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GAS HYDRATE AND FREE GAS DETECTION USING SEISMIC QUALITY FACTOR  
ESTIMATES FROM HIGH-RESOLUTION P-CABLE 3D SEISMIC DATA

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## ABSTRACT

Seismic attenuation in gas hydrate and free gas bearing sediments is estimated from high-resolution P-Cable 3D seismic data from Vestnesa Ridge on the Arctic continental margin of Svalbard. P-Cable data have broad bandwidth (20-300 Hz) which is extremely advantageous in estimating seismic attenuation in a medium. The seismic quality factor (Q), inverse of seismic attenuation, is estimated from the seismic dataset using centroid frequency shift method and spectral ratio method. Centroid frequency shift method establishes a relationship between the change in the centroid frequency of an amplitude spectrum and the Q value of a medium. Spectral ratio method estimates the Q value of a medium by studying the differential decay of different frequencies. Broad bandwidth and short offset characteristics of the P-Cable dataset are useful to continuously map Q for different layers throughout the 3D seismic volume. The centroid frequency shift method is found to be relatively more stable than spectral ratio method. Q values estimated using these two methods are in concordance with each other. The Q data documents attenuation anomalies in the layers in the gas hydrate stability zone above the BSR and in the free gas zone below. Changes in the attenuation anomalies correlate with small-scale fault systems in the Vestnesa Ridge suggesting a strong structural control on the distribution of free gas and gas hydrates in the region. We argue that high and spatially limited Q anomalies in the layer above the BSR indicate the presence of gas hydrates in marine sediments in this setting. Hence, the presented workflow to analyze Q using high-resolution P-cable 3D seismic data with a large bandwidth can be a potential technique to detect and directly map the distribution of gas hydrates in marine sediments.

43

## INTRODUCTION

44 Gas hydrates are crystalline ice like structures normally formed at certain temperature  
45 and pressure conditions (Brooks et al., 1986). The temperature and pressure conditions  
46 required for gas hydrates formation are available in continental slope and permafrost  
47 environments (Sloan, 1998). The presence of marine gas hydrates in continental margins has  
48 been confirmed from different drilling activities (Collett and Ladd, 2000; Collett et al., 1999;  
49 Riedel et al., 2010; Ryu et al., 2013; Liu et al., 2012; Zhang et al., 2007). Seismic methods  
50 are commonly used to remotely identify gas hydrates in the marine sediments. The presence  
51 of gas hydrates in the sediments is often indicated in seismic data by a bottom simulating  
52 reflection (BSR) (Shipley et al., 1979). It marks a sharp impedance contrast between hydrate-  
53 bearing and gas-charged sediments. The BSR occurs at the base of the hydrate stability zone,  
54 which is governed mostly by pressure and temperature conditions (Sloan, 1998). Due to this  
55 control, the BSR often mimics the seafloor; therefore, cross-cuts the sedimentary strata  
56 (Shipley et al., 1979). Since the presence of gas hydrates increases the velocities,  
57 concentration of gas hydrates in sediments is usually estimated using seismic velocity models  
58 (Lee and Collett, 2001; Gei and Carcione, 2003; Ecker et al., 1998; Chand et al., 2004). The  
59 presence of gas hydrates in sediments has a pronounced effect on amplitude and frequency  
60 characteristics of a seismic signal also (Guerin and Goldberg, 2002; Pratt et al., 2003; Chand  
61 and Minshull, 2004). Hydrates in sediments show contradicting amplitude characteristics in  
62 seismic sections like amplitude blanking (Korenaga et al., 1997) and amplitude enhancements  
63 (Nouzé et al., 2004; Yoo et al., 2013, Riedel et al., 2010) at different geological settings.

64 Seismic signal attenuates mainly due to extrinsic attenuation (due to factors like  
65 spherical divergence, obliquity factor, scattering etc.) and intrinsic attenuation (due to  
66 conversion of vibration energy into heat energy) (Mavko et al., 1998). At seismic

67 frequencies, analysis on attenuation normally refers to intrinsic attenuation (Mavko et al.,  
68 1998) which can be studied through spectral analysis (Jacobson et al., 1981). Since gas  
69 hydrate increases the stiffness of the matrix (Jung et al., 2012) and P-wave velocity, it was  
70 normally assumed that the sediments saturated with gas hydrates will show lower attenuation  
71 (Wood et al., 2000). Unlike P-wave velocity, no unique trend of seismic attenuation in gas  
72 hydrates can be observed from the literature; thus making attenuation characteristic of the gas  
73 hydrate bearing sediments a debatable topic (Guerin et al., 1999; Wood et al., 2000; Chand et  
74 al., 2004; Rossi et al., 2007; Sain et al., 2009; Sain and Singh, 2011; Jaiswal et al., 2012;  
75 Dewangan et al., 2014). Laboratory experiments in hydrate bearing sediments indicated  
76 increase of attenuation with hydrate saturation (Priest et al., 2006; Best et al., 2013) whereas  
77 attenuation estimates from field experiments on gas hydrates indicated contradicting results.  
78 For example, studies on well log data (Guerin and Goldberg, 2002; Guerin and Goldberg,  
79 2005; Matsushima, 2005), VSP data (Pratt et al., 2005; Bellefleur et al., 2007) and on cross-  
80 hole seismic data (Pratt et al., 2003; Bauer et al., 2005) indicated an increase in attenuation.  
81 Other studies, mainly on surface seismic data (Dewangan et al., 2014; Rossi et al., 2007;  
82 Matsushima, 2006) indicated a decrease in attenuation. The increase (Guerin and Goldberg,  
83 2002; Gei and Carcione, 2003; Chand and Minshull, 2004; Lee and Collet, 2006) and  
84 decrease (Dewangan et al., 2014; Sain and Singh, 2011) in attenuation has been explained by  
85 using different rock physics models depending on the assumed micro structure of the hydrate  
86 and also sediment-hydrate mixtures. Chand and Minshull (2004) suggested that the amount of  
87 attenuation not only changes with hydrate saturation but also with the frequency of seismic  
88 signal.

89           The seismic quality factor ( $Q$ ), inverse of seismic attenuation, can be estimated from  
90 the seismic dataset using different methods which includes the amplitude decay method  
91 (Badri and Mooney, 1987), the rise time method (Gladwin and Stacey, 1974), the centroid

92 frequency shift method (Quan and Harris, 1997), wavelet modeling (Jannsen et al., 1985), the  
93 pulse broadening method (Hatherly, 1986), the spectral ratio method (Jannsen et al., 1985;  
94 Båth, 1982) and the inversion method (Amundsen and Mittet, 1994). Tonn (1991) compared  
95 10 methods of attenuation estimation using VSP seismograms and concluded that no single  
96 method is suitable for all situations.

97         In the present study, we apply two different methods to investigate seismic  
98 attenuation in gas hydrate and free gas saturated sediments from Vestnesa Ridge, a deep-  
99 water gas hydrate system located offshore west-Svalbard (Figure 1). The quality factor (Q)  
100 has been estimated from P-Cable seismic data using the spectral ratio method (Jannsen et al.,  
101 1985) and the centroid frequency shift method (Quan and Harris, 1997). The centroid  
102 frequency shift method establishes a relationship between the change in the centroid  
103 frequency of an amplitude spectrum and the Q value of a medium (Quan and Harris, 1997).  
104 On the contrary, the spectral ratio method estimates the Q value of a medium by studying the  
105 differential decay of different frequencies (Båth, 1982). Due to limitation of seismic  
106 bandwidth in conventional seismic data, it is almost impossible to map Q with high accuracy.  
107 Low signal to noise ratio, short bandwidth, source/receiver array directivity and distinct  
108 raypaths in a CDP gather are the main problems encountered in Q analysis from conventional  
109 surface seismic data (Hustedt and Clark, 1999). But P-Cable surface seismic data is  
110 essentially zero-offset (offset varying from 97-143 m) in deep water and has broad bandwidth  
111 (20-300 Hz). Raypaths of different traces in a CDP gather of P-Cable data are approximately  
112 similar at deep water depth as offset is quite small. Stacked P-Cable data has high signal to  
113 noise ratio and the stacking process involves traces with almost similar raypaths. These  
114 characteristics of P-cable data match well with the characteristics of VSP data (Galperin,  
115 1985) and make P-Cable data suitable for subsurface Q analysis. Moreover, using P-Cable  
116 3D seismic data for estimating Q allows us to analyze the spatial distribution of Q which can

117 be integrated with 3D seismic interpretation. Thereby, we can link Q estimates with  
118 anomalies related to the presence of gas hydrate and free gas in the sediments.

## 119 STUDY AREA

120 Our study focuses on the active seeping segment of Vestnesa Ridge, a ~100 km long  
121 gas hydrate charged contourite drift developed over < 20 Ma oceanic crust offshore west-  
122 Svalbard (Figure 1) (Eiken and Hinz, 1993; Vogt et al., 1994, Bünz et al., 2012). The  
123 contourite drift is in close proximity to the Molloy and the Knipovich slow-spreading oceanic  
124 ridges and it is located between the Molloy and the Spitsbergen Transform Faults (e.g.,  
125 Ritzmann et al., 2004). Vestnesa Ridge consists of three main stratigraphic sequences named  
126 according to correlation with ODP sites at the Yermak Plateau (YP) (Eiken and Hinz, 1993):  
127 the oldest unit (YP1) is of Miocene age and consists dominantly of syn-rift deposits directly  
128 lying over the oceanic crust (Eiken and Hinz, 1993; Ritzmann et al., 2004); the middle  
129 sequence (YP2) consists of sediments deposited by migrating contour currents; and finally  
130 the youngest sequence (YP3), is dominated by margin parallel contour currents and by  
131 glacial debris flow deposits (Howe et al., 2008).

