Rock-physics Models for Gas-hydrate Systems Associated with Unconsolidated Marine Sediments

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ABSTRACT

Rock-physics models are presented describing gas-hydrate systems associated with unconsolidated marine sediments. The goals are to predict gas-hydrate concentration from seismic attributes, such as primary (P)- and secondary (S)-wave velocities, and to analyze compressional-wave (PP) and converted-shear-wave (PS) reflectivity at the base of hydrate stability zones. Elastic properties of gas-hydrate systems depend on elastic properties of the host sediments, elastic properties of gas hydrates, concentration of hydrates in the sediments, and geometrical details of hydrate morphology within the host sediments. We consider various scenarios for hydrate occurrence, including load-bearing gas hydrate, pore-filling gas hydrate, and two different thin-layered models of gas hydrate intercalated with unconsolidated sediments. We show that the geometrical details of how gas hydrates are distributed within sediments have a significant impact on relationships between gas-hydrate concentrations and seismic attributes. Therefore, to accurately estimate gas-hydrate concentrations from seismic data, we need to understand how hydrates are formed and distributed within marine sediments. The modeling results for thin-layered hydrated morphologies show significant S-wave anisotropy, which may be used to infer gas-hydrate distributions and concentrations in alternating thin layers of hydrate-bearing sediments if multicomponent seismic data are available.

We compare the theoretical predictions of the isotropic rock-physics models with published laboratory measurements on synthetic gas-hydrate formed in unconsolidated sands. We find good agreement between the rock-physics model of disseminated, load-bearing gas hydrate and laboratory measurements, which suggests that, in this case, gas hydrates may act as part of the mineral frame of the unconsolidated sediments.
INTRODUCTION

The main goal of rock-physics modeling of gas-hydrate systems is to infer gas-hydrate concentration from seismic measurements. Most gas hydrates around the globe are distributed in marine environments, especially in deep-water, near-sea-floor strata.

Many published relations between gas-hydrate concentration and seismic attributes are empirical (Pearson et al., 1983; Miller et al., 1991; Wood et al., 1994; Holbrook et al., 1996; Lee et al., 1996; Yuan et al., 1996; Collett, 1998; Lu and McMechan, 2002, 2004). Although empirical approaches are easy to implement, they do not have predictive power and are applicable only to the specific site where the relationships are derived. Moreover, empirical relations do not provide insights into the morphological character of gas-hydrates deposited within sediments. Other studies use physics-based effective-medium models of gas-hydrate systems to relate hydrate concentration to seismic properties (Helgerud et al., 1999; Carcione and Tinivella, 2000; Ecker et al., 2000; Chand et al., 2004; Waite et al., 2004; Winters et al., 2004; Kleinberg and Dai, 2005; Murray et al., 2006). Some of these effective-medium models are based on Dvorkin and Nur’s (1996) model of unconsolidated sediments, which uses Hertz-Mindlin’s theory (Mindlin, 1949; Mavko et al., 1998). These models apply well to relatively deep marine sediments (>300 m [984 ft] below sea floor [mbsf][fbsf]), but they cannot explain laboratory observations or in-situ observations within the first 200 mbsf (656 fbsf) where sediments have low shear strengths and large compressional wave velocity/shear wave velocity (Vp/Vs) ratios.

Consequently, improved rock-physics models to characterize highly unconsolidated sediments within deep-water, near-sea-floor strata where significant amounts of hydrates are known to be present are needed. Rock-physics models are also needed to help understand the way gas hydrates are distributed in relation to their host sediments. Are the hydrates disseminated and acting as part of the load-bearing frame? Do they fill the pores of the sediments without affecting the mineral frame? Are they layered bodies? Do they occur as nodules and veins?

In this chapter, we present rock-physics models for unconsolidated sediments in which gas hydrates are assumed to be present in the following geometrical occurrences: A) disseminated, load-bearing clathrates acting as part of the mineral frame of the host sediments; B) pore-filling clathrates that float in the porous space without changing the dry-mineral frame of the host sediments; C) thin layers of pure hydrates intercalated with unconsolidated sediments saturated with fluid; and D) thin layers of disseminated, load-bearing hydrates intercalated with unconsolidated sediments saturated with fluid.

The main objective of this article is to understand the relationships between gas hydrate concentrations and seismic properties, such as primary (P)- and secondary (S)-wave velocities, for different scenarios of gas hydrate occurrence. We also simulate amplitude variations with angle of incidence (AVA) for compressional wave (PP) and shear wave (PS) reflectivity from a hypothetical interface separating the base of the gas-hydrate stability zone (GHSZ) and the sediments below that contain free gas. We show that the geometrical details of how gas hydrates are distributed within sediments have a significant impact on the relationships between gas-hydrate concentrations and seismic attributes. Therefore, to accurately estimate gas-hydrate concentrations from seismic data, we need to understand how hydrates are formed and distributed within marine sediments. Our modeling results for two thin-layered hydrate-sediment morphologies (cases C and D listed above) show significant S-wave anisotropies. These S-wave-splitting effects may be used to infer gas-hydrate distribution and concentration in near-sea-floor environments having alternating thin layers of hydrate-bearing and non-hydrate-bearing sediments, if multi-component seismic data are acquired across targeted gas hydrate systems.

We compare the theoretical predictions of our isotropic rock-physics models with laboratory measurements by Yun et al. (2005) on synthetic gas hydrates formed in unconsolidated sands and clays. We find good agreement between our rock-physics model of disseminated, load-bearing gas hydrates (morphology option A listed above) and these laboratory measurements. This observation suggests that, in this case, gas hydrates may act as part of the mineral frame of the unconsolidated sediments. Winters et al. (2004) have other published laboratory measurements on naturally formed gas hydrates from the Japanese National Oil Corporation (JNOC)/Japan Petroleum Exploration Company (JAPEX)/Geological Survey of Canada (GSC) Mallik 2L-38 well, Mackenzie delta, which also show that gas hydrates are a part of the load-bearing frame of their host sediments.

ROCK-PHYSICS MODEL FOR HIGHLY UNCONSOLIDATED SEDIMENTS

The elastic properties of marine gas-hydrate systems depend on the elastic properties of their host
sediments. In the first part of this section, we present the background and theoretical aspects of rock-physics modeling of the elastic properties of unconsolidated marine sediments found in near-sea-floor, deep-water environments. In the second part, we summarize a practical methodology for estimating these elastic properties.

**Background and Theory**

Newly deposited sediments at the bottom of the ocean are in a pseudosuspension regime, and their shear strength is nearly zero. The \(V_p/V_s\) ratio is very large within the first few meters of sediments and tends to infinity at the boundary between ocean-bottom sediments and the water column.

