The Plio-Pleistocene seepage history off western Svalbard inferred from 3D petroleum systems modelling

Matthias Daszinnies[1,*], Andreia Plaza-Faverola[2], Øyvind Sylta[1], Stefan Bünz[2], Rune Mattingsdal[3], Are Tømmerås[1,] Jochen Knies[2,4]

[1]*Migris AS, NO-7462 Trondheim, Norway*

[2]*CAGE - Centre for Arctic Gas Hydrate, Environment and Climate; Department of Geosciences, UiT The Arctic University of Norway, NO-9037 Tromsø, Norway*

[3]*Norwegian Petroleum Directorate, NO-9488 Harstad, Norway*

[4]*Geological Survey of Norway, NO-7491 Trondheim, Norway*

[*]Corresponding author. Migris AS, NO-7462 Trondheim, Norway.

E-mail address: matthias.daszinnies@migris.no (M.Daszinnies).

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Abstract

Methane seepage off the coast of Svalbard is demonstrated by active gas flares in the water column today and through precipitation of methane-derived authigenic carbonates (MDAC) over the past 160 000 years. Though submarine discharge of thermogenic methane is well documented, the geological history of past leakage is still debated, largely due to unconstrained free gas sources and seepage trigger mechanisms. We have assembled a high-resolution 3D petroleum systems model to evaluate the impact and charge of potential thermogenic hydrocarbons on the Vestnesa Ridge, NW Svalbard. We show that gaseous hydrocarbons, originating from Miocene age terrigenous organic matter, accumulates largely in ~2 million-year-old (Ma) sedimentary sequences underneath the pockmark system on the ridge's crest. These traps are constantly charged until present day. The supply of free gas to the gas hydrate stability zone (GHSZ) initiated the hydrate formation at ~3.0 Ma. We also show that gas leakage to the seafloor is governed by fault corruption of the hydrocarbon traps and not by excess pore fluid pressure. The onset of episodic seafloor seepage on Vestnesa Ridge can be associated with the first shelf edge glaciation of the Svalbard-Barents Sea ice sheet (SBIS), ~1.5 Ma ago. The results of the modelled petroleum system are consistent with the notion that repeated forebulge uplift and subsidence, due to cyclic SBIS build-up and decay, can be advocated as a mechanism that repeatedly caused extensional fracturing of the eastern Vestnesa Ridge segment. This fault damage affected modelled gas accumulations and led to the formation of hydrocarbon migration pathways to the seafloor. Repeated tapping into hydrocarbon reservoirs due to (a) fracture formation promoted by glacial isostatic adjustments (GIA) and (b) fracture re-activation could explain

recent observations of multiple seepage events on Vestnesa Ridge during episodes of intense cooling over the past 160.000 years.

1 Introduction

The dynamics of Arctic methane seepage in the eastern Fram Strait has been linked to gas hydrate dissociation regulated by several mechanisms: long term and seasonal climatic changes (Westbrook et al., 2008; Berndt et al., 2014), retreat of the Svalbard-Barents Sea ice sheet (SBIS) (Portnov et al., 2016) and isostatic adjustment after the Last Glacial Maximum (LGM) (Schneider et al. 2018, Wallmann et al., 2018, Ferré et al., 2020). However, gas hydrates are not widespread in areas of abundant seepage near the shelf break; hence gas hydrate dissociation as sole explanation for seepage in the region remains elusive (Smith et al. 2014, Pape et al. 2019). Instead, a link between seepage sites, gas chimneys and near-surface faults from seismic data along the Vestnesa Ridge point towards a close relation between episodic seepage and fault dynamics (Plaza-Faverola et al., 2015).

While present day seepage distribution along the Vestnesa Ridge may be controlled by fault dilation under a tensile tectonic stress regime due to spreading at the Molloy and Knipovich ridges (Plaza-Faverola and Keiding, 2019), historical seepage periodicity through faults was possibly influenced by glacio-isostatic adjustments (GIA) of the SBIS (Schneider et al., 2018; Plaza-Faverola et al., 2019; Himmler et al., 2019). It has been hypothesized that since intensification of the Northern Hemisphere glaciation, ~2.7 million years ago, episodic release of Arctic methane at the seabed has increased (Plaza-Faverola et al. 2015); a fact that is corroborated by recent fluid migration modelling results in the Arctic-Atlantic gateway (Knies et al. 2018). On younger time scale, Schneider et al. (2018) recently summarized episodic seepage events in the gateway region during build-up and decay of the

SBIS during the LGM implying a close relationship between GIA and fault-reactivation to open, highly permeable conduits and enhanced methane flux. Recently, Himmler et al. (2019) suggested that frequent methane release events in the region have occurred since the Penultimate Glacial Maximum (PGM), about 0.160 Ma ago. Also, these events correlate to ice sheet dynamics of the SBIS. The geochronology of these events indicates that some episodes of methane release coincide with dramatic climatic shifts in the Northern Hemisphere expressed by so-called Heinrich Events (Hemming, 2004) – thick layers of ice-rafted debris deposited in the North Atlantic during glacial times because of massive discharges of icebergs from the Laurentide ice sheet – implying a causal relationship between large-scale ice sheet dynamics and methane release in the Arctic.

Although large uncertainties exist on the extent and volume of circum-Arctic ice sheet beyond the last glacial period (Patton et al. 2015, 2016), it is hypothesized here, that the source of seeping methane off western Svalbard is of pre-glacial origin (Miocene) and the transfer of methane into the seepage system may have been triggered and facilitated by various mechanisms during the Plio-Pleistocene. These mechanisms comprise (1) excess fluid pressure build-up, (2) hydrocarbon migration along large scale fault surfaces, (3) persistent upward hydrocarbon migration across the stratigraphic inventory, (4) hydrocarbon leakage loss from shallow gas accumulations or (5) various combinations of them. We are testing our hypothesis by applying an innovative 3D hydrocarbon migration modelling approach on one of the most active methane seep sites in the Arctic, the Vestnesa Ridge, NW Svalbard.

2 Geological setting

The Vestnesa Ridge is a ~100 km long, boomerang-shaped and pockmark covered submarine feature offshore western Svalbard (Fig. 1). It hosts sediments of >5 km in thickness and rest on top of hot (heat flow >115mW/m^2) and relatively young oceanic crust (<20 Ma), near the ultraslow spreading Molloy and the Knipovich ridges in the Fram Strait (Eiken and Hinz, 1993; Vogt et al., 1994; Hustoft et al., 2009 and Bünz et al., 2012). Slow spreading of the Molloy and Knipovich ridges led to the onset of the Fram Strait opening about 33 Ma ago (Engen et al., 2008). Deep marine conditions and water mass circulations were established in the Fram Strait in the Miocene, ca. 17 – 10 Ma ago (Jakobsson et al., 2007; Ehlers and Jokat, 2009), and shaped the environmental conditions for the evolution of bottom current-driven sedimentary drifts (Eiken and Hinz, 1993; Johnson et al., 2015).

Three main stratigraphic units (YP1-YP3) are designated according to their correlation with Ocean Drilling Program (ODP) sites at the Yermak Plateau (Eiken and Hinz, 1993, Mattingsdal et al. 2014). Miocene – Pliocene age syn and post-rift deposits overlying the oceanic crust (YP1) are superseded by contourites (YP2). By the onset of the Pleistocene around 2.7 Ma the contourites became intermixed with glaciogenic debris flow deposits sourced from the East (YP3) (Howe et al., 2008).

