Joint analysis of P-wave velocity and resistivity for morphology identification and quantification of gas hydrate

Tao Liu, Xuewei Liu, Tieyuan Zhu

School of Geophysics and Information Technology, China University of Geosciences, Beijing, 100083, China
Department of Geosciences, The Pennsylvania State University, State College, PA, 16802, United States
EMS Energy Institute, The Pennsylvania State University, State College, PA, 16802, United States

1. Introduction

Natural gas hydrate is of worldwide interest due to its widespread occurrence in oceans and permafrost regions (Kvenvolden, 1993), its effects on climate change and seafloor stability (Archer, 2007; Nixon and Grozic, 2007) and its potential as an energy source (e.g., Ruppel, 2007). Distributions of gas hydrate within host sediments are typically categorized as either pore-filling or fracture-filling, according to the scientific reports of national gas hydrate expeditions in China, India, America, Korea, and Canada (Holland et al., 2008). Pore-filling gas hydrate accumulates uniformly in pore spaces, whereas fracture-filling gas hydrate accumulates in fractures with different shapes such as veins, layers, or nodules (Holland et al., 2008) caused by over-pressured gas and fluid. Fracture-filling gas hydrate often yields anisotropic physical properties (e.g., seismic velocity), as widely reported in previous investigations, e.g., Hydrate Ridge offshore Oregon in the United States (e.g., Kumar et al., 2006), the Krishna-Godavari Basin in India (e.g., Lee and Collett, 2009; Cook et al., 2010; Ghosh et al., 2010; Sriram et al., 2013), the Ulleung Basin in Korea (Holland et al., 2008), and the eastern Pearl Mouth Basin in China (Qian et al., 2017; Liu and Liu, 2018a, b).

One of the major goals of gas hydrate expeditions is the quantification of in situ hydrate content, which is strongly dependent on the quantitative relationship between the physical properties of the sediment and gas hydrate content. Previous studies have shown that gas hydrates with different morphologies have significantly different effects on sediment properties (e.g., Lee and Collett, 2009). Thus, a knowledge of hydrate morphology is crucial to estimate hydrate saturation from field data. Effective and safe exploitation also requires the knowledge of gas hydrate morphology, to avoid procedures that may reduce seabed stability and lead to the release of vast amounts of greenhouse gases (e.g., Dickens, 2003; Sultan et al., 2004; Ruppel, 2011).

Current methods to quantify gas hydrate morphology and saturation have their strengths and limitations. Direct information may be obtained from pressure cores, examined using infrared scanners and pore water measurements. However, pressure cores are limited and rarely obtain continuous successions (Sain et al., 2010), making it difficult to...
characterize the spatial variation of a hydrate reservoir (Cook, 2010). Moreover, the deployment of pressure cores is technically difficult, and includes a high possibility of failure. For example, 59% of the pressure core deployments failed during the India National Gas Hydrate Program Expedition 1 (NGHP-01) (Collett et al., 2008). Resistivity at the bit (RAB) images can detect small-scale fractures filled with hydrate (Weinberger and Brown, 2006; Cook et al., 2008; Cook and Goldberg, 2008, 2014), but this becomes difficult when gas hydrate is present in veined or nodular structures (Cook, 2010). Moreover, RAB images are unable to detect fractures dipping at angles less than 10° (Cook et al., 2010). Recently, Liu and Liu (2018a, b) showed that amplitude variation with angle analysis (AVA) and P-wave velocity combined with density information can be used to distinguish the two types of gas hydrate-bearing sediments (GHBS), but their methods cannot provide accurate estimates of gas hydrate saturation.

P-wave velocity and resistivity well logs are the most commonly used tools in gas hydrate studies (e.g., Lee and Waite, 2008; Malinverno et al., 2008; Wang et al., 2011). They can provide estimates of hydrate saturation in a continuous depth profile by using predictive models that relate hydrate content to field measurements, such as Archie’s equation (1942) for resistivity and effective medium theory (EMT) for velocity (Helgerud et al., 1999). However, these models are based on the isotropic assumption, the application of which might cause serious errors in the case of anisotropic fracture-filling hydrates (Lee and Collett, 2009; Cook et al., 2010). To interpret anisotropic fracture-filling GHBS that has two unknowns, hydrate saturation and fracture dips, a combination of at least two types of measurements is required. In addition, neither P-wave velocity nor resistivity can be used to identify hydrate morphology independently, because they both increase in the presence of gas hydrates, regardless of the hydrate morphology (e.g., Collet and Ladd, 2000; Lee and Collett, 2009; Cook et al., 2010).

Herein, we propose a joint analysis of P-wave velocity and resistivity aiming to not only identify gas hydrate morphologies, but also provide reliable hydrate saturation estimations. We first employ rock physics modeling to investigate the effects of hydrate morphology on the velocity and resistivity properties of GHBS. Archie’s equation and EMT are used to characterize the isotropic resistivity and velocity of pore-filling GHBS, respectively. Further, we incorporate the theory developed by Kennedy and Herrick (2004) and the transversely isotropic theory (TIT) (Lee and Collett, 2009) to characterize the anisotropic resistivity and velocity of fracture-filling GHBS, respectively. The theoretical models chosen here have been proven to be effective for GHBS in this study area (Qian et al., 2017; Liu and Liu, 2018a, b) as well as in other areas containing gas hydrate (Lee and Collett, 2009, 2012; 2013; Wang et al., 2013). We then test and verify the method using well log data from the gas hydrate sites in the South China Sea. The method is then discussed in terms of its limitations and applicability in different porosities and sediment types.