Most rock-physics models for unconsolidated sediments are based on contact models, such as Hertz-Mindlin’s theory (Mindlin, 1949) for describing the elastic properties of granular materials. The porosity at which a granular composite ceases to be a suspension and becomes grain-supported is called the critical porosity. Dvorkin and Nur (1996) assumed that, at critical porosity, the effective elastic moduli of the dry-mineral frame of sediments can be calculated using Hertz-Mindlin’s contact theory for elastic particles. For porosity values smaller than the critical porosity, the elastic properties of the dry-mineral frame are estimated using the modified Hashin and Shtrikman (1963) lower bound. For porosity values larger than the critical porosity, Dvorkin et al. (1999) proposed to use the modified Hashin and Shtrikman upper bound to derive the elastic properties of the dry-mineral frame of granular materials. Gassmann’s (1951) theory is then used to derive the elastic properties of the sediments saturated with fluids. These modeling assumptions have been applied to marine sediments (Prasad and Dvorkin, 2001) in various areas around the globe. However, Hertz-Mindlin’s theory is not appropriate for sediments within the first 200 mbsf (656 ftbsf) where the S-wave velocity is very small and, implicitly, the \(V_p/V_s\) ratio is very large, as in-situ observations from four-component ocean-bottom-cable (4C-OBC) data suggest (Backus et al., 2006; Hardage et al., 2009). A very-low-S-wave-velocity behavior has also been observed in laboratory measurements on unconsolidated sediments (e.g., Zimmer, 2003; Yun et al., 2005), which show significantly lower \(V_s\) than that predicted by Hertz-Mindlin’s theory. The main reason for the discrepancy between S-wave measurements and theory is that the Hertz-Mindlin model overestimates the shear modulus for granular composites. The Hertz-Mindlin theory assumes infinite friction at grain contacts and does not allow grain rotation and slip at grain boundaries, which probably occur, especially at low effective pressures.

The new model we propose is based on the initial approach of Dvorkin and Nur (1996) and Dvorkin et al. (1999), with the distinction that, at critical porosity, the elastic properties of newly deposited deep-water sediments are described by Walton’s (1987) contact theory. Walton’s model, like other contact models (Digby, 1981; Mavko et al., 1998), assumes that the granular material is made up of an aggregate of perfect spheres. The elastic moduli of such a granular material derived using Hertz and Mindlin’s solutions for the displacement of two identical spheres in contact under normal and shear forces are as follows (Mavko et al., 1998):

\[
K_{\text{eff}} = \frac{C(1-\phi)}{12\pi R} S_n
\]

\[
G_{\text{eff}} = \frac{C(1-\phi)}{20\pi R} (S_n + 1.5S_t)
\]

In equation 1, \(K_{\text{eff}}\) and \(G_{\text{eff}}\) are the effective bulk and shear moduli, respectively, for the granular material; \(C\) is the coordination number, which represents the average number of contacts of a grain with its neighboring grains; \(\phi\) is the porosity of the aggregate of spheres and \(R\) is the radius of the identical spheres representing the grains; \(S_n\) and \(S_t\) are the normal and tangential stiffnesses, respectively, of the two-grain combination and depend on effective pressure, radius of contact of the two grains, and elastic bulk and shear moduli of the grain mineral. The standard Hertz-Mindlin theory’s expressions for the effective bulk and shear moduli expressed in terms of effective pressure are

\[
K_{\text{eff}} = \frac{\sqrt{C^2(1-\phi)^2G^2}}{18\pi^2(1-\nu)^2P}
\]

\[
G_{\text{eff}} = \frac{5 - 4\nu}{5(2-\nu)} \sqrt{\frac{3C^2(1-\phi)^2G^2}{2\pi^2(1-\nu)^2P}}
\]

In equation 2, \(G\) and \(\nu\) are the shear modulus and the Poisson’s ratio, respectively, for the mineral grains, and \(P\) is the effective pressure.

Walton (1987) derived a model that assumes no friction between the grains, referred to as a smooth
contact model. The physical meaning of this model is that grains are allowed to rotate and slip at grain boundaries during seismic excitation. This assumption reduces the effective shear strength of the granular material in comparison with the Hertz-Mindlin theory’s assumption that infinite friction between the grains exists. Mathematically, Walton’s model sets the tangential stiffness \((S_t)\) between two grains in contact to zero in the standard Hertz-Mindlin theory’s expressions in equation 1. Thus, the effective shear modulus \((G_{eff})\) is significantly reduced and becomes 60% of the bulk modulus. The final expressions for the bulk and shear moduli for a random arrangement of dry spheres using Walton’s approach are as follows:

\[
K_{eff} = \frac{C^2(1 - \phi)^2 G^2}{18\pi^2(1 - v)^2} P
\]

\[
G_{eff} = \frac{3}{5} K_{eff}
\]

If we compare equations 2 and 3, we observe that the effective shear modulus for Walton’s smooth model is smaller than that of Hertz-Mindlin’s model, and the ratio between the two shear moduli is simply a function of the Poisson’s ratio of the grain mineral:

\[
G_{eff}^{Walton} = \frac{2 - v}{5 - 4v} G_{eff}^{Hertz–Mindlin}
\]

In the case of quartz grains that have a Poisson’s ratio of 0.08 (Mavko et al. 1998), the effective shear modulus given by Walton’s smooth contact expression \((G_{eff}^{Walton})\) is only 41% of the effective shear modulus predicted by Hertz-Mindlin’s theory \((G_{eff}^{Hertz–Mindlin})\). Therefore, Walton’s model predicts a shear velocity less than half of the shear velocity predicted by Hertz-Mindlin’s model for granular materials with quartz grains.

Walton’s model is particularly appropriate for highly unconsolidated sediments at low effective pressure where grain rotation and slip along grain boundaries are most likely to occur. Walton’s model better explains the low shear strengths and higher \(V_p/V_s\) ratios observed in field data from deep-water, near-sea-floor strata, as well as in laboratory measurements on unconsolidated sediments.

**Methodology**

Elastic properties of unconsolidated deep-water granular materials can be estimated across the full porosity range using the following five steps:

1) Compute the bulk and shear moduli of the dry mineral frame at critical porosity using Walton’s theory (equation 3). If the solid grains have mixed mineralogy, then the bulk modulus and shear modulus \((K_{dry} \text{ and } G_{dry})\) for the solid phase can be computed from those of the individual mineral constituents using Hill’s (1963) average as follows:

\[
K = \frac{1}{2} \left[ \sum_{i=1}^{m} f_i K_i + \left( \sum_{i=1}^{m} f_i K_i \right)^{-1} \right]
\]

\[
G = \frac{1}{2} \left[ \sum_{i=1}^{m} f_i G_i + \left( \sum_{i=1}^{m} f_i G_i \right)^{-1} \right]
\]

In equation 5, \(m\) is the number of mineral constituents for the solid phase, \(f_i\) is the volumetric fraction of the \(i\)th constituent, and \(K_i\) and \(G_i\) are the bulk and shear moduli, respectively, of the \(i\)th mineral constituent.