Restricted to the youngest stratigraphic sequence (YP3), a gas hydrate system underlain by a free-gas zone occurs along-strike the entire Vestnesa Ridge. This system is permeated by a series of fault bound gas chimneys which terminate in pockmarks at the seafloor (Hustoft et al., 2009; Petersen et al., 2010; Plaza-Faverola et al., 2015, Panieri et al., 2017 and Plaza-Faverola and Keiding, 2019). Active methane discharge from pockmarks into the water column is recorded on the eastern ridge segment today (Bünz et al., 2012; Smith et al., 2014). Multiple proxies such as radiogenic and stable isotopes, foraminifera and methane-derived authigenic carbonate (MDAC) crusts in the geological record indicate episodic seepage over the past 160 ka (e.g. Ambrose et al., 2015, Panieri et al.,2017, Sztybor and Rasmussen, 2017, Schneider et al., 2018, Himmler et al., 2019). However, seismic data showing seepage features such as gas chimneys and buried pockmarks seem to indicate that methane release has been prevailed since the intensification of the Northern Hemisphere glaciation, about 2.7 million years ago (Plaza-Faverola et al., 2015; 2019). Contrasting, no seepage activity is observed on the western ridge today though foraminiferal records indicate methane seepage activities about 14 – 13 and 10 ka ago (Consolaro et al., 2014). Isotopic $\delta^{2}H$ and $\delta^{13}C$ signatures in methane and ethane from dissolved gas and hydrates samples (Faverola et al., 2017, Pape et al., 2019, Sauer et al., 2021) collected at the Lunde and Lomvi seep sites (Fig. 1) evidence a deep thermogenic source of the hydrocarbons charging the hydrate and seep system from below the gas hydrate stability zone (GHSZ) at the eastern ridge segment.

3 Material and Methods

3.1 Petroleum systems model

Petroleum systems modelling (PSM) is a well-established method in hydrocarbon exploration and numerically reproduced processes of hydrocarbon generation, migration, trapping and leakage within sedimentary basins over geological times. We employ the Migri simulator (http://www.migris.no/technology), a fast commercial PSM software. Modelled properties are treated dynamically, meaning their values change through geologic time unless stated otherwise. Details on the methods for modelling the individual processes are described in Sylta (2004) and Sylta (2005). The vertical pore entry pressure distribution within the layers is evaluated to determine whether primary migration out of the source rocks should be directed upward or downward. Over time, entry pressure values in the grid elements are calculated from the permeabilities, which are derived from the lithological composition and burial depths. Secondary migration is modelled as mostly buoyancy-driven within permeable systems underneath flow barriers where the direction of flow is governed by the steepest gradient of ascent. Lateral entry pressure differences within the flow unit may, however, cause the flow to be deflected. Lateral migration may also be stopped or deflected when the flow encounters a shale pinch-out barrier with higher entry pressures than more porous strata. Migrating hydrocarbons are modelled to accumulate in traps. Gas and subsequently oil are modelled to leak out of the trap according to the entry pressure distributions modelled in the sealing rocks (e.g. in shales). Both, the trapped hydrocarbon column heights, and the seal permeabilities determine the rates and areas of leakage from traps.

The 3D petroleum systems model and associated initial boundary conditions are of the same derivation as previously used to model the methane seepage along the 2D seismic line across the Vestnesa Ridge (Knies et al. 2018) where a more complete description of the model setup and implementation can be found. Details on the key changes to expand the 2D into a 3D-PSA model (Fig. 1) are given in the enclosed supplements. Key petrophysical input parameters are given in Tab. 1. The main elements of this PSM comprise the basin configuration and stratigraphy, burial and thermal history models as well as models for lithological and organic matter inventory in the source rocks and are briefly summarized here.

3.1.1 Basin configuration and burial history model

Four main depth horizons and the seafloor were interpreted from several 2D seismic lines (CAGE-VR2D-005, CAGE-VR2D-006, SV 3-97 , Svalex P11, Svalex P12) and one 3D P-cable seismic volume (Plaza-Faverola et al., 2015 and references therein, Knies et al., 2018). These include: Top basement (assigned age of 300 Ma to avoid impact on the basin modelling), the 6.2 Ma regional biostratigraphic datum for the Nordic Seas and Arctic Ocean (Matthiessen et al., 2009) and the chronostratigraphic framework established for the Yermak Plateau/western Svalbard including the Matuyama/Gauss chron boundary (∼2.6 Ma) and the ~1.5 Ma old regional seismic reflector (Knies et al., 2009; Mattingsdal et al., 2014).

Details on the depth conversion are given in the enclosed supplements. Additional depth horizons were constructed using lithology-age-depth models of ODP Hole 909C (Winkler et al., 2002; Wolf-Welling et al., 1996) and Hole 986D (Jansen et al., 1996) with assigned ages to the model of 0.02 Ma, 0.043 Ma, 0.148 Ma, 2.1 Ma, 3.2 Ma, 4.39 Ma, 12.05 Ma, 15.18 Ma, 17.37 Ma, and 18.5 Ma (Fig. 2). Burial history reconstruction was performed using a back-stripping approach (Allen and Allen, 2005) with paleo-bathymetries inferred to follow present-day marine trends, assuming constant tectonically induced subsidence. Thicknesses and depth of units over time are derived from decompaction modelling using the concept of Sclater and Christie (1980). The permeabilities of the units are calculated dynamically over time and range from impermeable to 5000 mD. They are based on the lithologies and porosity estimates from the decompaction calculation (Tab. 2). The modelling is performed over 10 timesteps between 12.1 Ma and present-day (Tab. 2).

3.1.2 Thermal history model

The thermal history model infers the highest geothermal gradients at the Molloy Transform Fault (MTF) location and decreasing gradients away from the MTF in SW and NE directions (Vanneste et al., 2005). Data from Crane et al. (1992) and ODP sites 908, 909, 910 and 912 (Stein et al., 1995) were jointly used to interpolate geothermal gradient maps across the area of interest (AOI). Geothermal gradient values range from 55 °C/km to maximum 120 °C/km at the MTF (Tab. 1 and Fig. S1). Seafloor temperatures are set uniformly to -1°C (Plaza-Faverola et al. 2017). Inferred geothermal gradients agree with heat flow values reported by Hustoft et al. (2009) and Smith et al. (2014) west of Svalbard, considering thermal conductivities of silt and shale lithologies. High sedimentation rates during Pliocene/Pleistocene times at the Vestnesa Ridge exerted a cooling effect in the upper sedimentary layers (Plaza-Faverola et al., 2017). We note that the resultant transient changes in the thermal regime challenge the assumption of a constant geothermal gradient over time and, hence, impact the correct approximation of the timing of source rock maturation. To account for this, we apply the equation of Palumbo et al. (1999) to the gradient model and thereby better assess the timing of the source rock maturation due to rapid sedimentation. Modelled burial and thermal histories for four locations across the AOI (including vitrinite reflection (VR) calibration data at ODP Site 909) are shown in the enclosed supplements (Figs S2 and S3).

