High-resolution 3D seismic study of pockmarks and shallow fluid flow systems at the Snøhvit hydrocarbon field in the SW Barents Sea

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[1] Abstract

The Barents Sea is an epicontinental shelf sea with a fragmented structure consisting of long fault complexes, basins and basement highs. Fluid leakage from deep-seated hydrocarbon accumulations is a widespread phenomenon and mostly related to its denudation history during the glacial/interglacial cycles. In this study, we aimed to better understand shallow fluid flow processes that have led to the formation of numerous pockmarks observed at the seabed, in this area. To achieve this goal, we acquired and interpreted high-resolution 3D seismic and multibeam swath bathymetry data from the Snøhvit area in the Hammerfest Basin, SW Barents Sea. The high-resolution 3D seismic data were obtained using the P-Cable system, which consists of 14 streamers and allows for a vertical resolution of ∼1.5 m and a bin size of 6.25 x 6.25 m to be obtained. The frequency bandwidth of this type of acquisition configuration is approximately 50-300 Hz. Seismic surfaces and volume attributes, such as variance and amplitude, have been used to identify potential fluid accumulations and fluid flow pathways. Several small fluid accumulations occur at the Upper Regional Unconformity separating the glacial and pre-glacial sedimentary formations. Together, these subsurface structures and fluid accumulations control the presence of pockmarks in the Snøhvit study area. Two different types of pockmarks occur at the seabed: a few pockmarks with elliptical shape, up to a few hundred meters wide and with depths up to 12 m, and numerous circular, small, “unit pockmarks” that are only up to 20 m wide and up to 1 m deep. Both types of pockmarks are found within glacial ploughmarks, suggesting that they likely formed during deglaciation or afterwards. Some of the larger normal pockmarks show columnar leakage zones beneath them. Pressure and temperature conditions were favourable for the formation of gas hydrates. During deglaciation, gases may have been released from dissociating gas hydrates prolonging the period over which active seepage occurred. At present, there is no evidence from the 3D seismic data of active gas seepage in the Snøhvit area. Low sedimentation rates or the influence of strong deep ocean currents may explain why these pockmarks can still be identified on the contemporary seabed.

Keywords: Fluid flow, pockmarks, ploughmarks, Barents Sea, P-Cable

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Citation: Tasianas et al., 2018. High-resolution 3D seismic study of pockmarks and shallow fluid flow systems at the Snøhvit hydrocarbon field in the SW Barents Sea. Marine Geology.

1. Introduction

Seabed fluid flow, which involves the flow of gases and liquids through the seabed, is a common phenomenon in sedimentary basins worldwide (Judd and Hovland, 2007; Mazzini et al., 2017; Mazzini et al., 2016). Fluid flow and escape is often indicated by the presence of sub-circular depressions at the seabed, commonly called pockmarks. They range in size from a few meters to a few kilometers in diameter and from a few meters to a few hundreds of meters in depth (Hovland et al., 2002; Judd and Hovland, 2007). A comprehensive study conducted above the Troll East gas field in the Norwegian North Sea revealed more than 7000 pockmarks on the seafloor, present in a ~600 km^2 area as isolated structures, on average ~35 m wide and up to 100 m in size (Mazzini et al., 2017; Mazzini et al., 2016). Pockmarks are evidence of past or active gas seepage and any observation of gas flares in the water column above the pockmarks suggests that they are active today (Bünz et al., 2012; Chand et al., 2012).

Some pockmarks correspond to gas-escape features that have also been linked to methane hydrate destabilization (Davy et al., 2010; Hovland, 1981; King and Maclean, 1970; Mazzini et al., 2017; Mazzini et al., 2016; Pau et al., 2014a; Pau et al., 2014b; Riboulot et al., 2016; Sultan et al., 2010). During the glacial maximum a large ice sheet covered the Barents Sea (Patton et al., 2016) and trapped gas within sediments beneath the ice in the form of gas hydrates. It has been suggested that the last deglaciation could have triggered gas hydrate dissociation causing methane seepage at the seabed and the formation of the extensive Troll gas field (Mazzini et al., 2017; Mazzini et al., 2016). However, not all pockmarks involve gas. They may correspond to erosive features formed by fluid escape when sediment is taken up by the escaping fluids (Judd and Hovland, 2007). Soft, fine-grained sediment that is brought into suspension can be transported by currents and thus constitute a necessary recording medium for pockmark formation, as illustrated by Rise et al. (2014).

Seepage phenomena have been found in many parts of the world's oceans and in various geological settings (Hovland, 1981; Paull et al., 1984; Suess et al., 1999; Zühlsdorff and Spieß, 2004). They can occur in association with various seabed features such as mud volcanoes, pockmarks or diatremes (Judd and Hovland, 2007). Understanding pockmarks and gas seepage phenomena is important for estimating the impact of the latter on global climate change (Judd et al., 2002), deep sea ecosystems (Sibuet and Olu, 1998) and seafloor stability (Evans et al., 1996).

Pockmarks in the Barents Sea are widespread with most pockmarks in the greater Snøhvit area measuring about 20-30 m in width and less than 3 m in depth (Rise et al., 2014). Their shapes and forms range from oval to elongated to even more complex ones. The elongated pockmarks have their long axis orientation parallel to the prevailing bottom current direction (Bøe et al., 1998; Farin, 1980).

The mode of activity in pockmark formation can be either continuous or periodic, during special external events such as storm surges (Hovland et al., 2002) and earthquakes (Reusch et al., 2016). Pockmarks also occur in a post-glacial setting in the presence of very hard sediments, where the mechanisms of pockmark formation may be less well understood compared to other settings. Some large pockmark-like depressions, however, may have been formed by icebergs impinging the seafloor (Bass and Woodworth-Lynas, 1988; Eden and Eyles, 2001). Such icebergs scoured the seabed during ice retreat in late Weichselian times (Judd and Hovland, 2007).The overall objective of this paper is to unveil the fluid flow pathways and better understand the driving mechanisms and fluid flow dynamics in the shallow subsurface leading to pockmark formation at the seabed in the vicinity of the Snøhvit gas field. Moreover, we will assess the age and duration of pockmark development. The paper thus aims to provide a better understanding of the shallow fluid flow processes that have led to the formation of pockmarks at the seabed. It will achieve this by collecting and analysing high-resolution 3D seismic data from the Snøhvit area in the Hammerfest Basin, SW Barents Sea (Figure 1). The P-Cable data have proven more useful than conventional 3D seismic for mapping fluid leakage systems, including seabed depressions interpreted as pockmarks (Figures 1b and 2a) and shallow gas and thus for better understanding fluid flow processes (Petersen et al., 2010; Rajan et al., 2013). With the high-resolution P-Cable system, the temporal resolution is improved by 3-5 times and the spatial resolution can be at least one order of magnitude higher than for conventional 3D seismic (Bellwald et al., 2018; Planke et al., 2009). For making the reproducibility of scientific findings possible and for reinforcing the validity of data gleaned from research, the precise location of any figures produced is indicated by Figure 1b.

