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Author(s)	Bui, Thi Ngoc Oanh
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Dissolved methane distribution in seawater and its controlling factors in the polar regions

(極域における海水中メタンの分布とその制限因子に関する研究)

Bui Thi Ngoc Oanh

Division of Earth System Science

Graduate school of Environmental Science

Hokkaido University

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Bùi Thị Ngọc Oanh

ABSTRACT

Methane (CH_4) is a potent greenhouse gas and plays significant roles in both tropospheric and stratospheric chemistry. During the last two centuries, mixing ratio of atmospheric CH_4 has increased by 1000 ppb and reached to 1834 ppb in 2015 as a result of anthropogenic activities. It is known that ocean acts as one of the sources of atmospheric methane and the methane flux depend on the oceanic region owing to the difference in biological activities and physical condition. For example, the region characterized by high productivity releases more CH_4 to the atmosphere than the other regions because CH_4 is produced via microbial methanogenesis in anaerobic environments from bacteria in organic particles or the gut of zooplankton. Until now, some works have implemented to measure CH_4 concentration in the polar regions where the impact of climate change is believed to be strongest in the world ocean. However, the data scarcity causes the uncertainty of the CH_4 distribution in the polar oceans, its controlling factors and the contribution of the polar oceans to global CH_4 sea-air fluxes. I, therefore, used the high-resolution dataset to detect oceanic CH_4 in seawater to reduce the uncertainty from the previous estimations.

The Southern Ocean is one of the most biologically productive regions among the world's oceans, where is characterized by high biomass of zooplankton, Antarctic krill, and salps, from which large amount of CH_4 production could be expected. A cavity ring-down spectroscopy (CRDS) system coupled with a shower-head type equilibrator was used to obtain quasi-continuous underway measurements of the mixing ratio of oceanic CH_4 in seawater during legs 2 and 3 of cruise MR12-05 of R/V *Mirai* in the Southern Ocean from late November 2012 to mid-February 2013. High CH_4 saturation ratio (SR , %) occurred in the lower latitudes including coastal area north of sub-antarctic front (SAF). SR decreased gradually and monotonously between SAF and southern limit of upper circumpolar deep water (SBDY), and occasionally decreased lower than 100% due to vertical mixing/upwelling in high latitudes south of polar front (PF). A good correlation between SR and apparent oxygen utilization (AOU) was found south of PF, $SR = -0.294 \times \text{AOU} + 100.093$ ($n = 1549$, $r^2 = 0.89$). On the basis of SR -AOU relationship, I have calculated the sea-air CH_4 flux during December to February in the entire of the Southern Ocean (south of 50°S). The sea-air CH_4 emission was estimated to be $0.027 \text{ Tg-CH}_4 \text{ yr}^{-1}$ in December, $0.04 \text{ Tg-CH}_4 \text{ yr}^{-1}$ in January, and $0.019 \text{ Tg-CH}_4 \text{ yr}^{-1}$ in February. From my estimation, south of 50°S , the average flux of CH_4 in austral summer is $0.024 \text{ Tg-CH}_4 \text{ yr}^{-1}$, which is account for a minor contribution (less than 1% of the global oceanic release of $4\text{-}15 \text{ Tg yr}^{-1}$) of CH_4 to the atmosphere.

The Arctic Ocean is particularly the most sensitive to global warming, for example, Arctic sea ice thickness and extent are declining, due to increasing sea-surface temperature. In the Arctic Ocean, a massive CH₄ hydrates from marine sediments release and/or methanogenic CH₄ producing in anaerobic environment with the presence of high organic matter on the seafloor and then diffusion into water column was examined. However, the vertical profile of CH₄ distribution is still scarce, so it is needed to be examined for comprehension of CH₄ dynamics in the water column. I examined vertical profiles of CH₄ distribution inside and outside of a warm-core eddy during Leg. 1 of the cruise MR15-03 of R/V *Mirai* in the Arctic Ocean in late summer/early fall 2015. Because a coherent mesoscale eddy is one of the possible mechanisms for transporting the shelf water from Chukchi Sea into the Canada Basin interior, I hypothesized that the warm-core eddy transported nutrients as well as dissolved CH₄ and impacted the distribution of CH₄. In this study, the existence of the warm-core eddy that it would carry nutrients as well as dissolved CH₄ and impact on CH₄ distribution. Thus, seawater samples were collected to obtain detailed vertical profiles of CH₄ distribution in the southwestern Canada Basin. High CH₄ concentrations were found at the bottom on the shelf, where located outside the warm-core eddy. Saturation of CH₄ in that water was high 947% respecting to atmospheric CH₄. I also found that sub-surface CH₄ peaks within the radius of maximum velocity of the eddy were broader spread in-depth than those outside of the eddy. The presence of the eddy was likely the controlling factor responsible for subsurface CH₄ peaks inside the warm-core eddy. This study indicates that eddies play a crucial role in controlling CH₄ distributions and transport it from shelf water to the southwestern Canada basin.

In addition, extent of sea ice in the Arctic has been decreasing while sea ice acts as physical barrier. Removal of this barrier directly affects sea-air exchange of CH₄. I investigated CH₄ distribution in the Arctic Ocean under two conditions: in the shelf water and sea ice cover region in order to evaluate the influence of sea ice coverage on CH₄ emission. During the cruise of IBRV *ARAON* (ARAON-07B) in August 2016, seawater samples were collected from the surface to the bottom to detect the vertical profiles of CH₄ distribution in the Bering, Chukchi Seas and the Arctic Ocean. Over the whole region, dissolved CH₄ concentration ranged from 3.5-23.1 nmol kg⁻¹ in the surface water, which was supersaturated (range: 124.7-604.6%) to the atmospheric CH₄. Observed CH₄ concentration increased with depth at the shallow water. High dissolved CH₄ concentration was found at the Chukchi shelf slope resulted from high *in situ* methanogenesis of organic matter. In the sea ice cover region, CH₄ concentration increased at the subsurface water (50-100 m) and decreased with depth to

the bottom. Low CH₄ was observed at the deeper water (> 200 m) in sea ice cover region. Dissolved CH₄ varied in the distinct water masses (PSW, PWW, ACW, MW and AW). Relations between CH₄ and nutrients as well as its precursors are compared under the two conditions. High CH₄ concentrations observed at surface seawater were consistent with high sea ice concentrations. This investigation allows us to gain some insights into the role of the sea ice on CH₄ concentration and the flux.