3.1.3 Lithological model

Sand-shale-carbonate ratios observed in ODP Hole 909C (Wolf-Welling et al., 1996) and Hole 986D (Butt et al., 2000; Knies et al., 2009) are combined into a composite lithology log whereby age and lithology information from ODP Hole 986D are projected onto the model's age-depth-lithology relation at the Vestnesa Ridge (Fig. 3). We expect that Pleistocene, glacially-derived sediments there are more coarse-grained compared to pre-glacial sediments (with age >2.7 Ma) and are comparable with the lithological composition of ODP Hole 986D (Butt et al., 2000). Consequently, lithostratigraphic information from ODP Hole 909C (for stratigraphic ages from 17.4 Ma to 2.6 Ma) and the projected ODP Hole 986D (for stratigraphic ages from 2.6 Ma to 0 Ma) were combined to populate the model with a laterally invariant sand-shale-carbonate ratio stratification, between the top basement and the seafloor reflectors. The entire lithological model is stratified according to borehole information and therefore accounts for vertically changing proportions of shale, sand, and carbonate (Fig. 3) Model porosities and permeabilities are dynamically derived from the lithology properties in the cells (Tab. 1) meaning that units with low shale and high sand contents are characterized by higher porosities and permeabilities compared to high shale content strata at a similar burial depth.

Table 1: Petrophysical input parameters to the petroleum systems model

Parameter/variable	Values/units/dependencies	Physical meaning	Reference
Spatial resolution	50 m x 50 m avg. 10 – 30 m, depends on $N_{sublayer}$.	Area represented by a grid cell in a 3D element. Height of a 3D element	
$N_{horizon}$	15 in total, 5 based on seismic interpretation, 10 constructed from seismic horizon- and well-data (ODP holes 909C and 986D)	Number of depths horizons employed in the layer model	
$N_{sublayer}$	2 – 20 per layer, depending on lithology variation and layer thickness	Number of discrete lithological subunits in geo-model layer between 2 depth horizons	
N_{time}	10 between 12.1 Ma and 0 Ma	Number of model time steps	
dT/dz	55 - 120 °C/km, laterally variable	Geothermal gradient, map in Fig. S1	Crane et al., 1992 Stein et al.,1995 Hustoft et al., 2009, Johnson et al., 2015, Smith et al., 2014,
V_{sh}, V_{ca}, V_{ss}	0 – 1, fraction	Fraction of shale, carbonate, and sand with $V_{sh}+V_{ca}+V_{ss} = 1$	
ϕ	0.0077 – 0.5780, fraction, depending on lithology and burial depth	Porosity of the modelled lithology	
K	$10^{-7} – 5 \cdot 10^{3}$ mD, depending on lithology and burial depth, cells in sealing gas hydrate zone = $1 \cdot 10^{-5}$ mD	Permeability of modelled lithology	
P_c	$2 \cdot 10^{-2} – 2.9 \cdot 10^{2}$ MPa, depending on hydrocarbon phase and K, cells in sealing gas hydrate zone = $2.1 \cdot 10^{1}$ MPa	Capillary entry pressure of pore space	

3.1.4 Source rock model

Two organically enriched and thermally immature siliciclastic intervals of middle Miocene age are encountered in ODP Hole 909C (Myhre et al., 1995). The organic matter is terrigenous nature and originated from the bordering basin flanks of AOI_{west} (Hovgaard Ridge) and AOI_{east} (Svalbard Platform) (Knies and Mann, 2002). Two source rock units with a type III kerogen were defined between the depth horizon intervals 17.37 Ma and 15.18 Ma (source rock unit 1) and 15.18 Ma and 12.05 Ma (source rock unit 2) from direct observation in ODP Hole 909C (Knies and Mann, 2002, Stein et al., 1995). Source rock unit 1 employs laterally variable total organic carbon (TOC) and hydrogen index (HI) values both in the AOI_{east} and AOI_{west} (Fig. 1) in accordance with the organic facies distribution model of Knies and Mann (2002). Employed TOC values range from 0 to 3.5 wt% and HI values span between 20 and 200 mgHC/gTOC with maximum effective thicknesses being limited to 75 m (Tab. 2). Further details on the model derivation and lateral value distributions are given in the enclosed supplements (Fig. S4). For the younger source rock unit 2, uniform values for TOC of 1.0 wt% and HI of 50 mgHC/gTOC are used, as observed in ODP Hole 909C (Myhre et al., 1995). Its effective thickness is delimited to a maximum of 100 m (Tab. 2).

Table 2: Source rock input parameters to the petroleum systems model

Parameter/variable	Values/units/type	Physical meaning	References
Source rock unit 1 – age: 17.37 – 15.18 Ma			
TOC_0 [1]	0-3.5 wt%	Total organic carbon in the source 1 prior to thermal alteration	Knies and Mann, 2002
HI_0 [1]	0-200 mgHC/gTOC	Hydrogen index of kerogen in source 1 prior to thermal alteration	Stein et al., 1995
Thickness [1]	up to 75 m	Effective thickness of source 1	
Source rock unit 2 – age: 15.18 – 12.05 Ma			
TOC_0 [2]	1 wt%	Total organic carbon in the source 2 prior to thermal alteration	Stein et al., 1995
HI_0 [2]	50mg/gTOC	Hydrogen index of kerogen in source 2 prior to thermal alteration	
Thickness [2]	up to 100 m	Effective thickness of source 2	
Source rock characterization			
Kerogen type [all]	Terrigenous, type III	Kerogen type classification of the organic matter in source rocks [1, 2]	Stein et al., 1995
Kinetic scheme [all]	Behar type III	Kinetic scheme for the kerogen to hydrocarbon transformation for all source rocks [1, 2]	Behar et al., 1997

3.1.5 Methane hydrate stability zone model

A model (for numerical details see supplements) to determine the depth of the base of the GHSZ and incorporate the GHSZ a sealing barrier to migrating hydrocarbons over geological times has been implemented. The concept employs the CSMHyd program (Sloan and Koh, 2007) for calculating the P-T phase boundary curves for gas hydrates with mixed source gas compositions. We employed two end members of source gas composition from Plaza-Faverola et al. 2017 (Scenario A: 100%; C_1 and Scenario B: 95.6% C_1, 2.06% C_2 and 2.32% C_3-C_6 from sample GC19, see Table S1 in supplements) to test the potential impact of variable GHSZ thicknesses on hydrocarbon migration patterns over time. Specifically, it was tested if the directions of lateral HC migration routes at shallow depth will be deflected by a GHSZ barrier and if such barrier is suited to accumulate continuous columns of gaseous hydrocarbons beneath it.

3.1.6 Faults and fault leakage model

Two types of fault sets were interpreted in the seismic dataset (Fig. 4). The first set (set 1) consists of fault outline polygons collated from fault sticks interpreted within 2D seismic lines and corresponding fault scarps exposed at seabed surface. Faults of this set have a regional extent in the AOI, occur on the southern flank of the Vestnesa Ridge towards the northward end of the Knipovich Ridge (Fig. 1) and trend N-S to NW-SE. They are steeply dipping at angles between 50 to 75 degrees between the seafloor and the 3.2 Ma depth horizon (Fig. 4). Except for their surface fault scarp, they do not exhibit any distinct vertical offset at horizons below the seafloor at the resolution of the available seismic surveys. Timing of the formation for these faults is unknown. However, based on the relative age of the horizons truncated by these faults, we assume that faulting and re-activation have occurred between ~3 Ma at the earliest and today evidenced by the seabed fault scarps.