2. Geological setting

China’s second gas hydrate expedition was conducted from June to September 2013 by the Guangzhou Marine Geological Survey (GMGS2) in the eastern part of the Pearl River Mouth Basin in the South China Sea. This is an active continental margin controlled by complex tectonic interactions among the Pacific, Eurasian, and Indian-Australian plates (Taylor and Hayes, 2011). Five sedimentary basins extend in a southwest-northeast direction in the northern South China Sea, including Pearl River Mouth Basin, Yingge-hai Basin, Taixinan Basin, Beibuwan Basin, and Qiongdongnan Basin. The Pearl River Mouth Basin experienced two main extension stages from the end of lower Cretaceous to Oligocene. The water depth in the study area ranges from 500 to 2500 m. Widely distributed bottom simulating reflectors are observed indicating the occurrence of gas hydrate and free gas (Wang et al., 2009; Li et al., 2013). A large number of erosional canyons are also observed (Li et al., 2013; Zhang et al., 2015). Mud diapirs and faults are well-developed to provide effective permeability pathways for fluid and gas migrations (Wu et al., 2005; Yan et al., 2006). The predominant sedimentary facies of this area are fine-grained marine clays interbedded with sand-rich turbidites (Sha et al., 2015).
3. Data and methods

3.1. Well data

During GMGS2, seismic surveys, logging-while-drilling (LWD), wireline-logging, and coring were undertaken to investigate the nature and distribution of the gas hydrate in the South China Sea. A total of 21 holes were drilled at 13 sites in the central part of the study area (Fig. 1). Different gas hydrate morphologies were discovered by core samples at Sites 05, 07, 08, 09, and 16 (Sha et al., 2015). At Site 09, core samples were only collected at hole 09B, and LWD data were not collected at this hole. At Site 07, no well log data is available. Therefore, for our study we choose the LWD data from Sites 05, 08, and 16 (Fig. 1).

The LWD data acquired during the GMGS2 expedition included caliper, gamma ray, neutron porosity, density, P-wave velocity, and ring resistivity. All these measurements are one-to-one correspondence, and are displayed in Fig. 2. S-wave velocity data were not collected. Pressure cores were collected using a Fugro pressure corer to evaluate gas hydrates at in-situ pressures. The chlorinity of pore-water was measured to provide a reference for gas hydrate saturations (Sha et al., 2015). The formation of gas hydrate excludes chlorine and causes it to diffuse through the sediments. On depressurization, gas hydrates dissociate and generate fresh water in the sediment. Hydrate saturation was derived from the degree of pore-water freshening (Ussler and Paull, 2001).

3.2. Previous work on gas hydrates from well data

Five gas hydrate-bearing intervals confirmed by LWD data and core samples (Sha et al., 2015) are marked by green zones in Fig. 2. At Site 05, one layer of pore-filling gas hydrate was found in the interval from 197 to 207 m below mudline (BML), with high resistivities and velocities (Fig. 2). According to pore water chlorinity measurements, gas hydrate occupied up to 31% of the pore space in this interval. At Site 08, two layers of fracture-filling gas hydrate were identified from 8 to 21 m and 65–98 m BML, respectively. Gas hydrate filled as thin veins in the upper layer and as thick layers in the lower intervals. The velocity of the upper hydrate interval does not show high velocity anomalies except for an extremely thin layer. This is likely because of the low hydrate saturation, which ranges from 0 to 14% (Sha et al., 2015). The lower hydrate interval demonstrates extremely high resistivities and velocities, implying the presence of thick hydrate layers.

At Site 16, both pore- and fracture-filling gas hydrates were discovered. The resistivity and velocity of these two types of GHBS show high anomalies. The upper fracture-filling hydrate layer is located at 10–23 m, and the hydrate saturation estimated from pore-water chlorinity measurements is less than 20% (Feng et al., 2015). Core samples indicate that gas hydrates occur as nodules or veins in this interval. The lower pore-filling hydrate layer is located at 189.5–230 m, and the hydrate saturation is as high as 50%. The velocity increases first and then decreases (Fig. 2), while the resistivity remains high, suggesting the possible existence of free gas (Sha et al., 2015). At 193.5 m, the velocity decreases, whereas the resistivity begins to increase, which means that the upper boundary of the gas-bearing zone is likely at this depth. Note that the effective velocity data recorded in this hole is only up to 208 m.

3.3. Isotropic models for pore-filling GHBS

3.3.1. Isotropic velocity model

The EMT (see Appendix A) that has been widely applied to both field and laboratory data (e.g., Kleinberg et al., 2005; Lee and Collett, 2009) is used to predict the P-wave velocity of isotropic pore-filling GHBS. In EMT, three different microscopic patterns of gas hydrate are assumed: hydrate cements the sediment grains; hydrate floats in pores without any grain contacts; and hydrate supports solid frame (load-bearing). The cementing EMT predicts considerably higher velocities than those observed in nature, and at times, even unphysical results (Ecker et al., 1998; Chand et al., 2004). In the non-contacting EMT, the gas hydrate is treated as a pore fluid, and the shear modulus is thus assumed to be zero, which can be inappropriate for solid hydrates (Sain et al., 2010). That the S-wave velocity predicted by this model is unrelated to gas hydrate saturation contradicts observations (Collett, 1999). Moreover, the predicted P-wave velocity is found to be considerably lower than field observations (Chand et al., 2004). The third load-bearing EMT predicts reliable gas hydrate saturations (Helgerud

