS42	0.490 ± 0.002	0.473 ± 0.003	0.421 ± 0.004	0.410 ± 0.005	NO	0.390 ± 0.003	0.435 ± 0.006	0.405 ± 0.005
OBS43	0.490 ± 0.002	0.479 ± 0.003	0.460 ± 0.004	0.450 ± 0.004	/	0.370 ± 0.005	0.360 ± 0.005	0.415 ± 0.006
OBS44	0.490 ± 0.002	0.475 ± 0.003	0.465 ± 0.004	0.448 ± 0.004	/	0.435 ± 0.004	0.410 ± 0.004	0.390 ± 0.006
OBS45	0.490 ± 0.002	0.487 ± 0.003	0.455 ± 0.003	0.435 ± 0.005	/	0.460 ± 0.006	0.445 ± 0.006	0.390 ± 0.007

arises from the choice of a reflector from below. We also present travel-times for acoustic head-waves that are S converted when emerging. Such rays may arise on shots fired up-dip of steeply dipping strata, presenting the fastest move-out, and could be observed particularly on the landward side of OBS43, 44 and 45. These arrivals also branch onto the P-S reflections, thus indicating once more the critical offset to which P-S reflections are predicted by the ray-tracing. All 4 OBS records present a clear drop in amplitudes of the P-S reflections, between offsets of 3 and 4 km, close to the ray tracing predicted critical

angle. Finally, at offsets greater than 5 km, poor component separation, multiples, and more complex wave propagation make the data interpretation complicated. Our investigation of the shear-wave velocities are therefore focused on the vicinity of the OBS, where P-S reflections are sufficient to estimate a meanfull Poisson's ratio in the strata above, at, and below the observed BSR.

Furthermore, at the sea floor, strong shear-wave impedance between the seawater and shallow sediments could induce P-S transmitted waves. However, such S-S reflections (type s_2 and s_6), characterized by slow move-out, are

not clearly observed in the seismic records, suggesting that the high-porosity clay-rich shallow sediments exhibit a weak shear-wave impedance in the survey area. Finally, downward transmitted S waves across the BSR, later S reflected at depth (type s_5), requiring a strong shear-wave impedance at the BSR, are difficult to identify with certainty among the deeper arrivals on our OBS records.

In the two upper-slope basins surveyed by OBS42 and 43, our model, purely constrained by the depth migrated reflection data, produces synthetic shear-wave arrivals in good agreement with the recorded data, both in terms of travel-times (move-out) and amplitudes. Thus, layers 1 and 2 mark distinctive boundaries in the lithology beneath these two stations (Figures 9c and 10c). The Poisson's ratio decreases relatively continuously with depth, although with a different gradient for each of the two basins. On OBS42, a reversal in the Poisson's ratio is inferred within layer 6. The fit of shear-wave arrival times to the seismic signal recorded by the horizontal instruments of OBS44 (Figure 11) is generally poorer than that of OBS42 and 43. As a matter of fact, the misfit was noted at the acoustic modeling stage, particularly below the BSR, suggesting that our migration velocity field is not accurate at this location. A relatively complex structure with eastward dipping strata lies to the east and beneath OBS44 (Figure 2a), that our layered model does not render. Thus, the velocity model weakly accounts for the significant asymmetry present in the wide-angle seismic data. Still, we are able to produce a reasonable fit of the P-S reflected arrivals on the westward portion of the seismograms in the offset range of -1 – 3 km, and the Poisson's ratio appears to decrease continuous with depth. OBS45 lies in a similar setting as OBS44. This station has landed on a narrow flat along the 14° -dipping slope near the deformation front of the submarine Taiwan accretionary wedge. It is rather difficult to assess the validity of our acoustic velocity field based on the vertical component of this station. However, the pre-stack depth migration provides a much better imaging of the structures beneath OBS45 than beneath OBS44. The shear-wave modeling partly renders the asymmetry present in wide-angle seismic data (Figure 12). The fit of the P-S reflections below the BSR is

better on the western portion of our model than the east side. P-S reflections from within the low acoustic velocity zone inferred from the reflection data are strongly attenuated, thus suggests a particularly high Poisson's ratio in layer 5.

Discussion

Much of the validity of our results relies on the acoustic velocity model generated from the seismic reflection data. Furthermore, our 9-layer models do not fully account for the complex geometry of the accretionary wedge. As a result, the reflections observed in the wide-angle seismograms do not necessarily occur where our boundaries predict them, and amplitudes are poorly constrained. Still, the layer-based forward modeling is advantageous as it allows us to identify the conversion modes based on their move-out and offset range on the horizontal components of the OBSs. Table 1 summarizes the Poisson's ratio values that we estimated at the vicinity of the four OBS stations analyzed in this study. The uncertainties in the Poisson's ratio estimations correspond to the range of reasonable travel-time fits.

