1 New insights into geology and geochemistry of the Kerch seep area in the
2 Black Sea

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5 Timo Zander¹, Matthias Haeckel¹, Ingo Klaucke¹, Jörg Bialas¹, Dirk Klaeschen¹, Cord
6 Papenberg¹, Thomas Pape², Christian Berndt¹, Gerhard Bohrmann²

7 ¹GEOMAR Helmholtz Centre for Ocean Research Kiel, Wischhofstraße 1-3, 24148

8 Kiel, Germany

9 ²MARUM – Center for Marine Environmental Sciences and Department of
10 Geosciences, University of Bremen, Post Box 330 440 D-28334 Bremen, Germany

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24 Corresponding Author: Ingo Klaucke, iklaucke@geomar.de

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Abstract

High-resolution 3D seismic data in combination with deep-towed sidescan sonar data and porewater analysis give insights into the seafloor expression and the plumbing system of the actively gas emitting Kerch seep area, which is located in the northeastern Black Sea in around 900 m water depth, i.e. well within the gas hydrate stability zone (GHSZ). Our analysis shows that the Kerch seep consists of three closely spaced but individual seeps above a paleo-channel-levee system of the Don Kuban deep-sea fan. We show that mounded seep morphology results from sediment up-doming due to gas overpressure. Each of the seeps hosts its own gas pocket underneath the domes which are fed with methane of predominantly microbial origin along narrow pipes through the GHSZ. Methane transport occurs dominantly in the form of gas bubbles decoupled from fluid advection. Elevated sediment temperatures of up to 0.3 °C above background values are most likely the result of gas hydrate formation within the uppermost 10 meters of the sediment column. Compared to other seeps occurring within the GHSZ in the Black Sea overall only scarce gas indications are present in geoacoustic and geophysical data. Transport-reaction modeling suggests that the Kerch seep is a young seep far from steady state and probably not more than 500 years old.

1. Introduction

Marine cold seeps are locations on the Earth's surface where cold (<100 °C) fluids (i.e. gas, liquid, liquidized solids or a mixture of these) emanate into the overlying water (Suess, 2014). The fluids are of non-magmatic origin and generally rich in methane although other light hydrocarbon gases and oil can also be part of the mixture if associated with an underlying hydrocarbon reservoir. Cold seeps have received much attention because the emanating fluids fuel diverse chemosynthetic ecosystems (Sibuet and Olu, 1998; Boetius et al., 2000), serve as indicators for underlying hydrocarbon accumulations (Mazzini et al., 2003; Løseth et al., 2009) with gas hydrate accumulations probably being the most important

resource (Wallmann et al., 2012), and play a distinctive but not fully quantified role in ocean chemistry (Aloisi et al., 2004; Reeburgh, 2007; Regnier et al., 2011) and climate change (James et al., 2016). While the surface structures of cold seeps are well described and understood, there is much less information on the fluid sources at depth and the different fluid pathways (Talukder, 2012). As most cold seep sites are dominated by methane (Judd, 2003, Bohrmann and Torres, 2006) their distribution on continental margins is largely controlled by the extent of the gas hydrate stability zone (GHSZ) (Milkov, 2004). In most marine basins, including the Black Sea, cold seeps are widespread in areas shallower than the top of the GHSZ (Egorov et al., 2003; Naudts et al., 2006; Römer et al., 2012) whereas seeps within the GHSZ are comparably rare and require exceptional circumstances allowing them to persist. These include localized elevated heat flux or the advection of deep, warm fluids at mud volcanoes (Bohrmann et al., 2003; Greinert et al., 2006; Sahling et al., 2009) or the presence of very high gas flux exceeding the rate of gas hydrate formation (Liu et al., 2019). The latter has been postulated for the cold seeps in the Black Sea offshore Georgia (Klaucke et al., 2006; Nikolovska et al., 2008; Pape et al., 2011; Reitz et al., 2011). Gas hydrate dissociation and subsequent gas seepage has also been described in the Black Sea as the result of landsliding (Hillman et al., 2018) and vice versa as gas seepage has been shown to weaken the mechanical stability of the slope (Xu et al., 2018). The effect of fresh porewater layers in the sedimentary column has also been suggested to be responsible for gas hydrate dissociation (Riboulot et al., 2018).

The Kerch seep area offshore of the Crimean peninsula (Römer et al., 2012) is one of the rather exceptional places where strong gas emissions into the water column have been observed although the area is located well within the GHSZ. Römer et al. (2012) suggested that gas migration along the base of the gas hydrate stability zone (BGHSZ) together with highly localized heat flux caused the occurrence of seepage at this site. Based on new high-resolution backscatter and seismic data, coupled with porewater geochemistry and gas flux

modeling, we suggest in the present paper that high temperatures close to the seafloor are the result of ongoing gas hydrate formation rather than by advection of warm fluids leading to gas hydrate dissociation. The Kerch seep area actually consists of three individual seeps, each having its own feeder system. The feeder channels are not merging at depth despite two of the seeps being located within coarse-grained surface sediments having likely good lateral permeability. Based on these observations the question to what extent the Kerch seep area and probably many other seeps are determined by underlying structures will be discussed.

2. Geological Setting and Previous work

The Kerch seep area is located on the western margin of the Don-Kuban deep-sea fan in the Black Sea (Fig. 1). The Black Sea is an epicontinental sea that formed by back-arc spreading in the wake of the closure of the Thetys Ocean (Nikishin et al., 2003). The Black Sea basin has undergone several phases of extension and compression and is composed of two sub-basins that are separated by the mid-Black Sea high (Rangin et al., 2002). The Don-Kuban deep-sea fan is part of the eastern Black Sea basin and developed during glacial sealevel lowstand when the Don and Kuban rivers drained sediments through Kerch Strait to the study area (Barg, 2007). During periods of low sealevel input of dense Mediterranean waters through the Bosphorus was interrupted and limnic conditions developed in the Black Sea (Degens and Ross, 1974). These fresh water conditions still persist in the porewater profiles and might have an influence of gas hydrate dissociation (Riboulot et al., 2018; Ker et al., 2019). At present sediment input from these rivers accumulates in the Sea of Azov (Fig. 1).