![Fig. 2. Well logs at Sites GMGS2-05, 08, and 16 including neutron porosity, density, P-wave velocity and ring resistivity. Blue lines represent measured well logs. Pink lines represent the baseline P-wave velocities (water-saturated sediment) calculated by effective medium theory (EMT) and baseline resistivities calculated by Archie's equation. Gas hydrate-bearing intervals indicated by previous studies are marked by green areas. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)](image-url)
et al., 1999; Dai et al., 2004; Petersen et al., 2007), and is more favored in gas hydrate modeling (Kleinberg et al., 2005; Lee and Waite, 2008; Sain et al., 2010). Therefore, we choose the load-bearing EMT to characterize the velocity of pore-filling GHBS.

The sediment porosity in the model is chosen to be 60% based on the porosity log (Fig. 2). We provide a validation of this method at different porosities in Section 6.1. Other parameters (e.g., critical porosity) are determined by fitting the measured velocity in hydrate-free zones using the EMT. As indicated in Fig. 2, the EMT-predicted velocities (pink lines) agree closely with the measured velocities (blue lines) in hydrate free zone apart from the interval below the GHBS at Site 05, which is likely caused by free gas. After obtaining the most suitable parameters for each site, we calculate averages to perform the forward modeling. Since core analysis results are not known, the sediment constituents are derived from NGHP-01 (Table 1), which shares the same sediment type with the study area: clay-dominated sediments (Collett et al., 2008). In discussion, we will show that the clay content has a minimal influence on our theoretical results. The elastic parameters of the sediment constituents are presented in Table 1.

3.3.2. Isotropic resistivity model

Archie’s equation is used to characterize the isotropic resistivity of pore-filling GHBS. First, the resistivity of brine-saturated sediments is written as:

\[ R_0 = aR_w \phi^{-m} \]  

(1)

where \( a \) and \( m \) are Archie’s constants, \( R_w \) is the connate water resistivity, and \( \phi \) is the porosity. \( R_w \) is calculated using the method of Arps (1953) with a background salinity of 3.5%, a seafloor temperature of 4 °C, and a temperature gradient of 45 °C/km (Wang et al., 2016). \( a \) and \( m \) are derived based on the cross plot between the porosity and formation factor, which is defined as \( FF = R_a/R_w = \phi^{-m} \) (Fig. 3). The data used to obtain \( a \) and \( m \) must be extracted from the hydrate-free, water-saturated interval adjacent to the hydrate-bearing interval (Cook and Waite, 2018). The resistivities calculated using the fitted \( a \) and \( m \) are shown in Fig. 2. The average Archie constants of the three sites are used to perform the modeling (\( a = 2.21 \) and \( m = 1.839 \)). Then, the measured resistivity \( (R_i) \) in terms of hydrate saturation \( (S_h) \) is expressed as

\[ R_i = \frac{aR_w}{\phi^{-m}(1 - S_h)^n} \]  

(2)

where \( n \) is a reservoir-dependent parameter varying based on the sediment properties and hydrate saturation itself (Spangenberg, 2001). Here we choose \( n = 2 \) for isotropic resistivity analysis (Lee and Collett, 2012; Cook et al., 2010; Wang et al., 2011).

3.4. Anisotropic models for fracture-filling GHBS

3.4.1. Anisotropic velocity model

The velocity of gas hydrate filled in fractures is modeled using an anisotropic model, which is divided into two parts. The first part is a fracture filled by 100% gas hydrate; thus, the hydrate concentration is controlled by the volumetric fraction of fracture. The second part is a 100% brine-saturated sediment, modeled by Lee and Collett (2009)

![Cross plots between porosity and formation factor at Sites (a) 05, (b) 08, and (c) 16. The best-fit Archie equations for brine-saturated sediments are shown as red lines. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)](image)

using a simplified three-phase equation (STPE). However, the parameter \( \varepsilon \) in STPE, which refers to the reduced effect of compaction caused by hydrate formation, is not known in this study area. Hence, we use EMT to characterize the second part of the fracture system. The anisotropic P-wave velocities \( (V_p) \) are expressed as

\[ V_p = \left( \frac{\sin^2 \phi + \cos^2 \phi + L + Q}{2\rho} \right)^{1/2} \]  

(3)

where \( \rho \) is density, and \( \phi \) is incidence angle of seismic wave. The parameters \( A, C, L, \) and \( Q \) are related to both two parts, and detailed derivations are presented in Appendix B.

3.4.2. Anisotropic resistivity model

The anisotropic resistivity caused by gas hydrate-filled fractures is modeled using a two-part resistivity model presented by Kennedy and Herrick (2004). Part I is a fracture filled by gas hydrate with a porosity of \( \phi_1 = 0.015 \) and a volume fraction of \( \eta \). Part II is a water-saturated sediment with a porosity of \( \phi_2 = 0.6 \) and a volume fraction of \( 1 - \eta \). Note that \( \phi_1 \) is a given parameter to avoid an unusual formation factor in the resistivity analysis, rather than an actual porosity in the fracture. The choice of \( \phi_1 \) is important for accurately characterizing the resistivity of GMGS2; the reason for this choice is presented in Section 4.2.