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LIST OF ABBREVIATIONS

CH ₄	Methane
N ₂ O	Nitrous Oxide
DMS	Dimethylsulphide
DMSP	Dimethylsulfoniopropionate
Pdw	Pacific-derived water
Adw	Atlantic-derived water
ACW	Alaskan Coastal Water
PSW	Pacific Summer Water
PWW	Pacific Winter Water
AW	Atlantic Water
MW	Meltwater and River water
CRDS	Cavity Ring-Down Spectroscopy
GC	Gas Chromatography
SR	Saturation

CHAPTER 1:

GENERAL INTRODUCTION

1.1 Generally introduction CH₄ in the ocean

1.1.1 Introduction to CH₄

After carbon dioxide (CO₂), methane (CH₄) is an effective greenhouse gas and plays an important role in the Earth radiative and chemical balance, contributing 15-20% of atmospheric radiative forcing (IPCC, 2013). CH₄ is the most abundant reactive trace gas in the troposphere, and its reactivity is important to both of tropospheric and stratospheric chemistry (Cicerone and Oremland, 1988). Thus, CH₄ is one of the factors in consideration of global climate change (IPCC, 2001). Average concentration of atmospheric CH₄ is determined mainly by the balance between emission from the surface and destruction by hydroxyl free radicals (OH) in the troposphere (Wuebbles and Hayhoe, 2002). There is a recent evidence that there has been a dramatical increase in the tropospheric CH₄ mixing ratio over 100-year timescale. During the last two centuries, this ratio has increased by 1000 ppb, reached up to approximately 1850 ppb as a result of anthropogenic activities (https://www.esrl.noaa.gov/gmd/ccgg/trends_ch4/; Dlugokencky et al., 1994). CH₄ is increasing at a variable rate that is not clearly understood which is reflected on changing of sources and sinks strength (Dlugokencky et al., 1994). Although the contribution of greenhouse gases to radiative forcing is clear, the cycle of this gas in various reservoirs is still poorly understood. Hence, understanding of the global CH₄ budget and CH₄ cycling is a primary research objective in the Earth Science.

1.1.2 CH₄ sources and sinks

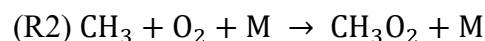
The sources can be organized into two main categories such as natural and anthropogenic sources. Natural sources consist of oceans, fresh water, gas hydrates, termites, natural forest fires, boreal and tropical wetlands including rice paddies. Another one is from anthropogenic activities including from fossil fuels, energy use, landfill, waste treatment, and ruminant animals, and biomass burning (Wuebbles and Hayhoe, 2002).

The ocean has only a minor role in the global CH₄ budget; oceanic emissions account for a few percentages of all natural and anthropogenic sources (Cicerone and Oremland, 1988; Conrad and Seiler, 1988; Crutzen and Zimmermann, 1991; Bange et al., 1994). On the other hand, a small region of the ocean is a sink of CH₄. Although major sources and sinks of

atmospheric CH₄ have been identified, significant uncertainties remain in estimations of the total earth and atmospheric budgets of CH₄ (Saunois et al., 2016).

CH₄ is produced through anaerobic methanogenesis of organic matters in sedimentary layer. Released CH₄ from sediments diffuses into seawater by bubbling and then oxidizes in the water column. Within the aerobic water column, CH₄ is produced by microbial methanogenesis in anaerobic environments and by bacteria in organic particles (Karl and Tilbrook, 1994), or within digestive tracts of organisms (de Angelis and Lee, 1994). CH₄ distribution is often controlled by biological production and oxidation. Oxidation processes and sea-air exchanges act to maintain CH₄ concentration near atmospheric equilibrium (Hanson and Hanson, 1996; Reeburgh, 2007; Grundward et al., 2009). Oxidation process in aerobic environments converts CH₄ into CO₂ and water (see **Fig. 1.1**). **Fig. 1.1** represents microbial processes of organic matters (OM) transformation in marine anoxic sediments and emission to an anaerobic/aerobic CH₄ oxidation in the water column. Other sources of CH₄ may come from freshwater runoff from regions with anaerobic sediments containing hydrocarbon seeps or destabilized gas hydrates (Naqvi et al., 2010). Recently, CH₄ is produced by the cleavage of methylated substrates by bacteria or phytoplankton (Karl et al., 2008; Damm et al., 2010). Dissolved CH₄ concentration in the ocean depends on the magnitude of aerobic/anaerobic microbial production and oxidation so that the ocean can be a net source or sink. Estimation of strength of both production and consumption of CH₄ are difficult to make precisely quantification because of predictions of changing the environmental conditions such as sea ice cover, ocean biogeochemistry and so on, so it is essential for evaluating of contribution of oceanic CH₄ fluctuation to the atmosphere.

On the contrary, there are one major and two minor sinks for tropospheric CH₄, which CH₄ is removed from the atmosphere through chemical reaction. Reaction with the hydroxyl radical (OH) is a main process of removal of almost 90% of the total sink (~90%) (R1) (IPCC, 1996; Wuebbles and Hayhoe, 2002). A lifetime of CH₄ in the atmosphere is determined by the rate of this loss process, which is about 12 years. It is much shorter than a lifetime of carbon dioxide (5-200 years) (IPCC, 2013; <https://www.ipcc.ch/ipccreports/tar/wg1/016.htm>). Oxidation of CH₄ in the atmosphere (in the troposphere and the lower stratosphere) through reactions with NO_x compounds lead to enhance ozone (O₃) formation as well as greenhouse gases from CH₄ emission (Isaksen et al., 2011) with the existence of N₂ molecule (M) and solar photon flux (hν) as follow.