Figure 13 summarizes the acoustic and shear-wave velocities (Figure 13a), the Poisson's ratio (Figure 13b), and estimates of the water-filled porosities (Figure 13c), at each OBS with respect to depth below sea-floor. On one hand, Poisson's ratios above the BSR (when observed) are similar within each layer, with an increasing gradient from OBS42 to OBS45. On the other hand, considerable variations are present below the BSR, as Poisson's ratio inversions in layer 7 for OBS43 and in layer 5 for OBS45 are observed, while the Poisson's ratio gradient is preserved for OBS44. Further more, both acoustic and shear-wave velocities below OBS 42 are relatively low when compared to the other stations (Figure 13a). OBS 43 displays acoustic velocities in agreement with OBS 44 and 45, but lower shear-wave velocities above the BSR. OBS 44 is characterized by a relatively continuous trend on both acoustic and shear-wave velocities. Finally at OBS45, while relatively high shear-wave velocities are observed above the BSR, both acoustic and shear-wave velocities drop below the BSR.

Figure 13. (a) Acoustic velocities (marked with lines) and shear-wave velocities (marked with dashed lines) versus depth below sea-floor. (b) Average Poisson's ratio below the OBSs and instantaneous Poisson's ratio versus depth below sea-floor. (c) Porosity computed from the instantaneous Poisson's ratio when no gas hydrate or free gas are considered in the pore-space (thick lines), and with 10% of hydrates (dashed lines). Porosities compiled from deep-sea basins and accretionary wedges, as well as the reference trend for this study are indicated in black.

The main potential of shear-wave information of gas hydrate bearing sediments lies in the combination of (1) the high V_s of 1890 m/s of pure methane hydrate (Waite et al., 2000) compared to the typical low V_s of 100–600 m/s for deep marine hydrate-free sediments and (2) the fact that V_s is hardly affected by free gas (Andreassen et al., 2003). Thus small amount of free gas in the pore-space will dramatically decreases the compressional velocity while leaving the shear-wave velocity relatively unaffected (Domenico, 1977). Moreover, when small amount of gas hydrate and free gas are present, the acoustic impedance contrast may be sufficient to produce a clear BSR on acoustic reflection profiles, while the P-S conversion would remain very weak so that the BSR is not observed on P-S stacks (Andreassen et al., 2003).

Numerous authors have investigated the variations of compressional velocity, shear-wave velocity, and density with respect to the composition, porosity, elastic coupling, and the concentration of hydrates and free gas of the host sediments (e.g. Lee et al., 1996; Tinivella and Accaino, 2000; Lee and Collet, 2001; Helgerud, 2001). Two end-member predictive models are

reviewed by Lee et al. (1996): Wood's model is representative of hydrates "floating" in the pore-space (no elastic coupling between the sediment frame and the hydrate component), while the time-average model corresponds to full elastic coupling between the sediment frame and the hydrate component (but no additional cementation). Lee et al. (1996) further assume that the shear-wave velocity of the aggregate is related to the weighted average of the velocity ratios by:

$$\begin{aligned} V_s/V_p = & \Phi(1 - S_h)V_{S_w}/V_{P_w} + \Phi S_h V_{S_h}/V_{P_h} \\ & + (1 - \Phi)V_{S_m}/V_{P_m}, \end{aligned}$$

where Φ is the porosity, S_h the hydrate saturation, and $V_{P_w}, V_{S_w}, V_{P_h}, V_{S_h}, V_{P_m}, V_{S_m}$ the compressional and shear-wave velocities for water, hydrate and sediment matrix, respectively. As the first term disappears since shear-wave velocity of fluid is zero, the shear-wave velocity is thus strongly dependant on the water-filled porosity. Also, this formula ensures that the Poisson's ratio is identical in the time average and Wood's equations, regardless of the mechanical coupling. Figure 14a shows the Poisson's ratio behavior as a function of the porosity, for a selected range of gas hydrate saturation S_h , when

Figure 14. (a): Poisson's ratio as a function of the porosity for a gas hydrate saturation S_h varying stepwise between 0 and 50%; (b) Poisson's ratio as a function of the porosity for a free gas saturation S_g varying stepwise between 0 and 10%. The dashed lines mark the ranges of observed Poisson's ratios and porosities from the OBS analyses.

for example the sediment matrix is composed of 75% clay and 25% sand. Following Domenico (1977), the Poisson's ratio behavior as a function of the porosity can be predicted for a selected range of free gas saturation, with a coupling factor of 2.5, density of 1.8 g/cm³ and 400 m below sea floor (Figure 14b). These mechanical models show that, when introducing gas hydrates in the pore-space, for a given observed Poisson's ratio, the predicted porosity increases by about the same amount, indicating clearly the predominant dependency of the Poisson's ratio on the water content in the pore space (Figure 14a). When introducing free gas, even in small amounts, the predicted porosity decreases drastically.