The parameters of the second component are identical to those applied in the isotropic resistivity analysis. The formation factors of the horizontal \( (F_h) \) and vertical fractures \( (F_v) \) are computed via the following equations (Kennedy and Herrick, 2004):

\[ F_h = \frac{1}{\eta \phi_1^m/\phi_1 + (1 - \eta) \phi_2^m/\phi_2} \]  

(4)

\[ F_v = \frac{(1 - \eta) \phi_1^m/\phi_1 + \eta \phi_2^m/\phi_2}{\phi_1^m \phi_2^m/\phi_1 \phi_2} \]  

(5)

Where \( \phi_1, m_1, a_2, \) and \( m_2 \) represent the Archie constants of the two components, respectively. Because the Archie constants appropriate for

Table 1

<table>
<thead>
<tr>
<th>Constituent</th>
<th>K (GPa)</th>
<th>G (GPa)</th>
<th>ρ (g/cm³)</th>
<th>Sources</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay</td>
<td>20.9</td>
<td>6.85</td>
<td>2.58</td>
<td>Helgerud et al. (1999)</td>
</tr>
<tr>
<td>Quartz</td>
<td>38.4</td>
<td>44.1</td>
<td>2.66</td>
<td>Pabst and Gregorová (2013)</td>
</tr>
<tr>
<td>Calcite</td>
<td>76.8</td>
<td>32</td>
<td>2.71</td>
<td>Helgerud et al. (1999)</td>
</tr>
<tr>
<td>Hydrate</td>
<td>8.4</td>
<td>3.54</td>
<td>0.924</td>
<td>Helgerud et al. (2009)</td>
</tr>
<tr>
<td>Water</td>
<td>2.3</td>
<td>0</td>
<td>1.02</td>
<td>Calculated by Cook and Waite (2018) using Batsle and Wang (1992)’s method</td>
</tr>
</tbody>
</table>
water-saturated sediments are also applicable to gas hydrate reservoirs (Kennedy and Herrick, 2004; Lee and Collett, 2009), we determine $a_1 = a_2 = 2.21$ and $m_1 = m_2 = 1.839$. By assuming that the resistivity was measured across the fracture and the fracture strike is in the $y$-direction in a Cartesian coordinate system with $z$-direction being vertical, the formation factor of the fracture with an arbitrary dip angle $\theta$ is calculated using equation (6) (Lee and Collett, 2012). Finally, the anisotropic resistivity can be calculated based on the formation factor $F_{xx}$.

\[ F_{xx} = F_0 \cos^2 \theta + F_1 \sin^2 \theta \]

(6)

4. Results of velocity and resistivity modeling of GHBS

4.1. Velocity modeling of GHBS

The computed P-wave velocities of two different GHBS as a function of gas hydrate concentration ($C_h$) at 60% porosity ($\varphi$) are displayed in Fig. 4. For pore-filling GHBS, $C_h = \varphi S_h$, where $S_h$ is the fraction of pore space occupied by hydrate (saturation). Thus, the largest hydrate concentration of pore-filling GHBS is 60% in this case. The $V_P$ of the two different gas hydrate accumulations indicate increasing trends versus $C_h$. Before gas hydrates begin to accumulate in pores or fractures, their $V_P$ are the same because both are 100% water-saturated. When $C_h = 100\%$, the $V_P$ is the $V_P$ of pure hydrate.

Because of anisotropy, the $V_P$ of gas hydrate-filled fractures varies with the incidence angle. For a vertical borehole, fractures rotating from horizontal to vertical can be modeled using the incidence angles from 0° to 90° (Lee and Collett, 2009). Note that the $V_P$ of fracture-filling GHBS does not increase monotonically with fracture dips, it decreases initially and then increases (e.g., Liu and Liu, 2018a). As shown in Fig. 4, the $V_P$ of pore-filling GHBS is basically higher than the $V_P$ of fracture-filling GHBS with different dips, and the difference increases as the hydrate concentration increases.

4.2. Resistivity modeling of GHBS

Fig. 5 shows the calculated resistivities as a function of hydrate concentration. The resistivity curves of the pore- and fracture-filling gas hydrate accumulations both increase with gas hydrate concentration. The resistivities of fracture-filling GHBS increase much more rapidly at first, except for the case of the horizontal fracture whose resistivity hardly increases at low hydrate concentrations. As can be observed, the resistivity of fracture-filling GHBS is higher than that of pore-filling GHBS for the majority of times, until the hydrate concentration exceeds a high value of about 48% (corresponding to saturation of 80% at 60% porosity in Fig. 5). Fig. 5 also reveals that the resistivity increases with the fracture dip, and reaches the highest value when the fractures are vertical, in agreement with the field observations (Cook et al., 2010).

The porosity of component I in anisotropic resistivity model is important, which is determined by comparing the theoretical resistivity with the measured resistivity. The theoretical resistivities of gas hydrate-filled fractures with various $\varphi_1$ are shown in Fig. 6, where $\varphi_1 = 0.035$ and 0.05 were suggested by previous studies (Lee and Collett, 2012, 2013). As indicated in the figure, $\varphi_1$ has almost no effect on the resistivity of horizontal fracture (green curves in Fig. 6), yet significant influence on the vertical fracture. The resistivities decrease a lot as $\varphi_1$ increases, because the space originally belonging to gas hydrate is occupied by water. The largest resistivities associated with $\varphi_1 = 0.05, 0.035,$ and 0.015 are approximately 100, 300, and 1000 $\Omega$m, respectively, which can be used to evaluate the rationality of the value of $\varphi_1$. As shown in Fig. 2, the highest measured resistivity occurs in the deeper hydrate-bearing layer at Site 08, which is about 1000 $\Omega$m. Based on this measurement, a judicious choice of $\varphi_1 = 0.015$ is determined.
4.3. Cross plots between P-wave velocity and resistivity

According to the theoretical modeling above, both P-wave velocity and resistivity properties of the two types of gas hydrate deposits increase with gas hydrate concentration, but there are clear differences. Compared to the fracture-filling GHBS, the P-wave velocity of pore-filling GHBS is higher and its resistivity is considerably lower. Based on this contrast, we hypothesize that the cross plot between P-wave velocity and resistivity can distinguish these two hydrate morphologies (Fig. 7). Different hydrate concentrations and fracture dips are involved. The cross plots with fracture dip angles higher than 45° are not presented because they nearly overlap the curve of vertical fracture.