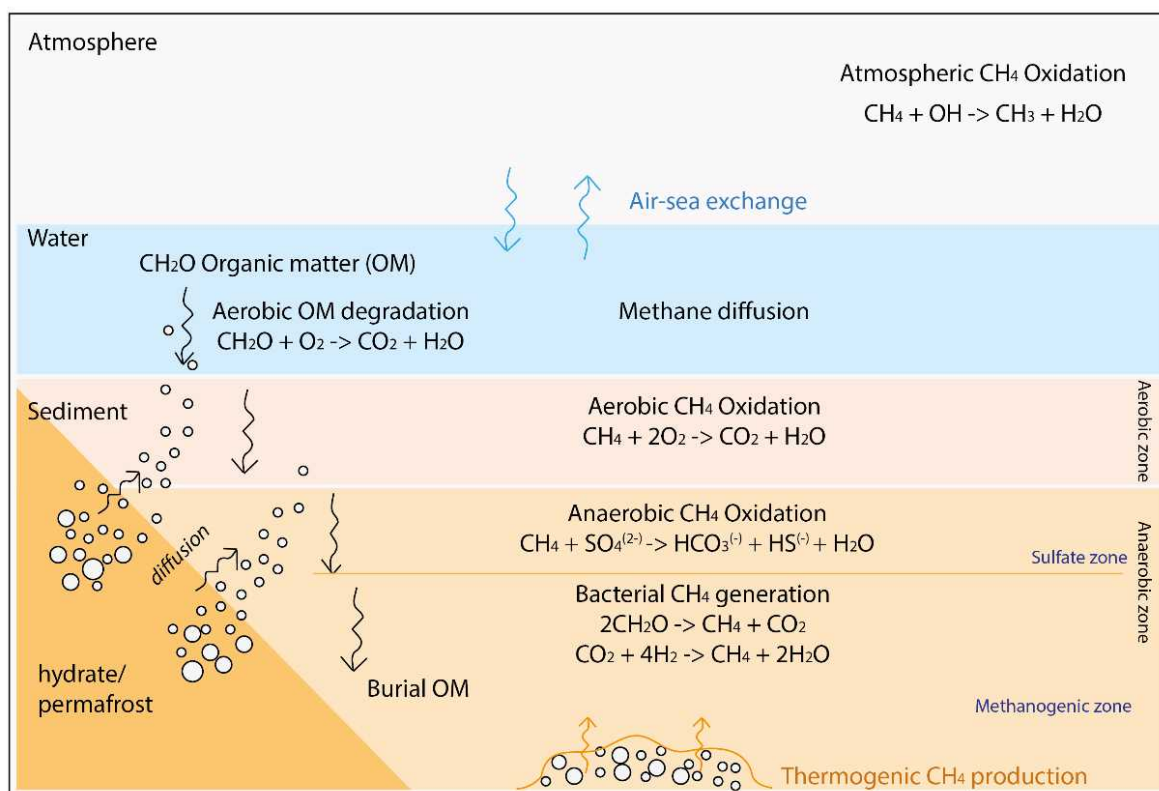
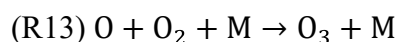
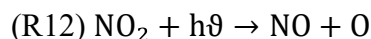
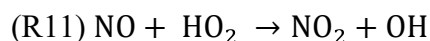
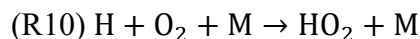
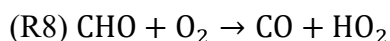
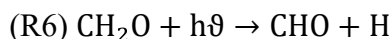
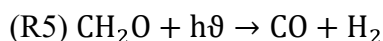
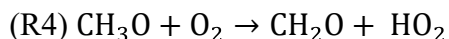


Fig. 1.1 The sketch is representing microbial processes of organic matter transformation in marine sediments. Organic matters (OM) is represented by CH_2O , deposited on the seafloor is aerobically degraded to CO_2 and H_2O . In the sulfate zone, OM is anaerobically transformed by fermentation to hydrogen (H_2), carbon dioxide (CO_2) and smaller organic compounds as acetate (CH_3COOH). CH_4 may be anaerobically and aerobically oxidized or emitted to the water column.

1.2 Polar regions under climate change

Polar regions include the Arctic Ocean and the southern on the continent of Antarctica, are well-known covered by sea ice and ice, respectively. This area is very low temperature in winter, even in summer, many areas remain below freezing. Most of the surface water remains frozen state year-round (IPCC, 2007). Two polar regions have experienced remarkably different climate changes in recent decades. The IPCC report indicated that some of the largest

changes which are expected to happen at high latitudes over the next century. Notably, over recent decades changing in sea ice extent has been observed different in the two polar regions. A marked reduction in sea-ice extent in the Arctic has reached minima in late summer, the extent of Antarctic sea ice has increased by a small amount (Turner and Overland, 2009) or remains steady (Maksym et al., 2012). In Western Antarctica, rates of warming are higher (Vaughan et al., 2003) with decreasing sea-ice (Stammerjohn et al., 2008) and retreating ice shelves (Wouters et al., 2015) than Eastern Antarctica. Further, a seasonal ice-free ocean could influence on Arctic ecology and climate, enhancement of solar irradiance, increase mixing, and reduce the albedo of the Arctic Ocean during boreal summer (IPCC, 2013).

