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An update on dissolved methane distribution in the subtropical North Atlantic Ocean

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Abstract. Methane (CH₄) is a potent greenhouse gas and plays a significant role in recent increasing global temperatures. The oceans are a natural source of methane contributing to atmospheric methane concentrations, yet our understanding of the oceanic methane cycle is poorly constrained. Accumulating evidence indicates that a significant part of oceanic CH₄ is produced in oxygenated surface waters as a by-product of phytoplanktonic activity. This study focused on the subtropical North Atlantic Ocean (26° N, 80° W and 26° N, 18° W) where the distribution of dissolved CH₄ concentrations and associated air–sea fluxes during winter 2020 were investigated. Water samples from 64 stations were collected from the upper water column up to depths of 400 m. The upper oxic mixed layer was oversaturated in dissolved CH₄ with concentrations ranging 3–7 nmol L^{−1}, with the highest concentrations of 7–10 nmol L^{−1} found to the east of the transect, consistent with other subtropical regions of the world's oceans. The high anomalies of dissolved CH₄ were found to be associated with phosphate-depleted waters and regions where the abundance of the ubiquitous picocyanobacteria *Synechococcus* and *Prochlorococcus* were elevated. Although other phytoplanktonic phyla cannot be excluded, this suggests that cyanobacteria contribute to the release of CH₄ in this region. The calculation of air–sea fluxes further confirmed the subtropical North Atlantic Ocean as a source of CH₄. This study provides evidence to corroborate the key role that picocyanobacteria play in helping to explain

the oversaturation of CH₄ found in surface mixed layer of the open ocean, otherwise known as the “ocean methane paradox”.

1 Introduction

Since the industrial revolution, the average global temperature has increased at the fastest rate in recorded history, primarily driven by growing emissions of greenhouse gases (GHGs). Among them, methane (CH₄) is considered the second-largest contributor to Earth warming after carbon dioxide (CO₂), with an atmospheric concentration of 1866 ppb (IPCC, 2021). Over the last 50 years, CH₄ concentrations have increased by 20 % (Karl et al., 2008; Rhee et al., 2009) – and are expected to rise further by approximately 2 % yr^{−1} (Dang and Li, 2018).

Oceans were generally considered to be a minor contributor to the total global CH₄ budget, yet recent calculations indicate that oceans could emit 6 to 17 Tg CH₄ yr^{−1}, i.e., 1 % to 10 % of the total natural emissions (Weber et al., 2019). This large variability reflects the great uncertainty in the contribution of natural sources due to a lack of data and understanding of the sources and controls of oceanic CH₄ emissions.

The marine flux of CH₄ results from the balance between production and oxidation processes; for instance, the microbial anaerobic oxidation of CH₄ (AOM) in sediments sig-

nificantly decreases CH₄ fluxes to the atmosphere, thus representing an important carbon sink in the ocean (Oppo et al., 2020). In fact, the marine methane release is dominated by shallow coastal environments including estuaries (up to 75 %; Weber et al., 2019). In marine sediments CH₄ can be released via microbial anaerobic methanogenesis as a consequence of the degradation of organic matter; it accumulates in the sediment, eventually forming gas hydrates, which may then be released into the overlying water column. Under the influence of pressure and temperature CH₄ diffuses out of the sediment and ebullition carries CH₄ to the atmosphere (Weber et al., 2019).

In open oxygenated waters, the primary mechanism controlling the CH₄ emissions is aerobic methanotrophy that converts CH₄ into CO₂ (Weber et al., 2019). However, this process could be overcome by in situ production of CH₄ in upper oxic waters that can significantly contribute to marine CH₄ fluxes to the atmosphere. Typical CH₄ depth distribution in the open ocean indicates a general oversaturation in the mixed layer (Reeburgh, 2007). In the surface waters of the Pacific Ocean (Weller et al., 2013), the Indian Ocean (Bui et al., 2018) and the Atlantic Ocean, values of 2–5 nmol L⁻¹ and a maximum of 10 nmol L⁻¹ were measured near the surface (Scranton and Brewer, 1977). These observations make the global ocean a net source of CH₄ to the atmosphere with a weighted supersaturation of 120 % (Kock and Bange, 2007). Exemptions are the Southern Ocean and the central Arctic Ocean, where surface waters are undersaturated in CH₄, either due to extensive upwelling supplying CH₄-depleted water to the surface (Bui et al., 2018) or the limitation of air–sea exchanges by ice cover (Weber et al., 2019).

The oversaturation of the surface mixed layer, commonly known as the “ocean methane paradox” (OMP), was explained by the result of methanogenic activity from *Archaea* living within anaerobic cavities of zooplankton guts and anaerobic micro-environments inside sinking particles of organic matter (Reeburgh, 2007). Initially, only microbes from the *Archaea* domain were thought to have the capability of producing CH₄ under strict anaerobic conditions. Although one cannot exclude this process to explain the methane paradox (Schmale et al., 2018) (Stawiarski et al., 2019), an increasing number of studies have revealed relationships between CH₄ anomalies in surface waters and the presence of certain phytoplanktonic groups such as coccolithophores (Lenhart et al., 2016a) and cyanobacteria (Bižić et al., 2020).