132 A gas hydrate system and associated free gas zone exists along Vestnesa Ridge  
133 (Hustoft et al., 2009; Petersen et al., 2010; Bünz et al., 2012; Plaza-Faverola et al., 2015).  
134 The system is restricted to the upper stratigraphic sequence (YP3) and has a series of gas  
135 chimneys and pockmarks associated along the full extent of Vestnesa Ridge. However, only  
136 pockmarks located towards the easternmost part of the ridge (where our 3D seismic survey is  
137 located; Figure 1) are actively seeping gas at present (Bünz et al., 2012; Smith et al., 2014).  
138 Gas chimneys towards the westernmost part of the ridge seem inactive at present but  
139 foraminiferal records indicated past activity at around 8000 my ago (Consolaro et al., 2014).

140

## DATA

141 We used high-resolution P-Cable (Planke et al., 2009; Petersen et al., 2010) 3D  
142 seismic data acquired in 2013 on board R/V Helmer Hanssen (Plaza-Faverola et al., 2015).  
143 The system consists of 14 streamers towed parallel behind the ship. The 25-m-long streamers  
144 contain 8 receiver groups each. The streamers are attached to a cross cable towed  
145 perpendicular to the vessel's streaming direction and spread by two large trawl doors. The  
146 spacing of streamers along the cross cable is 12.5 m. However, due to curvature of the cross  
147 cable, the distance between streamers is varying between 6-10 m. The high-resolution P-  
148 Cable system was used together with mini-GI gun (15/15 in<sup>3</sup>). The gun was fired at an  
149 interval of 6 s with a firing pressure of 170 bar. Source-receiver offset varies from 97-143 m.  
150 Traces have been recorded with 3 s record length at 0.25 ms sampling interval.

151 During seismic data processing, utmost care has been taken to avoid all the steps that  
152 can potentially distort the amplitude spectrum within the main seismic bandwidth. The  
153 processing of the high-resolution 3D seismic data mainly included navigational correction,  
154 static and tidal correction, binning, bandpass filtering (10-20-300-350), NMO correction and  
155 stacking. NMO correction can potentially distort the amplitude spectrum due to NMO  
156 stretching. But for short offset seismic data in deep water, this distortion will be too small and  
157 can be neglected. 3D Stolt migration was applied using a constant velocity of 1500 m/s. The  
158 spatial resolution of the seismic data is quite high with a bin size of 6.25 x 6.25 m. The  
159 seismic data covers an area of about 14 km<sup>2</sup>. The data has a broad frequency spectrum  
160 ranging from approximately 20 to 300 Hz (Figure 2a).

161

162

## METHODOLOGY

163 Amplitude, frequency and phase are three basic attributes of a seismic signal.  
164 Quantitative analysis of these attributes is done using different derivations and transforms.  
165 For example, quantitative estimation of frequency attribute involves different time-frequency  
166 transforms. Reine et al. (2009) discussed the robustness of seismic attenuation measurements  
167 using different time-frequency transforms. In the present study, short-time Fourier transform  
168 (Gabor, 1946) is used to transform a seismic signal into frequency domain. The data  
169 converted to frequency domain is used to study seismic attenuation. We use centroid  
170 frequency shift (Quan and Harris, 1997) and spectral ratio method (Jannsen et al., 1985) to  
171 estimate Q in gas hydrate and free gas saturated sediments.

172 **Centroid frequency method**

173 Centroid frequency of an amplitude spectrum ( $f_c$ ) is defined as:

$$174 \quad f_c = \frac{\sum A(f) \times f}{\sum A(f)} \quad (1)$$

175 Where  $A(f)$  corresponds to amplitude of frequency ( $f$ ) in an amplitude spectrum.

176 Centroid frequency of a signal gives an idea about the energy level of a signal. Since  
177 energy of a signal decreases as it propagates in the sub-surface, centroid frequency of an  
178 amplitude spectrum shifts towards lower values with further propagation into deeper layers.  
179 By plotting the centroid frequency for the entire seismic section, a broad overview about the  
180 subsurface seismic attenuation can be established. Quan and Harris (1997) proposed a  
181 method to estimate Q using centroid frequency shift analysis. They considered the amplitude  
182 spectrum of the received signal  $R(f)$  as a function of incident wave  $S(f)$  and  
183 instrument/medium response  $G(f)H(f)$ .



184 
$$R(f) = G(f)H(f)S(f) \quad (2)$$

185 Parameter  $G(f)$  includes geometrical spreading, instrument response, source/receiver  
 186 coupling, radiation/transmission coefficients, and phase accumulation effects caused by  
 187 propagation.  $H(f)$  is a factor which takes into account the effect of intrinsic attenuation on a  
 188 seismic signal. Since attenuation is proportional to frequency within the seismic bandwidth,  
 189 response  $H(f)$  can be written as (Johnston et al., 1979):

190 
$$H(f) = \exp\left(-f \int_{\text{ray}} \alpha_0 dl\right) \quad (3)$$

191 Where the integral is taken along the ray path, and  $\alpha_0$  is the attenuation coefficient  
 192 defined by (Johnston et al., 1979):

193 
$$\alpha_0 = \frac{\pi}{Qv} \quad (4)$$

194 Where  $Q$  is the quality factor and  $v$  is the velocity of the medium.

195 With the assumption that the amplitude spectrum follows Gaussian pattern of  
 196 distribution, Quan and Harris (1997) after rearranging the equations finally came to the  
 197 following equation:

198 
$$\int_{\text{ray}} \alpha_0 dl = \frac{f_s - f_r}{\sigma_s^2} \quad (5)$$

199 Where  $f_s$  is the centroid frequency of the source signal (Figure 2b),  $f_r$  is the centroid  
 200 frequency of the received signal, and  $\sigma_s^2$  is the variance of the source amplitude spectrum.

201 
$$\sigma_s^2 = \frac{\int_0^\infty (f - f_s)^2 A(f) df}{\int_0^\infty A(f) df} \quad (6)$$

202 Where  $A(f)$  is the amplitude spectrum of the source signal and other parameters are  
 203 same as described in the above equation. In order to account for the increase in the variance  
 204 of amplitude spectrum,  $\sigma_s^2$ , of seismic signal with arrival time, a trend line for  $\sigma_s^2$  at different  
 205 arrival times is estimated (Figure 2c). Straight line is fitted to the mean  $\sigma_s^2$  values.

206 If velocity and quality factor (Q) is assumed constant in a medium, the final  
 207 expression for quality factor (Q) can be written (Talukder, 2013) as:

$$208 \quad Q = \frac{\pi\sigma_s^2\Delta t}{f_s - f_r} \quad (7)$$

209 Where  $\Delta t$  is the total travel time and rest of the parameters are same as described in  
 210 above equations.

## 211 Spectral ratio method

212 Spectral ratio method is one of the most commonly used methods to estimate Q in a  
 213 medium. This method takes into account the differential decay of different frequencies.  
 214 Higher frequencies tend to decay at a much higher rate as compared to lower frequencies  
 215 while passing through an attenuating medium (Båth, 1982). Differential decay of different  
 216 frequencies depends upon Q of a medium.

217 Jannsen et al. (1985) discussed about the application of spectral ratio method to  
 218 estimate Q from seismic data. Amplitude spectrums ( $A_1(\omega)$  and  $A_2(\omega)$ ) of two reflections  
 219 from different depths ( $Z_1$  and  $Z_2$ ), can be written as:

$$220 \quad A_1(\omega) = A_0(\omega)G(Z_1)R_1e^{-2\alpha_1Z_1} \quad (8)$$

$$221 \quad A_2(\omega) = A_0(\omega)G(Z_2)(1 - R_1^2)R_2e^{-2\alpha_1Z_1}e^{-2\alpha_2(Z_2 - Z_1)} \quad (9)$$

222 Where  $A_0(\omega)$  is the amplitude spectrum of the incident wavelet at  $Z=0$ ,  $G(Z_1)$  and  
 223  $G(Z_2)$  accounts for the geometrical spreading and other factors leading to decay in  
 224 amplitudes,  $R_1$  and  $R_2$  are reflection coefficients for different boundaries, and  $\alpha_1$  and  $\alpha_2$  are  
 225 the attenuation coefficients. The spectral ratio (SR) of two spectra can be written as:

$$226 \quad SR(\omega) = C_1 e^{-2\alpha_2(Z_2-Z_1)} \quad (10)$$

$$227 \quad C_1 = \frac{G(Z_2)(1-R_1^2)R_2}{G(Z_1)R_1} \quad (11)$$

228 Where  $C_1$  is the ratio of factors related to geometrical spreading and reflection  
 229 coefficients. Assuming phase velocity  $\beta$  to be independent of frequency in spectral ratio  
 230 (Båth, 1982), natural log of spectral ratio can be written as:

$$231 \quad \ln(SR(\omega)) = \ln(C_1) - \alpha_2 \Delta T \beta \quad (12)$$

232 where  $\Delta T$  is the time difference between two reflections. Substituting the value of  $\alpha$   
 233 as  $\pi f / (Q\beta)$  (Johnston et al., 1979), we get linear relation between  $\ln(SR(\omega))$  and frequency,  
 234 i.e.,

$$235 \quad \ln(SR(\omega)) = \ln(C_1) - \left(\frac{\pi \Delta T}{Q}\right) f \quad (13)$$

236 Hence, the slope i.e.  $-(\pi \Delta T / Q)$  of the spectral ratio (in logarithmic scale) vs frequency  
 237 plot depends on the  $Q$  of a medium, and the intercept is related to the geometrical spreading  
 238 and reflection coefficients which are independent of frequency. Using this concept,  $Q$  value  
 239 can be estimated from the slope of the best fit line in spectral ratio (in logarithmic scale) vs  
 240 frequency plot. In real data, two wavelets can be picked by windowing two reflections  
 241 (Figure 3a) and then Fourier transform can be applied to get amplitude spectrum of these two

242 wavelets (Figure 3b). Spectral ratio method can be applied on these two amplitude spectrums  
243 and effective Q of a medium between these reflections can be estimated (Figure 3c).

