GEOPHYSICAL RESEARCH LETTERS, VOL. 32, L04308, doi:10.1029/2004GL021535, 2005

Imaging and quantification of gas hydrate and free gas at the Storegga slide offshore Norway

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Received 16 September 2004; revised 30 November 2004; accepted 18 January 2005; published 22 February 2005.

[1] Wide–angle reflection seismic experiments were performed at the Storegga slide offshore Norway in 2002 with the goal to quantify the amount of gas hydrate and free gas in the sediment. Twenty-two stations with Ocean Bottom Hydrophones (OBH) and Seismometers (OBS) were deployed for a 2D and a 3D experiment. Kirchhoff depth migration is used to transform the seismic wide–angle data into images of the sediment layers and to obtain *P* wave velocity–depth functions. The gas hydrate and free gas saturations are estimated from the elastic properties of the sediment on the basis of the Frenkel–Gassmann equations. There is 5–15% gas hydrate in the pore space of the sediment in the gas hydrate stability zone (GHSZ). The free gas saturation takes the value of 0.8% for a homogeneous distribution of gas in the pore water and 7% for the model of a patchy gas distribution.

Citation: Zillmer, M., T. Reston, T. Leythaeuser, and E. R. Flueh (2005), Imaging and quantification of gas hydrate and free gas at the Storegga slide offshore Norway, *Geophys. Res. Lett.*, *32*, L04308, doi:10.1029/2004GL021535.

1. Introduction

[2] The goal of the HYDRATECH project was to develop a method for the quantification of gas hydrates at European continental margins. Gas hydrates are gas molecules enclosed in cages of water molecules. Natural gas hydrates occur in marine sediments at continental margins, where appropriate pressure–temperature stability conditions prevail. The base of the gas hydrate stability zone can often be identified in seismic data as a Bottom Simulating Reflector (BSR), which marks the boundary between gas hydrate bearing and free gas bearing sediments [*Kaplan*, 1974; *Cox*, 1983; *Sloan et al.*, 1994; *Makogon*, 1997; *Henriet and Mienert*, 1998; *Sloan*, 1998; *Paull and Dillon*, 2001].

[3] The Storegga slide on the Norwegian margin is a large underwater slide, which was probably triggered by earthquakes, and the decomposition of gas hydrates may have contributed to the event. The occurrence of gas hydrates at the northern rim of the slide was first inferred from seismic data by *Bugge et al.* [1987]. Since then several seismic experiments for studying gas hydrates were performed. *Posewang and Mienert* [1999] concluded from seismic reflection and OBH data that a free gas layer exists within the GHSZ, and they explained a 'double BSR' by the occurrence of gases with different molecular weights. Other seismic reflection data sets were investigated by *Bouriak et al.* [2000] and *Bünz et al.* [2003]. A geotechnical borehole was drilled in 1997 up to 310 m depth. No gas hydrates were recovered, although the borehole penetrated the BSR depth, which could be caused by dissociation of gas hydrates during coring [*Andreassen et al.*, 2000]. Seismic measurements with an Ocean Bottom Cable (OBC) were performed at the location of the geotechnical borehole. *Andreassen et al.* [2003] reported that the base of the GHSZ did not generate a *P*–*S* converted wave, which was explained by the small amount of 1–1.5% gas hydrate in the pore space. The free gas saturation was estimated to 1–2%. *Bünz and Mienert* [2004] also investigated the OBC data, and determined the average gas hydrate saturation to 5% and the free gas saturation to 0.45%.

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[4] As a part of the HYDRATECH project two seismic experiments were performed at the Storegga slide in 2002 during a cruise with the RV Jan Mayen from the University of Tromsoe: A 3D experiment with an array of 22 OBH/OBS stations, centered at the location of the geotechnical borehole, and a 2D experiment with 22 stations along the profile. Two 40 in^3 sleeve guns were used as a seismic source at a shooting rate of 10 s. The seismic data from the 2D profile and from four stations of the 3D array are used for this study (Figure 1). Kirchhoff depth migration is applied to these data, and the amount of gas hydrate and free gas in the sediment is determined from *P* and *S* wave velocities, bulk density and porosity.

2. Kirchhoff Depth Migration

[5] Kirchhoff migration is a method, which transforms seismogram sections in the time domain to subsurface images in depth [*Bleistein et al.*, 2001]. A velocity model is obtained from the wide–angle data by repeating the migration and building the model from the top to the bottom of the structure. Twenty OBH/OBS images are shown in Figure 2 together with 1D velocity–depth functions, which were used to compute diffraction traveltimes.

