Modeling Hydrothermal Venting in Volcanic Sedimentary Basins: Impact on Hydrocarbon Maturation and Paleoclimate

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Abstract

Vent structures are intimately associated with sill intrusions in sedimentary basins globally and are thought to have been formed contemporaneously due to overpressure generated by gas generation during thermogenic breakdown of kerogen or boiling of water. Methane and other gases generated during this process may have driven catastrophic climate change in the geological past. In this study, we present a 2D FEM/FVM model that accounts for ‘explosive’ vent formation by fracturing of the host rock based on a case study in the Harstad Basin, offshore Norway. Overpressure generated by gas release during kerogen breakdown in the sill thermal aureole causes fracture formation. Fluid focusing and overpressure migration towards the sill tips results in vent formation after only few tens of years. The

size of the vent depends on the region of overpressure accessed by the sill tip. Overpressure migration occurs in self-propagating waves before dissipating at the surface. The amount of methane generated in the system depends on TOC content and the type of kerogen present in the host rock. Generated methane moves with the fluids and vents at the surface through a single, large vent structure at the main sill tip matching first-order observations. Violent degassing takes place within the first couple of hundred years and occurs in bursts corresponding to the timing of overpressure waves. The amount of methane vented through a single vent is only a fraction (between 5 and 16%) of the methane generated at depth. Upscaling to the Vøring and Møre Basins, which are a part of the North Atlantic Igneous Province, and using realistic host rock carbon content and kerogen values results in a smaller amount of methane vented than previously estimated for the PETM. Our study, therefore, suggests that the negative carbon isotope excursion (CIE) observed in the fossil record could not have been caused by intrusions within the Vøring and Møre Basins alone and that a contribution from other regions in the NAIP is also required to drive catastrophic climate change.

Keywords

sill intrusions in sedimentary basins; numerical model; hydrothermal vent complex; PETM; methane generation; hydrocarbons

Highlights

- Numerical model of sill intrusions in sedimentary basins
- Fracture and vent formation due to overpressure generation
- Methane fluxes through a single vent with upscaling to basin scales

- Additional regions in the NAIP required to correlate methane venting and the PETM

1 Introduction

Hydrothermal vent complexes are closely linked to igneous sheet intrusions in sedimentary basins where they form conduits for the release of fluids and gases generated in metamorphic aureoles. The vent complexes form pipe-like lower conduit structures, commonly observed to originate from near the tips of sills or other topological highs related to sill morphology, along with an upper part which manifests itself as a crater or dome at the paleosurface. Examples of this association have been observed globally and include the Vøring and Møre basins in the Norwegian Sea (Jamtveit et al., 2004; Planke et al., 2005; Svensen et al., 2004), the Karoo Basin in South Africa (Svensen et al., 2006), the Faroe-Shetland Basin (Grove, 2013; Hansen, 2006) and the Tunguska Basin in Siberia (Svensen et al., 2009). Extensive sill complexes are characteristic of Large Igneous Provinces (LIPs) emplaced in sedimentary basins, and considerably modify the thermal and hydrological properties of the host sedimentary strata. The understanding of this process is of importance not only to the scientific community but also to the industry as it can significantly affect petroleum exploration in volcanic basins (Monreal et al., 2009).

Two main overpressure mechanisms have been proposed to explain the formation of sill-associated hydrothermal vent complexes. The first mechanism suggests that boiling of pore fluid within the thermal aureole of the intruding sill and the release of additional fluids during metamorphic dehydration reactions would result in a large volume increase, thereby increasing the fluid pressure (Jamtveit et al., 2004). Hydrofracturing occurs once the fluid pressure exceeds the lithostatic and results in the formation of a vent. However, this mechanism would be limited to the upper 1 km of the sedimentary basin where pressures allow for the boiling of pore fluids. The second mechanism suggests that fluid

overpressure may be generated by the release of various gases such as CO₂, SO₂ and halocarbons during metamorphic dehydration reactions (Aarnes et al., 2010). Generation of large quantities of gaseous hydrocarbons such as CH₄ could result in large overpressures triggering catastrophic blowouts (Aarnes et al., 2012).

In a previous study, Iyer et al. (2013) developed a numerical model for a single generic sill that generated hydrothermal plumes without any need for overpressure mechanisms. These vents develop within few hundred years after sill emplacement and are controlled by flow patterns around the cooling intrusion. However, vent formation in this model is restricted to relatively high permeability systems. In this paper, we build on the model presented by Iyer et al. (2013) and study vent formation due to hydrofracturing associated with sill intrusives in low permeability systems. The study is based on a 2D transect in the Harstad Basin, offshore Norway, where multiple sills have been imaged in association with one main hydrothermal vent complex observed above the tip of the main sill. The model aims to recreate the observed vent structure and tracks the amount of thermogenic methane released into the system. Using this, we evaluate the amount of methane that vents from a single vent and extrapolate this volume up to the basin scale and reinvestigate its link to catastrophic extinction events.

2 Geological Setting

The Harstad Basin is located on the Norwegian continental shelf in the SW part of the Barents Sea, along the transition from the rifted Lofoten-Vesterålen margin to the sheared W Barents Sea margin (Figure 1a). It is bounded to the west by the continent-ocean boundary and to the east by the southernmost extension of the Troms-Finnmark Fault system (Olesen et al., 2002). The basin comprises a narrow rift basin forming a northerly extension of the NE Atlantic late Middle Jurassic-earliest Cretaceous rift system which formed the major depocentres of the Vøring and Møre basins further south (Faleide et al.,

2008). Minor magmatism occurred in the Harstad Basin during the Paleogene continental breakup between Norway and Greenland. To date, only one exploration well (7016/2-1) has been drilled into the far north of the basin making it a largely untested frontier basin. Well 7119/12-3, located ca. 100 km NE of the study area within the Ringvassøy-Loppa Fault Complex, was identified as the most suitable tie well outside the basin.

A seismic-stratigraphic interpretation of key horizons in the Harstad Basin is presented in Figure 1 and is based on interpretation of regional 2D and 3D seismic data and borehole ties. The basin has over 8 km of accumulated sediment in the deepest part and is dominated by a thick Lower Cretaceous succession. A number of laterally extensive sheet intrusions have been identified and mapped within the study area. Intrusions were identified by a combination of high amplitude, transgressive sections, saucer-shaped morphologies and associated hydrothermal vent complexes (Planke et al., 2005; Planke et al., 2014). The main sill body within the study area extends over ca. 380 km² in plane view and displays an elongated saucer-shaped morphology. An average thickness of ca. 0.1 km is inferred for the main sill based on interpretation of pre- and post-stack seismic reflection data (Planke et al., 2014), giving a total volume of ca. 4 km³ for the main sill complex.