Global average temperature increases up to 1-2°C over a few decades (Thorpe et al., 1996, <https://www.epa.gov/climate-indicators/climate-change-indicators-us-and-global-temperature>). Bracegirdle et al. (2008) found that temperatures across the Antarctic were expected to increase by about $0.33 \pm 0.1^\circ\text{C}$ per decade on land and $0.26 \pm 0.1^\circ\text{C}$ per decade in the ocean/sea-ice zone. Around East Antarctica sea ice is lost, the warming is predicted to be about 0.5°C per decade. Increase in annual mean temperature across the Arctic is expected with a value of $0.6\text{-}0.8^\circ\text{C}$ per decade over the northern parts of the continents and with a value of 1°C per decade over the Arctic Ocean (Solomon et al. 2007).

Those changes affect the global warming following in biology, chemistry, and ocean circulation of the ocean, so these changing may influence on the carbon cycles or biogeochemical cycles. Examination of Arctic and Antarctic CH_4 is vital to rapid climate changes. Besides, observed CH_4 concentration also increases in recent as shown in the previous studies. Increasing CH_4 emission helps to trigger climate change or how it responds to rising atmospheric temperature and sea surface temperature.

1.3 Dissolved methane distribution in seawater in the polar regions

Dissolved CH_4 in surface ocean water tends to be supersaturated with respecting to atmospheric CH_4 (Conrad and Seiler, 1988; Crutzen and Zimmermann, 1991; Bates et al., 1993; Tilbrook and Karl, 1995). Dissolved CH_4 concentrations are supersaturated throughout the equatorial and tropical regions of the Pacific Ocean (Bates et al., 1996; Yoshida et al., 2011), but both supersaturation and undersaturation are known in the extratropics (Kelly and Jeffrey, 2002) and at high latitudes (Bates et al., 1996; Yoshida et al., 2011). In the pelagic zone, CH_4 is supplied via biological production and removed from ocean water by microbial oxidation and by sea-air flux (Grunwald et al., 2009). In productive coastal areas, large amounts of CH_4 ,

which account in part for the high contribution of coastal areas and estuaries to the sea–air flux of CH₄ (Bange et al., 1994), are produced by methanogenesis in buried sediment layers. An enhancement of CH₄ fluxes in coastal upwelling regions likely contribute to the atmosphere on global scale (Rehder et al., 1999). CH₄ fluxes from the coastal areas occupy about 1/10 of the open ocean area, the degree of supersaturation is about 1 order of magnitude higher than that in the open ocean (Bange et al., 1994).

The Arctic area is well-known to be regarded as a great storage of CH₄ (IPCC, 2007; IPCC, 2013; McGuire et al., 2010). Large amount of CH₄ sources exists (e.g., sediment release, CH₄ hydrates, subsea permafrost), and they directly contribute to the water column as well as the atmosphere. Arctic continental shelves are sites of meaningful CH₄ production and emission, super-saturation CH₄ up to 125% was documented in the surface water in the Chukchi Sea (Savichev et al., 2007; Hirota et al., 2009). High levels of CH₄ in the atmosphere of the Laptev Sea and the East Siberian Sea with an average concentration were 2.97 ppm and 2.66 ppm, respectively (Shakhova et al., 2010), suggesting that unusually sizeable CH₄ source from here. The Laptev Sea and the East Siberian Sea are wide and shallow (averaging 50 m) (Jakobsson, 2002) and CH₄ is underlain by thawing permafrost (Nicolosky et al., 2012). Released CH₄ could originate from either deep thermogenic sources (Cramer and Franke, 2005) or released and transported from thawing permafrost on land (Charkin et al., 2011). The elevated CH₄ concentration of microbial CH₄ in sediment with enhanced bottom water CH₄ concentration were observed in the Beaufort Sea. Flux of CH₄ from the entire deep Arctic Ocean ($\sim 8.1 \times 10^6$ km²) resulted from 0.36 Tg CH₄ yr⁻¹ (Lorenson et al., 2016). High CH₄ flux of 18-48 Tg yr⁻¹ in shallow water from seabed is estimated (Hornafius et al., 1999) contributing CH₄ distribution in the water column. While an average annual CH₄ input into the atmosphere of Cicerone and Oremland. (1988) was about 560 Tg, their emission rates were lower than the number of that at eastern Siberian Arctic shelf of 3.7 mg CH₄ m⁻² d⁻¹ attributed to subsea permafrost degradation reported by Shakhova et al. (2010).

The Southern Ocean is the most biologically productive region with high biomass of zooplankton, Antarctic krill and salps (Knox, 2007) suggesting large amount of CH₄ in the Southern Ocean. High CH₄ concentration at high latitudes is suggested by enhancement of biological CH₄ production (Heeschen et al., 2004). Also, higher CH₄ concentration concurrently existed with an abundance of chlorophyll-*a* concentration (Yoshida et al., 2011). In the surface water of the Southern Ocean, oceanic CH₄ has been reported to be supersaturated and undersaturated with respecting to atmospheric CH₄ (Tilbrook and Karl, 1994; Bates et al.,

1996; Heeschen et al., 2004; Yoshida et al., 2011). The Southern Ocean has contributed minor sources and sinks to the atmosphere comparing to the Arctic Ocean. Air-sea exchange is a primary source of CH₄ in the West Southern Ocean and mixing between surface water and warm water from deep water upwell controls distribution of CH₄ in the surface water (Heeschen et al., 2004).

1.4 Relation of CH₄ versus precursors

Significant CH₄ production coincides with existence of high organic materials in sedimentary layer in the anaerobic environment (Reeburgh, 2007) while in oxygenated water methanogens produce CH₄ within suspended particulate matters or within the anoxic center of sinking particles (Sasakawa et al., 2008; Forster et al., 2009) which accumulate in the pycnocline (Karl and Tilbrook, 1994). Correlation between CH₄ and chlorophyll-*a* concentration (Damm et al., 2015a; Yoshida et al., 2011) has been found.