[6] A lense–shaped 20 m thick layer A in 250 m depth below the sea floor (bsf) is characterized by a low *P* wave velocity of 1300 m/s, which can only be explained by the presence of free gas in the pore space. The reflectors B, C and D have negative polarity and mark the top of thin layers with free gas. According to *Andreassen et al.* [2003] and *Bünz et al.* [2003] these are layers of mud and clay, which were deposited during interglacial periods, and which alternate with layers of coarse–grained sediments and debris flow deposited during glacial periods. The average *P* wave velocity in layers C is 1500 m/s. We assume that the velocity jumps to even lower values at the top of the layers, and then increases again gradually in the layers, which corresponds to a decreasing gas concentration. This model

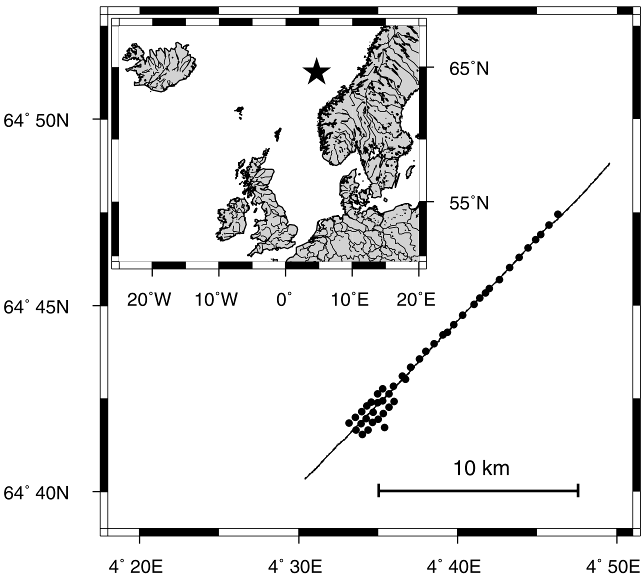

Figure 1. Seismic survey site offshore Norway. Shot profile with OBH/OBS stations.

explains the lack of reflections with positive polarity from the bottom of the layers. Reflections from below the free gas bearing layers have low amplitudes caused by attenuation during transmission through the gas bearing layers.

[7] The base of the GHSZ in 250–270 m depth bsf is indicated by the top of layer A and by the left ends of layers B and C, where free gas is laterally trapped by gas hydrate [*Bouriak et al.*, 2000]. A BSR can only be identified beneath station 687 [see also *Bünz and Mienert*, 2004]. A double BSR is found 60 m below the BSR. The raw data show that both BSRs have negative polarity, which supports the conclusion of *Posewang and Mienert* [1999] and of *Andreassen et al.* [2000] that they correspond to different types of gas composition. There are no low velocity layers with free gas in the GHSZ as reported by *Mienert and Posewang* [1999].

3. Gas Hydrate and Free Gas Saturation

[8] The Frenkel–Gassmann theory describes wave propagation in a medium, which consists of a solid and a fluid phase, for the case that the wavelength is much larger than the microstructure. This condition is fulfilled in marine seismic experiments, where the wavelength is about 10 m and the size of the sediment grains is between 10^{-3} and 10^{-6} m.

[9] A 'wet' (saturated) sediment consists of a solid phase and the pore fluid. The porosity ϕ is the fraction of the total volume, which is occupied by the fluid. The sediment is called 'dry' (unsaturated), if the pore fluid is substituted by voids. *Frenkel* [1944] and *Gassmann* [1951] showed that the shear modulus μ of the wet sediment and μ^* of the dry sediment are equal. They derived an equation, which connects the bulk moduli K of the wet and K^* of the dry sediment:

$$\frac{\sigma^2}{K-K^*}=\frac{\phi}{K_w}+\frac{\sigma-\phi}{K_s}. \tag{1}$$

Here K_s and K_w are the bulk moduli of solid phase and pore water, and $\sigma = 1 - K^*/K_s$. The bulk modulus of the wet sediment and the shear modulus can be calculated from P and S wave velocities, v_p and v_s, and density ρ, which are obtained from seismic or borehole measurements:

$$K=\rho\left(v_p^2-\frac{4}{3}v_s^2\right), \tag{2}$$

$$\mu^*=\mu=\rho v_s^2. \tag{3}$$

[10] The dry sediment is often represented by an aggregate of elastic spheres, and different theories of contact between the spheres were developed [*Wang and Nur*, 1992; *Mavko et al.*, 1998]. *Dvorkin et al.* [1999] used the Hertz–Mindlin contact theory and a Hashin–Shtrikman bound to compute the elastic moduli. The suspension domain is the interval of porosities between ϕ_c and 1, where $\phi_c \approx 0.36$ is the porosity of a dense random pack of spheres. By using the method of *Dvorkin et al.* [1999] it can be shown that the ratio of the bulk and shear modulus in the suspension domain is in the following interval (M. Zillmer et al., True amplitude Kirchhoff migration of OBH/OBS seismic data and quantification of gas hydrate at the Svalbard continental margin, submitted to *Geophysical Journal International*, 2004):