A well-defined hydrothermal vent complex is imaged above the sill tip in the central part of the study area (Figure 1b). The vent is characterized by a ca. 1.5 km wide mound-shaped structure above the Base Vent horizon, interpreted to be the paleo-seafloor at the time of venting. Disturbed reflections within a narrow region between the sill tip and the mound represent the lower conduit zone (Planke et al., 2005) surrounded locally by inward dipping and layer-parallel high-amplitude. The high-amplitude events are interpreted as gas-charged sediments. The age of the venting is interpreted to be of earliest Eocene age (ca. 55.8 Ma), similar to the age of similar hydrothermal vent complexes mapped in the Vøring Basin (Svensen et al., 2010; Svensen et al., 2004).

112

113 **3 Mathematical Model**

114 The 2D hybrid FEM/FVM model (Finite Element Method/Finite Volume Method) presented here builds
115 on the model presented by Iyer et al. (2013).

116

117 **3.1 Hydrothermal Convection**

118 The fluid density changes are spilt into the pressure and temperature equivalents to give the pressure
119 field for a compressible fluid (Eqn. 1). The pressure equation also accounts for the source term
120 associated with methane generated by thermal cracking of organic matter (see section 3.2). Fluid
121 properties are based on pure water corresponding to the IAPWS-84 steam tables
122 (<http://www.iapws.org/>).

$$123 \quad \phi \rho_f \beta_f \frac{\partial P}{\partial t} = \nabla \left[\frac{k \rho_f}{\mu_f} (\nabla P - \rho_f \vec{g}) \right] - \rho_f \frac{\partial \phi}{\partial t} + \phi \rho_f \alpha_f \frac{\partial T}{\partial t} + R_{CH_4} \quad (\text{Eqn. 1})$$

124 Table 1 lists all the notations and values used in the model.

125 The energy equation (Eqn. 2) describes diffusive and advective heat transfer including radioactive heat
126 production in the sediments.

$$127 \quad \left[\phi \rho_f c_{pf} + (1 - \phi) \rho_r c_{peff} \right] \frac{\partial T}{\partial t} = -\rho_f c_{pf} \vec{v}_f \cdot \nabla T + \nabla \cdot (\kappa \nabla T) + \lambda \quad (\text{Eqn. 2})$$

128

3.2 Thermal Maturation and Methane Generation

We use two different methods to calculate the amount of methane generated during thermogenic breakdown of organic material in sediments. The previously used, first method is generic and is based on a simple conversion of thermal maturity to methane generation (Aarnes et al., 2010; Iyer et al., 2013). Except for TOC content, this method does not account for any other sediment property and assumes a maximum of 85% TOC conversion to hydrocarbons. The second, more robust method is widely used within the petroleum industry to determine hydrocarbon generation and is based on the organofacies concept (Pepper and Corvi, 1995a, b; Pepper and Dodd, 1995). This method takes the depositional environment of the sedimentary layer into account and accordingly categorizes the kerogen that constitutes TOC available for hydrocarbon generation.

3.2.1 Generic Method

Vitrinite reflectance is a widely used indicator of thermal maturity and can be readily measured in the field. One of the most common methods used to calculate the synthetic thermal maturity of the source rock is the EASY%Ro method put forward by Sweeney and Burnham (1990). This model uses 20 parallel Arrhenius-type first order reactions to describe the process of kerogen breakdown due to temperature increase and the reacted fraction of kerogen (F) and corresponding vitrinite reflectance (%Ro) can be determined.

The amount of TOC that has reacted can be calculated by

$$\text{TOC} = \text{TOC}_o F \quad (\text{Eqn. 3})$$

and the rate of organic matter degradation R_{om} by

$$R_{om} = (1 - \phi) \rho_r \frac{\Delta \text{TOC}}{\Delta t} \quad (\text{Eqn. 4})$$

We assume that all of the hydrocarbons released are converted into methane if the maturity of the affected rock is within the gas window (%Ro>1). The amount of methane generated (R_{CH4}) for a time step affects the fluid pressure and is given by

$$R_{CH4} = R_{om} m_{CH4} \quad (\text{Eqn. 5})$$

where m_{ch4} is a stoichiometric conversion factor to transform carbon into methane.

3.2.2 Kerogen-based Method

This method uses kerogen kinetic parameters based on the gross depositional environment and stratigraphic age of the sedimentary layer (Pepper and Corvi, 1995a). Five kerogen organofacies are characterized: A – aquatic, marine, siliceous or carbonate/evaporite of any age; B – aquatic, marine, siliciclastic of any age; C – aquatic, non-marine, lacustrine of Phanerozoic age; D/E – terrigenous, non-marine, ever-wet, coastal for Mesozoic and younger age; and F – terrigenous, non-marine, coastal of late Paleozoic and younger age. Hydrocarbons are divided into initial petroleum present in the rock defined by the S1 values from pyrolysis and the amount of total kerogen present. The total amount of kerogen can be subdivided into reactive and inert kerogen. Inert kerogen does not produce any petroleum products. The amount of reactive kerogen is given by the S2 pyrolysis yield and is known as the hydrogen index (HI) when normalized to the TOC content. The reactive kerogen can be further subdivided into oil- and gas-producing kerogen based on the ratio (GOGI) or fraction (G^0) of the $C_{1.5}$ and C_{6+} products. Kinetic parameters such as the activation energies and frequency factors for the reactions are different for oil and gas kerogen fractions. The amount and rate of reaction occurring during thermal breakdown of kerogen is described by an Arrhenius-type of equation. Secondary cracking of generated oil to gas takes place as temperatures reach higher values and is described by another set of activation

energies and frequency factors (Pepper and Dodd, 1995). Generated oil and gas are adsorbed onto the remaining kerogen present in the source rock and expulsion of these products occurs once a maximum threshold is reached (Pepper and Corvi, 1995b). The thresholds for oil and gas expulsion are 100 and 20 mg/g kerogen, respectively. The models in this study assume that all of the expelled gas from kerogen breakdown is methane as it makes up the bulk of the gas released during this process (Tissot and Welte, 1984). Gas generated by this method (R_{CH4}) is also included in the fluid pressure equation.

3.3 Methane Transport

The model tracks fluid pathways in the domain and calculates the flow of methane out of the system at the surface. The single-phase model uses a simplified approach to track the release of methane: (1) methane generated moves with the fluid velocity and (2) methane and fluid within the pore space exist as a single, mixed phase. Methane advection is mass-conserving and is implemented using the Finite Volume Method (FVM) (Geiger et al., 2004).

$$\frac{\partial M_{CH4}}{\partial t} = -\nabla \cdot (M_{CH4} \vec{v}_f) + R_{CH4} \quad (\text{Eqn. 6})$$

where M_{CH4} is the mass of methane per unit volume. Increased buoyancy of the fluid due to methane content is accounted for by linearly varying the density of methane-bearing fluid. The density of fluids containing 120 kg/m³ CH₄ or more is equivalent to 10% of the density of pure water at similar conditions (Iyer et al., 2013).