Numerous sites of gas emission have been observed on the Don-Kuban deep-sea fan by acoustic observations in the water column (Egorov et al., 2003; Starostenko et al., 2010; Römer et al., 2012). These emission sites are most abundant in water depths shallower than

roughly 720 m water depth coinciding with the upper limit of the GHSZ for a bottom water temperature of $\sim 9^{\circ}\text{C}$ and a salinity of ~ 22 (Degens and Ross, 1974; Naudts et al., 2006). It should be noted that this boundary is shifted to a water depth of 665 m for the paleo-limnic conditions with a salinity of ~ 3 encountered below 20 mbsf (Zander et al., 2017). The regional thermal gradient at the Danube deep-sea fan has been modeled at $35^{\circ}\text{C}/\text{km}$ considering the depth of palaeo-bottom simulating reflectors (palaeo-BSRs; Zander et al., 2017). This thermal gradient is significantly higher than $23^{\circ}\text{C}/\text{km}$ measured with temperature probes in the upper ca. 7 meters of sediment in the Kerch seep area (station GeoB15519-1) that was assumed to represent a background value (Römer et al., 2012). Regional heat flow studies in the Dniepr paleo-delta (Kutas and Poort, 2008) showed average background heat flow of $41^{\circ}\text{C}/\text{km}$, but with strong local variations that appear to be common in areas with laterally changing conditions of erosion and deposition (Martin et al., 2004). Few gas emissions are observed in deeper water, with those from the Kerch seep area (Römer et al., 2012) being the most prominent and strongest ones in the region that are unrelated to mud volcanism.

The Kerch seep area is located in 890–940 m water depth at a paleo channel-levee system of the Don Kuban deep-sea fan (Fig 1.). The seep area was depicted on a map by Egorov et al (2003) and has been studied and described by Römer et al. (2012). Römer et al (2012) identified two mound structures with up to 10 m-high relief that showed high seafloor backscatter intensity in ship-borne multibeam data and that were interpreted to be the result of up-doming due to the formation of near-surface gas hydrates. Such gas hydrate pingos have also been described elsewhere (Paull et al., 2007; Sérié et al., 2012) with contrasting formation mechanisms including a form of gas hydrate diapirism (Paull et al., 2007) or localized gas hydrate accumulation above the BSR (Sérié et al., 2012). Römer et al. (2012) also recovered few cm-thick layers of massive as well as thin, platy gas hydrates and authigenic carbonates in gravity cores. However, they did not report on the presence of

extensive authigenic carbonate pavements and they measured strong positive temperature anomalies from which they deduced the advection of warm fluids from below the base of the GHSZ. This input of heat was believed to cause a shift in the GHSZ that favors the release of gas bubbles. Molecular and isotopic compositions indicated that methane in the emanating gas is predominantly of microbial origin.

The presence of methane hydrate pockets in the shallow subsurface at the seeps is believed to be responsible for up-doming of the seafloor (Römer et al., 2012). These authors explain the observed gas emissions into the water column by the advection of warm fluids dissociating the methane hydrates. In the present study we interpret the up-doming of the seafloor as the result of gas pockets in the shallow subsurface, following a rationale developed by Koch et al. (2015) for seeps offshore New Zealand. Gas would migrate laterally towards the flank of the structures, where it escapes into the water column explaining why gas emissions at the seep mounds were observed mainly at their margins.

Fig. 1: A. Location of the study area (red rectangle) in the northern Black Sea off the coast of Crimea. B. Seafloor pick of the 3D seismic survey. The outlines of three seeps in the channel-levee system of the Don-Kuban deep-sea fan are highlighted by black lines.

3. Materials and methods

This study is based on high-resolution sidescan sonar and sediment echosounder, as well as high-resolution 2D and 3D seismic reflection data obtained during RV Poseidon cruise P427 in February/March 2012 (Bialas, 2012). The geophysical data are complemented by porewater geochemistry data derived from gravity and multi-cores taken during RV Meteor cruise M84/2 in March 2011 (Bohrmann et al., 2011).

3.1 3D seismic data

The 3D seismic data were collected using the P-Cable system with 14 parallel 12.5 m-long streamers with a 1.5625 m group interval (Planke and Berndt, 2004). A 105/105 cubic-inch GI-gun (harmonic mode) was used as a seismic source. The 3D seismic dataset covers an area of approximately 3 x 9 km² oriented perpendicular to the channel (Fig. 1). Seismic processing involved geometry correction, trace cleaning, band-pass filtering, normal-moveout correction, and post-stack time migration. In order to fill several small gaps of the cube area that could not be filled satisfactorily during data acquisition, the cube was cut into time slices that were interpolated using the *surface* algorithm of the Generic Mapping Tool (GMT) software (Wessel and Smith, 1998). The time slices were then stacked resulting in a final bin size of 3.25 x 3.25 m. Unfortunately, for one band in the center of the 3D seismic cube the gaps are too large and could not be satisfactorily filled (artefact in figures 1 and 2). Due to the short length of the streamers subsurface velocity information is not available and the data were migrated using a constant velocity of 1500 m/s. Subsurface penetration of the seismic data is on the order of 800–1000 ms TWT beneath the seafloor. Vertical resolution depends on the dominant frequency and thus decreases with depth. A dominant frequency of 100 Hz for the upper 200 m of sediments results in a vertical resolution of approximately 4 m. Data

analysis involved calculating seismic coherence maps for time-gates ranging between 5 and 28 ms.

3.2 Sidescan data

Deep-towed sidescan sonar data were collected using GEOMAR's DTS-1 system consisting of an EdgeTech dual-frequency sidescan-sonar with integrated sub-bottom profiler and custom-built telemetry system. The chirp sidescan signals of 75 kHz center frequency provided a 1.5 km-wide swath and resulted in an across-track resolution of 5.6 cm. Survey speeds of 2.5–3 knots allowed processing the data to 1 m pixel size. Concurrently obtained sediment echosounder data have a frequency content of 2–8 kHz and a vertical resolution of a few decimeters for up to 40 m subsurface penetration in the Kerch area. Towfish navigation was difficult because ultra-short baseline (USBL) navigation was not available and the cable length counter was malfunctioning at times. Luckily, several features common to the sidescan data and other well-localized datasets such as 3D seismic data and the detailed bathymetry of Römer et al. (2012) allowed modeling the likely cable length and resulting in a reasonable layback navigation. We estimate the resulting error of the towfish navigation to be less than 100 meters. The interpretation of the sidescan sonar data and in particular the identification and location of gas bubble streams has been described in detail by Klaucke et al. (2010).

3.3 Hydrate stability modeling

In order to estimate the theoretical BGHSZ we calculated the methane hydrate phase boundary using the SUGAR Toolbox (Kossel et al., 2013) and applied a thermal gradient of 30 °C/km, a bottom water temperature of 8.9 °C, as well as a salinity of 3 (paleo-limnic conditions) for sediments below 20 metres below seafloor (mbsf) using the seafloor picked from the 3D seismic data. The resulting horizon was converted from depth into the time domain in order to compare it with the reflection seismic data. Due to the lack of seismic

velocity data from the Don-Kuban deep-sea fan, we applied a seismic velocity profile from the Danube deep-sea fan, which we consider a similar depositional setting in the Black Sea (Zander et al., 2017).