In the Arctic, distinct post-bloom nutrient limitation in Atlantic- and Pacific-dominated sectors are examined. Limitation of nutrient as a possible regulator of CH₄ in production in surface water has been investigated recently (Damm et al., 2010) while CH₄ excess in the ocean surface relative to the atmospheric equilibrium (Damm et al., 2015b). Different nutrient limitation can stimulate the growth of bacterioplankton assemblages and sequential organic matter, biogeochemical cycling of carbon (Thingstad et al., 2008). CH₄ is converted from methyl-phosphonate (MPn) with a phosphate-stress condition in subtropical gyres (Karl et al., 2008) but converted from dimethylsulfoniopropionate (DMSP) in polar oligotrophic water with nitrate-stress environment (Damm et al., 2010). Methanogens that produce CH₄ by catabolism of methylated compounds are microorganism mainly consisting of Archaea. Methylotrophic bacteria use methylated compounds including methylated sulfur compounds such as DMSP and dimethyl sulfide (DMS) (Neufeld et al., 2008; Florez-Leiva et al., 2013). Both DMSP and DMS are S- and C-containing organic compounds that are biosynthesized by phytoplankton, and assimilated and degraded by bacterioplankton (Kiene et al., 2000). DMSP transforms to DMS which influences global temperature with cooling effect (Andreae, 1990). The DMSP degradation supply precursors for the CH₄ formation suggesting methylotrophic methanogenesis as the principal pathway (Tallant and Krycki, 1997; Damm et al., 2010).

1.5 Physical and biogeochemical factors impact CH₄ distribution

The main factors controlling CH₄ supersaturation in tropical and extratropical regions are equatorial upwelling and seasonal variations of sea surface temperature (SST), respectively (Bates et al., 1996). SST affects CH₄ production rate and CH₄ oxidation rate (Hanson and Hanson, 1996). Maximum CH₄ growth rate is active in seawater temperature of 15°C. In the polar regions, seawater is cool; it may limit CH₄ production rate and oxidation rate compared to the warm seawater.

Water masses in the Southern Ocean are characterized by physical processes and biogeochemical processes (Orsi et al., 1995). In the Southern Ocean, maximum CH₄ concentration exists in a subsurface layer between 10°S and Sub-Tropical Front (STF, 47°S). In the south of STF, the maximum CH₄ concentration tends to be high by biogenic CH₄ production (Yoshida et al., 2011). A good correlation of CH₄ concentration and water density between 10°S and PF (Polar Front) in the Southern Ocean and in the eastern Arctic Ocean suggest a significant role of difference in density between particles having anaerobic microenvironment and seawater (Yoshida et al., 2011; Fenwick et al., 2017). CH₄ is produced by the abundance of Antarctic krill (Nicol et al., 2000) reported in the cool surface water in the south of Antarctic Circumpolar Current. Correlation between chlorophyll-*a* concentration and CH₄ concentration is found (Yoshida et al., 2011).

In the Arctic Ocean, high CH₄ concentration in colder and denser water from Pacific Winter Water has been found in the Chukchi shelf-water (Fenwick et al., 2017). Elevated CH₄ concentration is produced in anoxic sediment and released into the water column as bubble from sediment (Leifer and Patro, 2002). Then it is transported from shelf water into the basin by ocean current with mixing or diffusion (Shakhova et al., 2010; Matveeva et al., 2015; Lorenson et al., 2016).

1.6 Objective of the thesis

Global warming affects air temperature, SST, ocean biogeochemistry, sea ice cover, characteristics of water masses and so on. It leads to involve dissolved CH₄ distribution in seawater especially in the polar regions where are well-known the most sensitive to climate change. Atmospheric CH₄ concentration continues increasing over the past 150 years (Etheridge et al., 1992) due to increasing amount of CH₄ entering the atmosphere. Thus, this is a motivation to study CH₄ cycling with combining of physical and biogeochemical factors under climate change in the polar regions.

The sea-air CH₄ flux estimations are still large uncertainty, and the contribution of oceanic CH₄ production is still poorly understood because of the sparsity of available data in the previous studies. Examination of comprehension of CH₄ dynamics in surface seawater and water column is needed to investigate. Improvement of measurements and observations recently aims to enhance the accuracy of CH₄ flux estimation. Clarification of spatial distribution features of sea-air CH₄ flux with high-resolution data is one of the main purposes of this study to reduce these uncertainties. Almost previous studies focused on the influence of biochemical factors on dissolved CH₄ distribution in seawater than physical factors. Combining effects of physical and biogeochemical factors to investigate CH₄ distribution and sea-air CH₄ flux in the polar regions is different from the previous studies. I would like to know and understand how these factors control dissolved CH₄ distribution in seawater.

We conducted new continuous measurement of CH₄ in the eastern Southern Ocean in austral summer 2012-2013, and in the Arctic Ocean in summer 2015. Also, discrete samples of dissolved CH₄ were taken in the same cruise in 2015. Another Arctic expedition cruise conducted in the Arctic Ocean and extended to the northern Arctic Ocean in August 2016. Sampling, results, and discussion in each region are shown in each chapter with distinct factor impacting to dissolved CH₄.

In this thesis, contents include: General introduction of CH₄ is shown in Chapter 1. Chapter 2: Measurement system, new continuous measurement system coupling with an equilibrator and Gas chromatography - Isotope Ratios Mass Spectrometer, is used to detect dissolved CH₄ in seawater. Consideration of controlling factors impact to dissolved CH₄ distribution with high-resolution data in the Southern Ocean which is shown in Chapter 3. Influence of a warm-core eddy on CH₄ distribution in the Arctic Ocean is reported in Chapter 4. CH₄ distribution in the Arctic Ocean in the sea ice cover is displayed in Chapter 5. Finally, General Conclusions is described in Chapter 6.