Cyanobacteria are ubiquitous to every aqueous environment on Earth in both illuminated and dark water bodies (Percival, 2014). In the open ocean, small-sized picophytoplankton of the genera *Prochlorococcus* spp. and *Synechococcus* spp. account for ~ 80 % of the total phytoplanktonic chlorophyll *a* (Hickman et al., 2010) and could represent up to 8.5 % and 16.7 % of the ocean net primary production (ONPP), respectively (Flombaum et al., 2013). Generally, nutrient limitation sets the upper limit for primary production and the distribution of *Prochlorococcus* and *Syne-*

chococcus; the oligotrophic subtropical North Atlantic is in fact nitrogen–phosphorus (N–P) co-limited (Harvey et al., 2013), and hence cyanobacteria need to acquire these nutrients from alternative sources. New nitrogen to the surface ocean is introduced via biological nitrogen fixation by diazotrophs, which can supply up to 163.2 Tg of new nitrogen per year globally (Wang et al., 2019). With regard to P, *Prochlorococcus* and *Synechococcus* mostly depend on the remineralization of dissolved organic phosphorus (DOP) via hydrolytic enzymes (e.g., alkaline phosphatase; Muñoz-Marín et al., 2020). Additionally, evidence is mounting that cyanobacteria are major sources of semi-labile dissolved organic matter (DOM) phosphonates (Repeta et al., 2016). The bacterial degradation of methylphosphonates (MPn) releases CH₄, and therefore cyanobacteria are thought to play a key role in the global marine CH₄ flux. In situ production of CH₄ is a repercussion from different metabolic pathways, including the conversion of methylated substrates, and is induced by environmental stress (e.g., nutrient supply and variations in temperature, salinity and light attenuation). As such, CH₄ may be the by-product of the methylphosphonate remineralization in phosphate-stressed surface waters, i.e., the MPn way (Karl et al., 2008; Bižić et al. 2020). Due to the strong depletion of inorganic phosphorus in oligotrophic areas in the Atlantic and the Pacific Ocean, cyanobacteria use the organic phosphonates as a P source (Feingersch et al., 2012), leading to the release of methyl groups in the water that are rapidly converted into CH₄ (Beversdorf et al., 2010). In contrast, nitrate availability might control CH₄ production in phosphate-replete surface waters. While in P-limited waters cyanobacteria use methylphosphonate as a nutrient source and hence release CH₄, in N-limited waters, CH₄ may result from the breakdown of DMSP (dimethylsulfoniopropionate), a subsequent demethylation of DMS (dimethyl sulfide) and inherently the oxidation of dissolved methyl groups (Florez-Leiva et al., 2013). Other environmental parameters, such as variations in temperature or light attenuation, may also influence CH₄ formation, although data are lacking to fully understand the metabolic pathways leading to CH₄ production.

In the current context of climate change and warming ocean waters, we are likely to expect a shift in the community of primary producers towards smaller-sized cyanobacteria such as *Prochlorococcus* and *Synechococcus* (van de Waal and Litchman, 2020), with a concurrent decrease in total biomass in the open ocean (Marinov et al., 2010). Yet, with accumulating evidence on the importance of the cyanobacterial contribution to CH₄ production and given the important role of CH₄ as a potent GHG, it is crucial to intensify the monitoring and investigating of CH₄ production and fluxes in surface oceanic waters in order to feed a global database (Bange et al., 2009). There is a lack of data to comprehend the current and future role of the Earth’s oceans and primary production in relation to the global CH₄ budget.

In the present paper, we provide an update on the dissolved CH₄ concentrations and air–sea fluxes of CH₄ in surface wa-

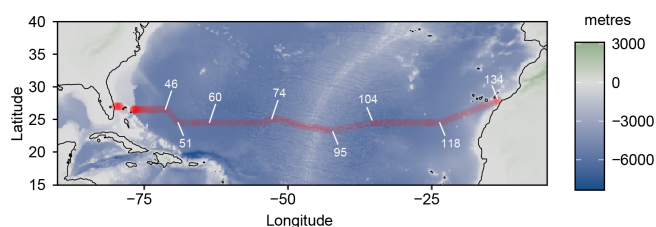


Figure 1. Cruise track of the JC191 24N expedition from Fort Lauderdale (Florida), USA, to Santa Cruz de Tenerife, Spain (January–March 2020). White numbers indicate some of the total 135 CTD stations.

ters of the subtropical North Atlantic Ocean as part of the JC191 hydrographical cruise (RRS *James Cook*, January–March 2020). Furthermore, we present data on the distribution of the two dominant cyanobacteria, *Prochlorococcus* and *Synechococcus*, to highlight the contribution of the planktonic community to the CH_4 flux. More specifically, our objectives were to confirm the presence of CH_4 anomalies in surface waters and the associated air–sea fluxes, as well as to propose possible mechanisms and sources controlling the CH_4 distribution by examining the relationships between physicochemical and biological parameters.

2 Methods

2.1 Water sampling and analysis for dissolved gas concentrations

Seawater samples were collected during the JC191 hydrographical cruise (as part of the GO-SHIP program, PI A. Sanchez-Franks; Sanchez-Franks, 2020) on board the RRS *James Cook* between January and March 2020 along a west–east transect in the subtropical North Atlantic from Fort Lauderdale (Florida), USA, to Santa Cruz de Tenerife, Spain, on the nominal 24°N parallel. 64 profiles (out of 135 stations occupied by CTD – conductivity, temperature, depth – casts in total) from the surface to 400 m depth (or full depth for the shallower continental margin stations) were sampled (Fig. 1).

Water samples for dissolved CH_4 measurements were collected into 20 mL headspace vials using a 24-Niskin-bottle rosette equipped with a SBE911+ CTD, an AquaTracka III fluorometer (Chelsea Technologies) and a dissolved oxygen sensor (Sea-Bird SBE 43). Dissolved gas samples were poisoned with $\text{Hg}(\text{Cl})_2$, then fitted with Teflon stopcocks and crimp-sealed under exclusion of any air bubbles. Samples were immediately stored at 4°C after sampling until analysis onshore at the Station Biologique de Roscoff.