As gas hydrate forms in sediments, the cross plots between the velocity and resistivity of fracture-filling GHBS exhibit extremely steep slopes at first, especially the high-angle fractures (Fig. 7). Then, the slopes flatten. However, the cross plot of pore-filling GHBS displays a nearly linear, moderate slope throughout the accumulation process. Clearly, these two different GHBS can be distinguished in most cases based on the cross plot between velocity and resistivity, by using theoretical cross plot as a template. With observations in Fig. 7, we can expect that hydrate saturation and fracture dip can also be estimated by a joint analysis of velocity and resistivity.

When fractures dip at very low angles (likely below 5°), and at the same time the hydrate contents of the two types of GHBS are both extremely low, identification becomes difficult. This is because their cross plots are too close to each other. However, according to many studies, gas hydrate in marine sediments tends to fill in high angle (or sub-vertical) fractures (Collett et al., 2008; Cook and Goldberg, 2008; Cook, 2010; Lee and Collett, 2012).

5. Application to LWD data from GMGS2

5.1. Identification of hydrate morphology in GMGS2

Velocity and resistivity log data within hydrate-bearing zones are extracted to generate cross plots for testing (Fig. 8). Theoretical cross plots are drawn as templates to assist the identification of the gas hydrate morphology. Because each hydrate layer has its own model parameters determined by fitting the data within the hydrate-free zone (Fig. 2), such as Archie constants, sediment porosity, and effective pressure, we fit the five hydrate layers individually. At Site 05, the cross plot of the pore-filling GHBS exhibits a small slope, resulting from modest velocity anomalies and small resistivity anomalies. At Site 08, the cross plot of the shallower hydrate layer shows a nearly linear, moderate slope throughout the accumulation process. Clearly, these two different GHBS can be distinguished in most cases based on the cross plot between velocity and resistivity, producing a steep slope (Fig. 8d).

For the lower pore-filling hydrate layer at Site 16, the logging data are divided into three parts based on their different features on cross plot: 189.5–193.5 m, 193.5–198 m, and 198–208.5 m. As shown in Fig. 9, the yellow and red dots that represent the field logging data below 193.5 m do not fit with the purple curve (associated with gas-free pore-filling GHBS). Thus, we hypothesize that the GHBS between 189.5 and 193.5 m may be gas-free (blue dots), whereas the GHBS below 193.5 m is gas-bearing. As we know, free gas is an electrical insulator and can significantly decrease the velocities of the host sediments (Collet and Ladd, 2000; Shankar and Riedel, 2011). Therefore, the velocity below 193.5 m (yellow and red dots) is low due to the occurrence of free gas, whereas the resistivity remains high throughout this interval (Fig. 2). To further validate this hypothesis, we re-compute a cross plot of pore-filling GHBS with a fixed gas saturation ($S_g$) of 3% (the green curve). This curve fits well with the field data from 198 to 208.5 m (yellow dots), indicating that the gas saturation within this interval is likely about 3%. However, we cannot fit well the field data from 193.5 to 198 m (red dots), which is
likely caused by the spatially varying gas saturation in this interval. In summary, this model generates reasonable crossplots compared with the field logging data from GMGS2, where hydrate morphology and occurrence of free gas are successfully identified.

5.2. Gas hydrate saturations of two pore-filling GHBS

Although saturations of gas hydrate-filled pore spaces could be estimated either by P-wave velocity or by resistivity alone, the estimated hydrate saturation could be non-unique when free gas and hydrate co-exist within the sediments. Based on previous observations, here we apply a joint analysis of P-wave velocity and resistivity to invert both hydrate and free gas saturations at Sites 05 and 16. The main uncertainty in the inversion comes from the Archie’s saturation exponent \( n \). Previous studies suggest \( n = 2 \) for pore-filling GHBS (e.g., Lee and Collett, 2012; Cook et al., 2010; Wang et al., 2011) and we applied \( n = 2 \pm 0.5 \) to quantify the uncertainty of our saturation estimations. The pore-water chlorinities of core samples provide a reference to test our estimations.

As shown in Fig. 10, calculated velocity and resistivity fit well with measured values. The saturations of gas hydrate and free gas estimated by this method are listed in Table 2. At Site 05, the estimated hydrate saturation is marginally higher than the pore-water chlorinity measurements (GH1). The average gas saturation is estimated as 0.4–0.9%. At Site 16, the estimated hydrate saturation is lower than the chlorinity-derived saturations (GH5). This is believed to be caused by the different depth ranges: the chlorinity-derived saturations correspond to 189.5–230 m, while our estimations correspond to 189.5–208.5 m because the effective velocity data is only up to 208.5 m (Fig. 2). The variation of the estimated gas saturation at Site 16 agrees well with the hypothesis in Section 5.1: it is almost negligible from 189.5 to 193.5 m, then slowly increases from 193.5 to 198 m. Finally, the gas saturation reaches 3% and become relatively stable (Fig. 10).