CHAPTER 2: METHODS

In this chapter, I describe the continuous underway measurement of oceanic CH₄ coupling with an equilibrator and Gas Chromatography (GC) measurement of CH₄ and N₂O. In addition, measurement of DMS/DMSP is also shown by GC equipped with Flame Photometric Detector (FPD).

2.1 Cavity-Ring Down Spectroscopy system (CRDS)

In this study, Cavity Ring-Down Spectroscopy (CRDS) system coupling with an equilibrator is using for detecting dissolved CH₄ in seawater.

CRDS is a highly sensitive technique using an optical cavity to measure absorption on the order of parts per million. Basing on tunable light sources were chosen, the spectroscopy and kinetics of chemicals can be detected. A laser is used as a device that emits light through a process of optical amplification and attenuation through reflected mirrors. When molecule gas is present, the rotational vibration modes provide additional loss mechanism and the ring-down times get shorter and is continuously repeated (~100 times per second). The loss of light is leakage through the mirrors produced a signal, the light intensity inside the cavity steady leaks out and decayed to zero in an exponential shape. This decay, or "ring-down", is measured in real-time by a photo-detector. The concentration of any gaseous species can be determined by measuring the strength of this absorption. The final concentration data is particularly robust because it is derived from the difference between these ring down times and is therefore independent of laser intensity fluctuations or absolute laser power. The ring-down profiles are transformed into an absorption curve with a line shape and are used to calculate the gas concentration. The gas concentration is determined by a multi-parameter fit to line-shape and is proportional to the area under the curve (Fig. 2.1) (<https://www.picarro.com>).

Light intensity is a function of time in CRDS system that has a resonant absorbance. The absorbance (A) can be calculated by the equation:

$$A = \ln \left(\frac{I_0}{I} \right) = \sigma L_{\text{abs}} N$$

where: A: absorbance; I₀: initial light intensity; I: transmitted light intensity; σ: absorption across section; L_{abs}: path length of light through the absorber; N: number density of absorber molecules.

If the loss process for light is leakage only through the mirrors, the ring down is characterized by:

$$I = I_0 \exp\left(-\frac{t}{\tau_0}\right)$$

$$\text{with } \tau_0 = \frac{t_r}{2(1-R)} = \frac{L_{\text{opt}}}{c(1-R)}$$

where: t_r : roundtrip time for light in the cavity; R : mirror reflectivity; L_{opt} : a distance between two mirrors; c : speed of light; τ_0 : 1/e decay time of the light (ring down lifetime).

Recently, CRDS is installed at atmospheric chemistry, is used for measuring concentrations of CH₄, CO₂, N₂O, NH₃ and water vapor. By combination with an equilibrator in order to extract CH₄ and CO₂ in seawater and is detected by CRDS.

Oceanic and atmospheric CH₄ and CO₂ measurements were made quasi-continuously onboard R/V *Mirai* (Japan Agency for Marine-Earth Science and Technology). Underway measurements of CH₄ and CO₂ mixing ratios in dry air equilibrated with seawater (xCH_4^{eq}) and ambient air (xCH_4^{air}) were obtained with a system consisting of a cavity ring-down spectroscopy analyzer (CRDS G2301, Picarro, Inc., California, USA). A shower-head type equilibrator has been used for measurement of the partial pressure of CO₂ in seawater and overlying air since the late 1960s (Kortzinger et al. 2000; Yoshikawa-Inoue 2000; Yoshikawa-Inoue & Ishii 2005) Fig. 2.2. Seawater was pumped from a water inlet ~5 m below the sea surface at the bow of the ship to an onboard laboratory for measurement of CO₂ and CH₄, salinity, dissolved oxygen and fluorescence. Sampled seawater was continuously introduced into the equilibrator at 10 L/min. Seawater temperature in the equilibrator was measured continuously with a PT-100 sensor. During the cruise, the increase of water temperature from the water inlet to the equilibrator was between 0.1 °C (at low latitudes) and 0.4 °C (at high latitudes). Air equilibrated with seawater was drawn continuously from the outlet of the equilibrator, then flowed into the CRDS analyzers (100 mL/min), and was then returned to the equilibrator. Sample air was circulated within a closed loop during measurements of oceanic CO₂ and CH₄. To avoid band broadening effects caused by water vapor (Bates et al. 1996; Nara et al. 2012), water vapor was removed from the sample air by two electric dehumidifiers that use the Peltier effect, Nafion tubing (Perma Pure LLC, NJ, USA), and a chemical desiccant column (Mg(ClO₄)₂). Equilibration pressure was assumed to be atmospheric pressure because the seawater inlet in the onboard laboratory was at atmospheric pressure.

The CRDS analyzer was calibrated daily with three CH₄ and CO₂ reference gases (1606.2, 1876.5, and 2075.1 ppb for CH₄; 249.23, 380.78, and 444.14 ppm for CO₂); the CH₄ data are

traceable to the World Meteorological Organization CH₄ mole fraction scale (Dlugokencky et al., 2005; Tsuboi et al., 2016). The CRDS analyzer produced about 100 digital analyses per minute from which we calculated and used one-minute means. We discarded data recorded for at least 30 min immediately after the changes in flow patterns during measurements of reference gases and ambient air. Based on replicated measurements of a sample gas in the cylinder, the precision of the analyses ($\pm 1\sigma$) was estimated to be better than 0.1 ppm and 2 ppb for CO₂ and CH₄, respectively.

Although both CO₂ and CH₄ were measured simultaneously during the cruise, I focused in this study on the CH₄ distribution, the factors controlling the CH₄ distribution, and the sea–air flux of CH₄. CH₄ concentrations in surface seawater (C_w) and the mixing ratio in dry air of CH₄ equilibrated with surface seawater (xCH_4^{sw}) were calculated by using the temperature and salinity dependence of CH₄ solubility in seawater (Wiesenburg and Guinasso, 1979) and the barometric pressure at the sea surface (P_{bar}). The partial pressure of CH₄ in surface seawater (pCH_4^{sw}) was calculated as

$$pCH_4^{sw} = xCH_4^{sw}(P_{bar} - W)$$

where W is water vapor pressure equilibrated with surface seawater.