2) Derive the elastic moduli of the dry frame for porosity values \(\phi\) smaller than critical porosity \(\phi_c\), using the modified Hashin-Shtrikmann lower bound, as follows:

\[
K_{dry} = \left( \frac{\phi/\phi_c}{K^{Walton} + \frac{4}{3} G^{Walton}} + \frac{1 - \phi/\phi_c}{K + \frac{4}{3} G^{Walton}} \right)^{-1}
- \frac{4}{3} G^{Walton}
\]

\[
G_{dry} = \left( \frac{\phi/\phi_c}{G^{Walton} + Z} + \frac{1 - \phi/\phi_c}{G + Z} \right)^{-1} - Z
\]

where

\[
Z = \frac{G^{Walton}}{6} \left( \frac{9K^{Walton} + 8G^{Walton}}{K^{Walton} + 2G^{Walton}} \right)
\]

Moduli \(K_{dry} \text{ and } G_{dry}\) from equation 6 correspond to the weakest possible option for combining the solid phase, which has moduli \(K\) and \(G\), and the critical-porosity material, which has moduli given by Walton’s theory \((K^{Walton} \text{ and } G^{Walton})\).

3) Derive the elastic moduli of the dry frame for porosity values \(\phi\) larger than the critical porosity \(\phi_c\),
using the modified Hashin-Strikman upper bound, by combining the critical-porosity material with the void, which has no bulk or shear strength:

\[
K_{\text{dry}} = \left( \frac{(1 - \phi)/(1 - \phi_c)}{K_{\text{Walton}} + \frac{4}{3}G_{\text{Walton}}} + \frac{(\phi - \phi_c)/(1 - \phi_c)}{\frac{4}{3}G_{\text{Walton}}} \right)^{-1} - \frac{4}{3}G_{\text{Walton}},
\]

\[
G_{\text{dry}} = \left( \frac{(1 - \phi)/(1 - \phi_c)}{G_{\text{Walton}} + Z} + \frac{(\phi - \phi_c)/(1 - \phi_c)}{Z} \right)^{-1} - Z.
\] (8)

Moduli \( K_{\text{dry}} \) and \( G_{\text{dry}} \) from equation 8 correspond to the stiffest possible combination of critical-porosity material and void space.

4) Compute the elastic moduli for the unconsolidated granular material saturated with fluid using Gassmann’s (1951) equation as follows:

\[
K_{\text{sat}} = K \frac{\phi K_{\text{dry}} - (1 + \phi)K_{\text{fl}}K_{\text{dry}}/K + K_{\text{fl}}}{(1 - \phi)K_{\text{fl}} + \phi K - K_{\text{fl}}K_{\text{dry}}/K},
\]

\[
G_{\text{sat}} = G_{\text{dry}}
\] (9)

In equation 9, \( K_{\text{sat}} \) represents the bulk modulus of the fluid-saturated granular material; \( G_{\text{sat}} \) is the shear modulus for the fluid-saturated sediment, which is the same as the shear modulus \( G_{\text{dry}} \) for the dry granular material; and \( K_f \) is the bulk modulus of the fluid, \( K_{\text{dry}} \) is the effective bulk modulus of the dry frame, and \( K \) is the bulk modulus of the solid grains.

5) Compute the bulk density of the unconsolidated sediments, which is given by the following relation:

\[
\rho = \phi \rho_{\text{fl}} + (1 - \phi)\rho_{\text{grain}}, \quad \text{with} \quad \rho_{\text{grain}} = \sum_{i=1}^{m} f_i \rho_i
\] (10)

In equation 10, \( \rho \) is the bulk density of the fluid-saturated sediments, \( \rho_{\text{fl}} \) is the density of the saturating fluid, and \( \rho_{\text{grain}} \) is the density of the solid phase. In case of mixed mineralogy, the density of the solid phase \( \rho_{\text{grain}} \) is given by the volumetric average of the densities of the individual constituents. In the above equation, \( m \) is the number of mineral constituents for the solid phase, \( f_i \) is the volumetric fraction of the \( i \)th constituent, and \( \rho_i \) is the density of the \( i \)th mineral constituent.

Assuming an isotropic medium, bulk modulus and shear modulus (equation 9) together with density (equation 10) completely characterize the elastic properties of unconsolidated, fluid-saturated sediments across the full porosity range. From these three quantities, we can also derive the P- and S-wave velocities of the unconsolidated sediments. Based on the methodology summarized in steps 1 to 5, we proceed to develop rock-physics models for unconsolidated sediments in near-sea-floor strata containing gas hydrates.

**ROCK-PHYSICS MODELS FOR GAS-HYDRATE SYSTEMS**

The effective elastic properties of gas-hydrate systems depend on (1) the elastic properties of the host sediments, (2) the elastic properties of pure gas hydrates, (3) the hydrate concentration, and (4) the geometrical details of how gas hydrates are distributed within the sediments. We analyze various physics-based models for gas-hydrate systems applicable to low-effective-pressure zones that occur within 300 mbsf (984 fbsf). We base our rock-physics modeling of deep-water gas-hydrate systems on the mathematical development of elastic properties of highly unconsolidated sediments presented in the previous section.

We consider the following four rock-physics models for gas-hydrate systems (Figure 1):

- **Model A** assumes gas hydrates are uniformly disseminated throughout the whole volume of rock and act as a part of the load-bearing frame of the host sediments.
- **Model B** assumes gas hydrates are also disseminated throughout the whole volume of rock, but they only float in the porous space and do not change the dry-mineral frame of their host sediments.
- **Model C** assumes an anisotropic, thin-layered medium with layers of pure gas hydrate intercalated with layers of unconsolidated sediments saturated with fluid.
- **Model D** is also an anisotropic, thin-layered medium. However, in this model, the gas hydrates are disseminated in some layers of sediments, occupying 99% of the porous space of these layers, and act as part of the load-bearing frame. These hydrate-bearing layers are intercalated with layers of unconsolidated sediments saturated with fluid.

The key input parameter in all of these models is gas-hydrate concentration. Our goal is to quantitatively...
relate gas-hydrate concentration to seismic attributes, such as P- and S-wave velocities and AVA, under different hypotheses for the morphological occurrence of gas hydrates within the sediments.

**Model A: Gas Hydrates as Part of the Load-Bearing Frame**

In this model, we assume that hydrates are disseminated throughout the volume of sediments and act as part of the load-bearing frame, as initially proposed by Helgerud et al. (1999). We start with the rock-physics model for unconsolidated granular materials presented in the previous section. Then, we derive the effective elastic properties of sediments containing gas hydrates in different concentrations by incorporating the gas-hydrate phase into the mineral frame (Helgerud et al., 1999) and by reducing accordingly the initial porosity of the host sediments. This initial porosity $\phi$ of the unconsolidated sediments without hydrates reduces to an effective porosity $\phi_{\text{eff}}$, as follows:

$$\phi_{\text{eff}} = \phi - c_{\text{gh}}$$  \hfill (11)

where $c_{\text{gh}}$ represents the volumetric gas hydrate concentration in the unconsolidated rock. When we incorporate gas hydrates into the system, the porosity ($\phi$) of our base model for unconsolidated sediments without gas hydrates is replaced in equations 6, 8–10 by the effective porosity defined in equation 11.