2. Geological setting

The Barents Sea is a ~300 m shallow shelf sea on the Norwegian continental margin (Breivik et al., 1998; Faleide et al., 1993). Typical water depths are in the range of approximately 315 m to 355 m (Figure 1). The Barents Sea is composed of a mosaic of platforms and basins, formed by two major continental collisions. The first event corresponds to the Caledonian orogeny, taking place ~400 Ma ago, and the second one to the collision between Laurasia and Western Siberia which led to the creation of the eastern margin of the Barents Sea ~240 Ma ago (Dore, 1995). The study area is located in the Hammerfest Basin (southwestern Barents Sea)(Figure 1), which is characterized by an uplifted reservoir and faults running in an E-W direction (Section 4.3 and related figures). The tectonic features of the Hammerfest Basin were created mainly by Upper Jurassic-Lower Cretaceous faulting (Berglund et al., 1986; Dore, 1995; Faleide et al., 1993; Gabrielsen, 1990).

The seabed in the Snøhvit area is characterized by generally straight or curved grooves (Bellec et al., 2008; Chand et al., 2009). Exceptionally, they can reach a depth of up to 15 m. These grooves were formed after the last glacial maximum and have been interpreted as iceberg ploughmarks (Andreassen et al., 2008; Winsborrow et al., 2010). Calving and drifting icebergs related to the collapse of the Bjørnøyrenna Ice Stream carved the seabed in multiple directions.

A major Upper Regional Uncomformity (URU) separates the glacial sediments from the underlying westward-dipping inclined layers (clinoforms) of the Torsk Formation of Paleocene-Eocene age (Sections 4.2 and 4.3 and related figures)(Linjordet and Olsen, 1992; Nagy et al., 1997).

Upper Jurassic and thick Cretaceous shales act as a cap rock for most of the structures in the Barents Sea region (Estublier and Lackner, 2009). In the study area the hydrocarbon source rocks correspond to the Upper Jurassic Hekkingen Formation, the Lower Jurassic Nordmela Formation and the Triassic Ingøydjupet Group. The Hekkingen Formation shales are mature for oil and gas generation at the western margin of the Hammerfest Basin and along the western fringe of the Loppa High and at the same time also correspond to the local cap rock (Dore, 1995; Mørk et al., 1999; Ohm et al., 2008).

The underlying lithostratigraphic formations, namely the Fruholmen, Tubåen, Nordmela and Stø formations, consist mainly of sandstones interbedded with thin shale layers (Estublier and Lackner, 2009). The lower unit of the Nordmela Formation forms the cap rock of the underlying Tubåen Formation, whereas the upper Nordmela Formation unit and the gas bearing Stø Formation are the main reservoirs in the area (Estublier and Lackner, 2009). The reservoir zone is located at depths of between ~2700-2800 m below sea surface (Linjordet and Olsen, 1992; Maldal and Tappel, 2004; Shi et al., 2013) and consists of Triassic to Jurassic delta plain deposits. Furthermore, the Tubåen Formation has been deposited in a marginal-marine fluvio-deltaic depositional environment. It is in the Tubåen and the Stø Formations where CO_2 has been injected as part of the CCS activities at the Snøhvit plant.

3. Data and methods

This paper uses high-resolution 3D seismic data acquired appoximately 600 m from the southern edge of the Snøhvit gas field in the Hammerfest Basin (Figure 1). In 2011, UiT the Arctic University of Norway acquired high-resolution P-Cable 3D seismic data aboard the R/V Helmer Hanssen research vessel (Figure 2a). Time was converted into meters using a velocity of 1500 m/s. Structure maps generated by the P-Cable technique have a higher resolution when compared to the multibeam swath bathymetry data (MBE)(Figure 2). The latter (Figure 2b) were acquired simultaneously to the 3D seismic data using a SIMRAD EM300 system (30 kHz) onboard the vessel. The final processed multibeam data, due to the close line spacing, have a bin spacing of 5 m also providing a high-resolution image of the seabed morphology.

The P-Cable system of this study consists of 14 streamers with a spacing of 12.5 m along a cross cable. Streamers measured 25 m long and contained eight channels each. The array of multi-channel streamers was used to acquire many seismic lines simultaneously, thus covering a large area with close in-line spacing in a cost efficient way. Due to the curvature of the cross cable, the streamers were slightly closer together (~10-12 m). One mini-GI gun (15in^3) was used as source and shot at a pressure of 170 bar and an interval of 4 s. The frequency bandwidth of this type of acquisition configuration is approximately 50-300 Hz. The 3D seismic data were processed in a workflow described by Petersen et al. (2010) that included post-stack time migration. The final processed 3D seismic data have a bin size of 6.25 x 6.25 m and the volume covers an area of approximately 8×2 km or 16 km^2 (Figure 1).

We used Schlumberger Petrel and Kingdom Seismic Geological Interpretation Software for interpreting the P-Cable 3D seismic data. In Petrel Software, we employed a horizon picking method that was based on manual interpretation, guided autotracking and on the use of various regional well formation tops. The horizons of the seabed and URU have been manually picked for every inline, and these picks have been gridded and snapped to a horizon afterwards. The maximum amplitude in a vertical window of 3 ms below and above the snapped horizon defined the final seismic surfaces. For both the seabed and the URU, we picked the general amplitude polarity corresponding to a seismic peak (red reflector). We used seismic surface and volume attributes, such as variance and amplitude, that aided in the identification of potential fluid accumulations and the pathways taken in the upper part of the overburden. The variance volume, for example, helped in identifying channels and faults (Bellman, 2014; Chopra et al., 2006; Gao, 2003).

In this paper, a Geographical Information System (GIS)-based methodology was implemented in order to digitise the unit pockmarks that are distributed within the study area. The location of the unit pockmarks, whether found inside normal pockmarks or inside or outside ploughmarks, was also recorded. For the successfull implementation of the methodology, spatial analysis tools were used in ArcGIS® software, in order to calculate various characteristics, such as the area, the diameter and the coordinates for each unit pockmark.

The resulting data were used as input for a statistical analysis on the unit pockmarks using the PAST software (Hammer et al., 2001). A density map was constructed using the kernel density method with a Gaussian kernel of width (standard deviation) of 80 m. An alignment map was produced using the method given by Hammer (2009). Other statistical tests performed include the Rayleigh's test for preferred direction, the spatial autocorrelation (Moran's I) of pockmark radii and the Ripley's K analysis using a 95% interval for random patterns.