#### 244 ANALYSIS USING CENTROID FREQUENCY METHOD

245 Analysis using centroid frequency plots has been done to study changes in the  
246 centroid frequency with depth. An inline has been selected from seismic data where a BSR is  
247 clearly identified by high-seismic amplitudes at about 1.9 s TWT in the seismic section  
248 (Figure 4a) (Bünz et al., 2012; Smith et al., 2014). The BSR separates hydrate-bearing  
249 sediments from a ~100 m thick free gas zone (Hustoft et al., 2009). Other notable features are  
250 vertical zones of acoustic transparency or chaotic seismic facies. These are interpreted as  
251 vertical fluid-flow features, so-called chimneys. They terminate in seafloor depressions  
252 known as pockmarks (Figure 4c) (Bünz et al., 2012).

253 The centroid frequency has been calculated for all the traces in the seismic section at  
254 an interval of 5 ms. Enough samples have been taken to ensure that the lowest frequency in  
255 the main seismic bandwidth will have at least one wavelength to sample. The derived  
256 centroid frequencies are shown in Figure 4b. The centroid frequency decreases significantly  
257 beneath the BSR. Prominent low centroid frequency anomalies are observed in the free gas  
258 zone especially in the southeastern part of the seismic section (Figure 4b). Gas chimneys are  
259 observed in the seismic section. Some gas chimneys also show low centroid frequency  
260 anomalies. Some of these anomalies appear to intrude the chimneys from the free gas zone  
261 beneath the BSR. Coincidentally, intrusion happens for gas chimneys that have active gas  
262 seepage at the seafloor (Figure 4c) (Bünz et al., 2012).

263

264 **Q estimation**

265 Variance of an amplitude spectrum ( $\sigma_s^2$ ) and reference centroid frequency ( $f_s$ ) were  
266 calculated to further estimate Q for different layers using centroid frequency shift method. In  
267 order to account for an increase in the  $\sigma_s^2$  with arrival time,  $\sigma_s^2$  of a seismic signal is plotted  
268 with respect to two-way arrival time. The trend line for  $\sigma_s^2$  at different arrival times is  
269 estimated (Figure 2c). The  $\sigma_s^2$  to be used in equation (7) is calculated from the linear fit  
270 parameters of the best fit line. The centroid frequency of a seismic signal at the sea floor is  
271 used as a reference centroid frequency for estimating 1-D Q model for every trace. The  
272 centroid frequency of the seismic signal in shallow sediments close to the seafloor (shown in  
273 Figure 2b) ranges from 150-185 Hz for almost all the traces except for few located in a  
274 pockmark where it showed lower centroid frequency values. The centroid frequency of the  
275 seismic signal at the seafloor is expected to be higher than the centroid frequency of a seismic  
276 signal in shallow sediments. After neglecting aberrations, we use 182 Hz as a reference  
277 centroid frequency (Figure 2b).

278 Prominent reflections observed in the seismic data have been picked and Q values for  
279 different layers between picked reflections have been estimated (Figure 5a and 6a). Q model  
280 derived from one of the traces in the seismic data is shown in Figure 5a-d. Centroid  
281 frequencies are calculated at a time interval of 5 ms for each trace using a sliding time  
282 window along the trace (Figure 5a-b). Fluctuations in centroid frequencies make Q estimation  
283 difficult. Therefore, in order to reduce these effects, centroid frequencies of traces falling  
284 within 31.25 x 31.25 m<sup>2</sup> bin have been stacked to get a centroid frequency trend/curve  
285 (Figure 5c). This process also reduces the size of the seismic dataset and makes it  
286 computationally convenient. Synthetic centroid frequency curves are generated for different  
287 possible Q models (Q varying from 1 to 600 in all the 5 layers). Synthetic centroid frequency

288 curves for 5 Q models out of  $600^5$  possible Q models are shown in green color in Figure 5d.  
289 These synthetic centroid frequency curves are matched with estimated centroid frequency  
290 curves (estimated from traces in  $31.25 \times 31.25 \text{ m}^2$  bin), and the misfit between the two curves  
291 has been computed. The  $L_1$  norm approach (Claerbout and Muir 1973) has been used to  
292 calculate the misfit and to pick the best possible Q model out of the possible range of Q  
293 models. The  $L_1$  norm approach has been adopted to reduce the significance of the spiky  
294 points in the centroid frequency curves. This process of estimating Q has been repeated on all  
295 the traces in the seismic data to generate a Q cube.

## 296 **Observations**

297 The Q values along one of the inlines (Figure 6a) indicate high Q anomalies in the  
298 layer above the BSR and low Q anomalies below the BSR (Figure 6b). Q slices for different  
299 layers give an idea about the lateral variation of Q within a layer. Figure 7 shows Q slices for  
300 different layers illustrating the lateral variation of Q anomalies within a layer. The BSR lies  
301 between Q slices in Figure 7c and 7d. We observe particularly low Q values in some areas  
302 within the first layer (Figure 7a) coinciding with the location of the chimney structures. Q  
303 estimates in the second layer follows a normal trend except for few small patches of high Q  
304 (Figure 7b). Some of these slightly elevated Q values seem to correspond particularly with  
305 the outer rims of the chimney structures (Figure 7b). Just above the BSR, we observe very  
306 high Q values, particularly in the southern half of the 3D seismic data (Figure 7c). These  
307 anomalies are found to be laterally continuous. On the contrary, Q values in the center part of  
308 this slice (Figure 7c) follow the trend of chimneys and Q values are comparatively lower than  
309 those in the slice above (Figure 7b). Extremely low Q values have been observed in Q slice  
310 corresponding to free gas zone beneath the BSR except for the locations corresponding to that  
311 of chimneys (Figure 7d).

## 312 ANALYSIS USING SPECTRAL RATIO METHOD

313 **Q estimation**

314 The spectral ratio method can be applied to estimate effective Q of a medium between  
315 two prominent reflections (Figure 3). We extend this method to estimate Q for the same 4  
316 layers between prominent reflections in the seismic data (Figure 6a) as used in centroid  
317 frequency shift method. Picked reflections are windowed and spectral ratio method is applied  
318 on adjacent reflections to estimate a subsurface Q model. Figure 8 shows different steps  
319 involved in the application of the spectral ratio method on one of the traces. The same  
320 procedure is repeated on all the traces in the seismic volume to generate a Q cube.

321 **Observations**

322 The Q pattern for one of the inlines (Figure 6a) shows high Q values in the layer just  
323 above the BSR (Figure 6c). Q estimates based on spectral ratio method in this layer (Figure  
324 6c) is comparable to the Q estimates from centroid frequency shift method (Figure 6b). Q  
325 slices for different layers are plotted to further analyze the results (Figure 9). In the plan view,  
326 Q estimates vary significantly within the first layer (Figure 9a). In the second layer, small  
327 patches of high Q values coincide with the chimney features similar to that obtained by  
328 centroid frequency shift method. Also, we observe high Q anomalies in the Q slice  
329 corresponding to the layer just above the BSR (Figure 9c). We find that these high Q  
330 anomalies are laterally continuous and match well with the anomalies observed in Q slice  
331 obtained from centroid frequency shift method (Figure 7c). Beneath the BSR, we observe  
332 predominantly low Q values (Figure 9d). However, we also observe some regions with high  
333 Q anomalies in the fourth layer roughly corresponding to chimney locations (Figure 9d).  
334 Here, results obtained through spectral ratio method (Figure 9d) and centroid frequency shift

335 method (Figure 7d) do not agree with each other. At the BSR depth, signal strength is  
336 significantly reduced and Q estimates from spectral ratio method are extremely unstable. This  
337 may be the reason for high Q values observed in some areas below the BSR.

## 338 UNCERTAINTIES AND LIMITATIONS

339 Estimating Q from seismic data is typically accompanied by some uncertainties and  
340 limitations. Contribution of reflectivity sequences in calculated amplitude spectrum directly  
341 affects Q estimates. In case of thin layers with some periodicity, Earth's reflectivity function  
342 contributes in shaping the spectrum of effective recorded signals. Weak reflectivity over a  
343 time window (Figure 5a) can also create bias in Q estimates if the noise spectrum is not  
344 white. Ning and Wen-kai (2010) discussed in detail about the effect of reflectivity sequences  
345 on Q estimates. Spectral ratio method is more sensitive to these effects as Q is estimated from  
346 the spectrum of two wavelets. Fluctuations observed in centroid frequency curves is also  
347 primarily due to the effect of reflectivity sequences in the recorded signal.