3.4 Fracture and Vent Formation

Vent complexes have been found in sedimentary basins associated with LIPs globally (Gay et al., 2012; Jamtveit et al., 2004; Planke et al., 2005; Svensen et al., 2006; Svensen et al., 2003; Svensen et al., 2009). 2D and 3D seismic imaging reveal a spatial relation between sill intrusions at depth and hydrothermal vents (Planke et al., 2005). These vents act as conduits for fluid and gas phases generated within the contact aureole of the associated sill intrusion (Planke et al., 2005). Explosive formation of vents may have occurred due to rapid buildup of overpressure as water is superheated near the hot sill and/or due to rapid generation and expulsion of hydrocarbon gases from thermally degraded sediments within the contact aureole (Aarnes et al., 2010; Jamtveit et al., 2004; Monreal et al., 2009; Planke et al., 2005; Svensen et al., 2006). Fracture formation in the model occurs once the fluid pressure, driven by gas generation, exceeds the lithostatic pressure. The permeability in this region is increased based on the overstep of the fluid pressure with respect to the lithostatic

$$k = k \left(\frac{P}{P_l} \right)^2 \quad (\text{Eqn. 7})$$

The increase in permeability is capped to a maximum value of 10^{-12} m^2 , comparable to the permeability of porous sandstones and in hydrothermal vent structures (Wilcock and McNabb, 1996). Rock permeability is returned to the initial value as fluid pressure drops using the same function.

4 Model Setup

The numerical model is based on the MA-I04 seismic section from the Harstad Basin, offshore northern Norway. The section has been flattened along the paleosurface corresponding to the timing of sill emplacement at 55 Ma just below the Eocene Unconformity. The modelled section is 50 km wide and

reaches a depth of 9 km. The model is divided into three sedimentary packages based on seismic interpretation which from top to bottom are: (1) Tertiary, (2) Upper Cretaceous and (3) Lower Cretaceous (Figure 2). A number of sills are emplaced within the Lower Cretaceous sedimentary layer. Three different sills are emplaced such that they form a large main sill that extends horizontally for about 35 km. Since there are no wells in this frontier region, initial TOC values are obtained from the nearby 7119/12-3 well which is northeast of the modelled section. The average TOC values obtained from well 7119/12-3 for the Tertiary, Upper and Lower Cretaceous sections are 0.65, 0.66 and 1.4 wt%, respectively (Figure 3A). These strata are lithologically similar (silty mudstones) and, therefore, various rock properties such as density, conductivity, etc. are assumed to be constant throughout the sedimentary section.

An exponential curve fitted to the sonic log derived porosity gives the relation of porosity in the sedimentary strata with depth (Figure 3B).

$$\phi = 0.4464 \exp(0.2353Z) \quad (\text{Eqn. 8})$$

Rock permeability is related to porosity and is described by a simplified Kozeny-Carman equation (Costa, 2006),

$$k = k_0 \left[\frac{\phi^{n+1}}{(1-\phi)^n} \right] \quad (\text{Eqn. 9})$$

where k_0 is the reference permeability.

The bottom boundary has a heat flux boundary condition that results in 35°C/km geotherm once temperature steady-state is reached in line with the measured vitrinite reflectance in the region (Figure 3C). Fluids can enter and exit the domain freely through the top boundary and no flow conditions are applied to the sides and the bottom. The top surface is kept at a fixed pressure of 22 MPa to suppress

phase changes so that overpressure generation in the model is restricted to gas generation by kerogen breakdown only. The sills are instantaneously emplaced at a temperature of 1150°C. The sill porosity and permeability are kept constant through time with values of 10^{-3} and 10^{-20} m^2 , respectively. Prior to sill intrusion, the model is run to steady-state temperature in order to equilibrate vitrinite reflectance and thermal degradation of organic matter in the sediments due to diagenesis. The models are run for a period of 10,000 years after sill emplacement corresponding to the time over which the CIE during the PETM is observed.

5 Results

5.1 Reference Simulations

The first set of simulations investigates the effect of permeability on the thermal structure and maturation of organic content within the host sediments. Fracturing and methane transport using the FVM model are not implemented in these simulations. The reference permeability, k_0 , is varied between 10^{-13} and 10^{-17} m^2 for the different models.

Models with a reference permeability less than or equal to 10^{-15} m^2 return a temperature solution that is conductive in nature (Figure 4). The low permeability structure suppresses vigorous fluid flow and advective heat transport is negligible. As a result the thermal aureole and kerogen maturation is largely symmetrical around the sills. Increased maturation occurs in regions where the thermal aureoles of multiple sills interact with each other. Models with a reference permeability greater than 10^{-14} m^2 result in vigorous convective flow which significantly impacts the thermal structure due to advective heat transport. Early plumes form at the tips of various sills after 150 years due to merging of hot fluids migrating from the upper and lower parts of these sills. Due to increased permeability, multiple

convection cells are also formed at the top of the sills after a sufficiently thick thermal boundary layer is formed. These cells merge with each other as they grow in height after around 1000 years with the largest plumes sustained at sill tips and local topographical highs (Figure 5). Permeability structures with a reference permeability between 10^{-15} and 10^{-14} m² result in a system that is transitional between conductive and advective dominated processes. The thermal aureoles in these models are largely conductive around the sill intrusion but show minor effects of advective heat transport in regions where fluid focusing occurs, i.e. sill tips and local topographic highs. The rate of methane generation increases drastically within the first 10 years as sediments directly adjacent to the sills are heated in the gas window. This rate gradually decreases as the thermal aureole increases in size and sediments further away are heated. The total amount of methane generated increases slightly from 405 to 460 Kt for models with reference permeabilities of 10^{-17} and 10^{-13} m², respectively, assuming the model domain extent is 1 m in the third dimension (all values for methane generation and venting given later make this assumption unless specified otherwise) (Table 2). This small variation is notable in spite of very different final thermal structures for the conductive and advective models. These results are similar to those reported by Iyer et al. (2013). However, they do not conform to observations in the Harstad Basin where only a single hydrothermal vent is detected at the main sill tip.

5.2 Vent Formation

None of the reference simulations tested in the previous section are successful in recreating the single vent observed at the main sill tip suggesting that some other process caused the formation of the vent structure. Building on the reference models, the models in this section also include fracture formation when fluid pressure exceeds the lithostatic (see section 3.4).