4. Results

4.1 Seabed

The seabed surface has been interpreted on both the multibeam data (Figures 1b and 3) and the seabed extracted from the P-Cable 3D seismic data (Figures 1b and 4). The seabed of the study area has a depth varying from 312 - 360 m and is characterized by highs on both sides and a depression in the center (Figures 3a and 4a). However, the difference in depth between the low-lying and high-lying regions does not exceed 10 m.

Grooves on the seabed are common features and cover the whole study area (Figures 3 and 4). The dominant groove orientation is ENE-WSW but the direction varies. The observed grooves reach a maximal depth of 5 m, are several kilometers long and up to 400 m wide. Most of them, however, measure 1-2 km in length and are up to 100 m wide (Figures 3 and 4). Vertical profiles across the grooves show that they have mostly a V-shape topography (Figure 3, profile 3), but that they can also have a U-shape (Figure 3, profiles 2, 7 and 8). They are highly erosive portraying a rugged shape. We interpret these grooves to be erosion-related features from iceberg keels scouring into the sediments, referred to as iceberg ploughmarks (Bellec et al., 2008; Hohbein and Cartwright, 2006).

The seabed is characterized by two classes of depressions. There are a few large and hundreds of smaller, widely distributed depressions (Figures 3, 4 and 5). Seven of these large depressions and 1539 smaller ones were picked for the statistical analysis on the unit pockmarks. These depressions can be located above vertical zones of low-amplitude chaotic seismic reflections (Figures 5.3, 5.5 and 5.8), above high-amplitude seismic anomalies (Figure 5.4) or above deep-seated faults (Figures 5.1-5.5; subchapter **Feil! Fant ikke referansekilden.** below). The depressions on the seabed are defined as pockmarks, which indicate ongoing and/or past fluid seepage at the seabed (Judd and Hovland, 2007).

The larger pockmarks, that will be referred to as “normal pockmarks” (NPs)(Hovland et al., 2002), have an elliptical or asymmetric morphology (Figures 2, 3, 4 and 5) and some of them are characterized by raised rims (see southern edge of NP6 and NP7, Figure 3; profiles 7 and 8, and 5.8). Since NPs do not usually have raised rims as an integral part of their structure, these features seem to correspond more to coincidental results of neighbouring ridges or highs. The depth and diameter of NPs decreases towards the eastern and western edges of the study area, where we encounter shallower waters (Figures 3a and 4a). We thus find the largest pockmark (NP6) in the center of the deep central zone, suggesting the existence of some regional trend that correlates with depth. This trend can be linked to existing faults below these features and a variable glacial sediment thickness with depth (Figures 3c, profiles 5 and 7 in Figure 3, 4a, c, 5.1 and 5.7).

The long axis orientation of the NPs is not constant, with the most common orientation being E-W- or NE-SW-wards. These NPs can also be referred to as “elongated pockmarks” (Hovland et al., 2002) with their long axes almost being double their short axes (Figures 2, 3 and 4). Their basin profiles are asymmetrical where the northeastern side is lower than the southwestern side (Figure 3, profiles 4 and 6). This elongated shape of the pockmarks is possibly influenced by the direction of the bottom currents (Figures 3b and c)(Ingvaldsen and Loeng, 2009; Ozhigin et al., 2011; Skagseth et al., 2008).

NPs appear to have sharp outlines with well-defined edges and steep slopes, and have depth-diameter ratios between 1:4 and 1:7. They are often aligned following a NW-SE direction (Cross section 1 on Figure 4a and Figure 5.1) and have diameters of up to 300 m and depths reaching 12 m (Figures 2, 3, 4 and 5). NPs often show several breaks in the angle of slope (Figure 3, profiles 1-8 and Figures 5.3, 5.4 and 5.7).

The proposed alignment of NP2-6 (Cross section 1 on Figure 4 and Figure 5.1) occurs in the deepest central zone of the study area between 447 ms and 468 ms (Figure 4) or below 334 m depth if using the multibeam swath bathymetry data (Figure 3).

We observe that, over the entire study area, unit pockmarks are more common within iceberg ploughmarks (141/Km^2) than outside (32/Km^2)(Figures 2, 3 and 4). NPs, such as NP7 for example, crosscut at their edges two ploughmarks, one being a 60˚N and the other one a 45˚N trending ploughmark (Figures 3a, d and 4a). However, some of the NPs, e.g. NP3, have no obvious relation to these ploughmarks.

The smaller pockmarks, that will be referred to as “unit pockmarks” (UPs)(Hovland et al., 2002), are widespread and usually of circular or elliptical morphology and measure up to 20 m wide and up to 1 m deep (Figure 3, profile 1). They appear to have smooth edges and gentler slopes. The features that we classified as unit pockmarks are illustrated explicitly in the density map in Figure 6a. They are concentrated at the extremities of the survey where they can reach densities of some hundreds to approximately 600-700 UPs/Km^2 (Figures 1b and 6a), according to the estimation carried out as part of the statistical analysis. The density map clearly shows broad regions of higher density in brighter colours. In the inner, deeper parts of the survey, however, their density has a tendency to be lower.

UPs are also associated in strings (Figures 4f and g), found within the ploughmarks (Figures 3a, b and profile 1 in Figure 3) that may or may not be extending out of some of the NPs. Figure 4f focuses on NP3, showing a string of UPs developing on the northeastern side of NP3. Figure 4g is focusing on the area around NP4 and NP5, with the variance map showing a string of UPs developing to the west of NP5 (Figure

4g). All strings observed here are roughly WNW-ESE oriented and the UPs are regularly spaced along the string (Figures 4f and g). The rose plot and Rayleigh's statistical test (Figures 6b and c) show a strong preferred orientation of UP alignment, of the same WNW-ESE and E-W orientation, thus validating the above observations. This preferred E-W orientation of pockmark alignment (Figures 6b and c) also coincides with the dominant ploughmark direction.

Most NPs seem to be composite ones, with smaller UP depressions found within the main larger structures (Figures 2a, 3b-d, 4b and c). The box plot showing the UP size distribution inside NPs and inside and outside ploughmarks (Figure 7a), reinforces the above observation. Furthermore, the box plot indicates that the mean UP size is statistically different between the three groups (one-way ANOVA, $p<0.001$). All three pairwise differences are significant (Tukey's post-hoc, $p<0.01$). The UPs located inside NPs are the largest (with a mean radius of 9.3 m), whereas the UPs found outside ploughmarks are the smallest (with a mean radius of 7.6 m)(Figure 7a). The differences between the three classes are not large, but significant.

UPs are mainly circular in map view (Figures 2a, 3b-d, 4b and d) occurring isolated or aligned within the ploughmarks (Figures 3 and 4). The spatial autocorrelation (Moran's I) of radii (Figure 7b) shows that UPs that are very close together (less than ca. 40 m) tend to have similar radii ($I=0.7$), but for larger distances the radius size is basically random ($I=0$). There is, therefore, little spatial smoothness in radius.