348 Scattering is another factor that will lead to reduction in amplitude of different  
349 frequencies. Different types of scattering can occur depending on the size of the particles  
350 which include Rayleigh, Mie and forward scattering (Mavko et al., 1998). Mie scattering is  
351 the type of scattering that will occur when the heterogeneity scale length is of the order of the  
352 seismic wavelength. The main difference between scattering and intrinsic attenuation is that  
353 scattering redistributes wave energy within the medium but does not remove the energy from  
354 the overall wavefield whereas intrinsic attenuation converts vibration energy into heat energy  
355 (Sato and Fehler, 1997). Intrinsic attenuation quantified using different methods also includes  
356 the contribution from scattering attenuation (Spencer et al., 1982). This will directly affect the  
357 Q estimates from different methods.



358 Processing of seismic data can be another source of error in Q estimation. Ideally, all  
359 the processing steps which can potentially alter amplitude spectrum of a seismic signal  
360 should be avoided. In P-Cable data, the potential of this problem is significantly reduced as  
361 P-Cable data is close to zero offset (97-143 m), particularly given the water depth in the  
362 present study. Frequency distortions due to NMO stretching are negligible for small offsets at  
363 deep water depths. Frequency distortions due to Stolt migration are also negligible as layers  
364 in the study area are essentially flat. Apart from these two processing steps, no other step has  
365 been involved which can potentially influence the analysis.

366 Travel time through a picked layer is a very important factor in estimating Q using  
367 centroid frequency shift method. Picking more reflections and using them as layer boundaries  
368 increases the number of layers for which effective Q model will be estimated but decreases  
369 the travel time of the layers. Effect of fluctuations in the centroid frequency curve on Q  
370 estimates is more pronounced for thinner layers. Thus, accuracy of Q estimates in thinner  
371 layers is poorer than thicker layers. Figure 5e shows best fit Q models for different number of  
372 layers. When the number of picked layers is increased from 5 to 7, the instability in the Q  
373 estimates can be clearly seen. Therefore, reflections need to be picked properly so that Q can  
374 be estimated for different layers with an acceptable accuracy.

375 Histogram of Q estimates from centroid frequency shift method (Figure 10a) and  
376 spectral ratio method (Figure 10d) in layer 3 (which lies just above the BSR) have been  
377 plotted to analyze the statistical distribution of Q estimates within a layer. The peak at Q=600  
378 observed in the histograms is due to the fact that only Q values up to 600 have been taken  
379 into consideration. All Q values greater than 600 will be estimated as 600 and it is extremely  
380 difficult to differentiate between different Q values for those higher than approximately 150.  
381 Accuracy of Q estimates decreases for high Q values where it changes very rapidly with

382 small change in seismic signal decay. This small amount of decay becomes comparable with  
383 the fluctuations caused by other factors which create a problem in Q estimation. Figure 5d  
384 shows the estimated Q for different layers. From layer 2 to layer 3, estimated Q changes from  
385 160 to 403, but there is a very small change in tilt of the best fit curve. This limits the  
386 accuracy of Q estimates for high Q values and due to this fact, only Q values up to 600 have  
387 been taken into consideration (Figure 10).

388           Given the two methodological approaches for estimating Q, their inherent limitations  
389 and the constraints of the 3D seismic, as earlier reported by Quan and Harris (1997) and  
390 Matsushima (2006), we also found that the centroid frequency shift method gave more stable  
391 Q estimates. Contribution of reflectivity sequences in calculated amplitude spectrum and  
392 scattering effects limited the vertical resolution of Q estimates. We tried to do high-resolution  
393 Q sampling but accuracy of Q estimates decreased when greater number of layers was used to  
394 estimate Q. Reflectivity sequences and scattering effects made the continuous mapping of Q  
395 unstable. Q estimates became unreliable for thinner layers. We observed trade-off between  
396 the accuracy and resolution.

## 397   DISCUSSION

398           There are different factors which contribute to intrinsic attenuation of a seismic signal  
399 (Toksöz and Johnston, 1981). Major factors which play a crucial role are lithology, fluid type  
400 and structural features (Walsh, 1966; Johnston et al., 1979; Toksöz et al., 1979; Toksöz and  
401 Johnston, 1981; Winkler et al., 1979; Spencer, 1979; Winkler and Nur, 1982; Murphy et al.,  
402 1986; Pointer et al., 2000; Parra et al., 2002; Prasad and Nur, 2003; Behura, 2009). The exact  
403 details of the sedimentary environment of the study area are not well known but it is believed  
404 to be composed of smoothly deposited layers of contourite deposits (Eiken and Hinz, 1993;  
405 Howe et al., 2008). Under such geological settings, the major factors which can prominently

406 change the intrinsic attenuation property of a medium are changes in fluid type and fluid  
407 saturation. Changes in gas hydrate saturation within gas hydrate stability zone will sharply  
408 change the intrinsic attenuation spatially. Several publications explained the relationship  
409 between seismic attenuation and fluid saturation (O'Connell and Budiansky, 1977; Mavko  
410 and Nur, 1979; Spencer, 1979; Murphy et al., 1986; O'Hara, 1989; Pointer et al., 2000;  
411 Prasad and Nur, 2003; Rapoport et al. 2004). In addition, structural features scatter the  
412 seismic signal and contribute significantly to the estimated intrinsic attenuation (Hamilton  
413 and Mooney, 1990). It is thus challenging to distinguish between scattering attenuation and  
414 intrinsic attenuation (Wennerberg, 1993). The Q parameter estimated for quantifying intrinsic  
415 attenuation of a medium also includes the effects from scattering attenuation (Spencer et al.,  
416 1982). Possible effects of gas hydrates and free gas on Q estimates is studied by estimating Q  
417 values for different layers in the gas hydrate stability zone and free gas zone. The spatial  
418 analysis of the Q estimates from the 3D seismic data then allows us to recognize structures  
419 and areas that can be related to the presence of gas hydrates in marine sediments even in the  
420 absence of seismic velocity control.

421 Q values have been estimated for different layers using centroid frequency shift  
422 method and spectral ratio method. Q values estimated in deeper layers (L2, L3, and L4) using  
423 these two methods are found to be in concordance with each other and Q values in layers just  
424 above the BSR (L2 and L3) are in good agreement with the Q values normally observed in  
425 the gas hydrate bearing marine sediments (Wood et al., 2000). Q estimates in the first layer  
426 (L1) do not correspond well. Noisy amplitude spectrum near the sea floor (Dewangan et al.  
427 2014) and fluctuating spectral ratio (Figure 8c) can be the possible reason for the unstable Q  
428 estimates from spectral ratio method in the first layer. However, in the context of this  
429 analysis, it is important to study relative changes in Q particularly along Q slices throughout

430 the whole volume as these might be related to the type of pore fluid and saturation in a given  
431 area or structure.

432 Both Q analysis methods estimate high Q values in a layer just above the BSR (Figure  
433 7c and Figure 9c). Below the BSR, the centroid frequency (Figure 4b) and Q values of both  
434 methods drop significantly (Figure 7d and Figure 9d). Very low Q values are observed below  
435 the BSR except for the locations below chimneys, where high Q is observed (Figure 7d and  
436 Figure 9d). High free gas concentration can be the reason for rapid attenuation of the seismic  
437 signal below the BSR. The strength of the BSR in the seismic data (Figure 4a) also gives  
438 some indication about the accumulation of free gas in the region which is estimated to be as  
439 high as 1.5-2% of pore space (Hustoft et al., 2009). In gas chimneys, seismic signal  
440 significantly attenuates due to scattering especially in shallow seafloor features like  
441 pockmarks. Low signal strength accompanied with seismic blanking in the gas chimneys  
442 make Q estimates in gas chimneys unreliable especially at deeper depths.

443 By analyzing the distribution of Q values in the layer L3 (Figure 10a and 10d), it can  
444 be stated that the background Q values in the marine sediments at the BSR depth is in the  
445 range of 60-90. If Q values in the layer L3 (layer above the BSR) above potentially gas  
446 saturated sediments ( $Q < 30$  below the BSR in the layer L4) are selectively picked (Figure 10b  
447 and 10e) and compared with the overall distribution of Q values in the layer (Figure 10c and  
448 10f), relatively higher Q values have been observed above potentially gas saturated sediments  
449 (Figure 10). Particularly the variable distribution of extended zones of high Q mapped on Q  
450 slices of the 3D data (Figure 11b), in comparison to adjacent areas with lower Q, points  
451 towards variable pore fluid type and/or saturation in this strata. There is no indication from  
452 the seismic data to expect significant lithologic changes in this rather homogeneous  
453 sedimentary environment. Therefore, we attribute this effect to the presence of gas hydrates

454 in the sediments and suggest that gas hydrate saturated sediments exhibit high Q values  
455 within the frequency range used in the study. This observation is supported by the fact that  
456 both Q analysis methods match well in the distribution of Q above the BSR. In contrast, areas  
457 with very low Q below the BSR indicate the presence of free gas (Figure 11c).

458         It is difficult to estimate accurate Q for high Q value areas as discussed earlier.  
459 Therefore it becomes difficult to state exact Q value in gas hydrate saturated sediments. But  
460 from the statistical analysis of the results obtained from both methods (Figure 10), it can be  
461 stated that high Q values are observed in gas hydrate saturated sediments. Earlier studies on  
462 seismic attenuation conducted in the nearby locations also indicated elevated Q values above  
463 the BSR (Rossi et al., 2007). Hence, we argue that Q analysis of high-resolution P-Cable 3D  
464 seismic data with a large bandwidth can detect and outline spatially limited areas of gas  
465 hydrate occurrence in marine sediments.