277 The reference permeability, k_0 , is varied between 10^{-13} and 10^{-17} m². Significant overpressure does not
278 develop in hydrothermally driven convection models with a reference permeability greater than or
279 equal to 10^{-13} m² as the generated overpressure is quickly dissipated within the host rock. The resulting
280 system is therefore similar to the high permeability models ($k_0 > 10^{-14}$ m²) in the previous section where
281 fracture formation is not implemented. Sustained overpressure from gas generation and, thereby,
282 fracture formation is observed only in models where the reference permeability is lower than or equal
283 to 10^{-14} m². This inverse relationship of permeability to the sustainability of overpressure and vent
284 formation is similar to findings from previous studies (Aarnes et al., 2012; Jamtveit et al., 2004).
285 Overpressure is produced around sills where large volumes of gas are generated by thermal cracking of
286 organic matter in the host sediments. The overpressure produced above the sills is dissipated upwards
287 in most regions except for the region between sills and near the left boundary as it cannot escape either
288 upwards through the impermeable sill or through the left boundary. Overpressure generated below the
289 sill cannot dissipate upwards through the sill and, therefore, migrates along the sill until the tip is
290 reached. This migration happens in waves and is due to the implemented fracture mechanism.
291 Overpressure is reduced where fractures are generated which in turn drives fluid flow into this region
292 resulting in increased fluid pressures and continued fracturing. This mechanism, therefore, self-
293 generates fluid pathways until the overpressure wave reaches the surface and leaves the system. This
294 process restarts once sufficient overpressure is built up again due to kerogen breakdown resulting in the
295 observed waves. The main sill tip acts as the main focusing point for the vast amounts of overpressure
296 generated over the large area under the main sill and results in the formation of the vent structure
297 there. The width of the vent formed at the main sill tip is approx. 1 km. The sill tips in other areas only
298 access small areas of overpressure and therefore do not form large, sustained vent structures. Very
299 small, transitional plumes are also formed close to the sill boundaries where overpressure generates
300 increased permeability. The model with a reference permeability of 10^{-15} m² recreates the single,

sustained vent structure observed at the main sill tip for the case study which forms within few tens of years of sill emplacement and is used as a base model for further simulations (Figure 6).

5.3 Effect of TOC Content

The model presented in section 5.2 successfully recreates the observed vent at the main sill tip. We use it in this section to investigate the effect of TOC content on the amount of methane generated within the contact aureole and its subsequent release at the surface. Transport of generated methane occurs with the fluid velocity using the FVM model. Four models are presented here where the TOC content of the Lower Cretaceous sediments is varied between 1.4, 2.5, 3.5 and 5 wt% TOC. The amount of methane generated by thermal breakdown in the sill aureole using the generic method increases linearly with increasing TOC content and ranges between 680 and 2670 Kt (Table 2). The value for the model with 1.4 wt% TOC is much higher than value for the reference simulation with the same setup except fracture generation. This is because intense fracturing quickly (within the first 30 years) drives hot fluid towards the edge of the thermal aureole, thereby rapidly expanding it. Thus, energy that would be otherwise lost in heating sediment with dead carbon as the aureole slowly expands heats sediment that is colder and further away liberating more methane. This is also reflected in the rate of methane generation that shows a distinct second peak after the initial thermal shock that corresponds to fracture generation and the outwards flushing of hot fluids. The methane transported through the vent and released at the surface increases non-linearly with increasing TOC content ranging from 70 to 420 Kt. The amount of methane vented relative to that generated in the system increases slightly from 10 to 16% for increasing TOC (Figure 7). Rapid methane release at the surface takes places within the first few hundred years. Vent formation and initiation of methane degassing occurs within the first 40 years for all models (Figure 7). In the model with 1.4 wt% TOC content, most of the violent degassing occurs between 40 and 100

years after sill emplacement during which approximately 50% of the total vented methane is lost. The rest of the methane is released at a much slower rate into the atmosphere through the rest of the 10,000 years of simulated time. Models with higher TOC contents show similar trends with half of the total methane vented lost in initial bursts while the rest is slowly vented at the surface. The time over which these initial bursts last increases to 200 years for TOC contents of 3 and 5 wt% as a larger amount of generated methane migrates towards the vent and is outgassed at the surface. The number of bursts in the initial phase increases with increasing TOC content and is due to the number of overpressure waves generated by gas released during kerogen breakdown. Increased gas generation corresponding to higher TOC content results in increased overpressure over a larger period of time. This results in an overall increase in the number of overpressure waves in the system. The time between methane outbursts at the surface is dependent on the time taken for overpressure waves to be generated, travel along the bottom of the sill and through the vent to subsequently degas at the surface.

5.4 Effects of Kerogen-type

The models presented in the previous section use the generic method to determine the potential amount of methane that can be generated by thermal degradation of organic matter. This method assumes a maximum transformation of 85% of the total TOC content to hydrocarbons during the heating process while the remaining 15% is inert and does not contribute to hydrocarbon production. However, the yield of hydrocarbons from the organic content present in the sedimentary layer is closely linked to the prevailing conditions during deposition of sediments (Pepper and Corvi, 1995a). In order to further constrain methane generation linked to sill emplacement, the models are adjusted such that the kerogen-type is accounted for in this section using the method outlined in section 3.2.2. The rest of the parameters are the same as the base model used in the previous section with TOC content of 1.4 wt%.

The oil phase in this system is not transported away from the contact aureole with the fluid and undergoes secondary cracking. The models, therefore, predict the maximum amount of methane that can be generated for a given kerogen type under the prevailing thermal conditions.

The methane generation potential in the system depends directly on the HI and TI of the kerogen, i.e. the amount of reactive kerogen and initial crackable oil present in the system. The total amount of methane generated in the models ranges from 100 to 360 Kt for organofacies F and B, respectively, for a TOC content of 1.4 wt% (Table 2). These values are 53 and 15% of the value obtained from the same model when the EASY%Ro method is used as a transformation ratio of 85% in the EASY%Ro method equates to an HI of approximately 977 mgHC gTOC⁻¹. Hence, the organofacies models return a much lower methane generation potential than the models based on the generic EASY%Ro method (Figure 8). The total amount of the methane vented for the models with different organofacies ranges from 4.5 to 30 Kt for the F and B-types, respectively. The total amount of methane generated and vented for the A and C organofacies is similar to the model with B organofacies as the HI values of the kerogen are similar, i.e. similar amounts of reactive kerogen to generate hydrocarbons.

6 Discussion

6.1 Vent Formation and Methane Venting

The reference simulations described in section 5.1 show that sill models can be largely grouped into two types based on their thermal evolution. The first type results in a conductive thermal aureole around the sill intrusion and is characteristic of low permeability systems while the second is dominated by advective heat transport driven by hydrothermal fluid circulation. Both of these systems have been

previously studied with respect to the effects of sill intrusions in volcanic basins (Aarnes et al., 2010; Galushkin, 1997; Iyer et al., 2013; Wang et al., 2013).