Gas extraction and analyses were performed using a Shimadzu headspace sampler (HS-20) connected to a gas chromatograph (Shimadzu GC-2030) fitted with a barrier discharge ionization detector (BID) and a 30 m SH-Rt-MSieve 5A column. With this setup, headspace extraction is entirely

automated with pressurization of the sample up to 2 bar, heating at 90°C and equilibration for 7 min. An aliquot of the gas sample was transferred to a 1 mL injection loop, maintained at 150°C and injected into a 50°C heated column. Calibrations were made by injecting a known volume of standard gas (Messer®, 1000, 500 and 100 ppm ± 1 ppm CH_4 in helium and 500 ppb H_2 in helium). All analyses were made in duplicate, and results are given as averaged values. The detection threshold of this method is 0.2 nmol for dissolved CH_4 , and variation between duplicates was 5 %.

2.2 Inorganic nutrient analysis

Samples for inorganic nutrient analysis (NO_2^- , NO_3^- , NH_4^+ , PO_4^{3-}) were collected unfiltered into sterile 15 ml centrifuge tubes (rinsed three times with water from the same Niskin). Samples were analyzed directly from the collection tubes within 1–8 h and measured from the lowest to the highest concentration (surface to deep) to reduce any carryover effects. Nutrients were analyzed on board using a four-channel Seal Analytical AA3 segmented flow autoanalyzer following GO-SHIP protocols (Becker et al., 2020). In order to test the accuracy and precision of the analyses, certified reference materials (CRMs) from KANSO Technos Co. (lots lot CD, CJ, CI and lot BW) were measured in triplicate in every run.

2.3 Cyanobacteria sampling and analysis

38 of the stations occupied during JC191 were sampled at six depths in the upper water column (max. sampling depth 375 m) following the live fluorescence profiles to determine the prevailing community of primary producers (unpublished data, Marx 2020). Bulk water samples (5 L) were collected from Niskin bottles, from which subsamples for the flow cytometric determination of *Prochlorococcus* and *Synechococcus* abundances were collected. 4 mL of sample water was immediately fixed with 40 mL glutaraldehyde solution (50 %) and stored at 4°C until transferred to a low-temperature freezer (-80°C) after 12 h.

Samples were analyzed at the University of Portsmouth on a CyFlow Cube 8 (Sysmex) flow cytometer immediately after defrosting and at a flow rate of $1\ \mu\text{L s}^{-1}$. The distinction between *Prochlorococcus* and *Synechococcus* was achieved by gating each group according to its fluorescence signals (red and orange fluorescence) against the size fractionation (forward and side scatter). For each of the stations sampled, the mixed layer depth (MLD, defined as the depth at which temperature decreased by 1°C from the surface) was determined, and the integrated average in abundance above said MLD was calculated.

2.4 Statistical analyses

In order to determine the biological and physicochemical parameters that influence the distribution of dissolved methane

in surface waters, a principal component analysis (PCA) was applied. This statistical tool simplifies the underlying structure of the multivariate dataset by converting a large number of variables into a smaller number of variables, i.e., components (PCs), with a minimum loss of information. Each PC is associated with an eigenvalue that indicates the variation in the data. Here, the PCA computes a singular value decomposition of the data matrix and does not incorporate the eigenvalue of the covariance matrix to maintain numerical accuracy. Then, factor loadings were calculated, with a high factor loading indicating a significant correlation between variables. Additionally, Kendall rank correlations were computed to evaluate associations between two variables (e.g., methane concentrations and fluorescence or phytoplankton abundances). All statistics were performed with RStudio and the R “stats package” (R Core Team, 2013).

2.5 Flux calculation

The flux of air–sea CH_4 was calculated following established methods (Kelley and Jeffrey, 2002; Wanninkhof, 2014):

$$F = k(C_w - C_a), \quad (1)$$

where F is the CH_4 flux ($\text{mol m}^{-2} \text{d}^{-1}$) from seawater to air, with k as a gas transfer coefficient under consideration of wind speed, C_w as the average CH_4 concentration measured in surface water and C_a as the equilibrated seawater–air CH_4 concentration. Wind speed data were obtained from on-ship measurements throughout the cruise transect, and the concentration of CH_4 in the air was assumed to be 1.9 ppm (based on NOAA, Global Monitoring Laboratory https://gml.noaa.gov/ccgg/trends_ch4/, last access: 28 August 2022). The gas transfer coefficient k was calculated following

$$k = 0.251 \langle U^2 \rangle (Sc/660)^{-0.05}, \quad (2)$$

where $\langle U^2 \rangle$ represents the average neutral stability of winds at 10 m height squared, and Sc indicates the Schmidt number, taking into account the kinematic viscosity of water and the molecular diffusion coefficient of the gas. From inverse modeling using CCMP winds and the Modular Ocean Model–General Circulation Model (MOM3 GCM), a value of 0.251 was obtained. Sc was calculated as

$$Sc = A + Bt + Ct^2 + dt^3 + Et^4, \quad (3)$$

with coefficients A , B , C , D and E obtained from a least square fourth-order polynomial fit (Wanninkhof, 2014).

3 Results

3.1 Spatial distribution of physicochemical parameters

CH_4 concentrations in surface waters of the subtropical North Atlantic along the 24°N transect were distributed

nonuniformly between 3 and 10 nmol L^{-1} , i.e., systematically above saturation of $\sim 2.7 \text{ nmol L}^{-1}$. The lowest concentrations of CH_4 of 3–4 nmol L^{-1} were found in the central gyre system above the mid-Atlantic ridge ($\sim 45^\circ \text{W}$), with increasing values of 8–10 nmol L^{-1} in both western and eastern boundaries closer to the continental shelf (Fig. 2a).