466         Hustoft et al. (2009) used 135 km east-west striking multi-channel seismic (MCS)  
467 profile to derive a velocity model. This profile lies approximately 10 Km southward to our  
468 study area (shown in Figure 1) and can be used to interpret the results of the Q analysis. High  
469 gas concentrations exist beneath the BSR towards the southwestern half of the Vestnesa  
470 Ridge (Hustoft et al., 2009). Similarly, low Q values beneath the BSR in the southwestern  
471 half of the 3D seismic data may indicate the presence of elevated gas concentrations at this  
472 location (Figures 7d and 9d). The gas chimneys that align at the crest of the Vestnesa Ridge  
473 separate this southwestern half from the northeastern half where Q values are generally  
474 higher. A similar behavior is observed above the BSR where high Q values in the  
475 southwestern half may indicate higher concentrations of gas hydrates than in the northeastern  
476 half. Hustoft et al. (2009) and Bünz et al. (2012) showed that the fluid flow system in the  
477 Vestnesa Ridge is topographically controlled and that gas migrates to the crest of the ridge

478 beneath the BSR, mostly from the southwestern half. Hence, gas availability may be higher in  
479 the southwestern half.

480 In a more recent study, Plaza-Faverola et al. (2015) showed that small-scale fault  
481 systems exist at the crest of the Vestnesa Ridge documenting a tectonic control of gas  
482 leakage. Fault systems mapped by Plaza-Faverola et al. (2015) at approximately the BSR  
483 depth coincide strikingly with the boundaries of abrupt Q changes within layer 3 and 4 above  
484 and beneath the BSR, respectively (Figure 11a-c). Changes in Q within a layer are attributed  
485 with a variable pore fluid fill. Hence, the Q analysis indicates that fluid distribution in the  
486 region is strongly controlled by fault systems in the Vestnesa Ridge. Fault 1 in Figure 11a  
487 delimits the southwestern part indicating higher fluid concentrations in both hydrates above  
488 the BSR and free gas below. This area also includes the two most active chimneys on the  
489 Vestnesa Ridge (Figure 4c) (Bünz et al., 2012; Smith et al., 2014) corroborating our results  
490 that gas is more readily available at this location to either leak to the seafloor or to be bound  
491 into gas hydrates. As gas migrate upslope in north-east direction (Figure 11d; Hustoft et al.,  
492 2009), they are trapped by the regional fault 1 and utilize the fault plane as migration  
493 pathway into the free gas zone beneath the BSR.

494 Q values between fault 1 and 2 indicate lower concentrations of gas hydrates (Figure  
495 11b) and free gas (Figure 11c). Coincidentally, several of the chimneys located in this fault  
496 block are inactive. The reduced availability of free gas in this fault block might explain this  
497 observation or that most gas has vented through the chimneys. Also other areas of the 3D  
498 seismic volume clearly indicate a relationship between Q values and the mapped fault system,  
499 e.g. to the northeast of fault 2 or between fault 2 and fault 4 (Figure 11a-c). Together, these  
500 results suggest that the availability of free gas is one of the major factors in the accumulation  
501 of gas beneath the BSR and the formation of gas hydrates above it, and that the availability of

502 free gas clearly seems to be controlled by the structural setting supporting the findings of  
503 Plaza-Faverola et al. (2015).

504 All the Q slices clearly exhibit the vertical fluid flow features in this area (Figures 7  
505 and 9). However, there are still some interesting subtleties that can be noted from the Q data.  
506 When centroid frequencies are plotted for a seismic section, they didn't drop rapidly at some  
507 places in the northwestern and central parts (Figure 4b) of the seismic section. It is possible  
508 that these frequency anomalies might be related to variable concentrations of gas within the  
509 free gas zone beneath the BSR. Lower amounts of free gas might indicate that the fluid flow  
510 features like gas chimneys in nearby locations may lack a gas source. Bünz et al. (2012)  
511 documented acoustic flares in the water column and shallow high amplitudes in upper 50 m  
512 of these fluid flow features. In their study they show that the chimneys in the central part of  
513 the 3D seismic volume and some chimneys in northwestern part are inactive as compared to  
514 chimneys in the southeastern part of the volume (Figure 4c). On the contrary, the active  
515 chimneys documented by Bünz et al. (2012) show low frequency anomalies in the lower part  
516 of the chimney just above the BSR (Figure 4b). It might indicate an active migration of gas  
517 from the free gas zone into the chimney structures supplying the seafloor seep with gas.

518 Low centroid frequencies have been observed in regions where pockmark features  
519 have been observed. This can be due to prominent scattering at pockmarks or attenuation of  
520 the seismic energy within 5-10 m of sediments below the sea floor possibly resulting from the  
521 presence of hydrates and/or carbonates. Prominent scattering in pockmarks and within  
522 chimneys significantly reduced the signal strength and made it difficult to image Q in gas  
523 chimneys at deeper depth. Low signal strength and seismic blanking in gas chimneys reduces  
524 the accuracy of Q estimates in gas chimneys. But still Q values with limited accuracy have  
525 been used to study gas chimneys. Both Q estimation methods show small patches of high Q

526 values associated with chimney features at medium depth beneath seafloor and BSR (Figure  
527 7b and Figure 9b). However, the centroid frequency shift method depicts high Q at the rim of  
528 the chimneys possibly indicating that chimneys are lined with hydrates, an interesting though  
529 speculative suggestion, although it would fit with theoretical models for chimneys structures  
530 (Liu and Flemings, 2007).

531

## SUMMARY

532 We applied the centroid frequency shift method and spectral ratio method to study  
533 seismic attenuation in gas hydrate and free gas saturated sediments using high-resolution P-  
534 Cable 3D seismic data from Vestnesa Ridge on the Arctic continental margin of Svalbard.  
535 We estimated Q values for different layers to develop a subsurface 3D Q model. We observed  
536 high Q values above the prominent BSR and low Q values ( $Q \approx 10-30$ ) below the BSR.  
537 Anomalies observed in Q slices obtained from two different methods are found in  
538 concordance with each other. But we got relatively more stable Q values from centroid  
539 frequency shift method.

540 After performing a statistical analysis, we found that an increase in Q values in  
541 certain, spatially limited areas above the BSR can probably be associated with the presence of  
542 gas hydrates. Under this premise, Q analysis of high-resolution P-Cable 3D seismic data is  
543 thus an effective method for the detection and mapping of gas hydrate occurrences in marine  
544 sediments. Q values estimated for the strata below the BSR are very low as a consequence of  
545 the occurrence of gas trapped in the free gas zone beneath hydrate-bearing strata.

546 Faults that exist throughout the Vestnesa Ridge coincide with the Q anomalies in the  
547 layers above and below the BSR corroborating recent findings and directly showing that the  
548 structural setting and tectonic activity in the region control the availability and spatial



549 distribution of free gas and gas hydrates in the Vestnesa Ridge. The availability of gas in  
550 certain spatially limited areas also might explain the present seepage from some of the  
551 chimneys on the Vestnesa Ridge whereas other chimneys are dormant. Low seismic signal  
552 strength accompanied by amplitude blanking makes it difficult to accurately image Q in gas  
553 chimneys. But still with limited accuracy, we observed high Q values in gas chimneys in Q  
554 slices hinting towards the presence of gas hydrates in gas chimneys.

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#### 561 REFERENCES

562 Amundsen, L., and R. Mittet, 1994, Estimation of phase velocities and Q-factors from  
563 zero-offset, vertical seismic profile data: *Geophysics*, **59**, 500– 517.

564 Badri, M., and H. M. Mooney, 1987, Q measurements from compressional seismic  
565 waves in unconsolidated sediments: *Geophysics*, **52**, 772– 784.

566 Båth, M., 1982, *Spectral Analysis in Geophysics*: Elsevier Science.

567 Bauer, K., C. Haberland, R. G. Pratt, F. Hou, B. E. Medioli, and M. H. Weber, 2005,  
568 Ray based cross-well tomography for P-wave velocity, anisotropy, and attenuation structure  
569 around the JAPEX/JNOC/GSC et al. Mallik 5L-38 gas hydrate production research well, *in* S.  
570 R. Dallimore, and T. S. Collet, eds., *Scientific results from the Mallik 2002 Gas Hydrate*

- 571 Production Research Well Program, Mackenzie Delta, Northwest Territories, Canada:  
572 Geological Survey of Canada, Bulletin 585, 1-21.
- 573 Behura, J., 2009, Estimation and Analysis of Attenuation Anisotropy: Doctoral thesis,  
574 Colorado School of Mines.
- 575 Bellefleur, G., M. Riedel, T. Brent, F. Wright, and S. R. Dallimore, 2007, Implication  
576 of seismic attenuation for gas hydrate resource characterization, Mallik, Mackenzie Delta,  
577 Canada: *Journal of Geophysical Research*, **112**, B10311, doi:10.1029/2007JB004976.
- 578 Best, A. I., J. A. Priest, C. R. I. Clayton, and E. V. L. Rees, 2013, The effect of  
579 methane hydrate morphology and water saturation on seismic wave attenuation in sand under  
580 shallow sub-seafloor conditions: *Earth and Planetary Science Letters*, **368**, 78-87.
- 581 Brooks, J. M., H. B. Cox, W. R. Bryant, M. C. Kennicutt II, R. G. Mann, and T. J.  
582 McDonald, 1986, Association of gas hydrates and oil seepage in the Gulf of Mexico: *Organic*  
583 *Geochemistry*, **10**, 221-234, doi: 10.1016/0146-6380(86)90025-2.
- 584 Bünz, S., S. Polyanov, S. Vadakkepuliambatta, C. Consolaro, and J. Mienert, 2012,  
585 Active gas venting through hydrate-bearing sediments on the Vestnesa Ridge, offshore W-  
586 Svalbard: *Marine Geology*, **332-334**, 189-197, doi:10.1016/j.margeo.2012.09.012.
- 587 Chand, S., and T. A. Minshull, 2004, The effect of hydrate content on seismic  
588 attenuation: a case study for Malik 2L-38 well data, Mackenzie delta, Canada: *Geophysical*  
589 *Research Letters*, **31**, L14609, doi:10.1029/2004GL020292.
- 590 Chand, S., T. A. Minshull, D. Gei, and J. M. Carcione, 2004, Elastic velocity models  
591 for gas-hydrate bearing sediments – A comparison: *Geophysical Journal International*, **159**,  
592 573-590.