Our study investigates both of these systems while also implementing rapid vent formation driven by fracture generation. We use the Harstad Basin case-study as an analogue for a typical sill complex with a single vent structure. Overpressure is generated by gas release during the thermogenic breakdown of organic matter in the thermal aureole around the sill. This overpressure is rapidly dissipated and cannot be sustained in systems with high permeability. In low permeability systems, this overpressure does not readily dissipate and results in sustained rock fracturing. Overpressure generated above sills migrates quickly upwards towards the surface and is depleted if no other barrier exists. Overpressure generated below the impermeable sill travels towards the sill tips and subsequently migrates upwards. The generated overpressure travels in a wave-like pattern due to the fracture mechanism which creates its own permeability for fluid and pressure to flow into resulting in increased pressure before dissipation. Fluid and overpressure focusing at sill tips results in vent formation there in few tens of years. The vertical extent of the vent formed at the sill tips depends on the amount of overpressure and fluid migrating towards it. As a result, a single, sustained vent is formed only at the main sill tip in the case study from the Harstad Basin due to access to the large amount of overpressure generated in the entire area below the main sill. Only a fraction of the methane released by kerogen breakdown, between 10 and 16%, is released through a single vent over a period of 10,000 years. Violent methane degassing through the vent initiates as soon as 30 years after sill emplacement and lasts over the first couple of hundred years. Bursts of methane coinciding with overpressure waves are vented at the surface during this initial period. The number of methane outbursts increases with increasing TOC content as the amount of generated overpressure increases. The total amount of methane generated in the model ranges from 680 to 2670 Kt for TOC content of 1.4 and 5 wt%, respectively, using the generic method to calculate methane release. This number is drastically reduced if the organofacies method is used as

typical amounts of reactive kerogen are lower. It ranges from 100 to 360 Kt for F- and B-type organofacies, respectively. The amount of vented methane also correspondingly decreases to 5 to 10% of the generated amount.

6.2 Climate Effects

Previous studies have shown that global warming, mass extinction and marine anoxia events are temporally linked to the formation of LIPs (Courtillot and Renne, 2003; Ganino and Arndt, 2009; Retallack and Jahren, 2008; Svensen et al., 2007; Svensen et al., 2004; Svensen et al., 2009; Wignall, 2001; Zhou et al., 2002). It has been suggested that the amount of gases such as methane generated in the contact aureoles around emplaced sills may be the driving force behind climate change (e.g. Aarnes et al. (2015); Svensen et al. (2004)). The Paleocene-Eocene Thermal Maximum (PETM) is temporally associated with the formation of the North Atlantic igneous province which contains the Vøring and Møre Basins. The area covered by sill complexes in these basins has been estimated to be around 85,000 km² (Svensen et al., 2004). In the current study, we determine the likely methane fluxes at the paleosurface in low permeability systems where vent formation at sill tips is the process that drives the out-flux of gases from the subsurface into the atmosphere. The modeled sill complex and associated vent is used as a template in sections 6.2.1 to 6.2.3 for upscaling gas generation and venting values using the two different methods outlined in section 3.2 thus providing a range for the Vøring and Møre Basins. A more realistic estimate that accounts for sediment properties in these basins is presented in section 6.2.4. It should be noted that the upscaled values assume that the sills were emplaced instantly within the entire basin. However, large uncertainties remain regarding the overall emplacement timing of large scale sill complexes which could span hundreds of thousands to millions of years (Jourdan et al., 2008; Magee et al., 2014; Svensen et al., 2012) further limiting the rate of methane degassing.

6.2.1 Basin-Scale Estimates of Methane Generated

The total sill length in our model is approximately 50 km when all of the sills in the complex are taken into account. The modeled amount of methane generated in sediment with a TOC content of 1.4 wt%, representative of the intruded Cretaceous sediments in the region, is 680 Kt using the generic method which assumes a maximum of 85% of TOC conversion. This value can be upscaled using the total modelled sill length, the modeled amount of generated methane and the area of the basin affected by sill complexes and results in a basin-scale methane generation potential of 1156 Gt (Table 2). The upscaled value decreases to between 170 and 612 Gt, depending on kerogen type, for a TOC content of 1.4 wt% when the organofacies method is used.

6.2.2 Basin-Scale Estimates of Methane Vented

A compilation of 734 mapped vents in the Vøring and Møre Basins shows that the average diameter of the various vent structures can vary from 1.5 to 7.5 km (Planke et al., 2005). The vents can be classified as small, intermediate and large based on their diameters (1.5, 3.5 and 7.5 km) and make up 62%, 35% and 3% of the total number of mapped vent complexes, respectively. The actual number of vents in these basins may, however, be as high as 2000 to 3000. The amount of methane fluxed out at the surface from the single vent in the models ranges from 70 to 420 Kt for TOC contents of 1.4 and 5 wt%, respectively, using the generic method for calculating methane generation. Therefore, the basin-scale amount of methane that may be potentially degassed from vent complexes can be determined by upscaling the values obtained from the Harstad Basin case study with a modeled vent width of 1 km and the vent complex vent width distribution for a total of 2000 vents. The total amount of methane that is vented within 10,000 years after sill emplacement adds up to 810 and 4860 Gt for TOC contents of 1.4 and 5 wt%, respectively. Note that the latter number exceeds the generated amount of methane in the basins for the same TOC content as the upscaling methods for the amount of methane generated and vented are different and independent from each other. This suggests that either the number of vents

chosen for upscaling is too high or that all of the methane may be potentially vented at the surface at higher TOC content. The upscaled amount of vented methane decreases dramatically to between 52 and 347 Gt when the organofacies method is used with a TOC content of 1.4 wt%.

6.2.3 Effect of Methane Venting on Carbon Isotope

The climate change events associated with LIP formation also coincide with negative carbon isotope excursions in the fossil record and may be explained by the release of massive quantities of $\delta^{13}\text{C}$ -depleted gases (Cohen et al., 2007; Dickens et al., 1997; Higgins and Schrag, 2006; Pagani et al., 2006; Retallack and Jahren, 2008). The PETM is characterized by a $\delta^{13}\text{C}$ incursion of -2 to -3‰ over 10,000 years. Thermogenic methane produced by the cracking of organic matter in sediments has a depleted carbon isotope ratio ($\delta^{13}\text{C} = -35$ to -50‰) with the average tending towards -35‰ (Higgins and Schrag, 2006; Svensen et al., 2004). Using the average carbon isotope ratio, the method proposed by Dickens et al. (1995) and the results from this study, the negative CIE during the PETM due to methane vented on a basin-scale is between -0.75 and -3.24‰ for TOC contents of 1.4 and 5 wt%, respectively, the latter of which corresponds to that observed in the fossil record. Kerogen-based calculations result in a much lower CIE between -0.05 and -0.33‰ for TOC content of 1.4 wt%.

6.2.4 Realistic Estimates for the Vøring and Møre Basins

The amount of methane that may be potentially generated in the Vøring and Møre Basins has been previously estimated at between 300 and 3500 Gt (!!! INVALID CITATION !!! (Aarnes et al., 2010; Svensen et al., 2004)). The Cretaceous sediments intruded by sill complexes in the Vøring and Møre Basins are dominantly composed of low TOC content (Aarnes et al., 2015) similar to that found in the 7119/12-3 well within our study area. Therefore, accounting for a more realistic scenario that includes a B-type kerogen, an average HI value of $200 \text{ mgHC gTOC}^{-1}$ and TOC content of 1.4 wt% results in a much lower methane generation potential of 215 Gt for the Vøring and Møre Basins using the organofacies method. Correspondingly, upscaling the amount of methane vented from these basins results in only 70

Gt of methane being released into the atmosphere triggering a reduction in carbon isotope values of only 0.07‰. Based on the calculations made above, a Kimmeridge Clay-type (Jurassic) source rock with an average HI of 600 mgHC gTOC⁻¹ and average TOC content of 8 wt% needs to be affected by sill intrusions with all of the generated methane being vented into the atmosphere in order for an area of 85,000 km² alone to trigger a -3‰ decrease in carbon isotope levels. Such high average values are not typical to the Cretaceous or indeed, even the Jurassic, sedimentary layers intruded by sill complexes in the region (Kubala et al., 2003). Therefore, these findings suggest that it is unlikely that the CIE observed during the PETM is driven by thermogenic methane vented from the Vøring and Møre Basins alone.