Chlorophyll a (Chl a , Fig. 2b) from real-time fluorescence profiles exhibits the highest concentrations near both the western and eastern shore systems and the lowest concentrations in surface waters throughout the central gyre system ($< 0.1 \mu\text{g L}^{-1}$, Fig. 2). A deep chlorophyll maximum (DCM) was observed between 100 and 130 m water depth and was consistently above the mixed layer depth (MLD). Concentrations of Chl a increase towards the Mauritanian upwelling off the North African coast to above $0.4 \mu\text{g L}^{-1}$, indicating higher primary production due to enhanced nutrient supply. Accordingly, light transmission is decreased due to higher content of suspended particles in the water column (Fig. 2h). Furthermore, fluorescence aligns with patterns of dissolved oxygen in the water column of this transect. The surface waters in the subtropical North Atlantic are well oxygenated with concentrations above $200 \mu\text{mol kg}^{-1}$ in the top 100 m of the water column, subsequently decreasing with depth. Increased concentrations of dissolved oxygen are observed towards the eastern boundary following enhanced lateral transport from the coastal upwelling (Fig. 2e).

Nitrate and phosphate concentrations were low (< 0.1 , $< 0.01 \mu\text{mol kg}^{-1}$, respectively) in surface waters throughout the transect with nutriclines $> 200 \text{ m}$ depth, shallowing ($< 150 \text{ m}$ in eastern basin and $< 100 \text{ m}$ in eastern boundary) towards the eastern boundary due to coastal upwelling off the eastern North African coast and associated enhanced mixing of deep, nutrient-enriched waters into surface waters (Fig. 2c and d).

3.2 Sea-to-air methane flux

At most stations, air–sea CH_4 flux was $1\text{--}2 \mu\text{mol m}^{-2} \text{d}^{-1}$, with maximum values of $5 \mu\text{mol m}^{-2} \text{d}^{-1}$ in the area of 24°N 25°W (Fig. 3). Only one station ($\sim 40^\circ \text{W}$) had a negative CH_4 flux, which indicates a sink of CH_4 . The overall average flux across the subtropical North Atlantic was $1.9 \mu\text{mol m}^{-2} \text{d}^{-1}$. However, it should be noted that the air–sea flux of CH_4 is dependent on the wind speed and can be locally different, varying depending on specific weather conditions.

Although the sea-to-air flux of CH_4 presented here broadly agrees with similar values of $1.6 \mu\text{mol m}^{-2} \text{d}^{-1}$ measured in the oligotrophic North Pacific, it can differ substantially both temporarily and spatially. For instance, the sea-to-air flux of CH_4 in the Sargasso Sea, also partly encompassed in the transect of JC191, fluctuates between 1.6 and $4.4 \mu\text{mol m}^{-2} \text{d}^{-1}$ depending on the season and weather conditions (Holmes et al., 2014). Furthermore, the base ecosystem can impact the sea-to-air flux of CH_4 if compared to