593 Claerbout, J. F., and F. Muir, 1973, Robust modeling with erratic data: Geophysics,  
594 **38**, 826-844.

595 Collett, T. S., and J. Ladd, 2000, Detection of gas hydrate with downhole logs and  
596 assessment of gas hydrate concentrations (saturations) and gas volumes on the Blake Ridge  
597 with electrical resistivity log data, *in* C. K. Paull, R. Matsumoto, P. J. Wallace, and W. P.  
598 Dillon, eds., Proceedings of the Ocean Drilling Program, Scientific results: Texas A & M  
599 University, College station Texas, **164**, 179-191.

600 Collet, T. S., R. E. Lewis, S. R. Dallimore, M. W. Lee, T. H. Mroz, and T. Uchida,  
601 1999, Detailed evaluation of gas hydrate reservoir properties using JAPEX/JNOC/GSC  
602 Mallik well 2L-38 gas hydrate research well downhole well-log displays, *in* S. R. Dallimore,  
603 T. Uchida, and T. S. Collett, eds., Scientific Results from JAPEX7/JNOC/GSC Mallik 2L-38  
604 gas hydrate well, Mackenzie Delta, North west Territories, Canada, Geological Survey of  
605 Canada Bulletin 544, 295-312.

606 Consolaro, C., T. L. Rasmussen, G. Panieri, J. Mienert, S. Bünz, and K. Sztybor,  
607 2014, Carbon isotope ( $\delta^{13}\text{C}$ ) excursions suggest times of major methane release during the  
608 last 14 ka in Fram Strait, the deep-water gateway to the Arctic: Climate of the Past  
609 Discussions, **10**, 4191-4227, doi:10.5194/cpd-10-4191-2014.

610 Dewangan, P., R. Mandal, P. Jaiswal, T. Ramprasad, and G. Sriram, 2014, Estimation  
611 of seismic attenuation of gas hydrate bearing sediments from multi-channel seismic data: A  
612 case study from Krishna–Godavari offshore basin: Marine and Petroleum Geology, **58**, Part  
613 A, 356-367, doi:10.1016/j.marpetgeo.2014.05.015.

614 Ecker, C., J. Dvorkin, and A. Nur, 1998, Sediments with gas hydrates: Internal  
615 structure from seismic AVO: Geophysics, **63**, 1659-1669.

- 616 Eiken, O., and K. Hinz, 1993, Contourites in the Fram Strait: *Sedimentary Geology*,  
617 **82**, 15-32.
- 618 Gabor, D., 1946, Theory of communication: *Journal of the Institution of Electrical*  
619 *Engineers*, **93**, no. 26, 429–457.
- 620 Galperin, E. I., 1985, *Vertical seismic profiling and its exploration potential*: Springer,  
621 doi: 10.1007/978-94-009-5195-2.
- 622 Gei, D., and J. M. Carcione, 2003, Acoustic properties of sediments saturated with gas  
623 hydrate, free gas and water: *Geophysical prospecting*, **51**, 141-157.
- 624 Gladwin, M. T., and F. D. Stacey, 1974, Anelastic degradation of acoustic pulses in  
625 rocks: *Physics of the Earth and Planetary Interiors*, **8**, 332– 336.
- 626 Guerin, G., and D. Goldberg, 2002, Sonic waveform attenuation in gas hydrate-  
627 bearing sediments from the Mallik 2L-38 research well, Mackenzie Delta, Canada: *Journal of*  
628 *Geophysical Research*, **107**, no. B5, 2088, doi:10.1029/2001JB000556.
- 629 Guerin, G., and D. Goldberg, 2005, Modeling of acoustic wave dissipation in gas  
630 hydrate-bearing sediments: *Geochemistry, Geophysics, Geosystems*, **6**, Q07010,  
631 doi:10.1029/2005GC000918.
- 632 Guerin, G., D. Goldberg, and A. Meltser, 1999, Characterization of in situ elastic  
633 properties of gas hydrate bearing sediments on the Blake Ridge: *Journal of Geophysical*  
634 *Research*, **104**, 17781-17795.

635 Hamilton, R. M., and W. D. Mooney, 1990, Seismic-wave attenuation associated with  
636 crustal faults in the new madrid seismic zone: *Science*, **248**, no. 4953, 351-354,  
637 doi: 10.1126/science.248.4953.351.

638 Hatherly, P. J., 1986, Attenuation measurements on shallow seismic refraction data:  
639 *Geophysics*, **51**, 250– 254.

640 Howe, J. A., T. M. Shimmiel, and R. Harland, 2008, Late Quaternary contourites  
641 and glaciomarine sedimentation in the Fram Strait: *Sedimentology*, **55**, no.1, 179-200.

642 Hustedt, B., and R. A. Clark, 1999, Source/receiver array directivity effects on marine  
643 seismic attenuation measurements: *Geophysical Prospecting*, **47**, 1105-1119.

644 Hustoft, S., S. Bünz, J. Mienert, and S. Chand, 2009, Gas hydrate reservoir and active  
645 methane-venting province in sediments on < 20 Ma young oceanic crust in the Fram Strait,  
646 offshore NW-Svalbard: *Earth and Planetary Science Letters*, **284**, no. 1-2, 12-24.

647 Jacobson, R. S., G. G. Shor, Jr., and L. M. Dorman, 1981, Linear inversion of body  
648 wave data- Part II: Attenuation versus depth using spectral ratios: *Geophysics*, **46**, 152-162.

649 Jaiswal, P., P. Dewangan, T. Ramprasad, and C. Zelt, 2012, Seismic characterization  
650 of hydrates in faulted, fine-grained sediments of Krishna-Godavari Basin: full waveform  
651 inversion: *Journal of Geophysical Research*, **117**, B10305, doi:10.1029/2012JB009201.

652 Janssen, D., J. Voss, and F. Theilen, 1985, Comparison of methods to determine Q in  
653 shallow marine sediments from vertical reflection seismograms: *Geophysical Prospecting*,  
654 **33**, 479-497, doi: 10.1111/j.1365-2478.1985.tb00762.x.

655 Johnston, D. H., M. N. Toksöz, and A. Timur, 1979, Attenuation of seismic waves in  
656 dry and saturated rocks: II. Mechanisms: *Geophysics*, **44**, 691-711.

657 Jung, J.-W., J. C. Santamarina, and K. Soga, 2012, Stress-strain response of hydrate-  
658 bearing sands: Numerical study using discrete element method simulations: *Journal of*  
659 *Geophysical Research*, **117**, B04202, doi: 10.1029/2011JB009040.

660 Korenaga, J., W. S. Holbrook, S. C. Singh, and T. A. Minshull, 1997, Natural gas  
661 hydrates on the southeast US margins: Constraints from full waveform and travelttime  
662 inversions of wide angle seismic data: *Journal of Geophysical Research*, **102**, 15345-15365.

663 Lee, M. W., and T. S. Collett, 2001, Elastic properties of gas hydrate-bearing  
664 sediments: *Geophysics*, **66**, 763-771.

665 Lee, M. W., and T. S. Collett, 2006, Gas hydrate and free gas saturations estimated  
666 from velocity logs on Hydrate ridge, offshore Oregon, USA, *in* A. M. Tréhu, G. Bohrmann,  
667 M. E. Torres, and F. S. Colwell, eds., *Proceedings of the Ocean Drilling Program, Scientific*  
668 *Results*, **204**: College Station TX (Ocean Drilling Program), 1-25.  
669 doi:10.2973/odp.proc.sr.204.103.2006.

670 Liu, B., Q. Yuan, K. H. Su, X. Yang, B. C. Wu, C. Y. Sun, and G. J. Chen, 2012,  
671 Experimental simulation of the exploitation of natural gas hydrate: *Energies*, **5**, 466-493, doi:  
672 10.3390/en5020466.

673 Liu, X., and P. B. Flemings, 2007, Dynamic multiphase flow model of hydrate  
674 formation in marine sediments: *Journal of Geophysical Research*, **112**, no. B3, B03101,  
675 doi:10.1029/2005JB004227.

676 Matsushima, J., 2005, Attenuation measurements from sonic waveform logs in  
677 methane hydrate-bearing sediments at the Nankai Trough exploratory well off Tokai, central  
678 Japan: *Geophysical Research Letters*, **32**, L03306, 1-5, doi: 10.1029/2004GL021786.