7 Conclusions

The 2D FEM/FVM model presented here provides insight into the effect of sill intrusions on hydrocarbon maturation and migration in low permeability systems. Although 3D multiphase models with an accurate CH₄-H₂O EOS and realistic fracture mechanics will help to further constrain these results, our model highlights the factors controlling methane venting while also successfully recreating observations and allows for easy upscaling to the basin scale. Overpressure is rapidly generated by gas generation within the sill thermal aureole by thermogenic breakdown of kerogen. This overpressure is sustained only in relatively low permeability systems and generates fractures around the sills. Vents are formed at sill tips due to fluid focusing and overpressure migration along the bottom of the impermeable sill till it reaches the tip. These vents are small at most of the sill tips as they only access a relatively small area over which overpressure is generated. A large, sustained vent is produced only at the main sill tip as the amount of overpressure generated below the sill is large and requires a relatively longer period of time to migrate towards the sill tip (ca. 1000 years) before it dissipates. The model, therefore, predicts the formation of

an ‘explosive’ vent structure at the main sill tip in the investigated sill complex in the Harstad Basin. It shows that only a fraction of the methane generated at depth escapes through a single vent. Upscaling these values to the Vøring and Møre Basins and using realistic host-rock carbon and kerogen content demonstrates that the amount of methane vented from these basins alone cannot explain the CIE observed during the PETM. The future testing of links between thermogenic methane release and the PETM should incorporate other areas of the NAIP such as East Greenland and the Faroe Shetland Basin and focus on the source rock properties of the intruded sediments.

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8 References

- !!! INVALID CITATION !!! (Aarnes et al., 2010; Svensen et al., 2004).
- Aarnes, I., Planke, S., Trulsvik, M., Svensen, H., 2015. Contact metamorphism and thermogenic gas generation in the Vøring and Møre basins, offshore Norway, during the Paleocene–Eocene thermal maximum. *Journal of the Geological Society*, 588-598.
- Aarnes, I., Podladchikov, Y., Svensen, H., 2012. Devolatilization-induced pressure build-up: Implications for reaction front movement and breccia pipe formation. *Geofluids* 12, 265-279.
- Aarnes, I., Svensen, H., Connolly, J.A.D., Podladchikov, Y.Y., 2010. How contact metamorphism can trigger global climate changes: Modeling gas generation around igneous sills in sedimentary basins. *Geochimica Et Cosmochimica Acta* 74, 7179-7195.
- Cohen, A.S., Coe, A.L., Kemp, D.B., 2007. The Late Palaeocene Early Eocene and Toarcian (Early Jurassic) carbon isotope excursions: a comparison of their time scales, associated environmental changes, causes and consequences. *Journal of the Geological Society* 164, 1093-1108.
- Costa, A., 2006. Permeability-porosity relationship: A reexamination of the Kozeny-Carman equation based on a fractal pore-space geometry assumption. *Geophysical Research Letters* 33, L02318.
- Courtillot, V.E., Renne, P.R., 2003. On the ages of flood basalt events. *Comptes Rendus Geosciences* 335, 113-140.

516 Dickens, G.R., Castillo, M.M., Walker, J.C.G., 1997. A blast of gas in the latest Paleocene: Simulating first-
 517 order effects of massive dissociation of oceanic methane hydrate. *Geology* 25, 259-262.
 518 Dickens, G.R., O'Neil, J.R., Rea, D.K., Owen, R.M., 1995. Dissociation of Oceanic Methane Hydrate as a
 519 Cause of the Carbon Isotope Excursion at the End of the Paleocene. *Paleoceanography* 10, 965-971.
 520 Faleide, J.I., Tsikalas, F., Breivik, A.J., Mjelde, R., Ritzmann, O., Engen, O., Wilson, J., Eldholm, O., 2008.
 521 Structure and evolution of the continental margin off Norway and the Barents Sea. *Episodes* 31, 82-91.
 522 Galushkin, Y.I., 1997. Thermal effects of igneous intrusions on maturity of organic matter: A possible
 523 mechanism of intrusion. *Organic Geochemistry* 26, 645-658.
 524 Ganino, C., Arndt, N.T., 2009. Climate changes caused by degassing of sediments during the
 525 emplacement of large igneous provinces. *Geology* 37, 323-326.
 526 Gay, A., Mourgues, R., Berndt, C., Bureau, D., Planke, S., Laurent, D., Gautier, S., Lauer, C., Loggia, D.,
 527 2012. Anatomy of a fluid pipe in the Norway Basin: Initiation, propagation and 3D shape. *Marine*
 528 *Geology* 332, 75-88.
 529 Geiger, S., Roberts, S., Matthäi, S.K., Zoppou, C., Burri, A., 2004. Combining finite element and finite
 530 volume methods for efficient multiphase flow simulations in highly heterogeneous and structurally
 531 complex geologic media. *Geofluids* 4, 284-299.
 532 Grove, C., 2013. Submarine hydrothermal vent complexes in the Paleocene of the Faroe-Shetland Basin:
 533 Insights from three-dimensional seismic and petrographical data. *Geology* 41, 71-74.
 534 Hansen, D.M., 2006. The morphology of intrusion-related vent structures and their implications for
 535 constraining the timing of intrusive events along the NE Atlantic margin. *Journal of the Geological*
 536 *Society* 163, 789-800.
 537 Henriksen, E., Bjørnseth, H.M., Hals, T.K., Heide, T., Kiryukhina, T., Kløvjan, O.S., Larssen, G.B., Ryseth,
 538 A.E., Rønning, K., Sollid, K., Stoupakova, A., 2011. Chapter 17 Uplift and erosion of the greater Barents
 539 Sea: impact on prospectivity and petroleum systems. *Geological Society, London, Memoirs* 35, 271-281.
 540 Higgins, J.A., Schrag, D.P., 2006. Beyond methane: Towards a theory for the Paleocene-Eocene Thermal
 541 Maximum. *Earth and Planetary Science Letters* 245, 523-537.
 542 Iyer, K., Rüpke, L., Galerne, C.Y., 2013. Modeling fluid flow in sedimentary basins with sill intrusions:
 543 Implications for hydrothermal venting and climate change. *Geochemistry, Geophysics, Geosystems* 14,
 544 5244-5262.
 545 Jamtveit, B., Svensen, H., Podladchikov, Y.Y., Planke, S., 2004. Hydrothermal vent complexes associated
 546 with sill intrusions in sedimentary basins, in: Breitzkreuz, C., Petford, N. (Eds.), *Physical Geology of High-*
 547 *Level Magmatic Systems*. Geological Soc Publishing House, Bath, pp. 233-241.
 548 Jourdan, F., Féraud, G., Bertrand, H., Watkeys, M.K., Renne, P.R., 2008. The ⁴⁰Ar/³⁹Ar ages of the sill
 549 complex of the Karoo large igneous province: Implications for the Pliensbachian-Toarcian climate
 550 change. *Geochemistry, Geophysics, Geosystems* 9, Q06009.
 551 Kubala, M., Bastow, M., Thompson, S., Scotchman, I., Oygard, K., 2003. Geothermal regime, petroleum
 552 generation and migration, in: Evans, D., Graham, C., Armour, A., Bathurst, P. (Eds.), *The Millennium*
 553 *Atlas: Petroleum geology of the central and northern North Sea*. The Geological Society of London, pp.
 554 289-315.
 555 Magee, C., Jackson, C.A.L., Schofield, N., 2014. Diachronous sub-volcanic intrusion along deep-water
 556 margins: insights from the Irish Rockall Basin. *Basin Research* 26, 85-105.
 557 Monreal, F.R., Villar, H.J., Baudino, R., Delpino, D., Zencich, S., 2009. Modeling an atypical petroleum
 558 system: A case study of hydrocarbon generation, migration and accumulation related to igneous
 559 intrusions in the Neuquen Basin, Argentina. *Marine and Petroleum Geology* 26, 590-605.
 560 Olesen, O., Lundin, E., Nordgulen, O., Osmundsen, P.T., Skilbrei, J.R., Smethurst, M.A., Solli, A., Bugge, T.,
 561 Fichler, C., 2002. Bridging the gap between the onshore and offshore geology in Nordland, northern
 562 Norway. *Norsk geologisk tidsskrift* 82, 243-262.