3.4 Calculation of gas pressure for circular seafloor domes

The gas column height required for creating a circular dome at the seafloor is calculated based on a thin-plate mechanical model which takes into account gravitational forces (Barry et al., 2012; Koch et al., 2015). In this model rising gas pushes a thin layer of sediments upward before being breached. This results in small doming effects at the seafloor (Fig. 2A).

We calculated the required gas column heights as follows:

$$h_g = \frac{(P_{def} + P_{lit})}{g(\rho_w - \rho_{CH_4})}$$

where h_g is the gas column height, P_{def} is the pressure accounting for sediment plate deformation, P_{lit} is the lithostatic pressure of the sediment plate $\rho_s g h$ (ρ_s is the sediment bulk density, h is the plate thickness), g is the gravitational acceleration, ρ_w is the density of water and ρ_{CH_4} is the density of methane.

According to Barry et al. (2012) P_{def} is calculated through

$$P_{def} = \frac{8}{3} \frac{E}{1 - \nu} \frac{h w_{max}}{a^4} \left(\frac{2h^2}{1 + \nu} + w_{max}^2 \right)$$

where E is Young's modulus, ν is Poisson's ratio, a is the plate radius, and w_{max} is the maximum vertical displacement.

3.5 Porewater analysis and transport reaction modeling

Porewater was extracted in 2011 onboard the research vessel in a cold room (4 °C) using a low pressure squeezer (argon gas at ~5 bar). The porewater was then filtered through

0.2 μm cellulose acetate Whatman filters and collected in recipient vials. Dissolved Cl and SO_4 were analyzed through ion chromatography, dissolved hydrogen sulfide by photometry, dissolved calcium by inductively-coupled atomic emission spectroscopy, dissolved methane (headspace technique) by gas chromatography, and dissolved total alkalinity by HCl titration. Porosity was determined by weight difference before and after freeze-drying of wet sediment subsamples. Analytical details and errors are described in Haffert et al. (2013) and Pape et al. (2014). Surface sediment loss of gravity cores was estimated based on solute gradients and profiles of solid constituents (i.e. porosity, organic carbon, calcium carbonate) by comparison with data from multiple cores. Coring locations are shown in Fig. 2A and Table 1.

Table 1: Sampling locations at the Kerch seep area. GC = Gravity Corer, MIC = Mini Corer

Core GeoB no.	Latitude (N)	Longitude (E)	Water depth [m]	Coring depth (cm)	Remarks
15519-1 (GC)	44°37.171'	35°41.763'	896	50–624	Background, Temperature-logger
15519-2 (MIC)	44°37.105'	35°41.759'	896	0–45	Background
15513-1 (GC)	44°37.386'	35°42.164'	878	120–678	Gas hydrates
15513-3 (MIC)	44°37.386'	35°42.164'	878	0–45	
15516-1 (GC)	44°37.230'	35°42.282'	889	220–724	Temperature-logger, Gas hydrates
15516-3 (MIC)	44°37.243	35°42.286	888	0–42	
15518-1 (GC)	44°37.182'	35°42.279	887	50–407	Gas hydrates

In order to determine local rates of methane sinks and sources at the reference and seep sites, we used a numerical transport-reaction model according to Haeckel et al. (2007, 2008) using the parameters and boundary conditions presented in Table 2. The model calculates concentrations of dissolved species chloride, methane, sulfate, calcium, hydrogen

237 sulfide, and total alkalinity (simplified as the sum of HCO_3^- and HS^-) as well as methane
 238 hydrate. According to Berner (1980) early diagenesis of solutes is governed by the partial
 239 differential equation:

$$\frac{\partial \Phi C}{\partial t} = \frac{\partial}{\partial x} \left(\Phi D_s \frac{\partial C}{\partial x} + u \Phi C \right) + \Phi \sum_j R_j$$

240 where Φ is the porosity, C is the concentration of a solute, t is time, $D_s = \frac{D_0}{1-2 \ln \Phi}$ is the
 241 molecular diffusion coefficient of the solute D_0 corrected for sediment tortuosity, u is the
 242 vertical advection velocity of the porewater, and $\sum R_j$ represents the biogeochemical reactions
 243 that are considered (methane hydrate formation, methane gas dissolution, anaerobic oxidation
 244 of methane (AOM), hydrogen carbonate removal by calcium carbonates, and sulfide removal
 245 by iron sulphides). Assuming steady state compaction the porewater advection velocity at
 246 cold seeps can be expressed by adding the contribution of the overpressure-driven upward
 247 fluid flow and the downward-directed burial velocity (Haeckel et al., 2004):

$$u(x, t) = \frac{\Phi_0}{\Phi(x, t)} u_0 - \frac{\Phi_\infty}{\Phi(x, t)} w_\infty$$

248 where u_0 is the velocity of the porewater expelled at the sediment surface ($x=0$) and
 249 w_∞ is the sediment burial velocity at infinite depth where sediment compaction is completed.

250 Steady-state compaction of the sediment is prescribed by an exponentially decreasing
 251 porosity profile (least-squares fitted to the measured data):

$$\Phi(x, t) = \Phi_\infty + (\Phi_0 - \Phi_\infty) e^{-\beta x}$$

252 where ϕ_0 is the porosity at the sediment surface ($x=0$), ϕ_∞ is the porosity at infinite
 253 sediment depth ($x=\infty$), and β is the attenuation coefficient.

254 Precipitation of pore-filling gas hydrate reduces the fraction of the pore volume
 255 available for the porewater, i.e. $\Phi_w(x, t) = \Phi(x, t) - \text{GH}(x, t)$. Consequently, the solid
 256 methane hydrate (GH) is expressed as a volume fraction of the sediment, which leads to the
 257 following transport-reaction equation:

$$\frac{\partial GH}{\partial t} = -w_{\infty} \frac{1 - \Phi_{\infty}}{1 - \Phi} \frac{\partial GH}{\partial x} + \Phi R_H \frac{M_h}{\rho_h}$$

where R_H is the rate of formation of methane hydrate having a molar weight of $M_h = 122.3$ g/mol (i.e. $CH_4 \cdot 5.9 H_2O$) and a density of ρ_h . The following reactions are considered in the transport-reaction model:

a) methane hydrate formation: $R_H = k_H \left(\frac{CH_4}{L_{GH}} - 1 \right)$,

b) methane gas dissolution: $R_G = k_G (L_{MB} - CH_4)$,

c) AOM: $R_{AOM} = k_{AOM} \frac{CH_4}{CH_4 + K_{CH_4}} \frac{SO_4}{SO_4 + K_{SO_4}}$,

d) hydrogen carbonate removal by calcium carbonates: $R_{CaCO_3} = k_{CaCO_3} \left(\frac{Ca \cdot HCO_3}{L_{CaCO_3}} - 1 \right)$,

e) sulfide removal by iron sulfides: $R_{FeS} = k_{FeS} \left(\frac{Fe \cdot HS}{L_{FeS}} - 1 \right)$

with a prescribed Fe profile of $Fe = 0.9e^{0.01(x-z)}$ to reproduce the observed sulfide-depletion depth.