Similar to C_w calculations, CH₄ concentrations in the ambient atmosphere (C_a) and the mixing ratios of CH₄ in the ambient air (xCH_4^a) were calculated by using equilibrium temperature and salinity dependence of equilibrium CH₄ solubility upper the surface seawater. The partial pressure of CH₄ upper the surface seawater (pCH_4^a) was calculated as

$$pCH_4^a = xCH_4^a(P_{bar} - W)$$

2.2 Gas chromatography: Isotope Ratios Mass Spectrometer (GC-IRMS)

2.2.1 Seawater sampling of CH₄ and N₂O

We took seawater samples during the cruise from a Niskin bottle attached to a conductivity-temperature-depth (CTD) rosette at a certain depth. Seawater was poured continuously into a 120 mL glass vial and had overflowed until at least 240 mL for avoiding contamination in the ambient air. After subsampling, 0.5-mL of a saturated solution of mercuric chloride (HgCl₂) was added to prevent biological activity in the vials. Then bottles were immediately sealed using butyl-rubber caps and aluminum caps with at least no headspace. The samples were then stored at about 4°C in the cooling room for later GC analysis in the

laboratory on land. These samples had been measured for dissolved CH₄ and N₂O in seawater by a gas chromatography - isotope ratios mass spectrometer (GC-IRMS).

2.2.2 Gas chromatography: Isotope Ratios Mass Spectrometer (GC-IRMS)

Both mixing ratios and stable isotopic composition of CH₄ and N₂O of seawater samples were determined by Gas Chromatography - Isotope Ratios Mass Spectrometer (GC-IRMS). Details of the analytical method we used have been reported by Tsunogai et al. (2000) and Hirota et al. (2009). The analytical system consisted of a purge-and-trap CH₄ and nitrous oxide (N₂O) extraction unit, a GC (HP6850; Agilent Technologies, Santa Clara, CA, USA), an IRMS (Finnigan MAT 252; Thermo Fisher Scientific, Waltham, MA, USA) with a combustion interface (Combustion III, Thermo Fisher Scientific), a gas dryer (dry ice temperature trap (-79°C)), a liquid N₂ temperature trap and liquid O₂ trap (Figure 2.2). Dissolved CH₄ and N₂O in the samples was extracted by an ultra-pure helium gas stream (>99.99995%) as a carrier gas with a flow rate of 80 ml/min. CH₄ was separated from other gases such as H₂O, CO₂, N₂ and non-condensable gases in seawater sample when going through this system. H₂O and CO₂ are removed by a chemical trap filled with magnesium perchlorate (Mg(ClO₄)₂; Wako Pure Chemical Industries, Ltd., Osaka, Japan) and Ascarite II (NaOH-coated silica; Thomas Scientific, Swedesboro, NJ, USA). The gas stream, including CH₄ and N₂O, was then passed through a dehydration unit (U-shaped stainless steel trap at dry-ice temperature). The extracted CH₄ and N₂O were collected in different trap and were maintained at -183°C and -195°C using liquid oxygen and liquid nitrogen bath, respectively. The non-condensable gases such as N₂, O₂, Ar, and CO could pass through these traps and could be released into the atmosphere. Trace gases, including N₂O and CH₄, were trapped in a cryotrap (Silicosteel tubing packed with Porapak-Q) at liquid-oxygen temperature. The concentrated gases were released from the trap by heating under helium flow and then re-concentrated on the head of a PoraPLOT-Q capillary column (Chrompack, Inc., Raritan, NJ, USA) maintained at liquid-oxygen temperature. Then, the liquid-oxygen trap was removed to release trace gases and introduced into the column of GC that maintained at 30°C. Trace gases were separated by using helium flow rate of 0.69 mL/min, and finally, they were converted to CO₂ by passing over the combustion interface maintaining at 960°C and then carried into a Finnigan MAT 252 for measuring carbon content and carbon isotope. Determining CH₄ and N₂O were individually indicated by the retention time at 390 s and 530 s, respectively. Then, the CO₂ was introduced into mass spectrometer in order to determine δ^{45} and δ^{46} of CO₂. The δ^{13} values of CH₄

calculated from δ^{45} and δ^{46} values of CO_2 . In contrast, both $\delta^{15}\text{N}_{\text{N}_2\text{O}}$ and $\delta^{18}\text{O}_{\text{N}_2\text{O}}$ calculated from the δ^{45} and δ^{46} values of N_2O .

After each single analysis, the column was heated to 120°C at a rate of $5^\circ\text{C}/\text{min}$ in order to remove contaminants from the column. During the time, seawater sample in the purging bottle was run out via the water vent. The purging bottle, water transfer line was washed out with pure water and flushed using carrier gas in preparation for the next sample. Each sample analysis took about 50 min.

2.3 Calculation of CH_4 concentration, saturation, and sea-air flux

2.3.1 Calculation of CH_4 concentration for CRDS system

The solubility of CH_4 is calculated by the method of Wanninkhof (2014) as:

$$\ln \beta = A1 + A2 (100/T) + A3 \ln (T/100) + S [(B1 + B2(T/100) + B3 (T/100)^2]$$

where β is the dimensionless Bunsen coefficient; K_o is the solubility expressed in $\text{mol L}^{-1} \text{atm}^{-1}$. T is sea water temperature in Kelvin and S is salinity in ‰. $A1$, $A2$, $A3$, $B1$, $B2$, and $B3$ are constant values which are shown in Table 2.1.