5.3. Gas hydrate saturations of the three fracture-filling GHBS

Without the information on fracture dips, the hydrate content in fractured reservoirs cannot be accurately estimated from well-logs. Considering that fracture dips cannot be easily obtained, most previous estimates for fracture-filling GHBS assume horizontal or vertical fractures (e.g., Lee and Collett, 2009; Sriram et al., 2013), which could lead to errors. By forcing the theoretical velocity and resistivity to match the measured values simultaneously, we can derive hydrate concentrations (Fig. 11). In the process of inverting hydrate concentrations, the main uncertainty comes from the porosity of Phase 2 in the fracture system.
For the fracture-filling hydrate-bearing zone, the neutron porosity log in Fig. 2 is not appropriate to represent \( \phi_n \) (Collett and Lee, 2011). Hence, the \( \phi_n \) of the three fracture-filling GHBS are chosen to be the averages of the porosity logs above and below the hydrate-bearing zones, which are \( 0.72 \pm 0.05 \), \( 0.6 \pm 0.05 \), and \( 0.68 \pm 0.05 \), respectively (Fig. 2). The inversion results are presented in Fig. 11, where estimated hydrate concentrations have been converted to saturations.

As shown in Table 2, the predicted hydrate saturations are mainly higher than those of chlorinity-derived values, especially in the lower GHBS at Site 08 (GH3). The pore-water chlorinity in this interval indicates a low saturation range of 0–33%, which is questionable because the extremely high velocity and resistivity and low density values in this interval all indicate the occurrence of thick hydrate layers. This is also confirmed by core samples (Sha et al., 2015). Thus, hydrate saturation as estimated by our proposed method (0–83%) is more reasonable. Several factors could account for the difference between hydrate saturations estimated from well logs and pore-water chlorinities. First, the number of core-based measurements is very limited, whereas hydrate concentration often exhibits strong heterogeneities both vertically and horizontally, making it difficult for core-based measurements to represent the full range of values. Second, core-based data likely underestimates hydrate contents due to its relatively low sampling frequency (Qian et al., 2017). Third, the difference could be partly due to the imperfect rock physical models and inaccurate sediment constituents used here.

According to the cross plots of Fig. 8b, c, and d, the logging data of the upper fracture-filling GHBS at Sites 08 and 16 are best modeled

### Table 2

<table>
<thead>
<tr>
<th>Depth range</th>
<th>( \phi_h ) range from pore-water</th>
<th>( \phi_h ) range from model</th>
<th>Average ( \phi_h ) from model</th>
<th>Average ( \phi_g ) from model</th>
<th>Average dips from model</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH1 197–207m</td>
<td>0–31%</td>
<td>0–40%</td>
<td>19–28%</td>
<td>0.4–0.9%</td>
<td>–</td>
</tr>
<tr>
<td>GH2 8–21m</td>
<td>0–14%</td>
<td>0–24%</td>
<td>5–8%</td>
<td>–</td>
<td>16°–27°</td>
</tr>
<tr>
<td>GH3 65–98m</td>
<td>0–33%</td>
<td>0–83%</td>
<td>36–41%</td>
<td>–</td>
<td>23°–29°</td>
</tr>
<tr>
<td>GH4 10–23m</td>
<td>0–20%</td>
<td>0–24%</td>
<td>8–13%</td>
<td>–</td>
<td>26°–42°</td>
</tr>
<tr>
<td>GH5 189.5–208.5m</td>
<td>0–50%</td>
<td>0–44%</td>
<td>24–35%</td>
<td>2–3.2%</td>
<td>–</td>
</tr>
</tbody>
</table>

Fig. 11. Gas hydrate saturations and fracture dips of the fracture-filling GHBS at Sites 08 and 16, estimated by matching P-wave velocity and resistivity simultaneously. Estimated hydrate saturations and fracture dips are displayed with three different \( \phi_h \) of each hydrate layer (67%, 72%, and 77% for the upper GHBS at Site 08; 55%, 60%, and 65% for the lower GHBS at Site 08; 63%, 68%, and 73% for the upper GHBS at Site 16). Calculated velocities and resistivities correspond to the middle \( \phi_h \) of each hydrate layer.
using fractures with dips larger than 15° and 20°, respectively. The inversion results shown in Fig. 11 provide more details, where gas hydrates in these two intervals are present dominantly in fractures of 20° and 30° dips, respectively. The lower hydrate layer at Site 08 shows highly variable fracture dips from 65 to 83 m depth, whereas the field data from 83 to 98 m depth is best modeled with fracture dip between 15° and 30°.

6. Discussion of limitations of the proposed method

The proposed method shows good promise in identifying hydrate morphology and predicting hydrate saturations. The estimated hydrate saturations are comparable to those from pore-water chlorinities. Our method is based on a two-part anisotropic model, which describes the velocity or resistivity of fracture-filling GHBS in a given direction of measurement. Previous studies have shown that such a model is able to interpret the similar field measurements even with incomplete information regarding the fracture characteristics (Lee and Collett, 2009, 2012; 2013; Wang et al., 2013). However, this method has limitations, because the anisotropic model is imperfect and there are discrepancies with the real structure of GHBS. Fracture systems in nature can be very complex, with shapes ranging from nodules to layers, and sizes ranging from several centimeters to tens of centimeters or even larger. The orientation of fractures may also be complex, varying from well oriented, aligned in one direction or chaotic, or striking in a range of directions without a dominant dip angle. There factors may account for the discrepancy between hydrate saturations estimated from our method and pore-water chlorinities. To accurately characterize the structure of small-scale fractures, therefore, the current anisotropic model needs to be improved considering these factors. However, in our study, hydrate morphologies are successfully identified, despite different fracture shapes and sizes.