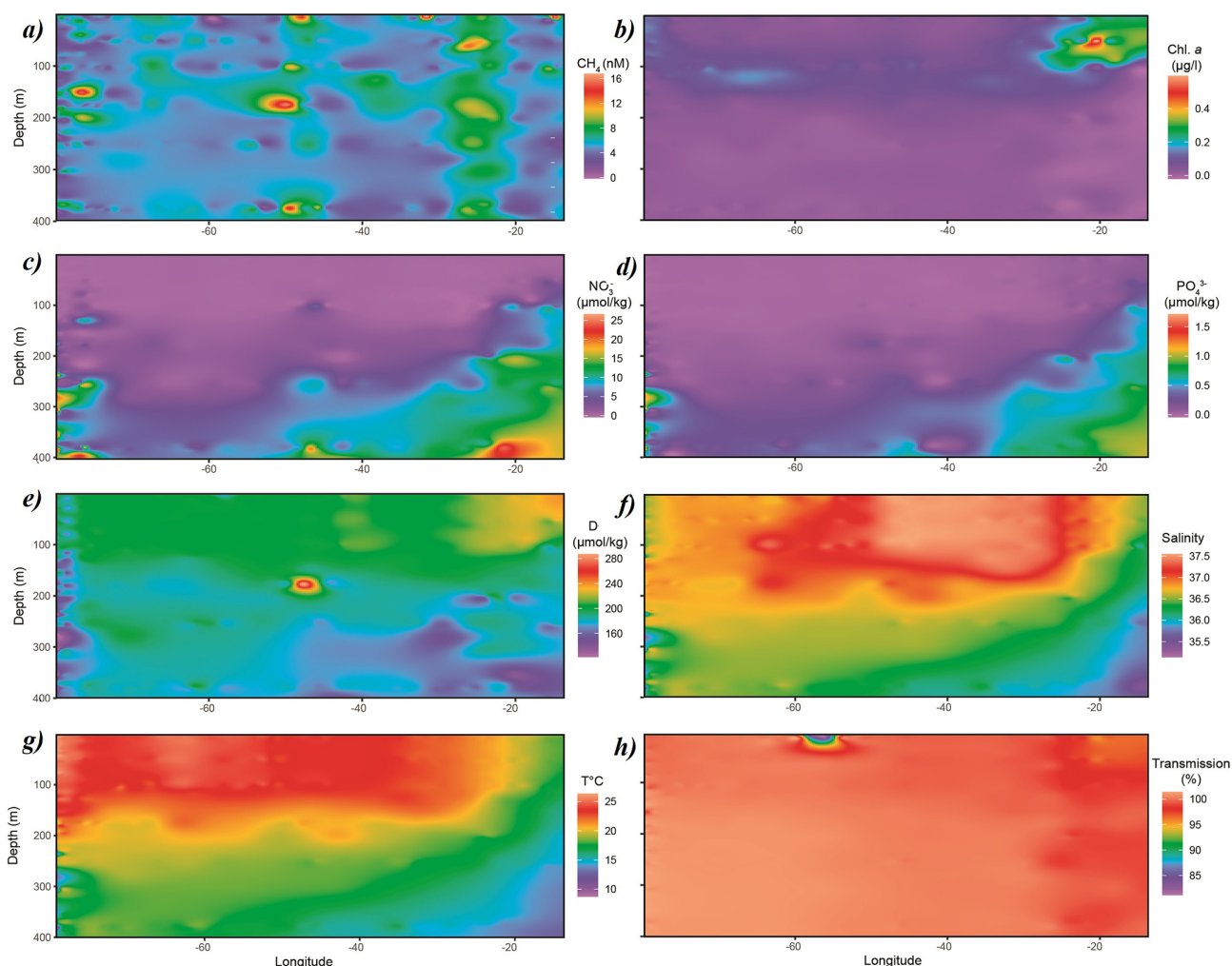


Figure 2. Distribution of (a) CH_4 , (b) Chl *a*, (c) NO_3^- , (d) PO_4^{3-} , (e) dissolved oxygen, (f) salinity, (g) temperature and (h) transmission along the cruise track. Chl *a*, dissolved oxygen, salinity, temperature and transmission data originate from sensors attached to the CTD. CH_4 , NO_3^- and PO_4^{3-} were measured analytically. Some of the data were taken from CCHDO (Sanchez-Franks, 2020).

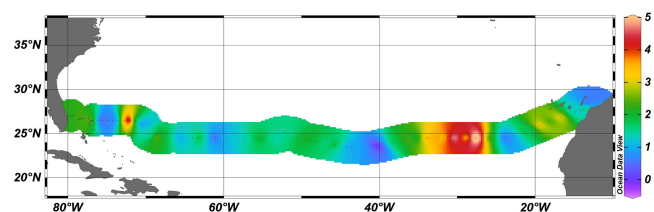


Figure 3. Sea-to-air CH_4 flux ($\mu\text{mol m}^{-2} \text{d}^{-1}$) along the cruise track of JC191. Dots indicate stations occupied by CTD casts.

other oceanic regions such as Belgian coastal zones with $1\text{--}160 \mu\text{mol CH}_4 \text{m}^{-2} \text{d}^{-1}$ (Borges et al., 2016), Red Sea mangroves with $13.3 \mu\text{mol CH}_4 \text{m}^{-2} \text{d}^{-1}$ (Sea et al., 2018), or surveys in the East China Sea ($6.5\text{--}7.4 \mu\text{mol m}^{-2} \text{d}^{-1}$, Ye et al., 2015) and the Gulf of Mexico ($0.38 \mu\text{mol m}^{-2} \text{d}^{-1}$, Kelley and Jeffrey, 2002).

4 Discussion

4.1 Methane distribution in surface waters of the subtropical North Atlantic Ocean

Our data across the subtropical North Atlantic Ocean unambiguously indicate an oversaturation in CH_4 of the surface layer (400 m) of 110 % to 370 %, which is in agreement with previous observations describing concentrations varying between 2 and 5 nmol L^{-1} and maximum values of 10 nmol L^{-1} (Scranton and Brewer, 1977; Conrad and Seiler, 1988; Forster et al., 2009; de la Paz et al., 2015; Leonte et al., 2020). This is also in line with previous observations describing the upper layer in subtropical areas of the global ocean as a source of CH_4 to the atmosphere (Reeburgh, 2007; Dang and Li, 2018) and also in some regions of the Arctic Ocean (Kitidis et al., 2010; Kudo et al., 2018). To date, only the Southern Ocean is undersaturated in CH_4 , although this