The Ripley's K analysis (using a 95% interval for random pattern, shown in red)(Figure 7c) shows that UPs tend to avoid each other (low K values) at very small scales. This signifies that very few pockmarks are closer to each other than ca. 40 m. At larger scales, there is clustering (indicated by the large K values); as also shown by the density map (Figure 6a). The tendency for clustering reduces at the largest scales (1 km and more)(Figure 7c).

There is an extremely weak ($R^2 = 0.04$) but significant ($p<0.001$) positive correlation between pockmark radius and water depth. This implies that UPs in deeper waters are slightly larger; which is also the case for NPs, as mentioned previously. According to the linear regression, from 336 to 348 m depth, the average UP radius increases by 3.2 m, but with a large scatter. This is probably not due to the water depth per se, but because of some regional trend that correlates with depth.

4.2 Upper Regional Uncomformity

The URU surface, which has a varying depth from 528 ms to 481 ms, but mostly between 500 ms to 510 ms under the NPs (Figure 5), is also characterized by a slight relative real deepening at locations underneath the seafloor pockmarks. The seismic reflections underlying some of these seafloor NPs are discontinuous, especially at the URU level, and seem to have been affected by a structural deformation (Figure 5.8).

The URU surface is characterized by curvilinear grooves and circular depressions in the western part of the study area, whereas the eastern part is dominated by elongated grooves (Figures 1b and 8a). The elongated landforms are interpreted as mega-scale glacial lineations (MSGLs), similar to long groove-rimmed features identified along the URU in the Hoop area (Bellwald et al., 2018). The curvilinear landforms are c. 5 m deep and crosscut each other, and are interpreted as iceberg ploughmarks (Figures 8a-d). The circular depressions, having radii of 25 m and depths of 5 m (Figures 8b and e), are interpreted as pockmarks, similar to the normal pockmarks at the seabed. Furthermore, fault junctions are mainly observed at the URU level (500 ms depth) in proximity to some of the upward dipping sedimentary strata of a clinoform system, namely Cl3, Cl5, Cl6 (Section 4.3.2), and NP4 and NP5 (Figures 1b and 9).

4.3 Glacial and sub-glacial sediment packages

4.3.1 Glacial sediment package

The glacial sediment package is contained between the seafloor and the URU. Commonly, reflection amplitudes below the center of the seabed pockmarks are weaker and more chaotic than elsewhere (Figure 5). The chaotic reflections can be followed into the Torsk Formation ~ 40 ms below seafloor (bsf) under NP7 (Figure 5.8), ~ 70 ms bsf under NP5 (Figure 5.6) or ~50 ms bsf under NP4 (Figure 5.5). Furthermore, under NP4 and above the URU (Figure 5.5) and above the buried pockmarks and ploughmarks at the URU level (Figure 8) the thin glacial package is very disturbed; with the disturbance being indicated by some boundaries locally bending up or down.

A very common NW-SE fault trend is identified above the URU surface (Figure 9). Below NP2 and NP3 (Figures 5.3 and 5.4), over a restricted area extending to the edges of the NPs, we find two normal faults, that are E-W and NE-SW oriented. They develop from a point at a depth of about 650 ms up to 520 ms, just below the URU.

In certain situations the disturbance in the glacial package can be due to a high-velocity "anomaly" such as halite, paleo-pockmarks, a carbonate reef or to the formation of methane derived authigenic carbonates (MDAC), causing strong reflections and up-bending of reflections or velocity pull-ups under the pockmarks (Figures 5.4, 5.5 and 5.6) and the ploughmarks. MDAC do not usually have enough thickness to produce visible pull-ups, as it’s the case here, but they can explain in some cases high amplitude positive reflections. Work carried out involving integrated geochemistry and geochronology of MDAC, coupled with gas hydrate modelling, provides evidence for methane seepage in the southwest Barents Sea, suggesting also that a main episode of carbonate crust formation in the Barents Sea took place after the collapse of the Scandinavian Ice Sheet (SIS)(Cremiere et al., 2016).

4.3.2 Sub-glacial sediment package

Fluid migration in the shallow subsurface (<400 m) occurs mostly along numerous, small-offset faults (Figures 5 and 8c) and laterally along upward-dipping sedimentary strata of a clinoform system in the Paleogene Torsk Formation. In order to determine potential fluid migration pathways in the Snøhvit subsurface and to explain the specific location of NPs in relation to the latter, we have mapped a multitude of these Torsk Formation clinoforms (abbreviated to Cl in Figures 9 and 10), dipping at an angle of about 10-20˚, under the URU.

We distinguish three main fault orientations: N-S, NE-SW and NW-SE trending faults (Figures 5 and 8c). Some of the normal pockmarks coincide with the location of faults and develop over them (Figures 5.1-5.5 and 5.8). Pockmarks NP2 (Figures 5.1 and 5.3), NP3 (Figures 5.1, 5.2 and 5.4), NP4 (Figures 5.1 and 5.5), NP5 and NP6, develop over normal faults which all terminate just below the URU.

The areas around and beneath the NPs are characterized by a fault network which is as dense or even denser than in the areas where major pockmarks are absent (Figure 9). We observe the very common NW-SE fault trend, which is parallel to many of the clinoform edges, very often developing under NPs (Figure 5). This observation suggests that both features could have been used as fluid migration pathways in the shallow subsurface to create the NPs at the seabed, which also have a NW-SE orientation in the study area (Figure 9).

In general, throughout the survey area and at various depths, fault junctions are occurring mainly between the N-S and NE-SW trending faults or between NE-SW and E-W trending faults. More rarely we observe N-S and E-W trending faults crossing each other. Although most fault junctions occur at the URU level, a few fault junctions are also observed at the 544 ms level close to Cl5 and NP6 and at the 555 ms level close to Cl6 (Figure 9).

We also observe how close some of these NPs form in relation to the clinoforms. Also, fault junctions have developed in proximity to the NPs, 200-500 m away from them, directly under certain pockmarks. More precisely, fault junctions at 544 ms depth developed a few hundred meters from Cl5 and fault junctions at 555 ms depth developed parallel to Cl6 and close to NP4 (Figure 9).

In some cases, we observe NPs to be forming at roughly the same distance i.e. 200-400 m from the edge of clinoforms and above the eastern edge of the upper Torsk Formation clinoforms (Figure 9a). NP1, for example, is located 187 m to the west of the edge of Cl6 (Figure 9a). We also observe how the edges of the clinoforms (Cl3-8) are usually N-S or NW-SE oriented (Figures 1b, 9a and 10). The eastern edge of Cl6 reaches the URU forming a line following a NW-SE orientation which is the same as the orientation of the line linking NP2-6, suggesting the existence of a link between potential fluid migration pathways, such as clinoforms, and the location of NPs.