Fault set 2 is restricted to the crestal zone of the Vestnesa Ridge and is partially imaged by the 3D P-cable dataset (Fig. 1). These faults have a very local occurrence and strike NNW-SSE to NW-SE, strongly paralleling the trend of the south-eastern Vestnesa Ridge segment. They have nearly vertical surfaces with dip angles >80 degrees and extend from the seabed to the YP3/YP2 boundary (~2.7 Ma) (Plaza-Faverola et al., 2015). According to Plaza-Faverola and Keiding (2019), they appear to be extensional in nature and have a linkage to seismic gas chimney expressions in the subsurface. Their formation age is unknown but relative relations to fractured horizons suggest an origin age between less than ~2.7 Ma age and today (Plaza-Faverola et al., 2015).

In the hydrocarbon migration model, fault set 1 was defined to permit up-dip hydrocarbon migration (fault leakage) from deeper into shallower carrier horizons. The fault leakage concept permits a pre-defined fraction of hydrocarbons (volume) from a migration stringer passing along a fault surface to move upward and exiting the fault at a shallower level if suited pore entry pressure and permeabilities for lateral hydrocarbon (HC) migration are met. If no favorable conditions are encountered at shallower levels, no fault leakage is modelled, and all HC volumes remain in the original migration stringers at depth.

Fault set 2 is too local with respect to the grid resolution to be included in the fault leakage model and hence no leakage model was applied .

3.2 Local stress considerations

A local stress analysis was performed to evaluate the stress impact of modelled gas accumulations on the overburden stability and their relation to documented methane seepage history at the western Vestnesa Ridge crest. To evaluate the local stresses in a passive margin setting, we assume that $\sigma_1 = \sigma_v$ (at or near vertical), and the minimum principle stress, $\sigma_3 = \sigma_h$ (at or near horizontal). We calculate σ_v (total vertical stress) from wet bulk density data of ODP Hole 986C and Hole 986D, and use the poro-elastic formulation of Engelder and Fischer (1994) to estimate the total horizontal stress, σ_h, considering the Poisson’s effect and the pore pressure increase as to:

$$\sigma_{\mathrm{v}} = \int_0^z \rho_{bulk} g \mathrm{d}z \qquad (3)$$

$$\sigma_{\mathrm{h}} = \left[\frac{\nu}{1-\nu}\right] \sigma_{\mathrm{v}} + \alpha \left[\frac{1-2\nu}{1-\nu}\right] (P_{\mathrm{water}} + P_{\mathrm{gas}}) \qquad (4)$$

where z is the overburden thickness above a hydrocarbon trap or at the BSR, ρ_{bulk} is the sediment bulk density, g is gravitational acceleration, v is the Poisson's ration of sediment at the top of a trap or at the BSR, α is the Biot-Willi poroelastic coupling constant of the effective stress, and P_{water} and P_{gas} are the fluid pressures of the water and gas phases, respectively. We allow v to vary between 0.41 to 0.45 (Hornbach et al. 2004) and α to range from 0.67 to 0.77 corresponding to 25% to 0% bulk hydrate in sediments, respectively (Helgerud et al., 1999). P_{water} is assumed to be hydrostatic as sediment porosities of ODP Hole 986C and Hole 986D show a continues decrease from around 60% near the seafloor to less than 30% at around 960 mbsf depth (Jansen et al., 1996). P_{water} is calculated as:

$$P_{water} = \rho_{water}(d_{seabed} + z)g \tag{5}$$

where ρ_{water} is assumed to be 1030 kg/m^3 and d_{seabed} is the water depth below mean sea level as obtained from the seismic data set. P_{gas} is calculated at the top of a gas column in a modelled trap or at the BSR as to:

$$P_{gas} = z_{gas}(\rho_{water} - \rho_{gas})g \tag{6}$$

where z_{gas} is the modelled gas column height in m and ρ_{gas} is the associated modelled gas density in kg/m^3 obtained from the petroleum systems model.

For the Mohr-Coulomb-Griffith failure criterion evaluation we assume a dip angle for normal faults of θ = 60° and calculate the coefficients of sliding friction μ = tan (90-2θ) = 0.577 and μ = 0.85 (Byerlee's law). For the reactivation of faults, we assume no cohesive strength (C = 0). During rock faulting we infer a cohesion of C = 2 MPa for normal faulting and a tensile strength of $T0$ = -1 MPa assuming compacted weak rock types (silty shales and silty sandstones). Effective stresses (σ´) are calculated by subtracting the cumulative fluid pressures from the estimated stress values (σ) at a given depth.

4 Results

4.1 Petroleum systems modelling

In this section we will present the results of the hydrocarbon generation and migration modelling for three different simulation setups. The first setup assumes no GHSZ whilst the second and third setup consider a GHSZ with gas compositions of scenario A and B as outlined in Tab. S1 (i.e., assuming 100% methane hydrates or a mix of measured higher order hydrocarbons in hydrates). At first, we summarize the results common to all three model setups. Thereafter, we focus on the key differences among the cases and use the non-GHSZ case as reference frame. In the non-GHSZ case the modelled hydrocarbon migration and trapping patterns are not affected by methane hydrates and thus reflect the likely patterns solely governed by the structural and lithological inventory of the model.

4.2 Similarities in the results of all GHSZ scenario runs

The modelling results show that hydrocarbon generation and migration started in the late Miocene at ~6 Ma, from a kitchen area in the AOI_{east} being located between the Vestnesa Ridge in the east and the MTF in the west (Fig. 5). In the AOI_{east}, 213 x 10^9 Sm^3 of gas and 169 x 10^6 Sm^3 of oil are modelled to be expelled cumulatively over time. Gas is the dominant hydrocarbon phase modelled to be released from the source rocks and subsequently migrated, trapped, spilled, and leaked across the basin model. Generated and released oil (> C_6) volumes are of minor magnitude and play a secondary role in the modelled hydrocarbon migration and leakage pattern.

With the progressive burial over time, hydrocarbons generated in the western kitchen migrate exclusively within the Miocene source rock stratigraphic layer in northern, western, and southern directions toward the physical model boundaries (Fig. 5). In contrast, hydrocarbons generated in the eastern kitchen migrate within different stratigraphic layers and in an eastward direction towards the crestal area of the Vestnesa Ridge (Fig. 5, lowest unit). During time interval 4.37 Ma to 3.2 Ma, eastward directed gas migration intensifies both within the younger source rock unit (15.18 – 12.05 Ma) and a middle Pliocene sedimentary sequence (4.37 – 3.2 Ma). After 3.2 Ma, lateral hydrocarbon migration shifts from the middle Pliocene into a sandier late Pliocene unit (3.2 – 2.7 Ma, Fig. 5). Within this late Pliocene unit, gas migration is directed towards the Vestnesa Ridge's crest where it prevails until today (Figs. 5 and 6). This lateral gas migration charges gas accumulations underneath the crest seeping gas by capillary leakage since ~3 Ma. Over time, gas traps at shallower stratigraphic levels form within two early Pleistocene age units: eP_A (~2.7 – 2.1 Ma) and eP_B , (~2.1 – 1.5 Ma) (Fig. 6).