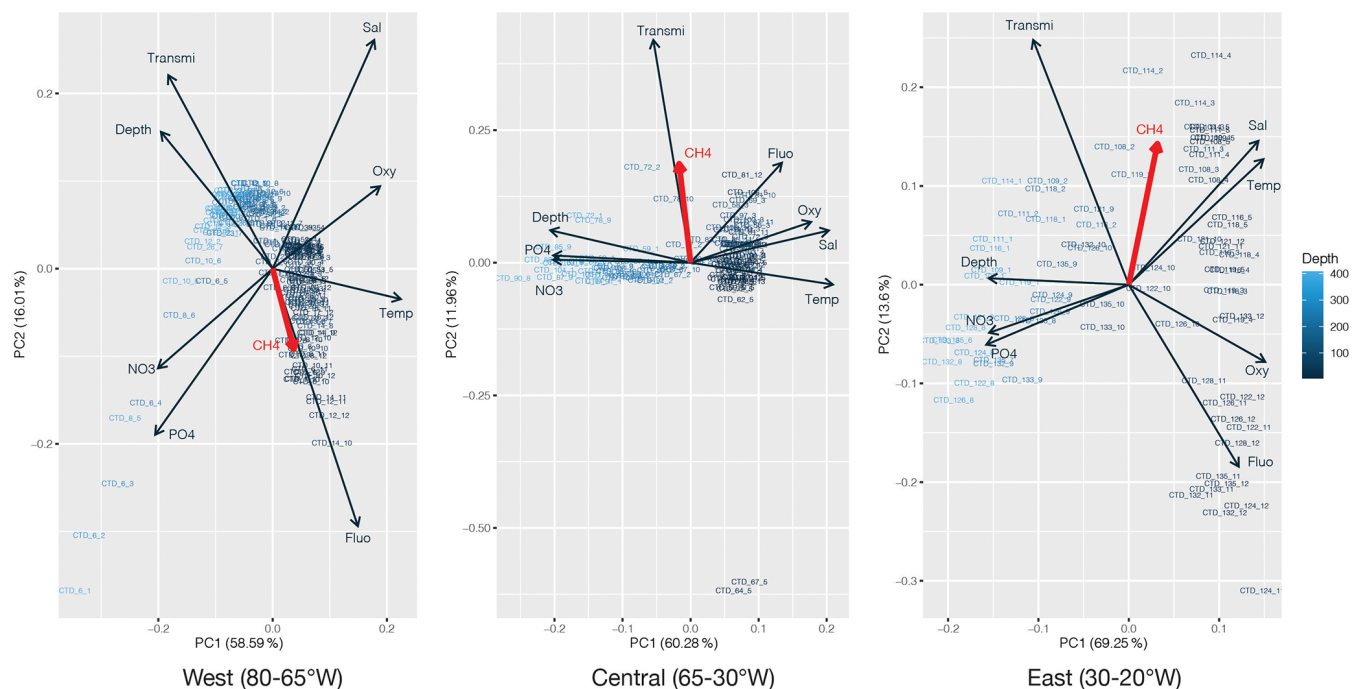


Figure 5. Principal component analysis of the regionalized dataset: west (80–65° W), central (65–30° W) and east (30–20° W). Numbers on the x and y axes indicate the factor loadings of each variable of each principal component (PC). The percentages show the explained variability in the dataset by each PC. CH₄ is highlighted in red.

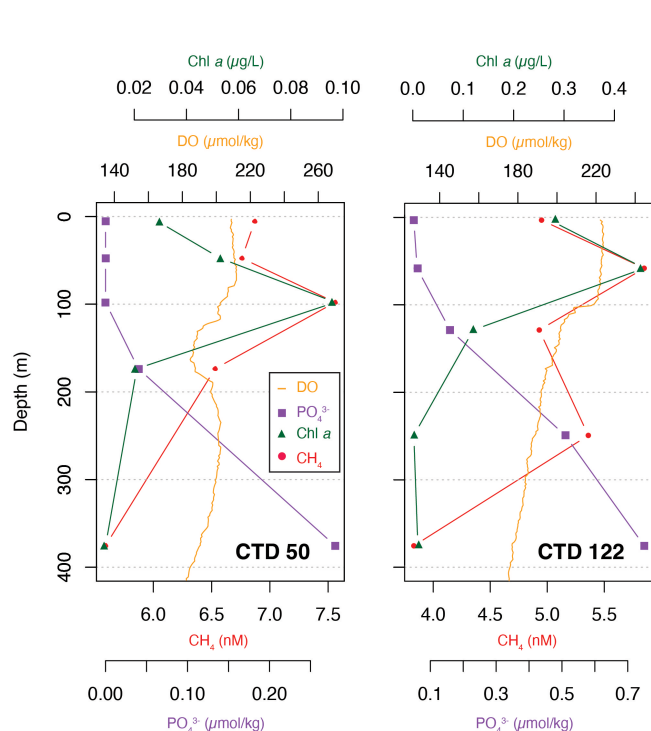


Figure 6. Selected data from CTD 50 (69.5° W, 24.9° N) and CTD 122 (20.8° W, 25.45° N) for CH₄ (red circles), Chl *a* (green triangles), phosphate (purple squares) and dissolved oxygen (orange).

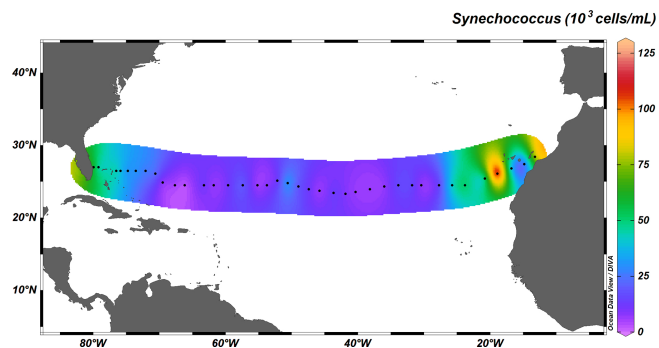


Figure 7. Depth-integrated abundance of *Synechococcus* (in 10³ cells mL⁻¹) above MLD. Black dots represent the 38 out of the total 135 stations sampled for cyanobacterial abundance.

coccus dominates the community throughout the transect (Fig. 8).

CH₄ distribution across the subtropical North Atlantic suggests that the influence of the cyanobacterial community is present; the highest CH₄ concentrations were found (1) at DCM in the central gyre system, where *Prochlorococcus* is the predominant genus, and (2) at the gyre boundaries, where higher abundances of both *Prochlorococcus* and *Synechococcus* were found. A PCA applied to the dataset including abundances of both cyanobacteria strengthens the conclusion that

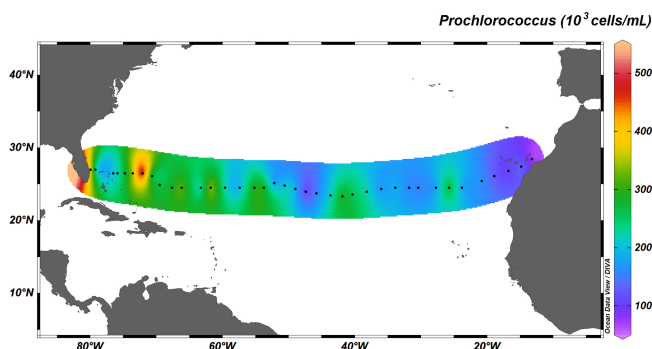


Figure 8. Depth-integrated abundance of *Prochlorococcus* (in 10^3 cells mL^{-1}) above MLD. Black dots represent the 38 out of the total 135 stations sampled for cyanobacterial abundance.

CH_4 production appears to be associated with primary production (Fig. 9).