Some clinoform edges reach the URU in locations which also correspond to the exact location of fault junctions, e.g. see point X, 300 m to the south of NP3 (Figure 9a). Otherwise, the fault junctions are located in close proximity to the clinoform edges, e.g. around 200 m to the NE of the clinoform edges at the URU level, see areas Y and Z (Figure 9a). In the last two examples we also observe how the NPs form a few hundred meters behind the clinoform edge. These NPs form around 200 m from the fault junction and all these three features are associated with a ENE-WSW trend (Figure 9). The proximity of NW-SE trending faults reaching the URU along with the same trending zones of weakness corresponding to the clinoforms and the fault junctions to the NPs, suggests that all these features have been used as migration pathways to facilitate the migration of fluid through the subsurfafce and the formation of NPs, of a similar trend, at the seabed.

5. Discussion

5.1 Geology of pockmarks

The datasets showed that normal and unit pockmarks tend to be more common features in some areas of the seabed in the Snøhvit area than in others. Pockmarks also form in the glacial-related ploughmarks, which are characterized by a thinner sedimentary cover (Figures 2-4 and 7a). Data acquired northeast of Nordkappbanken, for example, show a pockmark density that is usually higher in iceberg ploughmarks than in the surrounding areas (Rise et al., 2014). We usually find pockmarks restricted to areas with relatively soft sediments (Judd and Hovland, 2007), as gas can migrate more easily through such types of sediments, and preferentially create pockmarks there.

Although UPs are identified all over the seabed, both within and outside ploughmarks (Figures 2-4 and 7a), they mainly develop in areas at the extremities of the study area, most often within ploughmarks. Furthermore, UPs in deeper water are larger as indicated by the weak but significant positive correlation between UP radius and water depth. We have also noticed that NPs develop only in this central deeper water region (Figures 3 and 4). The higher relative abundance and size of both UPs and NPs in the deeper areas indicates the existence of a regional trend that correlates with depth and of zones of increased gas leakage there.

Previous studies using P-Cable 3D seismic data in the Hoop area (SW Barents Sea) show a strong link between the type and thickness of glacial sediments and shallow gas accumulations (Bellwald and Planke, 2018; Bellwald et al., 2018). An intraglacial reflection in the Hoop area has been mapped out and interpreted as a shear margin moraine, which is characterized by a soft bed at its base (Bellwald and Planke, 2018). Different types of glacial deposits in the Snøhvit area are indicated by a positive, high-amplitude reflection (Figure 5.8), which can affect fluid migration.

Furthermore, the enhanced high-amplitude reflections between the seabed and the URU underlying the pockmarks may correspond to a “push-down” in the reflectors. They are being pushed down by the possible presence of gas, characterised by low acoustic velocity (Figure 5.4). Such “pull-down” effects can be also due to the existence at the seabed of pockmarks, ploughmarks or paleo-channeling that has been infilled with reworked sediments (like muds) with a slower seismic velocity. Ray paths from the surface that go through the above-mentioned features will take longer to reach a certain flat event, because more of the path length is in the lower velocity water, and less of the path length is in the higher velocity rock. In the seismic data a layer that is "flat" in depth will seem to be "pushed down" in time, because the seismic raypaths go through more water than rock (Kearey et al., 2013; Lines and Newrick, 2004).

High-amplitude anomalies below pockmarks can be interpreted as shallow gas accumulations that through seeping feed the overlying pockmarks with gas (Figures 5.2 and 5.4). These anomalies can also be due to MDAC, causing strong reflections or velocity pull-ups under the pockmarks. Although the occurrence of likely relatively thin layers of MDAC, might not be clearly visible on the seismics and thus not associated with a clear pull-up, this scenario can be associated with gas hydrate decomposition, fluid-venting and carbonate crust formation following the collapse of the SIS that took place in the SW Barents Sea as proposed by other studies (Chand et al., 2008; Cremiere et al., 2016; Vadakkepuliyambatta et al., 2017).

The existence of normal faults developing under NP2-NP6 could suggest that large pockmarks have been formed by fluid flowing through discontinuities such as faults (Figures 5, 8 and 9). The existence of fault junctions below NP4 and between NP5 and NP6 as well as the development of a dense network of faults all around pockmarks NP1-NP6, suggests that such crossing points have also played a major role in facilitating fluid flow in these locations thus explaining the formation of UPs and NPs above them (Figure 9). The Rayleigh's test (Figures 6b and c) partly validates the above conclusions as it shows the existence of strong preferred orientations of UP alignment, along an E-W and NW-SE orientation, which also correspond to the predominant fault directions at the 555 ms and 544 ms levels and to a lesser extent at the URU level (Figure 9).

Creation of pockmarks at the surface is probably related to minor faults, micro fractures and disturbed sediments found below iceberg ploughmarks (Figures 4 and 5). All of those zones of weakness have contributed to the creation of migration pathways for gas. The iceberg ploughmarks can act as easy escape routes for the fluid flow and lead to the creation of pockmarks at the seabed (Haavik and Landrø, 2014; Rensbergen et al., 2007; Rise et al., 1999; Solheim and ElverhøI, 1985). The orientation and distribution of these glacial features would thus control the orientation and distribution of pockmarks that form preferentially within them.

The coincidence between micro faults and iceberg ploughmark orientations and the alignment of UPs suggests that iceberg ploughing is related to string pockmark formation (Figures 4d, 6, 7 and 8a-c). An iceberg-ploughed groove excavated in Scotland is a good example illustrating the above statement (Thomas and Connell, 1985). It was found to contain numerous minor faults, micro fractures and disturbed sediments below it. They were located in the influence zone being under stress during the passage of the iceberg. This leads to a probable localized increase in permeability and creation of migration pathways for gas (Thomas and Connell, 1985).

The regularly spaced pockmarks of similar size in the roughly E-W oriented pockmark strings (Figures 4d and 6) suggest some spatial correlation. This is validated by the spatial autocorrelation (Moran's I) of pockmark radii (Figure 7b) which shows that pockmarks that are very close together (i.e. less than ca. 40 m) tend to have similar radii (Moran’s I = 0.7). Fluid flow must have thus led to the creation of pockmarks at regular spaced intervals along the fault strike (Figures 4-7)(Ligtenberg, 2005). This coherence in the observations between seabed and deeper structures suggests that they are linked, that is to say, one contributes to the creation of the other.

The high amplitude anomalies encountered within the clinoform reflectors reaching the URU around NP1 (Figure 5.2) also indicate that there is a connection between faults, that were active at different periods in geological time, fluid migration and bright spots overlying and surrounding them. Faults in the shallow and deeper subsurface have thus allowed for fluids to migrate through them vertically and continue to migrate both laterally and vertically upwards through the clinoforms to reach the URU (Figure 5.2). Fluid flow can either take place through these clinoform surfaces alone or via an association of clinoforms and faulting (Figures 5.1-5.5 and 8c).