Between 3.2 Ma and 0 Ma, all three model scenarios show an increase in the average capillary gas leakage rates across all layers and traps from 10^8 to 10^{11} Sm^3 Ma^{-1}. This equates to gas flux estimates in the magnitudes 10^1 to 10^4 mg m^{-2} day^{-1} or 10^{-2} to 10^1 kg m^{-2} a^{-1} (Fig. 7). This gas flux increase is likely due to the progressive eastward maturation of the Miocene source rocks and thus more proximal supply of hydrocarbons below the Vestnesa Ridge.

Through time, many of the crestal gas traps hosted in the Pliocene-Pleistocene units are modelled to lose gas by (A) capillary leakage, even to the seafloor, and (B) along spill routes that broadly parallel the Vestnesa Ridge trend. All scenarios show an increase in the capillary leakage rates to the seafloor from 10^8 to 10^{11} Sm^3 Ma^{-1} (Fig. 8) from 3.2 to 1.5 Ma, respectively. The modelled long-distance spill routes may charge other traps and/or exit the physical boundaries of the AOI_{east} (Fig. 6). In all three scenarios, trap charge by spill is modelled to persist between 3.2 Ma and 0 Ma at average rates of 2 – 5 x 10^9 Sm^3 Ma^{-1} with maximum values reaching up to 155 x 10^9 Sm^3 Ma^{-1}. Taking the Lunde and Lomvi pockmark sites as the reference point, a north to north-west directed spill is modelled from traps located to the west of them. In contrast, a south-east directed spill is modelled from traps situated to the east of the pockmarks (Fig. 6).

Fault set 1 (Fig. 4) is cross cutting hydrocarbon migration routes in the late Pliocene carrier unit (3.2 – 2.7 Ma) which is charging gas to the Vestnesa Ridge since ~3 Ma (Figs. 5 and 6). However, the results show that the faults of set 1 are not chosen by the simulator as preferred leakage pathways into shallower stratigraphic levels. Instead, the hydrocarbons are modelled to migrate across these fault surfaces and continue within the same carrier unit (3.2 – 2.7 Ma). The reason for this is 2-fold. First, fault set 1 exhibit negligible fault throw (at detection limit in the seismic resolution) and hence no offset within the late Pliocene unit is implemented in the model. Consequently, similar pore entry pressures are present to either side of the faults in this unit. Second, this unit (3.2 – 2.7 Ma) is characterized by silty/sandy lithologies and hosts lower pore entry pressures than the shallower shale dominated units. Therefore, the simulator favors the 3.2 – 2.7 Ma unit for lateral hydrocarbon migration over fault leakage.

All three scenarios yield gas accumulations located directly underneath the Lomvi and Lunde pockmark seeps (Bünz et al., 2012) within the eP_A (Trap A) and eP_B (Trap B) units at present day (Fig. 6, and Figs. 10 – 13). These gas traps are modelled irrespective of the presence of a GHSZ (cf. Figs 11 – 13). They are stratigraphic traps that form underneath the ridge's crest whereby the depth of the accumulations is governed by the lithological distribution (sand-shale-carbonate) projected from ODP Hole 986D (Jansen et al., 1996). Both traps display a distinct NW-SE elongation (up to a few kilometers in length) and strongly parallel the Vestnesa Ridge crestal trend. Among the two units, unit eP_A hosts the largest amounts of gas and a major trap, Trap A, is filled to its maximum capacity since ~1.5 Ma. Gas migration and trapping within both units are modelled to occur at different sublayer levels. Hence both traps consist of several stacked sublayer accumulations which may have experienced independent charge-loss histories (Figs. 9 – 13).

4.2.1 Key differences in the results of the GSHZ scenario runs

Though all three model scenarios yield gas accumulations in the eP_A and eP_B units underneath the seep sites at Vestnesa Ridge since ~2.1 Ma to 1.5 Ma, distinct differences exist between the scenarios on how much gas is transferred into, lost from, and retained within both early Pleistocene age units. In the *no GHSZ scenario*, gases leak persistently to the seafloor until present day at rates between 10^8 to 10^{10} $Sm^3\ Ma^{-1}$ (Figs. 7 and 8). The maximum seafloor leakage rate occurs at 2.1 Ma (Fig. 7) and coincides with the first overall peak in capillary leakage (Fig. 8). At this time step, gases accumulate in eP_A unit in shallow buried traps, which can sustain only minor gas columns and hence lose gas to the seafloor. With progressive burial, traps in the eP_A and eP_B units accumulate larger column heights and capillary leakage rates to the seafloor are reduced but leakage remains (Fig. 8).

Both *GHSZ model scenarios A and B* show, that upon gas arrival underneath the paleo Vestnesa Ridge at ~3.2 Ma, the inferred GHSZ acts as lateral migration barrier deflecting the gas migration routes into deeper sublayers. This causes gas trapping and migration at levels, where, compared to the *no GHSZ scenario*, slightly higher gas columns can be attained whilst trap capacities are equally small. Therefore, an increased gas spill occurs in the *GHSZ model scenarios A and B* even though the bulk capillary leakage patterns remain similar for all three scenarios at rates of 10^8 Sm^3 Ma^{-1} (Figs. 7 and 8). With progressive burial, gas retention in deeper seated sublayer traps intensifies and more gas accumulates in eP_A and eP_B units over time.

Hence, the main effect of the presence of a GHSZ in the hydrocarbon migration and trapping model is to re-route and channel gas migration into the silty-sandy lithologies of the eP_A and eP_B units, while in the no GHSZ model scenario gas is constantly leaking upward to the (paleo-)seafloor. In the GHSZ model scenarios upward seafloor leakage is prevented by either spill toward the AOI_{east} model boundary or by storage in multi-sublayer traps, such as Trap A (Fig. 9) from -1.5 Ma and -0.148 Ma for scenarios B and A, respectively.

Of all three model scenarios, scenario B with the best present-day GHSZ fit (Tab. S1), yields the strongest confinement of gas within the eP_A and eP_B units underneath the Vestnesa Ridge. In this scenario, Traps A and B accumulate the largest gas volumes, yield the highest column heights, and are composed of the largest number of sublayer accumulations at present day (Figs. 9 – 13). Trap A is most prominent underneath the Lunde and Lomvi pockmarks and is consistently charged at average rates of 2 – 5 x 10^9 Sm^3 Ma^{-1} (0.6 – 1.4 kg m^{-2} $year^{-1}$) since ~1.5 Ma. Trap A is modelled to host 10.7 x 10^9 Sm^3 of gas, distributed over 14 sublayer accumulations at present day. Trap B (Figs. 9 – 13) is formed at ~0.15 Ma ago and charged by capillary leakage from the eP_A unit at average rates of 0.5 - 1 x 10^9 Sm^3 Ma^{-1} (0.14 – 0.3 kg/m^2/year). At present day, Trap B is modelled to host 0.4 x 10^9 Sm^3 within two sublayer accumulation with maximum gas column of 10 m. The shallowest sublayer accumulation of Trap B is in direct contact with the GHSZ along the crest of the Vestnesa Ridge (Fig. 13).