When gas hydrate is considered to be part of the mineral frame, its volumetric fraction within the solid phase, $f_{\text{gh}}$, is given as follows:

$$f_{\text{gh}} = \frac{c_{\text{gh}}}{1 - \phi_{\text{eff}}}$$  \hfill (12)

The volumetric fractions of the other various mineral phases change accordingly as

$$f_{i,\text{eff}} = \frac{1 - \phi}{1 - \phi_{\text{eff}}}$$  \hfill (13)

These new volumetric fractions for the mineral phases ($f_{i,\text{eff}}$) and for the gas hydrates ($f_{\text{gh}}$) replace the original volumetric fractions ($f_i$) in equations 5 and 10 respectively for the model of unconsolidated sediments without gas hydrates. In this way, we compute the new bulk modulus, shear modulus, and density of the dry-mineral frame containing gas hydrates with concentration $c_{\text{gh}}$.  

**FIGURE 1.** Graphical sketch of the four models of gas hydrate systems: load-bearing gas hydrates (model A), pore-filling gas hydrates (model B), thin layers of pure gas hydrate intercalated with unconsolidated sediments (model C), thin layers of disseminated, load-bearing gas hydrates intercalated with unconsolidated sediments (model D). Hydrates are represented in light blue and sediments in black. Figure used with permission of The Leading Edge.
Using equations 11–13 together with the model for unconsolidated sediments presented in the previous section, we compute the elastic properties of sediments with disseminated, load-bearing gas hydrates in different concentrations.

Model B: Gas Hydrates as Pore-Filling Clathrates

In this model, we assume that gas hydrates are distributed in the porous space, but they do not alter the dry-mineral frame of the sediments (Helgerud et al., 1999). This model implies that the gas hydrates are floating in the pores, away from the grains, and their effect is to modify only the elastic properties of the material filling the sediment pores without affecting the shear strength of the host sediments.

We assume the particles of gas hydrates are suspended in the fluid that saturates the pores. The elastic properties of the material filling the pores (a mixture of hydrate clathrates and fluid) are given by Reuss (1929) averaging, implying that this mixture does not have any shear strength. The bulk modulus of the mixture of hydrate clathrates suspended in fluid is thus given as:

\[ K_{gh_{\text{fluid}}} = \left( \frac{f_{gh} K_{gh} + f_{\text{fluid}} K_{\text{fluid}}}{K_{gh} + K_{\text{fluid}}} \right)^{-1} \tag{14} \]

In equation 14, \( K_{gh_{\text{fluid}}} \) is the bulk modulus of the mixture of gas hydrates and fluid in the porous space, \( f_{gh} \) represents the fraction of gas hydrates that occupies the porous space, which ranges from 0 to 1, and \( f_{\text{fluid}} = 1 - f_{gh} \) is the fraction of fluid that occupies the porous space.

The volumetric gas hydrate concentration, \( c_{gh} \), in the whole volume of sediments is related to the volumetric fraction of gas hydrates in the pores (\( f_{gh} \)) and is given by

\[ c_{gh} = \phi f_{gh} \tag{15} \]

where \( \phi \) is the porosity of the host sediments.

To summarize, we start with the rock-physics model for unconsolidated sediments presented in the previous section. Initially, these sediments with porosity \( \phi \) are saturated with fluid. Next, we predict the elastic properties of the sediments with gas hydrates in the pores using Gassmann’s (1951) theory. The bulk modulus of the mixture of gas hydrates and fluid (\( K_{gh_{\text{fluid}}} \)) from equation 14 then replaces the fluid bulk modulus (\( K_f \)) from equation 9 in the original model for unconsolidated sediments without gas hydrates.

This model predicts that the shear strength of sediments with hydrates in their pores is the same as the shear strength of sediments that are 100% saturated with fluid. Therefore, the shear modulus of the sediments with pore-filling gas hydrates is exactly the same as the shear modulus of the unconsolidated sediments in equation 9. However, Gassmann’s (1951) theory predicts that the bulk modulus of the sediments with gas hydrates in the pores will increase as gas hydrate concentration increases.

The bulk density of the sediment changes when gas hydrates replace part of the fluid in the porous space because the densities of fluid and gas hydrates are different. The adjusted bulk density is given by

\[ \rho = \left( f_{gh} \rho_{gh} + (1 - f_{gh}) \rho_{\text{fl}} \right) \phi + (1 - \phi) \rho_{\text{grain}} \tag{16} \]

In equation 16, \( \rho \) is the bulk density of the sediments with pore-filling gas hydrates, \( \rho_{gh} \) is the density of the gas hydrates, \( f_{gh} \) is the fraction of gas hydrates that replaces the fluid in the pores, \( \rho_{\text{fl}} \) is the density of the fluid, \( \phi \) is the porosity of the unconsolidated sediments, and \( \rho_{\text{grain}} \) is the density of the mixture of mineral grains (equation 10).

Model C: Thin-Layered Model with Pure Gas Hydrates

Our third model is represented by a layered medium made up of thin beds of pure gas hydrates intercalated with unconsolidated marine sediments. Backus (1962) showed that, in the long-wavelength limit, a stratified medium with individual isotropic layers is effectively anisotropic. If the thin layers are horizontal, this type of anisotropy is transversely isotropic with a vertical axis of symmetry (VTI). The anisotropic effective elastic properties of such a thin-layered medium depend on the elastic properties of the individual layered materials and their volumetric proportions in the rock (Backus, 1962).

In our case, the stratified medium is composed of two different materials: pure gas hydrates and unconsolidated marine sediments saturated with fluid. The elastic properties of pure gas hydrate are known (Sloan, 1998), and the elastic properties of unconsolidated sediments saturated with fluid can be estimated from the rock-physics model presented in the previous section. Therefore, using Backus averaging, we can
predict the elastic stiffness matrix for a layered medium having different gas-hydrate concentrations. The volumetric concentration of gas hydrate \( (c_{\text{gh}}) \) for the whole volume of sediments is equal to the volumetric fraction \( (f_{\text{gh}}) \) of gas hydrate layers in the stratified medium:

\[
c_{\text{gh}} = f_{\text{gh}}
\]  

Because this is an anisotropic model, the effective elastic properties will vary with direction. In particular, S waves polarized perpendicular to the layers (slow direction) will propagate with a slower velocity than S waves polarized parallel with the layers (fast direction). Also, P waves propagating orthogonal to the layers will have slower velocity than the P waves propagating along the layers.

The bulk density of the anisotropic thin-layered model is given by volumetric averaging of the densities of the two constituent materials: gas hydrates \( (\rho_{\text{gh}}) \) and unconsolidated sediments saturated with fluid \( (\rho) \) from equation 10.