Finally, the temperature, T , of the bulk sediment is considered by an equivalent partial differential equation:

$$\begin{aligned} \frac{\partial}{\partial t} \left([(\rho_w c_w)^{\Phi_w} (\rho_h c_h)^{\Phi_h} (\rho_s c_s)^{1-\Phi}] T \right) \\ = \frac{\partial}{\partial x} \left([\Phi_w \rho_w \lambda_w + \Phi_h \rho_h \lambda_h + (1 - \Phi) \rho_s \lambda_s] \frac{\partial T}{\partial x} + u \Phi \rho_w c_w T \right) + \Phi_w R_H \Delta H \end{aligned}$$

where ρ_i is the density of phase i , c_i its heat capacity, and λ_i its thermal conductivity with $i = w, h$, and s referring to the porewater, methane hydrate, and sediment grains, respectively. ΔH_j is the enthalpy of hydrate formation reaction. In the above equation, the bulk heat capacity is calculated as geometric mean and the bulk thermal conductivity as arithmetic mean of the volumetric contributions of the three phases.

Table 2: Model parameters and boundary conditions used in this study.

Parameter	Notation / Unit	Value
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Length of model domain	z / m	10
Bottom water temperature	$T_{\text{BW}} / ^\circ\text{C}$	8.95
Pressure (based on water depth)	p / MPa	8.85
Bottom water salinity	S	22.3
Porewater density	$\rho_w / \text{kg m}^{-3}$	^(a) 1021
Sediment grain density	$\rho_s / \text{kg m}^{-3}$	^(a) 2500
Methane hydrate density	$\rho_h / \text{kg m}^{-3}$	^(a) 900
Porosity at sediment surface	$\phi_0 / \text{m}^3 \text{m}^{-3}$	^(b) 0.915 (13)
Porosity at infinite depth	$\phi_\infty / \text{m}^3 \text{m}^{-3}$	^(b) 0.613 (19)
Exponential coefficient for porosity decrease	β / cm^{-1}	^(b) 0.0056 (12)
Sediment burial velocity at infinite depth	$w_\infty / \text{cm a}^{-1}$	^(c) 0.02
Advection velocity at sediment surface	$u_0 / \text{cm a}^{-1}$	^(d) 0 / 0
Solubility of methane gas	$L_{\text{MB}} / \text{mM}$	^(a) 113
Solubility of methane hydrate	$L_{\text{GH}} / \text{mM}$	^(a) 93
Effective solubility of calcium carbonates	$L_{\text{CaCO}_3} / \text{mM}^2$	^(d) 90 / 90
Solubility of iron sulfides	$L_{\text{FeS}} / \text{mM}^2$	^(d) 0.0005 / 0.0005
Heat capacity of porewater	$c_w / \text{J g}^{-1} \text{K}^{-1}$	^(a) 4.03
Heat capacity of sediment grains	$c_s / \text{J g}^{-1} \text{K}^{-1}$	^(a) 0.86
Heat capacity of methane hydrate	$c_h / \text{J g}^{-1} \text{K}^{-1}$	^(a) 2.19
Thermal conductivity of porewater	$\lambda_w / \text{W m}^{-1} \text{K}^{-1}$	^(a) 0.59
Thermal conductivity of sediment grains	$\lambda_s / \text{W m}^{-1} \text{K}^{-1}$	^(a) 1.2
Thermal conductivity of methane hydrate	$\lambda_h / \text{W m}^{-1} \text{K}^{-1}$	^(a) 0.58
Reaction enthalpy of methane hydrate formation	$\Delta H / \text{kJ mol}^{-1}$	^(a) -52
Rate constant for CH_4 hydrate formation	$k_H / \text{mM}^{-1} \text{a}^{-1}$	^(d) 0 / 17
Rate constant for CH_4 gas dissolution	k_G / a^{-1}	^(d) 0 / 0.1
Rate constant of AOM	$k_{\text{AOM}} / \text{mM}^{-1} \text{a}^{-1}$	^(d) 5 / 200
Monod constants of CH_4 and SO_4 in AOM reaction	$K_{\text{CH}_4}, K_{\text{SO}_4} / \text{mM}$	^(d) 4 / 4
Rate constant of alkalinity removal	$k_{\text{CaCO}_3} / \text{mM}^{-1} \text{a}^{-1}$	^(d) 0.05 / 5
Rate constant of sulfide removal	$k_{\text{FeS}} / \text{mM}^{-1} \text{a}^{-1}$	^(d) 0.1 / 0.1
<i>Boundary conditions</i>	$x = 0, t$	$x = z, t$
Chloride, Cl	360 mM	160 mM
Methane, CH_4	0.011 mM	^(d) 50 mM / L_{MB}
Sulfate, SO_4	18 mM	0 mM
Hydrogen sulfide, HS	0.3 mM	$dC/dx = 0$
Hydrogen carbonate, HCO_3	4 mM	$dC/dx = 0$
Calcium, Ca	7 mM	40 mM
Temperature, T	8.95 $^\circ\text{C}$	^(d) 9.25 / 9.60 $^\circ\text{C}$

277 ^(a) Values calculated according to algorithms in Kossel et al. (2013) and references therein, ^(b)
278 result of least-squares fitting of porosity equation to data (variance of fit is 0.001, χ^2 is 0.11,
279 and 2- σ standard deviation of fit parameters are given as last digits in brackets), ^(c) Jørgensen

et al. (2004), ^(d) adjusted values of the model runs “reference” / “bubble diss” reproducing the background / seep data, respectively. Differing values for particular simulations are stated in the text and figures.

4. Results

4.1 Seafloor morphology

The combined analysis of bathymetric, towed sidescan sonar and 3D seismic data showed that the Kerch seep area consists of three positive bathymetric features (seep mounds A–C) located at 890 – 940 m water depth (Fig. 2). Whereas seep A was already identified by Römer et al. (2012; ‘small mound’), seep B and seep C were considered as a single feature (‘large mound’) by these authors. Seep mounds A–C have a lateral extent of 150 – 700 m and a relief height of up to 10 m above the surrounding seafloor (Fig. 2A). They are located in a NW-SE trending channel-levee system of the Don-Kuban deep-sea fan (Fig. 1), which has a seafloor gradient of about 1.5 – 2.0° (Fig. 2B). The seeps (Fig. 2) developed on the crest of the western levee (seep A), on the flank of the western levee (seep B), and in the channel bed towards the eastern margin of the channel (seep C).