There is further evidence that clinoforms act as fluid flow pathways and determine the location of pockmarks at the seabed. This is indicated by the pockmarks forming at the same distance from the edge of clinoforms and the coincidence in orientation of clinoform edges and NP alignment orientation (Figures 9 and 10). Any fluid reaching the URU (Figure 8) can easily continue its upward migration further via a dense network of slight disturbances that one can recognize between a pockmark and the URU (Figures 5 and 8). However, heterogeneities in the glacial package, such as the moraines observed in P-Cable data of the Hoop area (Bellwald and Planke, 2018) could affect lateral fluid migration.

5.2 Pockmark formation mechanisms

Gas has been observed to leak from the seabed in the central Barents Sea around the upper limit for methane hydrate stability. Over 600 gas flares have been mapped in the water column of this area. Some of these gas flares derive from seabed mounds and craters, but most from their flanks and surroundings. Analysis of geophysical data provides a link between these gas flares, the craters and mounds, to seismic indications of gas advection from deeper hydrocarbon reservoirs along faults and fractures (Andreassen et al., 2015).

Gas has been observed to leak from the Barents Sea seabed not specifically from pockmarks, suggesting that pockmark formation in the recent past was followed by a phase of active fluid escape and then inactivity (Rise et al., 2014). The existence of a thin sediment cover in pockmarks and their penetration into the underlying glaciomarine sediments (Figures 4 and 5) suggests that they were formed after deposition of these sediments (Chand et al., 2012).

Additionally, the existence of sharply outlined pockmarks (Figure 3) suggests that they have been influenced by iceberg ploughmarks and could also have been formed recently as little time has passed for their shape to be modified or smoothed by other overlying sediments or water flow. The smaller, gentler-sloped pockmarks, however, could be of older age or might have been inactive for a longer period of time allowing more time for water or sediments to smooth them. Their age could be dated back to pre-ice retreat times (Figures 3-5).

The existence of undisturbed pockmark craters within and outside iceberg ploughmarks suggests that pockmark formation is synchronous to iceberg movement or that they were formed after the main phase of iceberg movement in the SW Barents Sea (Nickel et al., 2012). The difference in orientation of ploughmarks at the seabed may be related to different, multi-directional scouring events (Figures 3, 4a, b and 7a). The scouring events can be further separated by longer periods of time, e.g. seabed vs URU scours (Figures 3, 4a and 8a-d).

Interpretation and analysis of high-resolution seismic data from other areas of the Barents Sea even suggests that pockmarks formed late during postglacial sedimentation (Rise et al., 2014). However, gas seepage was not restricted to the time of pockmark formation (Pau et al., 2014b). A core sampled on the flank of a pockmark in the SW Barents Sea contained biozones characterized by *Nonionellina labradorica* acme, which indicates a deglacial event dated at 14.9 cal Kyr BP in this specific core (Pau et al., 2014b). Other sources date this deglacial event to 14.64 ± 0.186 cal kyr BP (Rasmussen et al., 2006), making it concurrent with the beginning of the Bølling warming (Pau et al., 2014b).

The above findings indicate that gas expulsion activity commenced after the collapse of the Barents Sea Ice Sheet and that unit pockmarks in the Barents Sea were formed in the early stage of the Bølling interstadial (~15 cal Kyr BP)(Pau et al., 2014b). Significant iceberg discharge led to the deposition of ice rafted debris on pre-existing glacial till in other pockmarks located in the SW Barents Sea (Pau et al., 2014b). Also the deposition of certain laminae observed in lithological subunits in pockmarks in this area is ascribed to an environment characterized by seasonal meltwater production close to the ice margin (O Cofaigh and Dowdeswell, 2001), thus reinforcing the idea that pockmark formation is related to ice retreat. At present many pockmarks studied in the SW Barents Sea (Pau et al., 2014b) appear as inactive seabed features, as no evidence for current upward methane flux has been detected.

A variety of mechanisms could have created the pockmarks within the Snøhvit study area. Several authors have suggested that pockmarks are a result of focused fluid flow and this is the most likely explanation of their occurrence within the Hammerfest Basin (Berndt, 2005; Judd and Hovland, 2007). This suggestion is reinforced by the existence of paleo depressions on the URU; interpreted to represent paleo pockmarks, which are likely to have a similar origin, where the venting of fluids has ceased (Figures 8b and e)(Judd and Hovland, 2007). A glacigenic origin for some of the paleo pockmarks cannot be ruled out, as the depressions on the URU are not as apparent in the seismic as those on the seabed. In this case, we can imagine icebergs, that are transported by winds and currents, creating sub-circular depressions when their keels occasionally hit the seabed (Bass and Woodworth-Lynas, 1988; King et al., 2016).

Mechanisms of pockmark formation include relating erosive glacial landforms to a reduction in overburden pressure facilitating shallow subsurface seal breaching, fluid flow and pockmark formation (Harrington, 1985). Another mechanism proposes the involvement of freshwater ice rafting on high latitude shelves. This phenomenon occurs when seeping freshwater freezes at the sediment-water interface when the bottom waters are below 0°C. Over time the frozen water can create a pockmark-like depression (Paull et al., 1999).

Pockmark formation is also most probably related to the seepage of gas that has been generated during the thermogenic and biogenic breakdown of organic material in the subsurface (Chand et al., 2012; Hovland, 1981), and the seepage of porewater through debris lobes. The deep-seated faults in the study area were probably acting as migration pathways for biogenic and thermogenic gas, which in turn allowed gas hydrates to form in response to the pressure and temperature conditions given underneath the ice sheet (Figure 11a). In this case, pockmark formation can be explained by a process where leaking and ascending gas, through a network of leaking faults, reaches the seabed and distributes the seabed sediments in the surrounding water column. Alternatively, seepage hinders settlement of sediment, so over time, sediment thickness grows outside the seepage area but not inside, forming a pockmark. This process, however, affects a small proportion of sediments and occurs to a very limited extent. This phenomenon creates depressions of various sizes and depths depending on the sediment thickness (Chand et al., 2009).

We have observed locations on the seabed with an absence of NPs but with underlying faults (Figures 5.1, 5.2 and 9), suggesting that there has either been no vertical fluid migration along these faults or that the migrated fluid has not reached the seabed to create NPs there. There is a need for both a fluid source and an open migration pathway, like an open fault or an association of an open fault and other migration pathways e.g. clinoforms, in order to form a pockmark at the seabed. In cases where we may observe bright spots overlying any faults, this would strongly suggest that these faults have acted as migration pathways at some point in the past. Absence of such seismic evidence would suggest that these faults have not acted as migration pathways.

Another hypothesis is put forward where changes in the ocean temperature may have led to the destabilization of methane hydrates which created a feedback process that significantly accelerated the shift of the climate system into an interstadial state (Pau et al., 2014b). These changes, linked to methane hydrate dissociation, may be responsible for pockmark formation as a genetical link can be proposed betweent the two.