**Model D: Thin-Layered Model with Disseminated Gas Hydrates**

The last model we consider is another thin-layered medium, with layers of gas hydrates disseminated in the unconsolidated sediments. The gas hydrates in these layers act as part of the load-bearing frame and occupy a certain fraction \( (f_{\text{gh}}) \) of the porous space of the host sediment. Layers containing gas hydrates are intercalated with layers of unconsolidated sediments that are 100% saturated with fluid. The elastic properties of the layers containing gas hydrates are estimated using model A (Figure 1), whereas the elastic properties of the unconsolidated sediments saturated with fluid are estimated using the rock-physics model presented in the previous section. In the assumption that the medium is thinly bedded, we again use Backus (1962) averaging to determine the elastic properties of the layered sediments for different gas-hydrate concentrations. The overall volumetric concentration of gas hydrate in the rock is given by

\[
c_{\text{gh}} = f_{\text{lr}} f_{\text{gh}} \phi
\]  

In equation 18, \( f_{\text{lr}} \) represents the volumetric fraction of layers containing disseminated gas hydrates, \( f_{\text{gh}} \) represents the proportion of gas hydrates occupying the porous space of these layers, and \( \phi \) is the porosity of the layers containing gas hydrates. For this model, we assume gas hydrates occupy 99% of the porous space \( (f_{\text{gh}} = 0.99) \). This model is also anisotropic, and P- and S-wave velocities will vary with direction as they do for the layered model presented in the previous subsection.

The bulk density of the thinly bedded medium is given by volumetric averaging of the densities of the two constituent materials: the sediments with load-bearing gas hydrate (see model A) and the unconsolidated sediments saturated with fluid (equation 10).

**RESULTS**

In this section, we present P- and S-wave velocities predicted by the four different rock-physics models of gas-hydrate systems described in the previous section. The elastic parameters used in the modeling are summarized in Table 1.

In the examples from Figures 2–4, we consider the unconsolidated sediments to be represented by pure quartz grains saturated with brine. The sediments are assumed to be at critical porosity of 37%. The coordination number \( C \) is considered to be 8, and the effective pressure is set at a low value of 0.01 MPa. This low value of effective pressure corresponds to a couple of meters (~6 ft) below sea floor where gas hydrates are known to be present. This low effective pressure value can also correspond to deeper strata within overpressured zones, commonly encountered in the Gulf of Mexico.

Figures 2–4 present the modeling results for P-wave velocity \( (V_p) \), S-wave velocity \( (V_s) \), and \( V_p/V_s \) ratio, respectively, as a function of gas-hydrate concentration \( (c_{\text{gh}}) \) for the four rock-physics models (A, B, C, and D) considered. For the two anisotropic layered models (C and D), we display two curves corresponding to the velocity of waves with their particle displacement vector parallel with the layering (solid line) and to the velocity of waves with their particle displacement vector orthogonal to the layers (dotted line).

### Table 1. Elastic properties of the sediment constituents.

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Bulk Modulus (GPa)</th>
<th>Shear Modulus (GPa)</th>
<th>Density (kg/m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>37</td>
<td>44</td>
<td>2650</td>
</tr>
<tr>
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FIGURE 2. Primary (P)-wave velocity as a function of the volumetric fraction of gas hydrate ($c_{gh}$) in pure quartz sediments for the four rock-physics models considered. Model A: load-bearing gas hydrates disseminated in the whole volume of sediments; model B: pore-filling gas hydrates disseminated in the whole volume of sediments; model C: layers of pure gas hydrates producing slow P waves (dotted line) and fast P waves (solid line); model D: layers of disseminated, load-bearing gas hydrates producing slow P waves (dotted line) and fast P waves (solid line). 1000 m = 3281 ft. Figure used with permission from The Leading Edge.

FIGURE 3. Secondary (S)-wave velocity as a function of the volumetric fraction of gas hydrate ($c_{gh}$) in pure quartz sediments for the four rock-physics models considered. Model A: load-bearing gas hydrates disseminated in the whole volume of sediments; model B: pore-filling gas hydrates disseminated in the whole volume of sediments; model C: layers of pure gas hydrates producing S waves with slow polarization (dotted line) and fast polarization (solid line); model D: layers of disseminated, load-bearing gas hydrates producing S waves with slow polarization (dotted line) and fast polarization (solid line). 1000 m = 3281 ft. Figure used with permission from The Leading Edge.
From all of these figures, we can observe that P- and S-wave velocities greatly depend on the geometrical details of how gas hydrates are distributed in their host sediments.

The results presented in Figure 2 show that, for all of the rock-physics models considered, the presence of gas hydrates increases the P-wave velocity in the sediments. The smallest increase in P-wave velocity with gas-hydrate concentration is obtained for the thin-bedded model with layers of pure gas hydrates (model C), whereas the largest increase in P-wave velocity is obtained for the models having disseminated, load-bearing gas hydrates (models A and D). The rate of change of $V_p$ with hydrate concentration is greatest if measured parallel with the thin layers of load-bearing clathrates (model D).

If we consider the two thin-layered models, models C and D, we see that the fast P-wave velocity propagating parallel with the layers (solid lines) and the slow P-wave velocity propagating orthogonal to the layers (dotted lines) are different for these two morphologies (Figure 2). For example, at 0.3 volumetric concentration of gas hydrate in the rock, fast P-wave velocity can range from 2000 m/s (6562 ft/s) for layers of pure gas hydrates (model C, solid line) to more than 3000 m/s (9842 ft/s) for layers of disseminated, load-bearing gas hydrates (model D, solid line). This huge difference is partly because the elastic moduli of layers having load-bearing gas hydrates are larger than the elastic moduli of pure gas hydrates considered in the models. In addition, the volumetric fraction of layers having disseminated gas hydrates is 2.73 times larger than the corresponding fraction of layers of pure gas hydrates for the same volumetric gas hydrate concentration in the two layered media. This value of 2.73 is obtained by comparing equations 17 and 18, and setting $\phi$, the porosity of the unconsolidated sediments, to a value of 0.37. When the volumetric fraction for the layers of disseminated gas hydrates in the thinly bedded medium is 1, the rock becomes isotropic, and the end point of the two curves for model D coincides with the end point of the curve for model A (Figure 2).

From Figure 3, we observe that S-wave velocity also increases with gas-hydrate concentration for all of the rock-physics models considered. However, for the model in which the gas hydrates float in the porous space without contributing to the load-bearing frame of the host sediments (model B), the increase in S-wave velocity is insignificant. In this model, the gas hydrates are not connected with the mineral frame, and although they are solid, they do not support the shear load. Gassmann’s (1951) theory, which is used in this model, predicts that the shear modulus
of the sediments does not change with gas-hydrate concentration. Only a minor difference in the densities of brine and gas hydrate is observed, and thus only a very small increase in $V_s$ occurs when gas hydrate replaces brine in the pores.