679 Matsushima, J., 2006, Seismic wave attenuation in methane hydrate-bearing  
680 sediments: vertical seismic profiling data from the Nankai Trough exploratory well, offshore  
681 Tokai, central Japan: *Journal of Geophysical Research*, **111**, B10101,  
682 doi:10.1029/2005JB004031.

683 Mavko, G., T. Mukerji, and J. Dvorkin, 1998, *The Rock Physics Handbook –Tools  
684 for Seismic Analysis in Porous Media*: Cambridge University Press.

685 Mavko, G. M., and A. Nur, 1979, Wave attenuation in partially saturated rocks:  
686 *Geophysics*, **44**, 161–178.

687 Murphy, W. F., K. W. Winkler, and R. L. Kleinberg, 1986, Acoustic relaxation in  
688 sedimentary rocks: Dependence on grain contacts and fluid saturation: *Geophysics*, **51**, 757–  
689 766.

690 Ning, T., and L. Wen-kai, 2010, Improve Q estimates with spectrum correction based  
691 on seismic wavelet estimation: *Applied Geophysics*, **7**, 217-228, DOI: 10.1007/s11770-010-  
692 0252-2.

693 Nouzé, H., P. Henry, M. Noble, V. Martin, and G. Pascal, 2004, Large gas hydrate  
694 accumulations on the eastern Nankai Trough inferred from new high-resolution 2-D seismic  
695 data: *Geophysical Research Letters*, **31**, L13308, doi:10.1029/2004GL019848.

696 O'Connell, R. J., and B. Budiansky, 1977, Viscoelastic properties of fluid-saturated  
697 cracked solids: *Journal of Geophysical Research*, **82**, no. 36, 5719–5735.

- 698 O'Hara, S. G., 1989, Elastic-wave attenuation in fluid-saturated Berea sandstone:  
699 Geophysics, **54**, 785–788.
- 700 Parra, J. O., C. L. Hackert, and P.-C. Xu, 2002, Characterization of fractured low Q  
701 zones at the Buena Vista Hills reservoir, California: Geophysics, **67**, 1061–1070.
- 702 Petersen, C. J., S. Bünz, S. Hustoft, J. Mienert, and D. Klaeschen, 2010, High-  
703 resolution P-Cable 3D seismic imaging of gas chimney structures in gas hydrated sediments  
704 of an Arctic sediment drift: Marine and Petroleum Geology, **27**, no. 9, 1981-1994, doi:  
705 10.1016/j.marpetgeo.2010.06.006.
- 706 Planke, S., F. N. Eriksen, C. Berndt, J. Mienert, and D. Masson, 2009, Spotlight on  
707 Technology: P-Cable High-Resolution Seismic: Oceanography, **22**, no. 1, 85.
- 708 Plaza-Faverola, A., S. Bünz, J. E. Johnson, S. Chand, J. Knies, J. Mienert, and P.  
709 Franek, 2015, Role of tectonic stress in seepage evolution along the gas hydrate charged  
710 Vestnesa Ridge, Fram Strait: Geophysical Research Letters, **42**, no. 3, 733-742, doi:  
711 10.1002/2014GL062474.
- 712 Pointer, T., E. Liu, and J. A. Hudson, 2000, Seismic wave propagation in cracked  
713 porous media: Geophysical Journal International, **142**, 199–231.
- 714 Prasad, M., and A. Nur, 2003, Velocity and attenuation anisotropy in reservoir rocks:  
715 SEG, Expanded Abstracts, 1652-1655.
- 716 Pratt, R. G., K. Bauer, and M. Weber, 2003, Crosshole waveform tomography  
717 velocity and attenuation images of arctic gas hydrates: SEG, expanded abstracts, 2255-2258,  
718 doi: 10.1190/1.1817798.



719 Pratt, R. G, F. Hou, K. Bauer, and M. Weber, 2005, Waveform tomography images of  
720 velocity and inelastic attenuation from the Mallik 2002 crosshole seismic surveys, *in* S. R.  
721 Dallimore, and T. S. Collet, eds., Scientific results from the Mallik 2002 Gas Hydrate  
722 Production Research Well Program, Mackenzie Delta, Northwest Territories, Canada:  
723 Geological Survey of Canada, Bulletin 585, 1-14.

724 Priest, J. A., A. I. Best, and C. R. I. Clayton, 2006, Attenuation of seismic waves in  
725 methane gas hydrate-bearing sand: *Geophysical Journal International*, **164**, 149-159, doi:  
726 10.1111/j.1365-246X.2005.02831.x.

727 Quan, Y., and J. M. Harris, 1997, Seismic attenuation tomography using the  
728 frequency shift method: *Geophysics*, **62**, 895-905.

729 Rapoport, M. B., L. I. Rapoport, and V. I. Ryjkov, 2004, Direct detection of oil and  
730 gas fields based on seismic inelasticity effect: *The Leading Edge*, **23**, 276–278.

731 Reine, C., M. V. D. Baan, and R. Clark, 2009, The robustness of seismic attenuation  
732 measurements using fixed- and variable-window time-frequency transforms: *Geophysics*, **74**,  
733 no. 2, WA123-WA135, doi: 10.1190/1.3043726.

734 Riedel, M., T. S. Collett, P. Kumar, A. V. Sathe, and A. Cook, 2010, Seismic imaging  
735 of a fractured gas hydrate system in the Krishna-Godavari basin offshore India: *Marine and*  
736 *Petroleum Geology*, **27**, 1476-1493, doi: 10.1016/j.marpetgeo.2010.06.002.

737 Ritzmann, O., W. Jokat, W. Czuba, A. Guterch, R. Mjelde, and Y. Nishimura, 2004,  
738 A deep seismic transect from Hovgård Ridge to northwestern Svalbard across the continental-  
739 ocean transition: A sheared margin study: *Geophysical Journal International*, **157**, 683-702.

740 Rossi, G., D. Gei, G. Böhm, G. Madrussani, and J. M. Carcione, 2007, Attenuation  
741 tomography: an application to gas-hydrate and free-gas detection: *Geophysical Prospecting*,  
742 **55**, 655-669.

743 Ryu, B.-J., T. S. Collett, M. Riedel, G. Y. Kim, J. H. Chun, J. J. Bahk, J. Y. Lee, J. H.  
744 Kim, and D. G. Yoo, 2013, Scientific results of the second gas hydrate drilling expedition in  
745 the Ulleung basin (UBGH2): *Marine and Petroleum Geology*, **47**, 1-20, doi:  
746 10.1016/j.marpetgeo.2013.07.007.

747 Sain, K., and A. K. Singh, 2011, Seismic quality factors across a bottom simulating  
748 reflector in the Makran Accretionary Prism, Arabian Sea: *Marine and Petroleum Geology*,  
749 **28**, no. 10, 1838-1843.

750 Sain, K., A. K. Singh, N. K. Thakur, and R. Khanna, 2009, Seismic quality factor  
751 observations for gas-hydrate-bearing sediments on the western margin of India: *Marine*  
752 *Geophysical Researches*, **30**, no. 3, 137-145, DOI: 10.1007/s11001-009-9073-1.

753 Sato, H., and M. C. Fehler, 1997, *Seismic wave propagation and scattering in the*  
754 *heterogeneous earth*: Springer, doi:10.1007/978-3-540-89623-4.

755 Shipley T. H., M. H. Houston, R. T. Buffler, F. J. Shaub, K. J. McMillen, J. W. Ladd,  
756 and J. L. Worzel, 1979, Seismic evidence for widespread possible gas hydrate horizons on  
757 Continental slopes and Rises: *The American Association of Petroleum Geologists Bulletin*,  
758 **63**, 2204-2213.

759 Sloan, E. D., Jr., 1998, *Clathrate Hydrates of Natural Gases*, 2<sup>nd</sup> edition: Marcel  
760 Dekker Inc.

- 761 Smith, A. J., J. Mienert, S. Bünz, and J. Greinert, 2014, Thermogenic methane  
762 injection via bubble transport into the upper Arctic Ocean from the hydrate - charged  
763 Vestnesa Ridge, Svalbard: *Geochemistry, Geophysics, Geosystems*, **15**, 1945–1959,  
764 doi:10.1002/2013GC005179.
- 765 Spencer, J. W., Jr., 1979, Bulk and shear attenuation In Berea Sandstone: The effect  
766 of pore fluids: *Journal of Geophysical Research*, **84**, 7521-7523.
- 767 Spencer, T. W., J. R. Sonnad, and T. M. Butler, 1982, Seismic Q — Stratigraphy or  
768 dissipation: *Geophysics*, **47**, 16–24, doi: 10.1190/1.1441275.
- 769 Talukder, M. W., 2013, VSP P-wave attenuation model study in elastic earth: spectral  
770 ratio method vs centroid frequency shift method: CSPG CSEG CWLS GeoConvention 2013,  
771 Integration: Geoscience Engineering Partnership, Calgary, Canada.
- 772 Toksöz, M. N., and D. H. Johnston, 1981, Seismic wave attenuation: *Geophysics*  
773 reprint series no. 2, Society of Exploration Geophysicists.
- 774 Toksöz, M. N., D. H. Johnston, and A. Timur, 1979, Attenuation of seismic waves in  
775 dry and saturated rocks: I. Laboratory measurements: *Geophysics*, **44**, 681-690.
- 776 Tonn, R., 1991, The determination of the seismic quality factor Q from VSP data: A  
777 comparison of different computational methods: *Geophysical Prospecting*, **39**, 1 – 27.
- 778 Vogt, P. R., K. Crane, E. Sundvor, M. D. Max, and S. L. Pfirman, 1994, Methane-  
779 generated(?) pockmarks on young, thickly sedimented oceanic crust in the Arctic: Vestnesa  
780 ridge, Fram strait: *Geology*, **22**, 255-258

781 Walsh, J. B., 1966, Seismic Wave Attenuation in Rock due to Friction: Journal of  
782 Geophysical research, **71**, 2591-2599.