$$\frac{2}{3}\le\frac{K^*}{\mu^*}\le\frac{5}{3}. \tag{4}$$

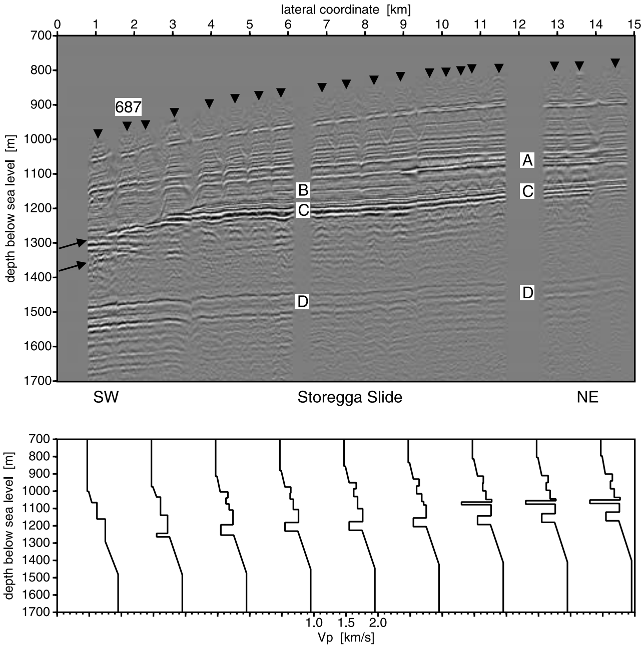

Figure 2. Images of 20 OBH/OBS stations in the vertical plane through the shot profile and corresponding velocity–depth functions. The Storegga Slide Scar is about 10 km to the SW. The water depth is between 800 and 1000 m. The station positions at the sea floor are marked by black triangles. The lense–shaped layer A contains free gas, and the reflectors B, C and D indicate the top of thin layers with free gas. A BSR can only be found beneath station 687 (arrow). The top of layer A and the left ends of layers B and C mark the base of the GHSZ in 250–270 m depth bsf.

K^* can be estimated from S wave velocity and density by using equations (3) and (4). A theoretical model, which describes the cementation of the sediment frame for high porosities, is not available. The results of *Dvorkin et al.* [1994] for low porosities indicate that the interval of equation (4) may still be valid in this case.

[11] In the frame of the HYDRATECH project, *Priest* [2004] performed laboratory experiments to study the elastic properties of gas hydrate bearing sand samples, and *Gei and Carcione* [2003] and *Chand et al.* [2004] developed and compared different effective medium theories. *Ecker et al.* [1998, 2000] and *Helgerud et al.* [1999] applied the Gassmann equation to gas hydrate and free gas bearing sediments. Gas hydrate as a component of the sediment frame reduces the porosity. The gas hydrate saturation S_h of the pore space can be written in explicit form (M. Zillmer et al., Seismic investigation of a bottom simulating reflector and quantification of gas hydrate in the Black Sea, submitted to *Geophysical Journal International*, 2003):

$$S_h = 1 - \left(\frac{1}{K_w} - \frac{1}{K_s}\right)^{-1} \frac{\sigma}{\phi}\left(\frac{\sigma}{K - K^*} - \frac{1}{K_s}\right). \qquad (5)$$

[12] Free gas in the pore water increases the compressibility of the pore fluid. The free gas saturation S_g of the pore space is given by:

$$S_g = \left(\frac{1}{K_w} - \frac{1}{K_g}\right)^{-1} \left[\frac{1}{K_w} - \frac{1}{K_s} - \frac{\sigma}{\phi}\left(\frac{\sigma}{K - K^*} - \frac{1}{K_s}\right)\right], \qquad (6)$$

where K_g is the bulk modulus of the free gas. K_w and K_g depend on temperature and pressure [*Batzle and Wang*, 1992].

[13] If patches of fully gas saturated and fully water saturated sediment occupy neighbouring regions on a length scale much smaller than the wavelength, then the result of equation (6) for S_g has to be multiplied by the factor

$$\frac{K - K^*}{\sigma^2\left(K + \frac{4}{3}\mu\right)}\left[\sigma^2 + \left(K^* + \frac{4}{3}\mu\right)\left(\frac{\phi}{K_g} + \frac{\sigma - \phi}{K_s}\right)\right], \qquad (7)$$

which leads to much higher estimates. This factor can be approximated by

$$\left[1 - \frac{7}{3}\left(\frac{v_s}{v_p}\right)^2\right]\left(1 + \frac{7}{3}\phi\frac{\rho v_s^2}{K_g}\right), \qquad (8)$$

because K_s is large and $K^* \approx \mu^* = \mu$. The factor μ/K_g, which is of size 10, can be used as a rule of thumb [see also *Nouzé et al.*, 2004].