For large hydrate concentrations (at the limit, for 100% hydrate concentration in pores), the pore-filling hydrate model (model B) is physically unrealistic, predicting that the gas hydrate would not contribute to the shear strength of the sediments. However, for the high-porosity sediments from deep-water near-sea-floor strata with a relatively small hydrate concentration, this model is physically possible. Porosities of the sediments from deep-water near-sea-floor strata can be very large (50–60% in the first 150 mbsf [492 fbsf]). For example, for sediments with an initial porosity of 60% and a hydrate concentration in pores of 50%, a 30% remaining porosity saturated with brine is observed. In this case, it is possible that the hydrate may float within the remaining porous space saturated with brine, without supporting the shear load of the host sediments. Therefore, when the hydrates are formed away from the grains, the pore-filling model is a good representation for the elastic behavior of sediments with a relatively small hydrate concentration in pores.

Once the gas-hydrate concentration grows larger, the hydrate may start supporting the shear strength as well, and a load-bearing model (model A) would be more appropriate.

The results presented in Figure 3 show that a large increase in S-wave velocity occurs when the waves are polarized parallel with the layers of a medium having thin beds of disseminated, load-bearing gas hydrates (model D, solid line). In this particular type of hydrate-sediment morphology, the velocity of S waves propagating with their displacement vector polarized parallel with the layering (fast direction) increases significantly for small gas-hydrate concentrations. This behavior suggests that for disseminated, load-bearing gas hydrates occurring in thin beds, we may be able to detect small hydrate concentrations using anisotropic S-wave information. The S-wave anisotropy for this model is very large. A huge difference between S-wave velocities polarized parallel (model D, solid line) with and orthogonal (model D, dotted line) to layers of disseminated, load-bearing gas hydrates is observed. S-wave anisotropy for a system of layers of pure gas hydrates (model C) is large as well. Therefore, if gas hydrates occur in thin layers within near-sea-floor strata, we should expect significant shear-wave anisotropy, and this anisotropy may be used with other seismic information to estimate gas-hydrate concentrations.

For model A, S-wave velocity increases very little at small gas-hydrate concentrations and then increases abruptly at gas-hydrate concentrations larger than 0.3 (i.e., when load-bearing clathrates occupy more than 80% of the pore volume). Thus, it will be more challenging to determine small gas-hydrate concentrations using S-wave information when load-bearing clathrates are uniformly disseminated within the sediments throughout thick intervals. Note again that the end points of the curves for model A and model D coincide, as they should.

Figure 4 presents the $V_p/V_s$ ratio for the four rock-physics models for gas-hydrate systems considered. This velocity ratio decreases with gas-hydrate concentration for all models except for model B, which assumes pore-filling gas hydrates. This anomalous behavior for model B occurs because P-wave velocity increases with gas-hydrate concentration (Figure 2) whereas S-wave velocity remains practically constant (Figure 3). As a result, the $V_p/V_s$ velocity ratio for model B increases with gas-hydrate concentration.

From Figure 4, we observe that the $V_p/V_s$ velocity ratio for slow P waves and S waves polarized orthogonal to layers of disseminated, load-bearing gas hydrates (model D, dotted line) is similar to the $V_p/V_s$ ratio of the rock with load-bearing gas hydrates uniformly disseminated in the whole volume of sediments (model A). For these two models, the velocity ratio decreases slightly at small gas-hydrate concentrations and more abruptly for larger gas-hydrate concentrations. This behavior suggests that it will be challenging to estimate small gas-hydrate concentrations using $V_p/V_s$ ratios in media represented by either of these curves. However, for anisotropic models C and D, a significant decrease in the $V_p/V_s$ ratio at low gas-hydrate concentrations for the waves with a particle displacement vector parallel with the layers (fast direction) occurs. This modeling result suggests that, for layered gas-hydrate morphologies, we can sometimes use $V_p/V_s$ ratios and anisotropy information to detect small gas-hydrate concentrations in sediments.

The theory presented in the preceding sections allows for mixed mineralogy as well as for different saturating fluids. Figures 5–7 are similar to Figures 2–4, respectively. However, more curves are displayed in each of the panels, corresponding to different clay content in the sediments, from 0% to 100%, with a 25% increment. The parameters used in the modeling, such as critical porosity and coordination number $C$, vary with mineralogy (Murphy, 1982). For clean quartz grains (0% clay content), the critical porosity is assumed to be 37% and the coordination number $C$ is
considered to be 8. For pure clay minerals (100% clay content), we use a larger critical porosity of 67% and a smaller coordination number equal to 4, as many geotechnical data suggest (Murphy, 1982). For each mixture of quartz and clay minerals, we derive the values for critical porosity and coordination number using a linear interpolation between the corresponding values for the two end members of pure quartz and pure clay. Also, at subsea-floor depths where hydrates are stable, the porosity of clay-rich sediments is larger than the porosity of pure quartz grains. Therefore, we compute $V_p$ and $V_s$ as a function of gas-hydrate concentration for sediments with different porosity values: 37% for pure quartz and 50% for pure clay minerals. For each mixture of quartz and clay, we use again a linear interpolation between the values for the two end members of pure quartz and pure clay. The effective pressure we use in these calculations is equal to 0.5 MPa, which corresponds to a depth below sea floor of approximately 60 m (197 ft).

As expected, the P- and S-wave velocities decrease with increasing clay content (and implicitly with increasing porosity), as we can observe from each panel in Figures 5 and 6. For model A of load-bearing gas hydrates, as porosity and clay contents of the sediments increase, it becomes more challenging to estimate small gas-hydrate concentrations, especially using shear-wave data. For layered model D, we observe that both P- and S-wave anisotropies decrease with increasing clay content, because the elastic properties of clay minerals are closer to those of hydrates than those of quartz minerals. Therefore, the modeling shows that we should expect larger anisotropy in P- and S-wave velocities if the hydrate layers are intercalated with clean sands than with clay-rich sediments. Models D and C consider only the anisotropy caused by thin layers and consider the clay-rich sediments to be isotropic. This assumption may hold for sediments immediately below sea floor. However, as the depth increases, the stress-induced anisotropy of clays is going to increase. Therefore, at larger depths, models C and D with clay-rich sediments should be adjusted to account also for anisotropy caused by the presence of clay minerals. The different behaviors of the $V_p/V_s$ ratio for the different rock-physics models presented (Figures 4, 7) suggest that amplitude-versus-incidence-angle analyses...
of P waves and converted PS waves may provide valuable insights into gas-hydrate morphologies and concentrations.

Figure 7 presents the $V_p/V_s$ ratio for the four rock-physics models for gas-hydrate systems considered, for which we also vary the clay content. Independently of the mineralogy, this velocity ratio decreases with gas-hydrate concentration for all models except for model B, which assumes pore-filling gas hydrates, which float in the pores and do not support the shear load. As expected, the $V_p/V_s$ ratio increases with increasing clay content.