Based on our high-resolution 3D seismic data from Snøhvit, pockmark formation was most likely associated with recent degassing and dewatering events, as explained in more detail by the proposed conceptual model for pockmark formation in section 5.3 below. The model illustrates the preferred mechanism of pockmark formation by illustrating the connection between ice retreat, gas hydrate formation, decomposition, methane and other gas release and migration and the formation of pockmarks at the seabed. The presence of pockmarks in ploughmarks (Figure 11c) suggests a formation mechanism related to iceberg scouring that has created zones of weakness in the seabed where gas subsequently escaped. Iceberg ploughmarks tend to contain a higher density of pockmarks than anywhere else on the seabed (Figures 3, 4 and 7a). A link between reflections from thin sandy, gas-charged layers and iceberg ploughmarks is suggested by Haavik and Landrø (2014).

Apart from thermogenic methane generation, we can also imagine a phase of anaerobic bacterial methane generation and fluid migration, e.g. rising methane, underneath the ice, leading to the formation of gas hydrates above a bottom simulating reflector (BSR)(Figure 11a). Ice then retreated at a later phase, leading to the formation of ice-free areas. With the consequent pressure change and gas hydrate dissociation (Figure 11b) that followed this probably led to the formation of normal pockmarks (Figure 11c). Gas hydrates were lost through their dissociation and bacterial consumption, with a proportion of the methane being lost through the seafloor.

In some places UP formation can be solely related to gas hydrate formation whereas in others it can be related to a mild, periodic venting of methane gas indicative of stratified diffusive flow. In some cases, methane gas can be the sole fluid responsible for the formation of these micro-pockmarks which periodically vent accumulated gas in small-scale events (Szpak et al., 2015).

UPs represent an expulsion event or events where seeping probably occurs between the URU and the seabed (Figures 5.4 and 9). This association of faulting and overlying high-amplitude anomalies may also suggest that there is a vertical pathway, that is not detectable by seismic, but that allows fluid to flow upwards. The fluid can reach the URU via the association of faults and clinoforms (Figures 5 and 8c). Further vertical migration can be explained by diffused fluid flow through loose sediments giving rise to a fairly even distribution of UPs across the study area (Figure 3, profile 1, and Figures 4d, 6 and 7c).

The action of marine currents and biological activity can also contribute to maintaining or transforming the depressional shape of the pockmarks (Pau et al., 2014a; Pau and Hammer, 2013; Pau et al., 2014b) (Figure 11c). The elongated shape of NPs can be attributed, for example, to the influence of strong bottom currents, which in the SW Barents Sea area have an E-W or SW-NE direction (Figures 3b, 3c and 11c), capable of eroding the newly formed pockmarks. These elongated NPs may indicate that the pockmarks were initially circular, but have been deformed by sediment transport, deposition and erosion (Bøe et al., 1998). Erosion by the prevailing bottom current will be most significant on the downstream side of the pockmark, resulting in an asymmetrical shape with longer shallower flanks upstream; in our case on the E or NE part of the NPs (Figure 3, profiles 4 and 6). Although creeping or other down-slope processes may also occur, it is most likely that these strong bottom currents existing in the area were capable of eroding and shaping these pockmarks. The orientation and shape of these elongated NPs could be controlled by a branch of the West Spitsbergen Bottom Current, namely the North Cape current, flowing northeastwards in the SW Barents Sea (Figure 11c)(Ingvaldsen and Loeng, 2009; Ozhigin et al., 2011; Skagseth et al., 2008).

5.3 Synthesis: conceptual model

A large ice sheet covered the Barents Sea during the last ice age (Andreassen et al., 2008; Knies et al., 2014; Navarro-Rodriguez et al., 2013)(Figure 11a). Gas leaking through fault systems and along the stratigraphic bedding was trapped beneath the ice and at appropriate pressure and temperature conditions it may also have formed gas hydrates (Figure 11a)(Chand et al., 2012; Solheim and Elverhøi, 1993).

Retreat of the ice sheet, scoured the seabed forming ploughmarks, then also caused a decrease in pressure and ensuing isostatic uplift. That might have led to the release of various gases, such as thermogenic and methane gas that was trapped in the shallow subsurface, and their migration along faults, clinoforms and via gas chimneys (Figure 11b). The gas that had potentially been trapped in gas hydrates would have been released due to the change in pressure and temperature (Figure 11b)(Rise et al., 2014). The marine/glaciomarine sediments underlying these pockmarks, deposited after the ice sheet retreated (Rise et al., 2014) seem to be rather undisturbed (Figures 5 and 8c), thus suggesting more of a slow process of pockmark formation rather than from an explosive release of gas. However, the release of gas from gas hydrate dissociation would have been delayed. Hence pockmark formation from gas hydrate dissociation (Figures 11b and c), is post-glacial, of Holocene age, and might have been going on several hundreds if not thousands of years after the ice-sheet retreat in a very similar mechanism as proposed by Mazzini et al. (2016) and Pau et al. (2014).

During the last stage of fluid leakage we have fluid reaching the seabed and formation of pockmarks of various forms and sizes there. Pockmarks formed in ploughmarks and in the rest of the seabed, either isolated or in association with other pockmarks. Pockmarks are observed to form also above or near faults, fault junctions, possibly MDAC, and clinoform edges and can be found aligned following orientations similar to those of local faults and clinoforms. Pockmark shape was probably modified by bottom sea currents flowing over them, which maintained a low sedimentation rate and allowed the pockmarks to be kept open up to today (Figure 11c).

Age datings of carbonate material from seepage areas in other parts of the Barents Sea (Cremiere et al., 2016) support a postglacial pockmark formation. No indications of present-day methane flux are found in the study area documenting that gas seepage may have been active until the recent past but not today (Figure 11c)(Pau et al., 2014b).

6. Conclusions

P-Cable high-resolution 3D seismic data allow to establish a link between the observed seabed morphology and pockmark structure, and shallow (<400m) subsurface phenomena and fluid flow mechanisms in the Snøhvit area in the Barents Sea in a much more comprehensive way than previous studies based on conventional 3D seismic data. Pockmarks at Snøhvit are now also better described through the interpretation of multibeam swath bathymetry data in association with sampling and ROV campaigns carried out elsewhere.

The P-Cable 3D seismic data provides evidence for a complex leakage system leading to the formation of two different types of pockmarks at the seabed; numerous smaller, circular or elliptical “unit pockmarks” or larger asymmetrical “normal pockmarks”. Larger than the ploughmarks, the unit pockmarks are often found within glacial ploughmarks, documenting that they likely started to form during deglaciation. Parts of the distribution of unit pockmarks is controlled by the orientation of the glacial ploughmarks.