[14] The upper/lower interval limit of equation (4) leads to more conservative estimates for the gas hydrate/free gas saturation. By using $K_g < K_w < K_s$, it follows from equations (5) and (6) that $S_h\, S_g \leq 0$. A fully water saturated sediment is indicated, if S_h and S_g are zero, but usually either gas hydrate ($S_h > 0$) or free gas ($S_g > 0$) seems to occur in the sediment. Thus it is important to take the error of S_h and S_g into account. The error of K_s hardly affects the result, which means that detailed information about the mineral composition of the sediment is not needed. Neglecting terms proportional K_s^{-1} in equation (5), one obtains

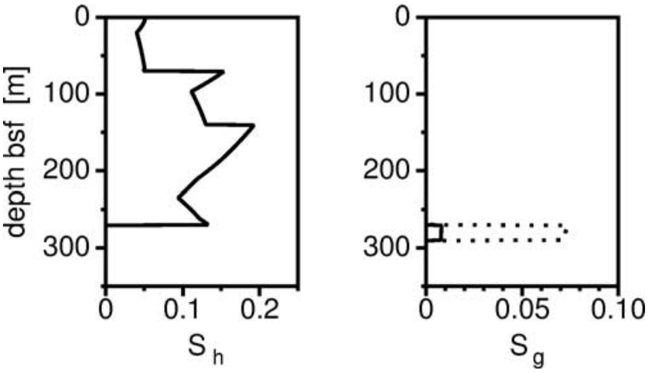

Figure 3. Gas hydrate S_h and free gas S_g saturation of the pore space beneath OBH/OBS station 687. A patchy distribution of free gas (dotted line) leads to much higher values than a homogeneous distribution of gas in the pore water (solid line).

$$\frac{\Delta S_h}{1 - S_h} \approx \frac{\Delta\phi}{\phi} + \frac{\Delta(K - K^*)}{K - K^*} - \frac{\Delta K_w}{K_w}. \qquad (9)$$

For small gas hydrate saturations the absolute error of S_h is given by the relative errors of porosity, bulk modulus and pore water bulk modulus. In general, an absolute error of 5–10% gas hydrate saturation is a realistic value.

[15] The gas hydrate and free gas saturations are computed for station 687, where a BSR is observed. This station is close to the geotechnical borehole, and a porosity–depth function from borehole measurements is available. The porosity ϕ varies between 63% and 48% [*Bünz and Mienert*, 2004]. The bulk density is computed from porosity by using a solid grain density of 2.65 g/cm^3. K_s is set to 40 GPa [*Bünz and Mienert*, 2004]. K_w and K_g are computed for a temperature of -1°C at the sea floor and a temperature gradient of 55°C/km [*Bouriak et al.*, 2000], and a salinity of 35000 ppm. K_w increases from 2.2 to 2.4 GPa in the first 300 m depth bsf and K_g takes a value of 0.03 GPa. S wave velocities were determined by *Andreassen et al.* [2003], and are of similar size as the empirical relations of *Hamilton* [1976]. The P wave velocity profile is obtained from Kirchhoff migration. The estimates for the gas hydrate and free gas saturations beneath station 687 are shown in Figure 3, where 5–15% gas hydrate is obtained between the sea floor and the base of the GHSZ. The details of the S_h profile are not interpreted due to possible large errors in the estimate. One might speculate that the porosity is 4% smaller for all depths, which shifts the gas hydrate saturation profile by 6% to smaller values. In this case no gas hydrate is obtained in the first 70 m depth bsf and up to 10% gas hydrate in a 200 m thick layer. The free gas saturation in layer A and the peak values in layers B and C take the value of 7 ± 3% for the patchy gas distribution and 0.8 ± 0.3% for the homogeneous distribution (Figure 3).

4. Conclusions

[16] Images of the sediment layers and P wave velocity models are obtained by applying Kirchhoff depth migration

to OBH/OBS common receiver gathers. Several layers with free gas are identified, which are characterized by high reflection amplitudes and low velocities. The free gas is laterally trapped at the base of the GHSZ, but a BSR is only locally observed. The gas hydrate and free gas saturation of the sediment is computed on the basis of the Frenkel–Gassmann theory. There is 5–15% gas hydrate in the pore space of the sediment in the GHSZ at the location of the BSR. The free gas saturation takes peak values of 7% for the patchy gas distribution and 0.8% for the homogeneous distribution.

[17] **Acknowledgments.** The authors thank the HYDRATECH project consortium for organising the experiments and providing the data. This work is part of the HYDRATECH project funded by the European Union (EVK3–CT–2000–00043). The free software packages GMT and SU were used in preparing this manuscript.

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