4.3 Local stress analysis for GHSZ scenario B

Analysis of the local stress situation on top of traps in units eP_A and eP_B and below the BSR are shown in Fig. 14. Stress state estimates are based on model values obtained from the GHSZ scenario B and are given in Tab. 3. At time step -1.5 Ma, a maximum gas column of 25 m is modelled for stacked sublayer accumulations of Trap A whilst Trap B has not been formed yet. At present day, maximum gas column heights have increased to 46 m and 10 m in Trap A and Trap B, respectively. These column height values, together with the modelled corresponding gas density values, are used in the Mohr diagram to evaluate the influence of gas column pressure on the local stress states (Fig. 14.) Only minor gas amounts are modelled to accumulate very localized underneath the GHSZ in upper sublayers of the eP_B unit (Fig. 13). Therefore, the local stress state at the base of the GHSZ is plotted without a free gas column (Fig. 14) but stress states assuming up to 10 m of gas, alike Trap B, are calculated and quoted in Tab. 3.

All trap top points are modelled to be in stable state of local effective stress as none of the stress circles are reaching the Coulomb failure criterion (green line in Fig. 14) and hence do not permit normal faulting nor tensile fracturing at either -1.5 Ma or present day (Fig. 14). This means that the gas columns of the traps modelled within eP_A and eP_B unit underneath the Vestnesa Ridge's crest do not exert enough pressure to damage intact rocks at any model time. For Poisson's ratios of $v \geq 0.45$, the gas column pressure is even insufficient to trigger reactivation by normal faulting, illustrated by the fact that none of the effective stress scenarios reaches the failure criterion on pre-existing faults (dashed circles in Fig. 14). However, when assuming Poisson ratio's of $v \leq 0.41$ (i.e., rocks that are more compressible) the local state of effective stress at both trap top points as well as the base of the GHSZ is decreased (i.e., the Mohr circles reach the fault reactivation criteria, Fig. 14) for all model time steps. This means that the modelled gas height can affect the weakened rocks by triggering normal faulting on pre-existing faults but only for fault planes dipping at ≥ ~60° (Fig. 14). The relation between this possible stress state and the plane orientation of fault set 2 cross cutting Trap A (Fig. 10) and Trap B will be discussed in the following section.

Table 3: Model results used for local stress stability.

Structure	Trap A	Trap B	at BSR
Hosting units (age base – top in Ma)	eP_A (2.7 – 2.1)	eP_B (2.1 – 1.5)	eP_B (2.1 – 1.5)
	At 0 Ma model time step		
Water depth (MSL in m)	-1216	-1216	-1222
Burial depth (MSL in m)	-1614	-1421	-1424
P_{gas} (MPa)	0.421	0.092	– (*0.092*)*
σ'_v (MPa)	2.976	1.457	1.569
σ'_H (MPa)	0.894 – 1.736	-0.031 – 0.626	-0.182 – 0.525 (*-0.245 – 0.445*)*
Gas column (m)	46	10	– (*10*)*
Gas density (kg m^{-3})	96.2648	94.2632	– (94.2632)*
	At -1.5 Ma model time step		
Hosting units (age base – top in Ma)	eP_A (2.7 – 2.1)	–	eP_A – eP_B (2.7 – 1.5)
Water depth (MSL in m)	-1236	–	-1241
Burial depth (MSL in m)	-1445	–	-1445
P_{gas} (MPa)	0.229		–
σ'_v (MPa)	1.401	–	1.630
σ'_h (MPa)	-0,067 – 0.526	–	-0.159 – 0.564
Gas column (m)	25	–	–
Gas density (kg m^{-3})	96.5552	–	–

σ´ are effective stresses and account for the influence of pore fluids (water and gas phases). Horizontal effective stress ranges (σ'_h) correspond to Poisson's ratio ranges of ν = 0.41 – 0.45. Gas densities are derived from the trap information calculated by the Migri simulator. The water depth values between above the traps and BSR vary as spatial locations with maximum column heights differ from the locations where the GHSZ extents downward into the units. *) Stress state parameters at the BSR are calculated for an assumed 10 m gas column underlying the base of the model GHSZ. This setting is chosen as a comparison option for the shallowest sublayer gas traps in the eP_B unit, which are partially overlain by the model GHSZ though the GHSZ is not the primary trap seal (Fig. 13).

5 Discussion

5.1 Thermogenic gas accumulations – the primary source of active methane seepage

Hydrocarbon migration modelling results show that accumulation of gas is possible in dedicated traps underneath the Vestnesa Ridge. Gas started to accumulate in late Pliocene units, ca. 3 Ma ago, while persistent gas traps (Trap A and B) sustaining seepage at the Lund and Lomvi pockmarks at present day since the early Pleistocene (~2 Ma) (Fig. 6). Furthermore, the gas trapped since ~1.5 Ma is confined to structures that are situated distinctly deeper than the modelled base of the concurrent GHSZ and its BSR expression (Scenario B; Tab. S1) at present-day (Fig. 13). Below the Vestnesa Ridge, the petroleum systems model suggests that the presence of a GHSZ (Scenario B) deflects and channels gas into early Pleistocene units rather than having a sealing effect for substantial amounts of gas at its base. This corroborates observations by Hushoft et al (2009) along the Svalex P11 seismic line in the AOI (Fig. 1) and Goswami et al. (2015), which show an overall, albeit outside the gas chimneys, low pore space saturation (less than 30% of the porosity) with hydrates within the GHSZ and low free gas contents (less than 5%) directly underneath the BSR. Across the ridge, results of Hushoft et al. (2009) point towards a discontinuous BSR in fine-grained clay lithologies with higher hydrate saturations in the pore space (up to 11%) near the vents at the crestal area and lesser values (up to ~6% at the base of GHSZ) away from chimneys and faults.