The bordering Gulf Stream to the west and the Canary Stream to the east provide a resupply of nutrients (Williams et al., 2011) and therefore support a greater abundance of cyanobacteria; the Canary Stream also presents a pathway for the horizontal transfer of organic matter from the North African coast to the open ocean. Previous studies confirm high OM production in the eastern subtropical North Atlantic as a result of the high rates of primary production fueled by the Mauritanian upwelling (Reynolds et al., 2014). The transmission data (Fig. 2h) suggest the export of organic matter throughout the eastern basin, which could contain MPn and thus a potential source of CH_4 . Elevated CH_4 concentrations are mainly limited to the highly productive boundary systems, whilst the central gyre, which lacks a sufficient resupply of bioavailable nutrients due to its downwelling nature, results in a decreased abundance of both *Prochlorococcus* and *Synechococcus* and hence decreased CH_4 concentrations (Fig. 2a). In this especially P-limited region, alternative nutrient sources such as the degradation of DOM to access organic phosphorus compounds become increasingly important in order for the cyanobacteria to meet their nutrient needs. The degradation of dissolved organic phosphorus via alkaline phosphatases (AP) such as *phoA* or *phoX* and the overexpression of the phosphate binding protein (pstS) have been believed to be the main adaptations to P stress (Luo et al., 2009; Cox and Saito, 2013; Sebastian and Ammerman, 2009). Recently, however, evidence was brought forth that some strains of *Prochlorococcus* can also oxidize MPn and other higher phosphonate compounds while releasing formate and potentially CH_4 as a by-product (Sosa et al., 2019a). Phosphonates are notably abundant and enriched in DOM (Sosa et al., 2019b), and their degradation releases considerable amounts of CH_4 . *Prochlorococcus* and *Synechococcus* are the most abundant primary producers in the oligotrophic ocean and as such produce considerable amounts of semi-labile DOM; both can synthe-

size phosphoenolpyruvate mutases (*pepM*), and therefore the DOM produced carries an enriched pool of MPn (Repeta et al., 2016; Sosa et al., 2019a). However, the metabolism of MPn is heavily regulated by bioavailable phosphate, and thus the metabolic pathway of *pepM* might be heavily down-regulated in the subtropical North Atlantic, whereas under replete conditions, i.e., in the North Pacific, *Prochlorococcus* can allocate up to 40 % of its internal P quota to phosphonate synthesis (Acker et al., 2022). Yet, the trait to produce phosphonates is located on genomic islands, is subject to horizontal gene transfer and can be frequently exchanged among marine microbial communities. Hence, proteobacteria such *Pelagibacter* spp. in the SAR11 clade also obtain *pepM* and are able to produce phosphonates (Acker et al., 2022). Similarly, the trait of phosphonate consumption is subject to horizontal gene transfer, and high-light strains of *Prochlorococcus* carry both production and consumption traits. They can therefore also utilize MPn as an alternative P source (Acker et al., 2022). The demethylation of MPn is mostly attributed as occurring under P-limiting conditions, such as in oligotrophic oceanic regions. Further, most MPn oxidation and subsequent release of CH_4 are due to bacterial degradation of DOM and the breakdown of highly energetic carbon–phosphorus bonds via C–P lyases, which are encoded by the *phn* operon, with transport systems including *phnC*, *phnD*, *phnE* and *phnJ* responsible for the cleavage of the C–P bond (Sosa et al., 2020). C–P lyases are abundant among *Pelagibacter* spp. and other alpha and gamma proteobacteria and can be found in ~ 50 % of organisms in the North Atlantic, where DOP concentrations are 4-fold lower with respect to the North Pacific (Sosa et al., 2020). Carini et al. (2014) estimated a potential production of $0.01\text{--}0.05$ nM CH_4 d^{-1} by *Pelagibacteriales* via C–P cleavage within the MPn pathway (Carini et al., 2014). However, due to complex biogeochemical and physical processes involved, in situ production of CH_4 is difficult to assess, and quantifying the individual contribution of primary producers to CH_4 production was not the primary scope of the work presented. Here, the focus was to link the distribution of CH_4 in the subtropical North Atlantic to two vastly abundant cyanobacteria as exemplary primary producers, *Prochlorococcus* and *Synechococcus*, in order to draw conclusions which might explain the ocean methane paradox. Many mechanisms and precursors of oceanic in situ production of CH_4 remain unknown, and it is unclear whether cyanobacteria such as *Prochlorococcus* or *Synechococcus* are responsible for the release of CH_4 , mainly via the degradation of MPn or indirectly by producing semi-labile DOM containing MPn cleaved by the bacterial community. Furthermore, recent laboratory-based studies also confirmed that planktonic organisms produce CH_4 per se under high light intensities; *Prochlorococcus* showed higher potential in CH_4 production than *Synechococcus*, with $0.8\text{--}110$ and $0.01\text{--}0.6$ pmol CH_4 per 10^6 cells per hour, respectively. Larger cyanobacteria such as *Microcystis* spp. ex-