When attempting to deduce gas hydrate and free gas content in the vicinity of the OBSs, we are faced with proposing a normal porosity trend. von Huene and Sholl (1991) compiled porosity measurements from DSDP and ODP drill sites and proposed to relate the porosity to depth below sea-floor with: $\Phi(z) = \Phi_{\min} + (\Phi_o - \Phi_{\min}) e^{-\beta z}$, with $\Phi_{\min} = 0.$, $\Phi_o = .63$, and $\beta = 8.6 \cdot 10^{-4}$ for deep-sea basins, and $\Phi_{\min} = 0.1$, $\Phi_o = .67$, and $\beta = 8. \cdot 10^{-4}$ for accretionary wedges. These two porosity trends show significant discrepancies with the results of our analysis, for which $\Phi_{\min} = 0.$, $\Phi_o = 0.72$, and $\beta = 190. \cdot 10^{-4}$ provides a general reference trend that fits best the porosities at OBS43, 44, and 45.

If we assume that no gas hydrate or free gas are present in the pore-space, our estimates of the Poisson's ratio yield a decrease in porosity from 80% at the sea-floor to about 40% at the BSR, for OBS 43, 44, and 45 where the BSR is clearly observed (Figure 13c). About 10% of gas hydrates in the pore-space for OBS43, and up to 20% for OBS44 and 45 are necessary for our inferred porosities to match the trend for accretionary wedges. Less than 5% of gas hydrate volumes (i.e. $S_h = 10\%$ for 50% porosity) are sufficient to account for the departure from the reference trend of the inferred porosities of OBS43, 44 and 45. Furthermore, relatively high shear-wave velocities (600–700 m/s) are displayed mostly above the BSR at OBS44 and 45. In the vicinity of OBS43 and 44, the negative P-impedance from AVA analysis is restricted to a thin layer just above the BSR, suggesting that a higher resolution than this OBS survey provided would be necessary to quantify the gas hydrate saturation more accurately. In the two layers below the BSR, the Poisson's ratio drops from 0.43 to 0.36 at OBS43, yielding estimates of the free gas saturation up to 2% for a porosity of 20% (Figure 14b). Moreover, the relatively high shear-wave velocities observed at this location, that are inconsistent with velocities predicted by mechanical models, are indicating lower acoustic velocities than those yield by the reflection data analysis. A porosity decrease from 30 to 20% results in a Poisson's ratio increase from 0.4 to 0.425, when considering $S_g = 0$, that is consistent with the values below the BSR observed at OBS44 and our reference trend. Finally, the increase in Poisson's ratio from 0.435 to 0.46 across the BSR at OBS45, subject to large uncertainty and a complex geological setting, is accounted for by a decrease of the porosity from 20 to 10% for $S_g = 0$ (Figure 14b).

Conclusion

Combined analyses of reflection and refraction seismic data prove to be appropriated for the characterization of gas hydrate and free gas bearing sediments (e.g. Mienert and Posewang, 1999; Tinivella and Accaino, 2000; Tréhu and Flueh, 2001). The layer stripping streamlined Deregowski (1990) loop coupled with the horizon

based depth error automatic picking implemented in this study allows us to propose fair estimates of the compressional velocities along the seismic reflection profile. Large variations in the pre-stack depth migrations velocities are observed and several low velocity zones are imaged below the BSR. Forward modeling of the acoustic arrivals with models derived from the migration velocities generally fit the data at the four OBS stations analyzed in this study. Then, forward modeling of the shear-wave arrivals shows the preponderance of reflection P-S conversion. P-S transmission at the sea floor, then S reflected, characterized by slow move-out and large offset range, are not identified on the horizontal components of the OBS records. Upward P-S transmissions across the BSR, generated at larger offsets, and characterized by faster move-out when compared to P-S and S-S reflections, could be observed. Finally, S waves that emerge from head waves are observed, characterized by the fastest move-out, at several locations.

Data in the vicinity of OBS42 does not suggest the occurrence of either gas hydrates nor free gas. Below OBS43, weak evidence for substantial gas hydrate saturation is observed, but the Poisson's ratio pull-down, as well as the AVA attribute analysis indicate a wide zone of sediment bearing up to 2% of free gas in the pore-space, possibly underestimated during the reflection data analysis. While a weak BSR is observed on the seismic section just below OBS43, a strong AVA pseudo-Poisson anomaly is present on either side of this OBS. Given the complexity of the structures below OBS44 and 45, our velocity models poorly account for the significant asymmetry present in the wide-angle seismic data. From the analysis of OBS44 records and the reflection data, the free gas zone just below the BSR is poorly saturated and relatively thin at this location. We infer that less than 10% gas hydrate saturation is present in the two layers above the BSR near OBS44 and 45.

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