The vertical resolution of resistivity measurement is significant for the fracture dip estimations in this study. Lee and Collett (2013) have demonstrated that estimated fracture dips can differ when using ring resistivity and P40H resistivity that are collected with different logging tools of different vertical resolutions. Therefore, the fracture dips we estimate are more like apparent dips that depend on the measurement scale. We were not able to test the method using different types of data as we only have ring resistivity measurements. The accuracy of the fracture dip estimations cannot be verified due to a lack of RAB data. Therefore, we suggest caution when using the dip information estimated by this method. Nonetheless, the estimation of hydrate saturation appears to be independent of measurement scale. According to Lee and Collett’s (2013) results, saturations estimated using different types of resistivity logging tools are almost identical. Sediment porosity and sediment type may also limit the method, which are discussed in the following sections.

6.1. Effects of sediment porosity

The porosity of pore-filling GHBS, which is also the porosity of component II in the fracture system (p_e), has remarkable effects on sediment elastic properties. To investigate this, we calculate the theoretical cross plot at porosities of 40% and 80% to demonstrate how the method would behave at extreme low and high porosities (Fig. 12). Porosities of 40%–80% basically cover the range observed in marine sediments. Compared to Fig. 7, we find that porosity mainly influences the absolute values of velocity and resistivity, both of which decreases when the porosity increases. In both scenarios, the relationship between velocity and resistivity appears to be unchanged, with the cross plots continuing to indicate completely different slopes for pore- and fracture-filling gas hydrate reservoirs. Moreover, Fig. 12 also suggests that the identification of hydrate morphology is easier at high porosity because the cross plot of pore-filling GHBS is more clearly separated from that of a low-angle fracture (e.g., dip = 10°). In summary, sediment porosity in marine environment has minimal effect on the proposed approach.

6.2. Effects of host sediments

It is believed that host sediments play a decisive role in gas hydrate formation and distribution. Typically, pore-filling gas hydrate tends to

![Fig. 12. Cross plots between theoretical velocity and resistivity of two different gas hydrate accumulations with porosities of (a) 0.4 and (b) 0.8. Different fracture dips are marked in the figure. Gas hydrate concentrations of 0% and 30% are marked by blue circles and asterisks, respectively.](image1)

![Fig. 13. Cross plots between theoretical velocity and resistivity of the two types of GHBS in sandy sediments. Different fracture dips and gas hydrate concentrations are marked.](image2)
reside in coarse-grained sediments (sands), whereas fracture-filling gas hydrate is usually observed in fine-grained sediments (clay) (Clennell et al., 1999; Jain and Juanes, 2009; Cook et al., 2010). GMGS1 and 2 are the only drilling expeditions where pore-filling gas hydrate was recovered in fine-grained sediments (Zhang et al., 2007; Sha et al., 2015). Therefore, the modeling above is based on the fine-grained clays. To understand the applicability of the proposed method in coarse-grained sands, we recalculated the cross plot between velocity and resistivity of the two kinds of GHBS in sandy sediments (Fig. 13). Compared to the results for fine-grained sediments (Fig. 7), the velocities of the two types of GHBS increase, especially the pore-filling GHBS. Consequently, the slope of the cross plot associated with pore-filling GHBS is marginally flatter compared to our previous plots. However, the difference between two morphologies remains clear.

7. Conclusions

Gas hydrate morphologies have a major influence on the physical properties of gas hydrate-bearing sediments. Based on rock physics modeling, we found that gas hydrate morphology can be easily identified by the cross plot between P-wave velocity and resistivity; gas hydrate saturation can also be estimated by a joint analysis of these two properties. LWD data from GMGS2 expedition in the South China Sea are used for to validate the proposed method. Overall, the cross plots of the LWD data fit well with theoretical cross plots, and the results of the morphology identification agree with observations of core samples. The gas hydrate saturations estimated from the joint analysis method are comparable to those estimated from pore-water chlorinities, except for the lower hydrate layer at Site 08, which could be due to the limited amount of chlorinity data, different measuring frequencies, and the imperfect models used here. In addition, the variation of the predicted gas saturation is consistent with the characteristics of well logs. Fracture dips are also estimated, but we suggest readers exercise caution because fracture dips may depend on the resolution of different logging tools. Experiments to investigate the influence of the host sediments on the proposed method suggest possible applications to both fine- and coarse-grained sediments. The method is also applicable to a wide range of sediment porosities according to our test.