Based on the opening history of the Fram Strait (Engen et al., 2008), it is reasonable to assume that the AOI_{east} has been at water depths of >=1000m for the last ~3 Ma. Consequently, the upper strata have been permanently exposed to the temperature and pressure conditions suited for gas hydrate formation (Sloan and Koh, 2007). This implies that any thermogenic gas migrating at low rates through these strata would have likely transformed into hydrates, given an appropriate pore water salinity and gas-pore water dissolution rate are present (Darnell and Flemings, 2015; Spangenberg et al., ~~2005,~~ 2015). Modelled capillary leakage fluxes to the paleo-seafloor layer between 3.2 Ma and 1.5 Ma range from ~ 0.05 kg m^{-2} $year^{-1}$ up to 10 kg m^{-2} $year^{-1}$ (~10^{1} – 10^{4} mg m^{-2} day^{-1}, Fig. 8) for the GHSZ (Scenario B). Such flux rates are comparable to natural micro- to macro-seepage flux rates (Etiope and Klusman, 2010) where free, buoyancy driven gas movement occurs from underground hydrocarbon reservoirs to the earth's surface (Brown, 2000, ~~and~~ Etiope and Klusman, 2002 and Etiope et al., 2006). At 3.2 Ma, low leakage fluxes are modelled to pass through shaley lithologies and are captured within the concurrent GHSZ (Liu and Flemings, 2007). For the model times steps from 2.7 Ma to 1.5 Ma, capillary leakage fluxes correspond to gas flux magnitudes of ~1 kg m^{-2} $year^{-1}$ that could enable gas chimney formation in the silty-sandy lithologies (Liu and Flemings, 2006 and Liu and Flemings, 2007) of eP_A (2.7-2.1 Ma) and eP_B (2.1-1.5 Ma) units. However, paleo-pockmarks are rather restricted to the time interval between 1.5 Ma to present (Plaza-Faverola et al., 2015) implying that gas charges between 2.7 Ma and 2.1 Ma are captured in gas hydrates while the highest modelled gas fluxes through silty-sandy layers at ~1.5 Ma could indicate an initial chimney formation event and the onset of seafloor seepage (Fig. 8). Seepage periodicity through gas chimneys or pipes along continental margins has been documented from various settings and using diverse approaches (e.g., Hovland, and Sommerville 1985; Baraza and Ercilla 1996; Chand et al., 2012; Davy et al., 2010; Plaza-Faverola et al., 2011; Riboulot et al., 2014). Along the Vestnesa Ridge vertical fluid migration features (i.e., gas chimneys) are characterized by highly fractured strata with high amplitude anomalies concentrated at specific stratigraphic intervals (Fig. 15). High amplitude anomalies are usually present inside seafloor pockmarks (e.g., Panieri et al., 2017) as well as in buried cavities interpreted as paleo-pockmarks (Fig. 15). When the sedimentary layers are not entirely masked by shallow gas accumulations, high-resolution 3D P-Cable seismic data reveal the presence of these paleo pockmarks and mounds in the form of erosional surfaces with thicker sedimentary infill (Fig. 15). These seismic observations have been conceptually explained by the formation of authigenic carbonate concretions during active seepage events and further burial during less intense seepage events (e.g., Plaza-Faverola et al., 2011; Plaza-Faverola et al., 2015). Recently, Himmler et al. (2019) showed a correlation between interpreted paleo features and authigenic carbonate in deep sediment cores from the Lunde pockmark on the Vestnesa Ridge. The results of this study validate the conceptual model of buried seepage features and supporting the inference of an increased number of paleo-seepage events, along the Vestnesa Ridge, from ca. 1.5 Ma (Plaza-Faverola et al., 2015).

Gas migration and trapping models in GHSZ Scenario B show no capillary leakage from traps to the paleo-seafloor layers for model time steps younger than 1.5 Ma (Fig. 8). This implies that the multiple methane seepage events recorded during late Pleistocene times at the Lunde and Lomvi seep sites (Himmler et al., 2019 and Schneider, 2018) are not sourced by gas leakage due to an overcome of the capillary seal capacities (i.e. seal displacement pressure sensu Watts, 1987) of traps A and B. Both traps constitute the primary thermogenic gas sources for the documented seafloor seepage between the late Pleistocene and today, yet a driving mechanism is required to transfer thermogenic gases from both traps towards the seafloor and to bypass the GHSZ. We argue that the steeply dipping faults of set 2 crosscutting Trap A (Fig. 10) presumably constitute the most potential pathways for gas migration across the GHSZ and thus fueling the late Pleistocene methane seepage events. We exclude an excess of pore fluid pressure as a cause of mechanical cap rock failure (i.e. hydraulic fracturing, Watts, 1987) above Traps A and B for two reasons. (I) The depth – porosity trend given for reference site ODP 986 (Jansen et al., 1996) shows a continuous porosity decrease with depth and provides no indication of deviating compaction patterns prone to fluid overpressure build up. (II) Local states of stress below the GHSZ and for Traps A and B do not exert sufficient pore-fluid pressure to trigger mechanical rock failure of intact rocks (Fig. 14). Modelled gas column heights of up to 46 m in Trap A (Tab. 3) may permit normal fault reactivation for pre-existing faults with dip angles larger than ~60° (ν= 0.41, Fig. 14) yet require faults to be in place. We, therefore, advocate tectonically controlled faulting as the most likely governing mechanism to initiate gas leakage from Trap A since ~1.5 Ma and propose a plausible conceptual model of the leakage system below.

5.2 Extensional faulting and episodic seepage since ~1.5 Ma

A temporal correlation between seepage events and key glacial periods in the region have motivated cross-disciplinary studies of the effect of glacial dynamics on the Vestnesa Ridge's seepage system (Plaza-Faverola et al., 2015, Schneider et al., 2018, Plaza-Faverola and Keiding 2019, Himmler et al. 2019). It has been suggested that the dynamics at the SBIS glacial forebulge and/or GIA may have been geological drivers of episodic methane release along the Vestnesa Ridge (Plaza-Faverola and Keiding 2019; Himmler et al., 2019). Although uncertainties exist on the extent and volume of circum-Arctic ice sheet beyond the last glacial period, there is sufficient evidence for the last glacial period that Vestnesa Ridge was always located to the northwest of the Svalbard-Barents Sea ice sheet (SBIS) margin (Patton et al. 2015, 2016); most likely in a position where a glacial forebulge would be expected to form in response to SBIS build-up (Fjeldskaar, 1994). Thus, potential impacts of glacial cyclicities underneath the ice shield on the thermal regime, hydrocarbon generation, phase composition and migration routes as inferred e.g., on the southwestern Barents Sea shelf (Ostanin et al., 2017) are considered less likely for this study area on the upper slope. The extensional faults of set 2 are terminating and crosscutting Trap A located directly underneath the ridge's crest (Fig. 4 and Fig. 10) and coincide with a position where a glacial forebulge would be expected to form in response to SBIS build-up (Fjeldskaar, 1994). Forebulge location, amplitude and bulging response time to glacial loading are sensitive to the mantle's viscosity as a function of the mantle's temperature (Whitehouse, 2018). Fjeldskaar (1994) argued for low upper mantle viscosities in the order 10^{19} Pa s^{-1} and rapid mantle relaxation times of a few thousand years during GIA of the Fennoscandian mainland after the last glaciation. High heat flow values in the study area linked to the nearby mid-ocean ridge system (Crane et al., 2001; Plaza-Faverola et al., 2017; Hushoft et al., 2009) suggests that a low asthenosphere viscosity and associated rapid mantle GIA relaxation times (in the order of decades) are reasonable to assume for the SBIS after the LGM as well. The post-LGM Fennoscandian forebulge was pinned in its location (~100km from the ice margin) and responded to the waning ice sheet by continuous reduction in its amplitude and curvature (Fjeldskaar, 1994). For the SBIS and the Vestnesa Ridge site, we hypothesize that the glacial forebulge evolution was rather locally fixed and repeatedly changed amplitude and curvature as a crustal flexure in response to glacial cyclicity (Patton et al. 201~~0~~5, 201~~0~~6). Moreover, these responses may have happened within distinctly less than few thousands of years and even in non-linear fashion (Whitehouse, 2018).