**Figure 6.** Secondary ($S$)-wave velocity as a function of the volumetric fraction of gas hydrate ($c_{gh}$) in sediment mixtures with variable clay content for the four rock-physics models considered. Model A: load-bearing gas hydrates disseminated in the whole volume of sediments; model B: pore-filling gas hydrates disseminated in the whole volume of sediments; model C: layers of pure gas hydrates producing slow $S$ waves (dotted lines) and fast $S$ waves (solid lines); model D: layers of disseminated, load-bearing gas hydrates producing slow $S$ waves (dotted lines) and fast $S$ waves (solid lines). Different curves correspond to different clay content (from 0 to 100% with 25% increment), different critical-porosity values, and different coordination numbers. The curves are computed at increasing porosity values as the clay content increases. The arrows indicate increasing clay content. 1000 m = 3281 ft.

**PP AND PS AVA MODELING: BASE OF HYDRATE STABILITY ZONE**

To evaluate the potential value of AVA technology for studying gas-hydrate systems, we simulate the AVA response for PP and PS reflections from an interface between the base of the GHSZ and sediments immediately below that interface that contain free gas (Figure 8). The gas-hydrate systems considered in this AVA modeling are represented by isotropic rock-physics models A and B (Figure 1), which have gas hydrates disseminated in the host sediments. In model A, the gas hydrates act as part of the load-bearing frame of the sediments, whereas in model B, the gas hydrates fill the pores and do not change the elastic properties of the dry-mineral frame.

The sediments below the GHSZ are assumed to contain free gas. Their elastic properties are estimated based on the rock-physics model for unconsolidated sediments presented in this article.

First, we consider the pores to be 80% saturated with free gas, and the remaining pore space to be occupied by brine. We use Reuss (1929) averaging to derive the bulk modulus of the mixture of gas and liquid. We then use Zoeppritz’s (1919) equations to derive PP and PS reflectivity as a function of the incidence angle.

Figure 9 presents the results for AVA modeling of PP (left panel) and PS (right panel) reflectivity as a function of the incidence angle for the model with load-bearing gas hydrates (model A). Each curve corresponds to a different gas-hydrate concentration in the rock. The arrow indicates increasing gas-hydrate concentrations in the pores, from 0% to 99%.
The magnitude of the normal-incidence PP reflectivity increases with increasing gas-hydrate concentration, because the difference between the elastic properties of the gas-hydrate system and the free-gas sediments increases with gas-hydrate concentration. For this model, we observe that PP reflectivity at small incidence angles (near offsets) can better differentiate gas-hydrate concentrations than can PP reflectivity at larger angles of incidence (far offsets). At small incidence angles, PP reflectivity becomes brighter as hydrate concentration increases. For small angles, the increase in PP reflectivity between 0 and 99% hydrate concentrations is a little more than 60%, which should be measurable with good quality seismic data. However, PP amplitudes increase by only 12% when hydrate concentration changes from 0 to 40%, which would be difficult to detect from seismic data.

In contrast to PP reflectivity, if multicomponent seismic data are acquired so that PS reflectivity can be measured at incident angles of about 30°, PS amplitude increases by a factor of about 3 as hydrate concentration increases from 0 to 40% or by a factor of more than 30 if hydrate concentration changes from 0 to 99%. These amplitude variations should be measurable with reasonable-quality, multicomponent seismic data.

For gas-hydrate concentrations between 0 and 0.7, the PP reflectivity curves in Figure 6 are the type associated with class 3 reservoirs in the Gulf of Mexico, reservoirs for which the reflectivity at zero offset is negative, and its AVA gradient is also negative (Rutherford and Williams, 1989; Ruger, 2002; Roden et al., 2005). As hydrate concentration increases beyond 0.7, the PP AVA behavior shifts to a response associated with class 4 reservoirs, reservoirs for which the reflectivity at zero offset is again negative but have a positive AVA gradient. The hydrate-concentration

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**Figure 7.** Primary-wave velocity/secondary-wave velocity (Vp/Vs) velocity ratio as a function of the volumetric fraction of gas hydrate (Cgh) in sediment mixtures with variable clay content for the four rock-physics models considered. Model A: load-bearing gas hydrates disseminated in the whole volume of sediments; model B: pore-filling gas hydrates disseminated in the whole volume of sediments; model C: layers of pure gas hydrates producing slow primary (P) waves (dotted lines) and fast P waves (solid lines); model D: layers of disseminated, load-bearing gas hydrates producing slow P waves (dotted lines) and fast P waves (solid lines). Different curves correspond to different clay content (from 0 to 100% with 25% increment), different critical-porosity values, and different coordination numbers. The curves are computed at increasing porosity values as the clay content increases. The arrows indicate increasing clay content. 1000 m = 3281 ft.

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**Figure 8.** Model assumed for the base of hydrate stability zone (BHSZ).
range that produces these class 4 reservoir responses corresponds to the hydrate-concentration range in Figure 3, where a significant increase in $V_S$ in the layer above the gas reservoir occurs (layer 1 of Figure 8). This $V_S$ behavior is required for a class 4 PP AVA response (Castagna and Backus, 1993).

Figure 10 shows the results for AVA modeling of PP (left panel) and PS (right panel) reflectivity as a function of the incidence angle in the hypothesis of pore-filling gas hydrates (model B). Again, each curve corresponds to a different gas hydrate concentration in the rock. The arrow indicates the increasing gas hydrate concentration in the pores, from 0% to 99%.

In the hypothesis of pore-filling gas hydrates, only PP reflectivity exhibits any sensitivity to gas-hydrate concentration. All PP AVA reflectivity curves in Figure 10 are class 3 responses. Using PP reflection attributes to identify hydrate concentrations could be challenging because PP reflection amplitude increases by only 7% when hydrate concentration increases from 0 to 40%, and by a little more than 30% if concentration ranges from 0 to 99%. For this model, the shear strengths of the sediments containing gas hydrates do not change with gas-hydrate concentration. Moreover, shear velocities in sediments containing pore-filling gas hydrates and in sediments containing free gas are similar. The only change in $V_S$ is caused by a small density effect. Therefore, in this case, all PS reflections are weak and change very little as gas-hydrate concentration increases.

The modeling results presented in Figures 9 and 10 show that PP reflectivity cannot differentiate between the two hypotheses of gas-hydrate occurrence. However, PS reflectivity is very different for load-bearing gas hydrates (model A) and pore-filling gas hydrates (model B). Figures 11 and 12 are similar to Figures 9 and 10, respectively. The only difference is that, in the examples from Figures 11 and 12, the free gas below the...
GHSZ is 10% instead of 80%. The presence of residual gas of 10% may be a more plausible scenario than having a well-developed gas reservoir immediately below GHSZ. As expected, the AVA responses for 10% free gas from Figures 11 and 12 are very similar with the AVA responses for 80% free gas below GHSZ from Figures 9 and 10. The explanation is that very small amounts of free gas have the same effect on the seismic velocities as economical quantities of gas. Therefore, it may be challenging to estimate the free-gas saturation below the GHSZ.