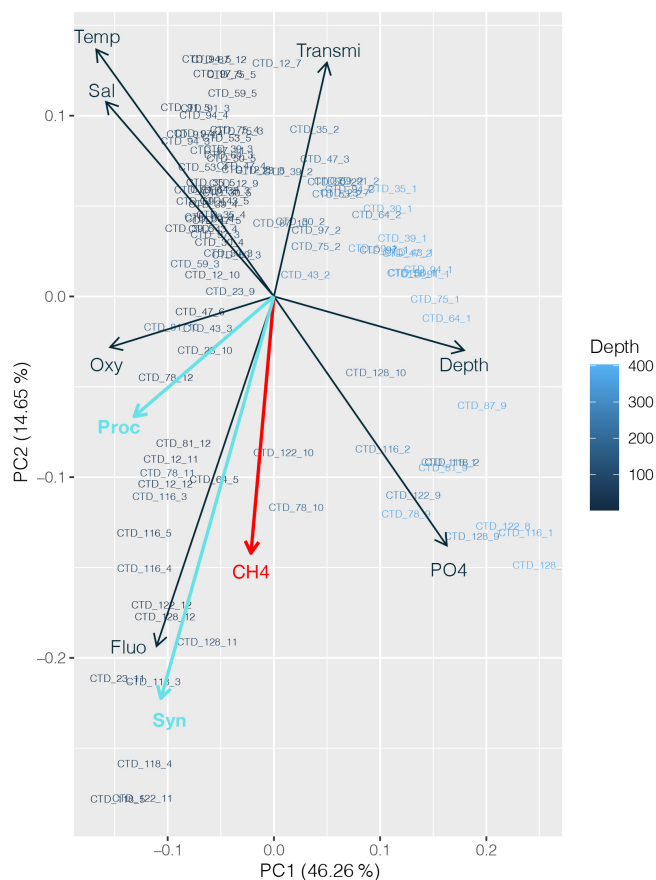


Figure 9. Principal component analysis (PCA) between dissolved CH_4 concentrations (red), *Prochlorococcus* and *Synechococcus* abundances (light blue), and other physicochemical parameters (nutrients, depth, fluorescence, temperature, salinity and turbidity). Numbers on the x and y axes indicate the factor loadings of each variable of each principal component (PC). The percentages show the explained variability in the dataset by each PC.

ceeded this by up to $0.51 \pm 0.26 \mu\text{mol g}^{-1} \text{h}^{-1}$ (Bižić et al., 2020).

Nonetheless, *Prochlorococcus* alone produces $\sim 22 \text{ nM CH}_4 \text{ d}^{-1}$ and therefore substantially contributes to the supersaturation of oligotrophic oceanic regions (typically between 7 and 25 nM) as well as to the oversaturated state measured here (up to 10 nM) in the subtropical North Atlantic (Bižić et al., 2020). Cyanobacteria can greatly influence and potentially control CH_4 production in the marine environment, although other marine organisms such as different bacterial clades or coccolithophores are believed to also contribute to overall CH_4 production. Nitrogen-fixing diazotrophs such as *Trichodesmium* spp. or *Rhodopseudomonas palustris* may contribute to an even higher degree (Bižić-Ionescu et al., 2018; Zheng et al., 2018), and, although spatially limited with *Trichodesmium* spp. dominating the western basin and *Hemilaulus* associated *Richelia* symbionts more so in the eastern basin (Luo et al., 2009), they are

also abundant throughout the subtropical North Atlantic. *Trichodesmium* has higher nutrient requirements and can therefore outcompete cyanobacteria in uptake of inorganic nutrients and degradation of alternative sources; yet, energy-intensive diazotrophic nitrogen fixation is controlled by micronutrients such as iron (Macovei et al., 2019) and primarily occurs in the western basin, an area of high iron input by aeolian plumes originating from the Saharan desert (Ratten et al., 2015; Reynolds et al., 2014). Lastly, coccolithophores such as the ubiquitously abundant *Emiliania huxleyi* can also produce CH_4 (between 0.7 and $3.1 \pm 0.4 \text{ CH}_4$ per gram POC per day; Lenhart et al., 2016b) from OM degradation and are believed to increase in abundance in the subtropical North Atlantic with increasing CO_2 concentrations at the air–sea interface due to further anthropogenic perturbation of the atmospheric CO_2 budget (Krumhardt et al., 2015).

Nonetheless, the data presented here suggest that the cyanobacterial community most likely plays a significant key role in the CH_4 flux in surface waters and the degradation of MPn from semi-labile DOM, helping to explain the methane paradox and hence the sea–air flux of CH_4 (Sosa et al., 2019a). Further investigation needs to focus on gathering in situ data and should also include future scenarios, considering future climate and whole-ecosystem community responses to consequences of altered climate conditions. The base ecosystem, specifically in P-limited regions, is ever changing, and highly adapted and horizontal transfer of genomic traits of P acquisition via various genomic pathways might become inherently important. Further work should also be considered in regards to mesocosm incubation studies with in situ communities of natural composition to improve our understanding of the various processes and interactions involved in oceanic CH_4 production linked to primary producers to explain the ocean methane paradox.

5 Conclusions

Our study demonstrates that the subtropical North Atlantic Ocean does indeed act as a source of CH_4 to the atmosphere, most likely controlled by cyanobacteria, which are the dominant primary producers in the surface waters. Yet, anomalies found at depths below 200 m could also be attributed to the degradation of sinking organic material. The concentrations of dissolved CH_4 in this study were considerably higher near shelf regions and in the eastern boundary under the influence of the Mauritanian upwelling. The accumulation of organic matter and nutrients in these areas provides favorable conditions for both aerobic and anaerobic CH_4 production. It is expected that with increasing stratification and subsequent reduction in nutrient supply to the surface oligotrophic North Atlantic Ocean, the prevailing P limitation will be further exacerbated, whereas coastal and shelf regions with increasing anthropogenic inputs of nutrients could experience more

frequent cyanobacterial blooms, which will in turn enhance CH₄ production (Dang and Li, 2018).

Data availability. All data are available from the authors upon request.

Author contributions. AK carried out dissolved gas sample collection, measurements and data analysis and drafted the final paper. LM carried out phytoplanktonic sample collection, measurements and data analysis and drafted the final paper. SR helped in preparation of the cruise, measurements and data analysis and supervised biological part of the survey. TC carried out nutrient sample collection, measurements and data analysis. EM carried out nutrient sample collection, measurements and data analysis. CB helped to plan and prepare for the cruise as well as result evaluation and provided overall project supervision. All authors contributed equally to the paper writing and made substantial input to the final version.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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