783 Wennerberg, L., 1993, Multiple-scattering Interpretations of coda-Q Measurements:  
784 Bulletin of the Seismological Society of America, **83**, 279–290.

785 Winkler, K. W., and A. Nur, 1982, Seismic attenuation: Effects of pore fluids and  
786 frictional-sliding: Geophysics, **47**, 1–15.

787 Winkler, K., A. Nur, and M. Gladwin, 1979, Friction and seismic attenuation in rocks:  
788 Nature, **277**, 528–531.

789 Wood, W. T., W. S. Holbrook, and H. Hoskins, 2000, In situ measurements of P-wave  
790 attenuation in methane hydrate and gas bearing sediments on the Blake Ridge, *in* C. K., Paull,  
791 R. Matsumoto, P. J. Wallace, and W. P. Dillon, eds., Proceedings of the Ocean Drilling  
792 Program, Scientific Results, **164**, 265-272.

793 Yoo, D. G., N. K. Kang, B. Y. Yi, G. Y. Kim, B. J. Ryu, K. Lee, G. H. Lee, and M.  
794 Riedel, 2013, Occurrence and seismic characteristics of gas hydrate in the Ulleung basin, East  
795 Sea offshore Korea: Marine and Petroleum Geology, **47**, 236-247,  
796 doi:10.1016/j.marpetgeo.2013.07.001.

797 Zhang, H., S. Yang, N. Wu, X. Su, M. Holland, P. Schultheiss, K. Rose, H. Butler, G.  
798 Humphrey, and GMS-1 Science Team, 2007, Successful and surprising results for China's  
799 first gas hydrate drilling expedition: Fire in the Ice, **7**, no. 3, 6-9.

800

## 801 LIST OF FIGURES

802 Figure 1 Bathymetry map showing the location of our study area at Vestnesa Ridge,  
803 on the west-Svalbard continental margin. The inset figure shows a seafloor time-structure  
804 map derived from the 3D seismic data. Key inlines and crosslines are indicated on this  
805 seafloor map. Two small boxes (B1 and B2) show the location of centroid frequency curves  
806 plotted in Figure 5. A velocity model was derived from the multi-channel seismic (MCS) line  
807 by Hustoft et al. (2009) (see also Figure 11d).

808 Figure 2 a) Amplitude spectrum of the seismic signal near the seafloor. b) Centroid  
809 frequency of seismic signal in shallow sediments near the seafloor for different traces in  
810 seismic data. The black line shows the assumed centroid frequency at the seafloor. c) The  
811 blue curve shows the mean variance of the amplitude spectrum ( $\sigma_s^2$ ) at different arrival times.  
812 The red line shows the best fit line for changes of mean variance with two way travel time.

813 Figure 3 a) Sea floor reflection and BSR picked from a trace located at 26<sup>th</sup> inline and  
814 260<sup>th</sup> crossline (see Figure 1 for location). b) Amplitude spectrum of the picked wavelets. c)  
815 Plot of spectral ratio (in logarithmic scale) vs frequency. The red curve shows the best fit line  
816 ( $L_1$  norm) in this plot. Q is derived from the slope of the best fit line.

817 Figure 4 a) Selected seismic section (Inline 115, see Figure 1 for location) from the  
818 3D seismic data. b) Centroid frequency plot corresponding to the seismic section. Black  
819 arrows indicate the possible sub-surface fluid migration through gas chimneys and  
820 subsequent seafloor gas seepage. c) Acoustic flares documenting active seepage (modified  
821 from Bünz et al., 2012). The black line shows the location of the inline shown in a) and b).

822 Figure 5 a) Seismic trace corresponding to 183<sup>rd</sup> Inline and 1093<sup>rd</sup> crossline (location  
823 of the trace lies in the center of B1 as shown in Figure 1). Q has been estimated for the  $L_1$ ,

824 L2, L3, L4 and L5 intervals. b) Centroid frequencies calculated in the 5 ms sliding time  
825 window (Figure 5a). c) Centroid frequency trend/curve for 25 traces in a 31.25 x 31.25 m<sup>2</sup>  
826 small box B1 (Figure 1). d) The black curve shows the centroid frequency curve. The green  
827 curves show the synthetic centroid frequency curves corresponding to Q model 1 ( $Q_{L1}=500$  ,  
828  $Q_{L2}=400$ ,  $Q_{L3}=300$ ,  $Q_{L4}=200$ , and  $Q_{L5}=100$ ), Q model 2 ( $Q_{L1}=70$ ,  $Q_{L2}=100$ ,  $Q_{L3}=200$ ,  
829  $Q_{L4}=300$ , and  $Q_{L5}=500$ ), Q model 3 ( $Q_{L1}=150$ ,  $Q_{L2}=125$ ,  $Q_{L3}=100$ ,  $Q_{L4}=75$ , and  $Q_{L5}=50$ ), Q  
830 model 4 ( $Q_{L1}=50$ ,  $Q_{L2}=75$ ,  $Q_{L3}=150$ ,  $Q_{L4}=30$ , and  $Q_{L5}=200$ ), and Q model 5 ( $Q_{L1}=90$ ,  
831  $Q_{L2}=65$ ,  $Q_{L3}=50$ ,  $Q_{L4}=40$ , and  $Q_{L5}=30$ ). The red curve shows the synthetic centroid frequency  
832 curve for the best fit ( $L_1$  norm) Q model. b) 5 layer best fit Q model (red) and 7 layer best fit  
833 Q model (green) for a centroid frequency curve (location shown by a small box B2 in Figure  
834 1).

835 Figure 6 a) Seismic section of Inline 69 (see Figure 1 for location) with picked  
836 seafloor and three major subsurface reflections. L1, L2, L3, and L4 show the layers for which  
837 Q is estimated. b) Overlay of seismic section and Q estimates obtained from centroid  
838 frequency shift method. c) Overlay of seismic section and Q estimates obtained from spectral  
839 ratio method.

840 Figure 7 Q slices for different layers estimated using centroid frequency shift method.  
841 Q slices (a), (b), (c), and (d) correspond to layers L1, L2, L3, and L4 (Figure 6a) respectively.

842 Figure 8 a) Seismic trace corresponding to 98<sup>th</sup> inline and 686<sup>th</sup> crossline (see Figure 1  
843 for location). L1, L2, L3, and L4 are four layers between five reflections ( $R_1$ ,  $R_2$ ,  $R_3$ ,  $R_4$ , and  
844  $R_5$ ). b)  $AS_1$ ,  $AS_2$ ,  $AS_3$ ,  $AS_4$ , and  $AS_5$  are amplitude spectra calculated over time windows  $R_1$ ,  
845  $R_2$ ,  $R_3$ ,  $R_4$ , and  $R_5$  respectively. c), d), e), and f) show spectral ratio vs frequency plot. The red  
846 lines show the best fit line derived using  $L_1$  norm.  $Q_{L1}$ ,  $Q_{L2}$ ,  $Q_{L3}$ , and  $Q_{L4}$  are the derived Q  
847 values for layers L1, L2, L3, and L4 respectively.

848           Figure 9 Q slices for different layers estimated using spectral ratio method. Q slices  
849 (a), (b), (c), and (d) correspond to layers L1, L2, L3, and L4 (Figure 6a) respectively.

850           Figure 10 a) Histogram of Q values obtained using centroid frequency shift method in  
851 the layer L3. b) Histogram of Q values obtained using centroid frequency shift method in the  
852 layer L3 with  $Q < 30$  (high concentration of free gas) in the layer L4. c) Ratio of histogram b  
853 and histogram a. d) Histogram of Q values obtained using spectral ratio method in the layer  
854 L3. e) Histogram of Q values obtained using spectral ratio method in the layer L3 with  $Q < 30$   
855 (high concentration of free gas) in the layer L4. f) Ratio of histogram e and histogram d.  
856 Layers L1, L2, L3 and L4 are shown in Figure 6a.

857           Figure 11 a) Variance map obtained from a time slice at BSR depth showing several  
858 faults (Plaza-Faverola et al., 2015) and gas chimneys (Bünz et al., 2012) piercing through the  
859 subsurface. b) Q slice obtained by overlaying (through 50% transparency) Q slices from  
860 spectral ratio method (Figure 9c) and centroid frequency shift method (Figure 7c) in the layer  
861 L3 (layer above the BSR). c) Q slice obtained by overlaying (through 50% transparency) Q  
862 slices from spectral ratio method (Figure 9d) and centroid frequency shift method (Figure 7d)  
863 in the layer L4 (layer below the BSR). d) P-wave velocity cross-section derived using multi-  
864 channel seismic profile (see Figure 1 for location) across Vestnesa Ridge (modified from  
865 Hustoft et al., 2009). Arrows in the Figure show upslope gas migration and its leakage from  
866 the Vestnesa Ridge.