However, independently of the free-gas saturation below the GHSZ, we can use multicomponent AVA behavior to quantify gas-hydrate concentration. Also, the PS-wave mode can help to differentiate between the two different hydrate morphologies: load-bearing gas hydrates (model A) and pore-filling gas hydrates (model B).

Therefore, multicomponent seismic technology, and the use of PS-mode AVA behavior in particular, can be a powerful tool to understand the way gas hydrates are distributed in relation to their host sediments. Using combinations of Vp/Vs ratios and AVA analyses of P waves and converted PS waves should improve estimates of gas-hydrate concentrations.

COMPARING ROCK-PHYSICS MODELING RESULTS WITH LABORATORY MEASUREMENTS

We compare our rock-physics modeling results for model A with laboratory measurements on synthetic
gas hydrates formed in unconsolidated fine sands. Yun et al. (2005) performed laboratory measurements on sediments containing gas hydrates in concentrations ranging from 0 to 100% gas hydrate in the porous space. For gas-hydrate concentrations smaller than 100%, the remaining pore space was occupied by brine. The porosity of the sand samples was 0.37. The laboratory measurements were performed at low effective pressure, smaller than 0.01 MPa.

In Figures 13 and 14, we compare their laboratory measurements for P- and S-wave velocity as a function of gas-hydrate concentration in the pores with our results for rock-physics model A (load-bearing gas hydrates). The left panels in these figures show the laboratory measurements, and the right panels show our rock-physics modeling results. The unconsolidated sediments in the rock-physics model are represented by quartz grains at critical porosity, assumed to be 0.37, the same as the porosity of the sand samples used in the laboratory measurements.

Figure 13 shows that P-wave velocity increases with gas-hydrate concentration for both laboratory measurements and the rock-physics model of load-bearing gas hydrates. The increase in P-wave velocity is non-linear and is larger when gas-hydrate concentrations in pores exceed 50%. For gas-hydrate concentrations smaller than 50%, the increase in the P-wave velocity caused by the presence of gas hydrates is relatively small. We observe a good agreement between the laboratory measurements of P-wave velocity as a function of gas-hydrate concentration and our rock-physics modeling results.

Figure 14 shows that S-wave velocity increases with gas-hydrate concentration for both laboratory measurements and the rock-physics model of load-bearing gas hydrates. The increase in S-wave velocity is larger when gas-hydrate concentrations in pores exceed 50%. For gas-hydrate concentrations smaller than 50%, the increase in S-wave velocity caused by the presence of gas hydrates is very small. We observe a relatively good agreement between the laboratory measurements of S-wave velocity as a function of gas-hydrate concentration and our rock-physics modeling results. Based on the agreement between the laboratory data and rock-physics modeling results, we conclude that, in this case, the gas hydrates act as part of the load-bearing frame of their host sediments.

Laboratory measurements by Winters et al. (2004) on samples containing natural gas hydrates from the Mallik 2L-38 well, Mackenzie delta, show that these naturally formed gas hydrates also act as part of the load-bearing frame of their host sediments. Winters et al. (2004) also performed measurements on Ottawa sands with methane gas hydrate formed in the laboratory. These measurements on Ottawa sands suggested that the gas hydrates cement their host sediments.

![Figure 13. Comparison of laboratory and model A results for primary (P)-wave velocities as a function of gas hydrate concentration in the pores. (left) Yun et al. (2005) laboratory results. (right) Rock-physics modeling results when gas hydrates are assumed to be part of the dry frame of the sediments (model A). Figure used with permission of The Leading Edge.](image1)

![Figure 14. Comparison of laboratory and model A results for secondary (S)-wave velocities as a function of gas hydrate concentration in the pores. (left) Yun et al.’s (2005) laboratory results. (right) Rock-physics modeling results when gas hydrates are assumed to be part of the dry frame of the sediments (model A). Figure used with permission of The Leading Edge.](image2)
Cementation hypothesis implies a large increase in P- and S-wave velocities for small gas-hydrate concentrations (Dvorkin et al., 1994; Ecker et al., 1998). Such behavior is not commonly observed in published field data acquired across gas-hydrate zones. Winters et al.'s (2004) laboratory study shows an apparent contradiction between naturally formed gas hydrates, which appear to act as part of the load-bearing frame and do not cement the grains, and their laboratory-formed gas hydrates, which do appear to cement the grains of the host sediments.

Based on our rock-physics results from model A, we observe that Yun et al. (2005) laboratory measurements on synthetic gas hydrates (Figures 13, 14, left panels) are in qualitative agreement with Winters et al.'s (2004) laboratory observations on naturally formed gas hydrates, both results suggesting a load-bearing frame hypothesis for hydrates within their host sediments.

CONCLUSIONS

We have presented rock-physics models for highly unconsolidated sediments that host gas hydrates in deep-water, near-sea-floor strata and that are characterized by large Vp/Vs ratios. We describe four possible rock-physics models for such gas-hydrate systems in which we assume various morphologies for hydrates in relation to their host sediments. Modeling results show that the elastic properties of gas-hydrate units depend on the geometrical details of the gas-hydrate distribution within the sediments. Therefore, seismic data and especially multicomponent seismic data that provide both P- and S-wave information can help investigators understand how gas hydrates are distributed in marine sediments and aid in the determination of gas-hydrate concentration.

Comparing the rock-physics models and laboratory measurements on synthetic gas hydrates in unconsolidated sands by Yun et al. (2005) suggests that, for this example, gas hydrates act as part of the load-bearing frame of their host sediments. We find good agreement between the theoretical predictions for P- and S-wave velocities corresponding to this rock-physics model (our model A) and laboratory measurements on synthetic gas hydrates in unconsolidated sands. Other laboratory data (performed on naturally formed gas hydrates from the Mallik well, Mackenzie delta [Winters et al., 2004]), which also suggest the load-bearing frame hypothesis for gas-hydrate occurrence within sediments, exist. We can conclude that, in some natural environments, gas hydrates may act as part of the dry-mineral frame of their host sediments.

More laboratory and field studies are needed to understand the complex mechanisms of gas-hydrate formation in sediments, mechanisms that may lead to various hydrate-sediment geometries depending on the specific in-situ conditions.

In some deep-water environments, there may be layered types of hydrate morphologies. Our modeling shows that, in these media, the effective elastic properties of stratified near-sea-floor sediments containing gas hydrates may be highly anisotropic, and the acquisition of fast and slow components of multicomponent seismic data has great value.

The AVA modeling indicates that multicomponent seismic technology can be important for understanding the way gas hydrates are distributed in relation to their host sediments and for estimating gas-hydrate concentrations. Using combinations of PP and PS AVA reflectivity, P- and S-wave interval velocities, and Vp/Vs ratios should improve our understanding of deep-water hydrate conditions.

We conclude that rock-physics modeling provides powerful tools for understanding gas-hydrate formation and deposition in deep-water environments and for quantifying the amount of gas hydrate present in deep-water, near-sea-floor sediments under different geological conditions.

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