Fig. 2: A. Shaded relief map based on bathymetry data from the seafloor horizon of the 3D seismic data. Flow direction of inferred sediment-gravity flows is from NW to SE. Red dots show coring locations from Römer et al. (2012). B. Slope angle calculated from this seafloor horizon. C. Seismic amplitudes derived from the seafloor horizon. Location is shown in figure 1.

The seismic amplitudes of the seafloor reflection are higher for the channel-floor deposits compared to the adjacent southwestern levee sediments, but they are not significantly increased for the seep areas except for the eastern part of seep B (Fig. 2C). The levees produce a homogeneous low backscatter return in sidescan sonar images (Fig. 3A). Closer to the channel bed, the seafloor backscatter increases and shows an inhomogeneous distribution combined with patches of very high backscatter. This is broadly consistent with the seismic reflection amplitudes.

The sidescan sonar data indicate that seep B has the strongest relief of the three seep mounds. This is indicated by the presence of shadow zones and strongly increased backscatter intensity of the slopes facing towards the sidescan towfish (Fig. 3A). The central, elevated area of seep B is characterized by low backscatter, similar to the backscatter signal of the surrounding levee. The northern, western and southwestern margins of seep B are marked by a thin zone of homogenous medium backscatter.

Fig. 3: A. Processed sidescan mosaic of the study area. The location of this image is shown in figure 1. High backscatter intensity is white. B. Sidescan image overlain with the outlines of seafloor doming indicating the seeps presented in figure 2. Location of the near-surface bright spots interpreted in the 3D seismic dataset is shown in red. The extent of acoustic blanking in the subbottom profiler records is shown by blue lines, and hydroacoustic anomalies in the water column indicative for rising gas bubbles ('flares') are highlighted by yellow arrows.

Seep C does not produce a clear image in the sidescan sonar data, especially at its western margin in the center of the channel (Fig. 3). It is characterized by sharp medium to low backscatter contrast of the eastern and southern margins towards the surrounding sediments. In general, seep C shows less indication for topographic relief compared to seep B, and less elevation above the surrounding sediments. The surface inclination of seep C is remarkably uniform with inclination angles of less than 1.5° (Fig. 2B). The smallest seep of the study area, seep A on the crest of the levee, is characterized by four northwest to southeast trending linear features of high backscatter intensity with patches of low backscatter in between. Two minor elevated backscatter features could be identified in the sidescan mosaics: One on the western margin of seep B, and one on the northeastern margin of seep C (Fig. 3).

4.2 Subsurface analysis

Four main seismic units can be identified in the seismic data, but the absence of well data limits the geological interpretation of the area. Seismic unit 1 is situated in the bathymetric depression bounded by seismic unit 2 and comprises the channel deposits of the channel-levee system (Figs. 4C, D). It has a width of about 1500 m in SW to NE direction. The thickness of the unit ranges between 50 ms two-way traveltime (TWT) towards the margins and 120 ms TWT at the center. The seafloor amplitudes at the top of unit 1 are the highest observed in our data (Fig. 2C). The reflectors are discontinuous and often chaotic but some reflectors onlap onto the surrounding seismic unit 2. Their amplitudes are variable, and several scattered reflections are of very high amplitude with either the same or reversed polarity compared to the seafloor.

Seismic unit 2 comprising the levee sediments is found on either side of the channel (Figs. 4C, D). It consists of numerous well-stratified continuous reflections that dip towards the central axis of the ridge-like structures. On the western side of the channel, unit 2 is about 150 ms TWT below seafloor (bsf) thick. Unit 2 features numerous near-vertical displacements

of reflectors with offsets of up to 10 ms. Some of these discontinuities reach lateral extensions of up to 500 m (Fig. 4E), and the general orientation follows the strike direction of the levees (NW to SE). The shallow subbottom profile shows that the discontinuities reach up to the seafloor as close as 2–5 m (Fig. 4F).

Seismic unit 3 is situated below units 1 and 2. This unit varies in thickness between 10 ms TWT at the intersection point between units 1 and 2, and a maximum of about 40 ms TWT below unit 2. The seismic amplitudes are generally high and the polarity of the major reflections in unit 3 is reversed compared to the seafloor reflection.

Seismic unit 4 comprises the sediment column underneath unit 3. Unit 4 is generally seismically transparent, but several isolated and elongated reflectors can be identified. These reflectors are mostly parallel to the reflectors in unit 3 and the seafloor, and some are characterized by a reverted polarity compared to the seafloor with elevated amplitudes (Fig. 4A-D).

4.3 Doming seep morphology

The seafloor domes sit atop shallow high amplitude patches with reversed polarity (Fig. 4C, D). The lateral extent of these patches is limited to the seafloor domes (Fig. 3B) except for seep A, of which the outer margins are difficult to map due to the smaller bathymetric elevation compared to seeps B and C. The thickness of the high amplitude patches is in the range of 20–30 ms TWT (Fig. 4C, D). The subbottom profile shows blanking of the acoustic signal in 10 m depth below the seep sites (Fig. 4F). Below seep A two NW to SE trending linear features of about 100–150 m length are observed (Fig 4E) and extend vertically through unit 3. In the similarity time slices, these two anomalies differ from the typical narrow fault signatures as they both consist of two parallel lines of low similarity and a line of high similarity in between. This attribute signature is caused by an upward bending of the well-stratified reflectors, which is most likely a velocity effect (pull-up) and does not reflect the stratigraphy.

Fig. 4: A. 3D inline across the channel-levee system and seeps A and C. B. 3D inline across seep B. For profile locations refer to figure 2. Vertical exaggeration is roughly 15 times. C. Interpreted section of A highlighting three main seismic units: Channel facies (unit 1, blue), levee facies (unit 2, red), unit 3 underlying the channel-levee system (orange), and unit 4 (green). Bright spots underneath seeps A and C are highlighted in yellow (lateral extent shown in figure 3B) and the most likely migration pathway (vertical anomaly) for the free gas is shown as yellow arrows. Several faults were identified in the levees (black lines). In the absence of a BSR, the theoretical depth of the BGHSZ was calculated for the regional geothermal gradient (red line). D. Interpreted section of B. E. Similarity time slice of the area.

Narrow elongated anomalies of low similarity are present within all of the seeps (yellow arrows). The faults in the levee facies are apparent as sharp elongated zones of low similarity (white ellipse). The location of the time slice is shown as a dashed line in C and D. High similarity appears white. F. Subbottom profile across the seeps A and C showing acoustic blanking and levee deposits showing a prominent fault. Lateral extent of the blanking is shown in figure 3B. Vertical exaggeration is 10 times.