Crustal flexure in a forebulge causes minimum horizontal stresses (tensile stresses) in the upper crust and are orientated perpendicular to the bulge's axis and the ice sheet limits (Keiding et al., 2015, Lund, 2015). Various modelling approaches on the LGM history of the

SBIS show persistent glacial boundaries along the NW-SE running shelf edge, similar to the trend of the Vestnesa Ridge (Schmidt et al 2014, Patton et al., 201~~6~~5, 2016~~7~~, Fjeldskaar and Amantov, 2018), and the extensional faults of set 2 (Plaza-Faverola and Keiding, 2019). Lund et al. (2009) show that during the Weichselian glacial cycle, these minimum horizontal stresses within the glacial forebulge may attain values up to -10 MPa in the upper 2.5 km of the crust. Assuming this stress regime applies for the SBIS and the area offshore western Svalbard, we obtain a reduction in σ_h and a rapid shift of the Mohr circles in the Mohr diagram to cross the tensile failure criterion (Figs. 14 and 16), however, not the Coulomb failure criterion for normal faulting (Fig. 14). This would result in the formation of near vertically dipping faults with no detectable offset and a ~NW-SE strike orientation paralleling both to the western Svalbard margin and the south-eastern Vestnesa Ridge trend, analogous to the observed faults of set 2 (Fig. 4).

Extensive glacial loading on Svalbard is well documented since the first shelf-edge glaciations at ~1.5 Ma (Laberg et al., 2010, Mattingsdal et al., 2014) and could have provoked extensional faulting at Vestnesa ridge’s crestal area. We therefore argue that the faults of set 2 constitute such extensional faults that could have developed in response to forebulge formation over the Vestnesa Ridge during periods of intense cooling and extensive SBIS build-up at ~1.5 Ma. Hence, we suggest that the formation of the faults of set 2 as consequence of the first shelf-edge glaciation promoted the primary damage of Trap A and led to the formation of leakage pathways to the paleo-seafloor.

Multiple, intense methane seepage events occurred during four time intervals of intense cooling and SBIS build-up: (1) during the Penultimate Glacial Maximum (PGM) between 0.16-0.13 Ma (Himmler et al, 2019, Jakobsson et al., 2016, Alexanderson et al., 2018), (2) between 0.05 and 0.04 Ma including Heinrich events H5 and H4 (Himmler et al., 2019, Kindler et al. 2014, Andrews and Voelker, 2018), (3) Heinrich Event H2 (0.025 Ma) (Ambrose et al., 2015, Schneider et al., 2018) and (4), Heinrich Event H1 (0.017 Ma) (Sztybor and Rasmussen, 2017, Schneider et al., 2018). Yet, some correlative matches with late Pleistocene to Holocene interglacial events rather than intense cooling periods are documented at ~0.014 Ma, and ~0.010 – 0.005 Ma as well (Himmler et al., 2019 and Schneider et al., 2018). Still, the distinct episodic nature of the recorded seepage events during extreme glacial periods clearly suggests that one of the major forces controlling leakage is associated with extensional faulting at the forebulge. All three migration scenarios yielded persistent gas charges into both the early Pleistocene age eP_A unit and model Trap A between the model timesteps 1.5 Ma and present day. This might indicate that episodic nature of the seepage events was not governed by a charge fluctuation to the primary gas reservoirs but rather modulated by episodic faulting and/or changes in the transmissibility of the fault pathways as a consequence of extensive glaciation on Svalbard. Intermittent interglacial methane seepage episodes may involve reactivation or dilation (i.e., pore fluid pressure increase and subsequent negative effectives stresses) at the set2 faults (Figs. 14 and 16) induced by tectonic forcing derived from the mid-ocean ridges (e.g., Plaza-Faverola and Keiding et al., 2019).

6 Implications and conclusions

The dynamics of Arctic methane seepage in the eastern Fram Strait appear to be governed by fault re-activation linked to glacio-isostatic adjustments (GIA) due to advancing and retreating circum-Arctic ice sheets during the Pleistocene (Cremiere et al., 2016, Himmler et al., 2019). The results of the modelled petroleum system are consistent with the notion of repeated forebulge uplift and subsidence during glacial cycles and associated extensional fracturing of fault segments at the Vestnesa Ridge since the first shelf edge glaciation of the Svalbard-Barents Sea ice sheet, ~1.5 million years ago. This fault damage affected the modelled shallow hydrocarbon reservoirs. This led to the formation of hydrocarbon migration pathways to the seafloor and repeated tapping of the shallow gas accumulations ultimately caused the widespread seepage.

Similarly, shallow thermogenic gas reservoirs have been confirmed beneath glaciated continental margins across the circum-Arctic. An area of 33 million km^2 of confirmed hydrocarbon reserves was directly influenced by grounded ice sheets during the last glaciation (LG) (Gautier et al., 2009, Andreassen et al., 2017). Thus, glacially driven forebulge formation too was extensive and ice wastage during numerous Pleistocene glacial-interglacial cycles have potentially caused large-scale methane release upon deglaciation due to fracture formation and fault re-activation. Moreover, seafloor seepage linked to hydrocarbon loss from deeper seated thermogenic hydrocarbons traps during episodes of deglaciation, erosion, and uplift (Ostanin, et al., 2017) buttress ice shield dynamics as a major driver forcing thermogenic hydrocarbon release from geological storages. Given the relatively widespread occurrence of large-scale fracture systems on formerly glaciated continental margins in the circum-Arctic, it remains to be seen whether methane release from these systems have attained the atmosphere. Recent analysis of ice cores from Antarctica points towards a minor contribution of geological methane to the global carbon inventory during the last deglaciation (Dyonisius et al., 2020). We note, however, that gas emissions on Vestnesa Ridge is not equivalent to its original old carbon source signal, but rather biodegraded due to microbial methane formation (Pape et al., 2019, Sauer et al. 2021~~0~~). A clear source identification of seeping gas in the atmosphere is therefore difficult.

More investigations are needed on this topic to explore all the controlling factors of abrupt natural methane emissions, including re-activation of faults, gas hydrate dissociation and biodegradation that allow methane emissions at the seafloor and eventually to the atmosphere.

Either way, the 3D petroleum systems model at the Vestnesa Ridge gas hydrate system reveals that a key process sustaining Arctic methane seepage is the repeated tapping of hydrocarbon reservoirs due to (a) fracture formation promoted likely by glacial isostatic adjustments (GIA) and (b) fracture re-activation of existing fault patterns. Our model results show that migrating and capillary leaking hydrocarbons underneath the Vestnesa Ridge started to reach the palaeo-GHSZ between 3.2 and 1.5 Ma ago and were focused along-strike the crestal trend of the paleo-ridge. This implies gas hydrate formation on Vestnesa Ridge as early as ~3 Ma ago. However, we do not find evidence that substantial amounts of gas were accumulated beneath the base of the GHSZ. Rather, the gas is trapped in silty-sandy lithologies of early Pleistocene age and occurs directly below the currently active Lunde and Lomvi seepage sites where they are crosscut by steeply dipping faults. These traps are inferred to constitute the most likely thermogenic gas reservoirs sourcing the episodic seafloor seepage events documented since the first shelf edge glaciation of the Svalbard-Barents Sea ice sheet, ~1.5 million years ago.

Local stress analysis suggests that crosscutting faults underneath the seep site do not originate from rock fracturing by excess pore fluid pressures. Maximum gas columns heights up to 46 m in in the modelled traps are too low to trigger neither normal nor extensional rock faulting. Hence, we suggest tectonic stresses as the main geological controlling factor, modulating both faulting activity and gas transmission to the seafloor seepage system. We propose that episodic methane seepage is particularly well coupled to extensional faulting corrupting the modelled thermogenic gas traps underneath the Vestnesa Ridge crest due to forebulge development during repeated Svalbard Barents Sea shelf edge glaciations since ~1.5 Ma.

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