563 Pagani, M., Caldeira, K., Archer, D., Zachos, J.C., 2006. An ancient carbon mystery. *Science* 314, 1556-
 564 1557.
 565 Pepper, A.S., Corvi, P.J., 1995a. Simple kinetic models of petroleum formation. Part I: oil and gas
 566 generation from kerogen. *Marine and Petroleum Geology* 12, 291-319.
 567 Pepper, A.S., Corvi, P.J., 1995b. Simple kinetic models of petroleum formation. Part III: Modelling an
 568 open system. *Marine and Petroleum Geology* 12, 417-452.
 569 Pepper, A.S., Dodd, T.A., 1995. Simple kinetic models of petroleum formation. Part II: oil-gas cracking.
 570 *Marine and Petroleum Geology* 12, 321-340.
 571 Planke, S., Rasmussen, T., Rey, S.S., Myklebust, R., 2005. Seismic characteristics and distribution of
 572 volcanic intrusions and hydrothermal vent complexes in the Vøring and Møre basins, in: Doré, A.G.,
 573 Vining, B.A. (Eds.), *Petroleum Geology: North-western Europe and global perspectives - Proceedings of*
 574 *the 6th Petroleum Geology Conference*. Geological Society, London.
 575 Planke, S., Svensen, H., Myklebust, R., Bannister, S., Manton, B., Lorenz, L., 2014. Geophysics and
 576 Remote Sensing, *Advances in Volcanology*. Springer Berlin Heidelberg, Berlin, Heidelberg, pp. 1-16.
 577 Retallack, Gregory J., Jahren, A.H., 2008. Methane Release from Igneous Intrusion of Coal during Late
 578 Permian Extinction Events. *The Journal of Geology* 116, 1-20.
 579 Svensen, H., Corfu, F., Polteau, S., Hammer, O., Planke, S., 2012. Rapid magma emplacement in the
 580 Karoo Large Igneous Province. *Earth and Planetary Science Letters* 325, 1-9.
 581 Svensen, H., Jamtveit, B., Planke, S., Chevallier, L., 2006. Structure and evolution of hydrothermal vent
 582 complexes in the Karoo Basin, South Africa. *Journal of the Geological Society* 163, 671-682.
 583 Svensen, H., Planke, S., Chevallier, L., Malthé-Sørensen, A., Corfu, F., Jamtveit, B., 2007. Hydrothermal
 584 venting of greenhouse gases triggering Early Jurassic global warming. *Earth and Planetary Science*
 585 *Letters* 256, 554-566.
 586 Svensen, H., Planke, S., Corfu, F., 2010. Zircon dating ties NE Atlantic sill emplacement to initial Eocene
 587 global warming. *Journal of the Geological Society* 167, 433-436.
 588 Svensen, H., Planke, S., Jamtveit, B., Pedersen, T., 2003. Seep carbonate formation controlled by
 589 hydrothermal vent complexes: a case study from the Voring Basin, the Norwegian Sea. *Geo-Marine*
 590 *Letters* 23, 351-358.
 591 Svensen, H., Planke, S., Malthé-Sørensen, A., Jamtveit, B., Myklebust, R., Rasmussen Eidem, T., Rey,
 592 S.S., 2004. Release of methane from a volcanic basin as a mechanism for initial Eocene global warming.
 593 *Nature* 429, 542-545.
 594 Svensen, H., Planke, S., Polozov, A.G., Schmidbauer, N., Corfu, F., Podladchikov, Y.Y., Jamtveit, B., 2009.
 595 Siberian gas venting and the end-Permian environmental crisis. *Earth and Planetary Science Letters* 277,
 596 490-500.
 597 Sweeney, J., Burnham, A.K., 1990. Evaluation of a simple model of vitrinite reflectance based on
 598 chemical kinetics. *AAPG Bulletin* 74, 1559-1570.
 599 Tissot, B.P., Welte, D.H., 1984. *Petroleum Formation and Occurrence*. Springer-Verlag, Berlin Heidelberg.
 600 Wang, D.Y., Song, Y.C., Xu, H.S., Ma, X.J., Zhao, M.L., 2013. Numerical modeling of thermal evolution in
 601 the contact aureole of a 0.9 m thick dolerite dike in the Jurassic siltstone section from Isle of Skye,
 602 Scotland. *Journal of Applied Geophysics* 89, 134-140.
 603 Wignall, P.B., 2001. Large igneous provinces and mass extinctions. *Earth-Science Reviews* 53, 1-33.
 604 Wilcock, W.S.D., McNabb, A., 1996. Estimates of crustal permeability on the endeavour segment of the
 605 Juan de Fuca mid-ocean ridge. *Earth and Planetary Science Letters* 138, 83-91.
 606 Zhou, M.-F., Malpas, J., Song, X.-Y., Robinson, P.T., Sun, M., Kennedy, A.K., Leshner, C.M., Keays, R.R.,
 607 2002. A temporal link between the Emeishan large igneous province (SW China) and the end-
 608 Guadalupian mass extinction. *Earth and Planetary Science Letters* 196, 113-122.