Seep B on the levee flank is also situated above well-stratified levee sediments. A vertical narrow low-amplitude anomaly of about 150 m length in NW to SE direction extends vertically into unit 3 (Fig. 4E). Seep C in the channel bed is of relatively low similarity compared to the other two seeps. Between seep C and unit 3, the chaotic layered reflectors of the channel bed show several patches of high amplitudes and inversed polarity. Towards the eastern margin of seep C, the similarity slices reveal a similar vertical narrow anomaly of about 100 m length which connects the shallow bright spot with unit 3. The lower termination of these vertical anomalies cannot be determined unequivocally with the available seismic data.

A BSR could not be identified in the seismic dataset. However, the modeled BGHSZ shows that at least for seep A (and to a lesser extent for seep B), the vertical anomaly appears to extend down below the GHSZ (Fig. 4C).

4.4 Porewater geochemistry

Two cores taken at a reference site (GeoB15519-1, 15519-2, Fig. 2A) show a decrease in Cl^- concentration from ~360 mM at the sediment/water interface to ~225 mM in 600 cm depth (Fig. 5, black symbols). This decrease in Cl^- is typical for Black Sea sediments where the progressing diffusion front of the salt ions from the marine bottom waters with a salinity of 22.3 has currently reached a sediment depth of 20–25 mbsf in limnic deposits with a porewater salinity of ~3 (Soulet et al., 2010). Sulfate decreases almost linearly from ~18 mM at the sediment surface until it is completely consumed in about 300 cmbsf. At this depth methane concentrations under ex situ conditions start to increase from less than 0.011 mM to

values of more than 1–2 mM, sometimes having preserved even up to 8 mM in the collected sediments (Fig. 5, black symbols). Total alkalinity and hydrogen sulfide show a maximum around the depth of sulfate depletion (i.e. HS^- : ~2 mM, TA: ~14 meq/l), while dissolved calcium concentrations are gently increasing below this depth (Fig. 5). At station GeoB15519-1 a background temperature gradient of 30 °C/km has been calculated based on thermometers attached to the gravity core barrel (Fig. 5).

A good match of all the background data was achieved by simulating a steady-state situation (Fig. 5, solid line), where the methane that is primarily produced by microbial degradation of organic matter below the depth of the model domain is diffusing upwards and gets oxidized by the downward diffusing sulfate via AOM. Subsequently, the produced alkalinity induces calcium carbonate precipitation and the produced sulfide is removed by iron sulfide precipitation. The downward diffusion of the conservative tracer chloride was mimicked by fixing the lower boundary condition at a concentration of 160 mM at 10 mbsf. Hence, the downward progressing diffusion front was not simulated explicitly in this study, but an adequate boundary condition was determined that allows a later comparison to the modeled Cl profiles affected by different modes of seepage, i.e. dissolution of gas bubbles and upward fluid flow, or gas hydrate formation.

Fig. 5: Depth profiles of temperature (taken from Römer et al., 2012) and porosity as well as solute concentration of Cl^- , SO_4^{2-} , CH_4 , total alkalinity (TA), HS^- and Ca^{2+} cores taken at a reference station (black symbols) and the seep sites (red, magenta and blue symbols). Simulation results of the background situation (solid line) and model scenarios of the seep situation (dashed and dotted lines) are also shown (see description in the text). Römer et al. (2012) determined lower geothermal gradients of 23 °C/km (GeoB15519-1) and 49 °C/km (GeoB15516-1) for the same temperature measurements. Note that for CH_4 and SO_4 only the upper 400 cmbsf are shown to improve the presentation of the near-surface gradients at the seep sites. Cores GeoB15513-1 and GeoB15516-1 were sampled for gas hydrates and not analyzed for CH_4 distribution. Core locations are shown in figure 2A.

Cores taken at the seeps A (GeoB15516-1, 15516-3, 15518-1; Fig. 2A) and seep B (GeoB15513-1, 15513-3; Fig. 2A) show a shallow sulfate penetration depth of ~10 cmbsf (Fig. 5) and a steep methane increase just below this depth. TA and HS^- peak around the same depth and are also significantly higher (TA of up to ~27 mM, HS^- of up to ~12 mM) than at the reference station. Chlorinity is significantly increased compared to background values down to a depth of about 700 cm. Calcium concentrations decrease from ~7.5 mM at the sediment surface to about 4 mM in the upper 50 cmbsf before they gently increase again to about 15 mM at the base of the gravity cores. At all seep sites, methane hydrates in the form of platy chips and veins were recovered in different sections below ~200 cmbsf in cores 15513-1, 15516-1, and below ~350 cmbsf in core 15518-1. The much shorter (<400 cm long)

core GeoB15518-1 behaves slightly different, because it showed some gaps in the sediment sequence that are possibly the result of gas expansion during core recovery. Although showing similar geochemical trends data from this core are considered less reliable except for in-situ temperature measurements taken at the outside of the core barrel. From core GeoB15516-1 at the margin of seep A a temperature gradient of ~ 65 °C/km has been calculated (Fig. 5, magenta triangles).

We modeled upward methane transport via fluid flow of methane-saturated porewater (Fig. 5, dotted line) and compared it to methane dissolution from a free gas phase (Fig. 5, dashed line) in order to determine whether seepage is governed by gas bubble ascent or water expulsion. For both scenarios a match of the dissolved sulfate and methane profiles is possible, however, the other solutes (particularly Cl and Ca) can only be reproduced in the gas bubble dissolution scenario (Fig. 5). Upward porewater advection expectedly leads to concave/convex-shaped profiles that do not match the other solute data. In both model scenarios, an increased overall heat flow from depth, realized by increasing the lower boundary value from 9.25 °C to 9.6 °C, is needed to reproduce the temperature measurements of core GeoB15516-1.

At the seeps the increased methane flux raises the maximum AOM reaction rate (~ 49 mM/a at ~ 6 cmbsf) by almost 500 times compared to the background situation (~ 0.1 mM/a at ~ 265 cmbsf), thereby inducing a carbonate precipitation rate of 0.94 mM/a that is ~ 150 times higher than at the reference site (i.e., 0.006 mM/a). Interestingly, the maximum sulfide removal rate stays at the same level (0.04–0.05 mM/a), but the rate profile broadens at the seeps.

5. Discussion

5.1 Free gas distribution in the subsurface

Both the geoacoustic data of this study and the acoustic flare data published by Römer et al. (2012) show relatively little gas venting at the Kerch seep area, although the latter authors have come to a different conclusion. These authors consider the Kerch seep a high flux based on the number of observed gas flares and many hundreds of small-sized depressions, separated only by several centimeters from each other that they interpret as individual gas bubble release spots. Gas venting is a temporally highly variable process on timescales of days, weeks, or even years (Greinert et al., 2006; 2010; Klaucke et al., 2010) and, hence, gas bubble release could have been significantly increased or reduced during the 4 days of hydroacoustic survey/coring in 2011 and the 9 days of geophysical survey in 2012. However, the gas bubbling evidence for the Kerch seep is much less than for the high-flux areas offshore Georgia in the Black Sea (Klaucke et al., 2006; Nikolovska et al., 2008).