Appendix A

The bulk and shear moduli of dry marine sediments are given by Dvorkin et al. (1999):

\[
K_{\text{dry}} = \left[ \frac{(1 - \varphi)K_{\text{IM}}}{K_{\text{IM}} + \frac{4}{9}G_{\text{IM}}} + \frac{(\varphi - \varphi_c)/(1 - \varphi_c)}{\frac{4}{9}G_{\text{IM}}} \right]^{-1} - \frac{4}{3}G_{\text{IM}},
\]

\[
G_{\text{dry}} = \left[ \frac{(1 - \varphi)K_{\text{IM}}}{G_{\text{IM}} + Z} + \frac{(\varphi - \varphi_c)/(1 - \varphi_c)}{Z} \right]^{-1} - Z,
\]

\[
Z = \frac{6G_{\text{IM}}}{9K_{\text{IM}} + 8G_{\text{IM}}} \left[ \frac{1}{K_{\text{IM}} + 2G_{\text{IM}}} \right]^{-1} K_{\text{IM}} = \left[ \frac{(1 - \varphi)G_{\text{dry}}^2}{18(1 - \varphi)v^2\rho^2} \right]^{\frac{1}{2}}, \text{ and } G_{\text{IM}} = \frac{5 - 4\psi}{2(2 - v)} \left[ \frac{3(1 - \varphi_c)^2n^2G_{\text{dry}}^2}{2(1 - \psi)v^2n^2} \right]^{\frac{1}{2}}.
\]

where \( \varphi \) is sediment porosity, \( \varphi_c = 0.4 \) is critical porosity, and \( \varphi \geq \varphi_c \). The case of \( \varphi < \varphi_c \) is not shown since it is not applicable for the marine sediments in our study. \( n \) denotes the number of contacts per grain (Murphy, 1982); \( \psi = (K - 2G/3)/(K + G)/2 \) denotes the Poisson’s ratio; \( P = (1 - \psi)(\varphi_d - \varphi_c) \) denotes the effective pressure, where \( \varphi_d \) and \( \varphi_c \) denote the densities of frame and pore water, respectively; \( g \) is gravity acceleration, and \( h \) is the depth below seafloor. The modulus of the mineral constituents are calculated by: \( X = \left[ \sum_{i=1}^{n} f_i X_i + \left( \sum_{i=1}^{n} f_i \right) / X \right] / 2 \), where \( X \) represents \( K \) or \( G \), \( m \) represents the number of mineral constituents, and \( f_i \) denotes the volume fraction of the \( i \)th constituent.

The moduli of brine-saturated sediments are computed by the Gassmann (1951):

\[
K_{\text{sat}} = K_{\text{dry}}/\left(1 + \varphi K_{\text{dry}}/K_f \right) K_f/\left(1 - \varphi \right) K_f + K_f/\left(1 - \varphi \right) K_f K_{\text{dry}}/K_f \text{ and } G_{\text{sat}} = G_{\text{dry}}
\]

A-2

where \( K_f \) denotes bulk modulus of brine. Since gas hydrate acts as a part of the frame, the porosity reduces and the moduli of the frame should be recalculated. The reduced porosity is given as \( \varphi' = \varphi(1 - S_h) \), where \( S_h \) is hydrate saturation, and the modified moduli are given by

\[
K = (f_h K_h + (1 - f_h)/K_f) K_f/K_{\text{dry}}, \text{ and } G = (f_h G_h + (1 - f_h)/G_f) G_f/K_{\text{dry}}
\]

A-3

\[
f_h = \frac{\varphi(1 - S_h)}{1 - \varphi S_h}
\]

where \( K_h \) and \( G_h \) are the primary moduli, and \( K_{\text{dry}} \) and \( G_{\text{dry}} \) are those of gas hydrate. The elastic velocities can be expressed as:

\[
V_p = \sqrt{(K_{\text{sat}} + 4G_{\text{sat}}/3)/\rho_b}, \text{ and } V_s = \sqrt{G_{\text{sat}}/\rho_b}
\]

A-4

where \( \rho_b \) is the bulk density of the whole sediments.

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Declarations of interest

None.
Appendix B

Transverse Isotropic Theory for Fracture-filling GHBS.

The parameters in equation (3) are given by White (1965):

\[ A = \frac{\mu_{1}(\lambda + \mu) + \Lambda}{(\lambda + 2\mu)(\lambda + 2\mu)} = \frac{1}{2}\left(\frac{1}{\lambda + 2\mu}\right)^{-1} = \frac{1}{2}\left(\frac{1}{\lambda + 2\mu}\right)^{-1} = \frac{1}{2}\left(\frac{1}{\lambda + 2\mu}\right)^{-1} = \frac{1}{2}\left(\frac{1}{\lambda + 2\mu}\right)^{-1} \]

with

\[ G = \frac{G_1 + G_2}{2}, \quad \text{and} \quad G' = \frac{G_1 + G_2}{2} \]

where \( G_1 \) and \( G_2 \) are any elastic constant of part I and part II, with the volume fractions of \( G_1 \) and \( G_2 \), respectively. \( \lambda \) and \( \mu \) are Lame constants derived from the properties of pure gas hydrate (part I) or EMT model (part II).

Note that the velocity in equation (3) is phase velocity. To perform the anisotropic analysis, group velocity should be used (Lee and Collett, 2013). Thomsen (1986) expressed the relationship between group and phase velocities as \( V_g(\phi) = V_p(\phi) \), where \( V_p \) denotes the group velocity. \( \phi \) denotes the angle between ray direction and fracture axis symmetry, which is given by

\[ \tan \phi = \tan \phi [1 + 2\delta + 4(\epsilon - \delta)\sin^2\phi], \quad \delta = \frac{(F + L)^2 - (C - L)^2}{2C(C - L)}, \quad \text{and} \quad \epsilon = \frac{A - C}{2C} \]

See Lee and Collett (2009) for more details.

Appendix C. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.marpetgeo.2019.104036.

References