Symbol	Description	Units	Values
C_{peff}	Effective rock heat capacity	$\text{J kg}^{-1} \text{K}^{-1}$	
C_{pf}	Fluid heat capacity	$\text{J kg}^{-1} \text{K}^{-1}$	EOS
C_{pm}	Melt heat capacity	$\text{J kg}^{-1} \text{K}^{-1}$	820
C_{pr}	Rock heat capacity	$\text{J kg}^{-1} \text{K}^{-1}$	880
$\vec{g}$	Gravitational acceleration	m s^{-2}	9.81
k_0	Reference permeability	m^2	$10^{-12} - 10^{-17}$
k	Rock permeability	m^2	
L_c	Latent heat of crystallization	KJ kg^{-1}	376
M_{CH_4}	Mass of methane per unit volume	kg m^{-3}	
m_{CH_4}	Methane conversion factor		1.34
n	Kozeny-Carman factor		1.5
P	Fluid pressure	Pa	
P_l	Lithostatic pressure	Pa	
R_{CH_4}	Rate of methane generation	$\text{kg m}^{-3} \text{s}^{-1}$	
R_{om}	Rate of organic matter degradation	$\text{kg m}^{-3} \text{s}^{-1}$	
t	Time	s	
T_L	Liquidus temperature	K	1423
T_S	Solidus temperature	K	1223
T	Temperature	K	
$\vec{v}_f$	Fluid velocity	m s^{-1}	
Z	Depth	km	
α_f	Fluid thermal expansion	K^{-1}	EOS
β_f	Fluid compressibility	Pa^{-1}	EOS
$\Delta t_{\log}$	Acoustic log	$\mu\text{s ft}^{-1}$	
Δt_{matrix}	Acoustic matrix	$\mu\text{s ft}^{-1}$	55.5
ϕ	Rock porosity		
ϕ_s	Sonic log derived porosity		
κ	Conductivity	$\text{W m}^{-1} \text{K}^{-1}$	2.1 (sills) 1.8 (sediments)
λ	Radioactive heat production	W m^{-3}	10^{-6} (sediments)
μ_f	Fluid viscosity	Pa s	EOS
ρ_f	Fluid density	kg m^{-3}	EOS
ρ_m	Melt density	kg m^{-3}	2600
ρ_r	Rock density	kg m^{-3}	2400

611 *Table 1. List of symbols and values used in the numerical models.*

Reference Simulations (m ²)	Methane Generated (Kt)	
10 ⁻¹⁷	405	
10 ⁻¹⁵	420	
10 ⁻¹⁴	430	
10 ⁻¹³	460	
TOC content (wt%) – Generic Method	Methane Generated Vented (Kt)	Basin-Scale Methane Generated Vented (Gt)
1.4	680 70	1156 810
2.5	1190 140	2023 1621
3.5	1710 240	2907 2778
5.0	2670 420	4540 4862
Organofacies Method (TOC 1.4 wt%)	Methane Generated Vented (Kt)	Basin-Scale Methane Generated Vented (Gt)
A	340 30	578 347
B	360 30	612 347
C	350 30	595 347
D/E	210 10	357 116
F	100 4.5	170 52

613 Table 2. Methane generated in the modeled domain and vented at the surface for varying TOC content
 614 (generic method of methane generation) and kerogen type (organofacies method of methane
 615 generation) in the Lower Cretaceous sediments. Note: the amount of methane generated in the models is
 616 calculated based on a 1m extension of the model in the third dimension. The values of generated and
 617 vented methane are also upscaled for the Vøring and Møre Basins (see text for details).

Figure 1. A. Location map showing the study area (3D survey) within the Harstad Basin, the main sill area (red) and the vent location (yellow dot). B. Seismic characteristics of the interpreted hydrothermal vent complex within the study area. C. Depth converted seismic stratigraphic interpretation across the model profile flattened on the Base Vent horizon.

624

625 *Figure 2. Modelled domain with sedimentary strata ranging in age from Tertiary to Lower Cretaceous.*

626 *Sills (red) intrude into the Lower Cretaceous sediments.*

627

628

629 Figure 3. Data from well 7119/12-3. A) Histogram plot with TOC data (NPD database). The average
 630 values of the different sedimentary layers are shown (dashed lines). B) Curve (Eqn. 8) fitted to the
 631 porosity values obtained from the sonic log. C) Fit of the computed vitrinite reflectance for a 35°C/km
 632 geotherm with data from well 7119/12-3. Note that the well data depth has been adjusted for erosion of
 633 450m (Henriksen et al., 2011).

634

635 *Figure 4. Temperature and maturity of the reference simulation with a reference permeability of 10^{-15}*
 636 *m^2 . The low permeability structure inhibits fluid flow and energy transfer from the hot sill to cold*
 637 *sediments is largely conductive. The rate of methane generation increases drastically as the*
 638 *sediment directly adjacent to the sill is initially heated up. The rate then decreases as the thermal*
 639 *aureole gradually widens.*

640

641 *Figure 5. Temperature and maturity of the reference simulation with a reference permeability of 10^{-13} m^2 .*

642 *The high permeability structure induces vigorous fluid flow which results in a number of hydrothermal*

643 *plumes. The resulting thermal structure is considerably different from a conductive solution. The rate of*

644 *methane generation increases drastically as the sediment directly adjacent to the sill is initially heated*

645 *up. The rate then decreases as the thermal aureole gradually widens and hydrothermal plumes are*

646 *formed.*

647

648 Figure 6. Temperature, maturity and permeability plots at 93 years for the model with a reference
 649 permeability of 10^{-15} m^2 in section 5.2 shows a single, long-lived vent formed at the main sill tip due to
 650 fracture generation (red overlay in the permeability plot) by overpressure. This vent is sustained for a
 651 relatively long period of time (a few hundred years) and reaches the surface as it accesses a large region
 652 of overpressure developed under the main sill which migrates in bursts towards the sill tip. This results in
 653 increased temperatures and maturity in the vent due to enhanced fluid flow. Smaller vents are formed at
 654 the other sill tips but do not reach the surface as not enough overpressure is generated below them. Note
 655 that the color axes for the temperature and permeability plots have been limited for better visibility.

658 Figure 7. The rate of methane generated and subsequently vented at the surface through time for
 659 models with varying lower Cretaceous TOC content. The arrow on the right axis shows the value of the

corresponding curve after 10,000 years. Note: the amount of methane is calculated based on a 1m extension of the model in the third dimension. Also note that not all of the peaks denoting methane venting (blue curve) may be resolved in the figure as the time resolution of the plot data is larger than that during which bursts may occur. This does not affect the amounts calculated for generated and vented methane.

Figure 8. Methane generated and vented at the surface for different kerogen organofacies compared to the model using the generic method. Note: the amount of methane is calculated based on a 1m extension of the model in the third dimension.