The subsurface data also show a contrasting picture with only little evidence for gas accumulations. In particular there is no accumulation of gas below the inferred BGHSZ, although a few areas of elevated amplitudes are present below this zone. Unequivocal seismic evidence for free gas, i.e. bright spots with reversed seismic polarity, are only visible right underneath the individual seeps, which, together with the documented microbial origin of methane in the venting and hydrate-bound gas (Römer et al., 2012), points to gas pockets within the uppermost 10 metres of the sediment column. Such shallow gas pockets are a common phenomenon at cold seeps (Schwalenberg et al., 2010; Reitz et al., 2011; Koch et al., 2015). Acoustic blanking in the subbottom profile indicates that the gas is trapped as shallow as 10 m below the seafloor at the small mound on the crest of the ridge (Fig. 4F). The different bright spots are not interconnected and they do not extend beyond the surface expression of the seeps. Each seep appears to have its own feeder system despite their close proximity. Clear migration pathways are only imaged for seep A, where gas migrates along almost vertical faults through the levee sediments (Fig. 4A, C). A faint acoustic anomaly is

also present below seep B (Fig. 4B), but it is not clear whether this vertical zone of reduced amplitudes is real or a seismic imaging artefact caused by the overlying free gas.

5.2 Gas transport mechanism

The background sediments at sites GeoB15519-1 and -2 are characterized by upward diffusion of dissolved methane at a flux of $\sim 7 \mu\text{mol cm}^{-2} \text{ a}^{-1}$ that is consumed by downward diffusing sulfate via AOM. In contrast, the methane flux at the seeps (GeoB15513-1, 15516-1, 15518-1) is 50-fold increased ($\sim 350 \mu\text{mol cm}^{-2} \text{ a}^{-1}$) and has shifted the AOM reaction front to a sediment depth of ~ 6 cmbsf. Our numerical transport-reaction simulations (Fig. 5) clearly demonstrate that this methane is delivered by rising gas bubbles and not by upward advection of porewater (as initially suggested by Römer et al. (2012) on the basis of acoustic water column surveys and sediment temperature measurements alone), because as a physical process the latter would produce concave/convex-shaped depth-profiles for all solutes (Fig. 5, dotted line), not only for sulfate and methane. However, dissolved Cl, Ca, total alkalinity, sulfide, and other constituents, exhibit much more gentle curvatures than an upward advection velocity of 5 cm/a would predict. Such advection rates are required to match the methane flux at the seeps. In contrast, at the seeps the increased AOM rate is fueled by the dissolution of the rising gas bubbles, which is highest (up to 10 mM/a) close to the seafloor in the diffusive mixing zone of bottom- and porewater, where porewaters are highly undersaturated with respect to methane. Transport of methane by individual gas bubble streams is also supported by the morphology of the seafloor at the Kerch seeps, where large areas are covered by small, cm-size depressions created by the individual emitted bubble streams (Römer et al., 2012). In the absence of bioturbation in the deep anoxic Black Sea and low bottom current velocities these structures can prevail for extended periods of time.

5.3 Gas doming

Two different dome geometries were tested to calculate the pressures required for the dome structures and the height of the gas column underneath. These geometries are circular shaped in order to closely match seeps B and C and the smaller-sized seep A (Figs. 2A, 6). For the plate thickness we assumed a maximum of 10 m according to the observed blanking in the subbottom profile at this depth, but also tested shallower plate thicknesses between 1–10 m. According to Römer et al. (2012) the smaller sized seep A is elevated by “a few meters” above the surrounding seafloor, and we consequently used an estimated elevation $w_{\max}$ of 5 m. For the larger seep, Römer et al. (2012) noted an elevation $w_{\max}$ of 10 m. The Young’s Modulus E was estimated to range between 140 and 3000 kPa for fine-grained, near-seabed marine sediments (Barry et al., 2012). Koch et al. (2015) further noted that after Hamilton et al. (1971), E can be approximated with 350 MPa for silty clays in ~1 km water depth. We consequently used this value as an upper limit in our calculation. All parameters used in this calculation are given in Table 3.

Fig. 6: A. Geometry of the plate bending after Barry et al. (2012) for a circular structure with radius a , plate thickness h , and a maximum vertical displacement $w_{\max}$. The doming is forced through buoyancy of a rising gas column from underneath. B. Circular radii as approximation of seeps A and C.

Table 3: Parameters for the calculation of the gas column height which is required to explain sediment up-doming

Parameter	Notation	Value/range	Reference
Poisson’s ratio	ν	0.49	-
Sediment bulk density	ρ_s	2500 kg/m ³	clayey sediments
Porewater density	ρ_w	1030 kg/m ³	-

Density of methane	ρ_{CH_4}	75.93 kg/m ³	Calculated for a water depth of 900 m and a temperature of 9 °C (Kossel et al., 2013)
Plate thickness	h	1–10 m	Gas accumulations as shallow as 10 mbsf indicated by subbottom profile
Radius	a	75 m (representative for seep A); 240 m (representative for seep C)	-
Maximum elevation	w _{max}	5 m (seep A), 10 m (seeps B & C)	Römer et al. (2012)
Young's Modulus	E	140 kPa–350 MPa	Koch et al., 2015, Barry et al. (2012), Hamilton et al. (1971)

In order to produce the observed doming effect gas columns below seep A and seep C require column heights on the order of 3–27 m for both seep geometries, assuming a Young's Modulus of 140 kPa (Fig. 7). The use of a larger Young's Modulus (350 MPa) requires larger gas column heights of 4–86 m in case of the seep A geometry (Fig. 6), and 3–28 m for the seep C geometry (the gas column height is thereby increasing marginally compared to the lower Young's Modulus of 140 kPa).

Fig. 7: Gas column height versus plate thickness for seeps A and C, shown for two different Young's Moduli E.