Most of the pockmarks can be associated with leakage pathways through a shallow fault system or along inclined bedding planes. The stratigraphic dip related to the Upper Torsk clinoforms also shows indications of controlling fluid movement. Some of the larger normal pockmarks show columnar leakage zones beneath them. The most likely source of the gas is from deep-seated hydrocarbon reservoirs. During the last ice age a large ice sheet covered the Barents Sea and trapped gas within sediments beneath the ice. Appropriate pressure and temperature conditions may have led to the formation of gas hydrates. During deglaciation gases may have been released from dissociating gas hydrates prolonging the period over which active seepage occurred. At present, there is no active seepage of gas observed in the P-Cable data from the Snøhvit area in the Barents Sea.

7. Acknowledgements

We acknowledge the participants of the cruise carried out in July 2011 for helping acquire the 3D P-Cable data. We would also like to thank Sergei Polyanov for processing the 3D P-Cable data. The research was supported by the ECO2 project funded by the European Union's Seventh Framework Programme (grant no. 265847) and by the Norwegian Research Council through CAGE-Center for Arctic Gas Hydrate, Environment and Climate (grant No. 223259). The dataset was collected in this particular location as part of the ECO2 project that focused on CO_2 storage activites in the Sleipner and Snøhvit CO_2 storage sites offshore Norway. We also acknowledge Schlumberger for providing the Petrel interpretation software.

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9. Figure Captions

Figure 1: a) Geographic location map of the Snøhvit field study area indicating the extent of the high-resolution 3D seismic volume. The real CO_2 injection point, indicated by the blue circle, is located in the 110 m thick Tubåen Formation (between 2560-2670 m depth below sea surface). Neighbouring hydrocarbon fields and wells are also shown. b) Map showing the precise location of and spatial relationship between the different figures.

Figure 2: Figure comparing P-Cable high resolution seismic data in a) with MBE data in b) over normal pockmarks (NPs) 5 and 6 of the study area at depths of 330-350 m. The P-Cable dataset was converted and presented in m with both parts a) and b) using the same colour bar. For location see Figure 1b.

Figure 3: Pockmark characterization through multibeam data. a) A seabed surface map and b), c) and d) are zoomed in images of a) showing the location and extent of normal pockmarks (NPs) 1-7, of unit pockmarks (UPs) and ploughmarks on the seafloor with associated cross sections 1-8. Profiles 1-8 illustrate the structure of the above-mentioned features. White arrows indicate iceberg ploughmarks and red arrows the bottom current direction. For location see Figure 1b.

Figure 4: a) Overall view of the seabed surface map obtained from high resolution P-Cable data with b) and c) corresponding to zoomed sections. d) and e) are RMS amplitude maps, of the same sub-sections as previously mentioned, obtained by extracting values from the RMS amplitude cube using the seabed as

horizon with a search window of 12 ms below the event and a horizontal offset of 0 ms. f) and g) are variance maps, of the same subsections as in subfigures b) and d), obtained with a search window of 12 ms below and specifying “Closest trough”, signifying that the attribute computation is done on the trace segment, around a specific horizon, corresponding to a “trough”. This is a maximum negative amplitude, which is located the closest to the selected horizon. The dotted lines in part a) correspond to the seismic cross sections 1-8 represented in Figure 5. For location see Figure 1b.

Figure 5: Seismic cross sections, for location see figures 4a and 1b, illustrating the underlying stratigraphy of the following features from the P-Cable study area: 1) Normal pockmarks (NPs) 2-6 with underlying faulting, 2) NP1, NP3, shallow and deep faults with associated high amplitude anomalies, 3) NP2, 4) NP3, 5) NP4, 6) NP5, 7) NP6 and 8) NP7 and intraglacial reflection. The Normal Pockmarks correspond to large, asymetrical, sub-circular depressions at the seabed whereas the Upper Regional Unconformity (URU), which features in all subsections, represents an erosional surface at the base of the Quaternary deposit and is the oldest preserved glacigenic surface in the southern Barents Sea. Clinoforms are indicated by blue/red dashed lines whereas faults by blue solid lines.

Figure 6: a) Density map using the kernel density method with a Gaussian kernel of width (standard deviation) of 80 m, showing broad regions of higher density, b) alignment detection map (Hammer, 2009), showing pockmark alignment along preferential directions shown by the rose plot in c) and the statistical Rayleigh's test showing a strong E-W preferred orientation of alignments. For location see Figure 1b.

Figure 7: a) Box plot with outliers representing the three groups of pockmarks indicating a mean size that is statistically different, b) Moran’s I spatial autocorrelation of pockmark radii and c) Ripley's K analysis of pockmark distances, using a 95% interval for random pattern shown in red, with the K values represented by the y axis, indicating the tendency for pockmarks to either avoid each other (low K values) or to cluster together (large K values). A number of variations of Ripley's original K-Function have been suggested. Here we implement a common transformation of the K-Function, often referred to as L(d) (y axis).

Figure 8: a) URU surface map with Mega-scale Glacial Lineations (MSGLs), b) associated zoomed in section showing cross cutting ploughmarks and buried pockmarks, visible through seismic cross sections X and Y respectively, both derived from the P-Cable high-resolution data, c) seismic cross section depicting fluid flow pathways and a pull-up effect in the Snøhvit subsurface (seismic profile from Inline 239), d) seismic profile X showing the URU being interrupted by a ploughmark and e) seismic profile Y showing the development of a pockmark on the URU. For the location of part a) see Figure 1b, and for parts b) and c) see box and dotted line, respectively, in part a). For the locations of parts d) and e), corresponding to seismic profiles X and Y respectively, see part b. A different colour scale is used between parts a) and b) of the figure. Clinoforms are indicated by red dotted lines and faults by black dashed lines.

Figure 9: a) Sketch map of the study area combining various interpretations, for location see Figure 1b, with zoomed sections b) and c) corresponding to variance maps at b) 555 ms and c) 544 ms depth. Interpretations include the eastward maximal extent of clinoforms (Cl), location of normal pockmarks (NPs) and faults and fault junctions at the Upper Regional Unconformity (URU) and at other levels.

Figure 10: RMS amplitude map along the Upper Regional Unconformity (URU) surface, using a search window of 7 ms below the URU horizon and 0 ms horizon offset, showing at this depth the contrast in amplitude between the eastward maximal extend of clinoforms (Cl), indicated by the dashed lines, and other areas of weaker amplitudes. For location see Figure 1b.

Figure 11: Proposed model of pockmark formation at the following stages: a) before denudation, b) during erosion and uplift and c) during the last stage of fluid leakage. The difference in the colour of the circles representing fluid type/migration is related to the variety of fluid origins in the local subsurface, in pink for the thermogenic gas and in orange for methane gas, and the migration pathway used e.g. along faults, in blue, and via gas chimneys, in purple.