The methane transport by gas bubbles is consistent with sediment up-doming due to buoyancy of shallow seated gas at the three individual seeps. This is further supported by side scan sonar data and ROV photographs (Römer et al., 2012; this study) that did not show signs of carbonate build-ups in order to explain the dome structures with similar observations made at gas domes offshore New Zealand (Koch et al., 2015). The required gas column heights for the seep domes are in the order of 3–27 m for both of the tested geometries. Seep C formed within a submarine channel with most likely non-cohesive, coarse-grained channel-fill deposits (Klaucke et al., 2004) that would have values of Young's Modulus more likely close to the low end within the 140 kPa to 3000 kPa range discussed by Barry et al. (2012). A variation of Young's Modulus within this range changes the resulting gas column heights only on a centimeter-scale. Compared to the thickness of the gas signatures identified in the seismic data (20–30 ms TWT, which corresponds to a thickness of 15–22.5 m using a velocity of 1500 m/s knowing that with the presence of gas seismic velocity is reduced by an unknown factor), these calculated gas column heights (3–27 m assuming $E = 140\text{--}3000\text{ kPa}$) are consistent and plausible. Fine-grained sediments on the western levee would have a higher Young's Modulus requiring a thicker gas column. Seismic data (Fig. 4A) show high amplitudes below seep A and velocity pull-ups restricted to two faults. The velocity pull-ups do not cover the entire extent of seep A. If the velocity pull-ups are the result of gas hydrates, the hydrates would be restricted to the faults. The results of the gas column modeling strongly suggest that the gas flux driven by buoyancy is causing the sediment up-doming at the seep sites. Layers of massive gas hydrates would probably support the permeability barrier and result in sealing at the center of the seeps, even if the hydrate layers are thin. In fact, only small amounts of gas hydrate have actually been recovered. This possible sealing by gas hydrates would force the free gas towards the margins of the domes as suggested by Römer et

al. (2012) and observed with the sidescan sonar data showing small flares at the margin of the seeps (Fig. 3), again equivalent to previous findings in a seep area offshore New Zealand (Klaucke et al., 2010; Koch et al., 2015). The sealing and sediment strengthening effect already takes place with thin gas hydrate layers, whereas seafloor doming due to gas hydrates requires much thicker gas hydrate layers, for which we do not see any evidence.

5.4 Gas hydrate formation

For thermodynamic reasons methane hydrates start to form, once the methane gas enters the GHSZ. However, under conditions of high gas fluxes as they occur at permeable chimney and pipe structures, the free gas can break through the GHSZ and is emitted into the water column, thus forming a cold seep. Liu et al. (2019) demonstrated that in this situation high hydrate saturations are expected to be built up in the upper few tens of meters below the seafloor, thereby producing a sub-seafloor salinity peak and elevated temperatures.

While the data at the Kerch seep sites do not exhibit significant salinity enrichments, increased thermal gradients of up to 65 °C/km are observed within the surface sediments, which is significantly higher than background values (Fig. 5; Römer et al., 2012, their Fig. 10). As discussed in section 5.2, these elevated thermal gradients cannot be explained by upward fluid flow, and variations in background heat flow in such magnitude on such small spatial scales is not supported by the overall geological setting in the area and in particular the absence of a heat source at depth. We therefore investigated the possibility of latent heat released by gas hydrate formation to explain the observed temperature profiles. Our numerical simulations indicate that methane hydrate formation rates of at least 2.1–2.4 mM/a are required to raise the temperatures by 0.2–0.4 °C above background values (Fig. 8). The produced temperature elevation would prevail for almost 500 a (blue line in Fig. 8) before conductive heat dissipation starts to dominate as the hydrate formation rates decline. This decline (Fig. 8) is invoked by the reduction in porosity because until this time gas hydrates

have filled up ~20 % of the available pore space (i.e. ~13 % of the bulk sediment volume). In parallel, salt exclusion during hydrate formation also elevates the salinity of the porewater. At the modeled hydrate formation rates and palaeo limnic porewater only a minor chloride enrichment is produced, which lies within the observed data scatter (Fig. 8) and consequently does not contradict the model predictions by Liu et al. (2019).

Of course, the observed temperature elevation can also be produced by applying faster hydrate formation rates. However, this would drastically shorten the time window, during which the data could be observed, as the resulting temperature anomaly would further increase. Hence, the presented simulation appears to be much more likely than any other scenario.

Fig 8: Time-dependent evolution of simulation results for gas hydrate formation coupled to methane gas bubble dissolution in comparison to the observed data of dissolved chloride concentration, temperature, and porosity. Data are shown with the same symbols as in figure 5. Note: modeled temperature gradient increase for time spans <100a then decrease again.

5.5 Age of the seeps

The simulations also allow constraining the age of the seeps. Cold seeps can be relatively long-lived features that persist over tens of thousands or even hundreds of thousands of years (Teichert et al., 2003; Liebetrau et al., 2010) with probably more intense seep activity during times of global sealevel lowstands. For the Black Sea, Gulin et al. (2003)

suggested that carbonate structures growing at methane seeps originate from 5300–2900 years before present based on ^{14}C -dating of seep carbonates. However, authigenic carbonates at seeps incorporate ^{14}C -dead carbon originating from the seeping CH_4 (Winckler et al. (2002). ^{14}C dates consequently have to be taken with care as they define a maximum age. In addition, seeps lacking authigenic carbonate pavements such as the Kerch seep area are likely even younger and should be considered seeps that are still developing. The observation of a substantial gas pocket underneath the seep, active gas hydrate formation and limited gas bubbling into the overlying waters would corroborate this interpretation. Our gas hydrate formation simulations suggest that the Kerch seeps are younger than ~500 years (Fig. 8). Shorter time spans of less than 100 years match observed porosities even better but fail to reproduce the measured chloride concentrations, while longer time spans better reproduce chloride concentrations but generate hydrate saturations (up to 25 % core volume; Fig. 8) that exceed those observed by Römer et al. (2012). In addition, if the seeps are older than 500 years excess heat from gas hydrate formation would start to dissipate and the observed temperatures would be lower than observed.

6. Conclusions

The combined analyses of the Kerch seep area carried out in this study is based on high-resolution 3D seismic data, deep-towed sidescan sonar data, and porewater analysis. The seep area is located in around 900 m water depth well within the GHSZ and comprises of three seafloor domes. Each of the domes sits on top of isolated gas pockets which are fed with methane migrating along individual narrow pathways through the GHSZ. Our calculations show that gas overpressure is likely the driver for the up-doming of the sediments, and the required gas column heights are of the order of 3–27 m, which fits well with the observed gas signatures in the seismic data. Methane is transported through the sediment column towards the seeps in the form of gas bubbles and not through fluid flow. Consequently, elevated

temperatures in the surface sediments of the seeps are the result of active gas hydrate formation and not due to elevated heat flow from below. Based on our observations of the surface expression of the seeps and geochemical modeling, we also suggest that the Kerch seep area is a rather young seep system with a maximum age of 500 years. Consequently, the Kerch seep site shows gas hydrate formation in very shallow subsurface depth and not at BSR depth. However, the geochemical database at our disposal is limited and requires additional future sampling.

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Conflict of Interest

The authors declare no conflict of interest

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