Dissolved concentration of CH_4 in seawater (C_w) is calculated as

$$C_w = K_o \times p\text{CH}_4^{\text{sw}} \quad (2)$$

Table 2.1 Coefficients for temperature and salinity dependence of solubilities of CH_4 (Wanninkhof, 2014).

	A1	A2	A3	B1	B2	B3
CH₄	-68.8862	101.4956	28.7314	-0.076146	0.04397	-0.006872

2.3.2 Saturation and Sea-air flux calculation

The saturation ratios (SR) of dissolved CH_4 , which is the ratio of the C_w to the CH_4 concentration in seawater equilibrated with ambient air (C_a), is given by:

$$\text{SR (\%)} = \frac{C_w}{C_a} \times 100 \quad (3)$$

SR < 100% indicates water that is undersaturated relative to the atmosphere, and SR > 100% indicates supersaturation; SR is not affected by barometric pressure. The sea–air flux of CH₄ (F) was calculated as

$$F = k \times K_o \times (pCH_4^{sw} - pCH_4^{air}) = k \times (C_w - C_a), \quad (4)$$

where K_o is CH₄ solubility in seawater and k is the gas transfer piston velocity calculated by the method of Wanninkhof (2014) as

$$k = 0.251 \langle U_{10}^2 \rangle (Sc/660)^{-0.5} \quad (5)$$

Here, $\langle U_{10}^2 \rangle$ is wind speed 10 m above the sea surface, and Sc is the Schmidt number of CH₄ in seawater.

Total CH₄ emissions in the Southern Ocean were calculated as

$$\text{Total CH}_4 \text{ emissions} = \sum_{i=1}^n (F_i \times SA_i) \quad (6)$$

where subscript i is grid number, and SA is the surface area within a grid where the sea surface is free of sea ice. In this study, I assumed no CH₄ exchange across the air-sea interface where there is sea ice.

2.5 Gas chromatography: Flame Photometer Detector (GC-FPD)

2.5.1 Seawater sampling of DMS/DMSP

We took DMS samples during the cruise from a Niskin bottle attached to CTD rosette from at a certain depth. Seawater was directly passed through a 47 mm GF/F filter (pore size 0.7 μ m, Manufactured by Whatman) via a tubing, which is fixed by filter holder (PP-47 type, Advantech), into a 30 mL brown glass vial without contacting to the air. Then bottles were sealed by using rubber caps and aluminum caps. Another two DMSP samples were collected in each 30 ml brown glass vial without filtration. Then in the laboratory, 300 mL of magnetic filter funnel using a GF/F wetter filter with a small amount of Mili-Q water, gravity filtration was carried out and took 4 mL of dissolved DMSP (DMSPd) comparing to a testing-standard volume which was prepared. 0.5-mL of 5M HCl (hydrochloric acid) was added into sub-sample of 4 mL seawater by using 1000 μ L micropipette to prevent a transformation of DMSP to DMS and biological activities. These bottles were sealed with a rubber cap and aluminum cap and stored in a dark container. For total DMSPt (DMSPt) sample, 8 ml seawater was sucked out and added into 10 ml brown vial by using 5000 μ L micropipette. Then 1 mL of 5M HCl was

added into sub-sample of 8 mL by using 1000 μ L micropipette and was sealed with a rubber cap and aluminum cap and stored in a dark container.

2.5.2 Gas chromatography: by Flame Photometer Detector (GC-FPD)

DMS and DMSP concentrations were measured by Gas Chromatography (GC-2014, Shimadzu) equipped with Flame Photometer Detection (FPD). GC-FPD is a technique used to analyze sulfur or phosphorus compounds. FPD is a type of detector which is used to separate components by burning with hydrogen flame and emitted specific wavelength for phosphorus and sulfur. Thus, GC-FPD is known useful for DMS measurement.

Before using measurement system, the lines were needed to purify by flowing Helium carrier gas through connecting line and GC, and the column was heated to 140°C for a few hours. DMS and DMSP standard solutions need to calibrate to the amount of signal intensity. 4 mL of NaOH 5M was added to 30 mL of DMSP standard solution of each concentration and then leave it in a refrigerator at 4°C for a day. During that time, DMSP was converted to DMS, show the correlation between them with a slope of 3124.3 ($R^2 = 0.997$) Fig. 2.4. The range of detection of DMS concentrations is from 0-32.3 (nmol L^{-1}).

Before doing an experiment, 1 ml and 2 ml of alkalinity (NaOH) 5M was added into DMSP bottles in order to permit the cleavage of DMSP into gaseous DMS. Then, bottles were shaken and stored at 4°C for at least 12 hours to complete the cleavage before analysis. For DMS samples, 30 mL of filtered seawater was transferred to glass purging bottle from syringe connected directly to sample. The dissolved DMS was extracted from sample water and transformed into the gas phase by purging with pure nitrogen gas for 12 minutes. The extracted DMS and water vapor were needed to separate them. Water vapor was trapped at 8 mm glass U-shape tube (SIBATA) in the 99.99% ethanol with dry-ice bath (-78°C). The extracted DMS was passed through H₂O trap and was concentrated in another glass U-shape tube with chemical substance TENAX TA 60/80 mesh for absorbing DMS in a dry ice-ethanol bath (-78°C). After the extraction completed (about 14 minutes 30 seconds), the liquid in the purging bottle is left via water vent. Thereafter, the DMS gas trapped in the U tube was warmed with the boiling water of about 95-100°C (hot trap) and released gas was introduced into a GC equipped with FPD detector for separation and detection. Sulfur compounds including DMS were detected by FPD (see Fig. 2.5).

Seawater samples for measuring DMS concentrations were measured onboard as soon as possible after filtering treatment. However, seawater samples for measuring DMSP whenever

on board or in the laboratory after adding acid solution and storing in a dark place. For measuring DMSP, it is needed to convert to DMS firstly, and then leaving it in a cool dark place for at least a half-day, and measured it. Due to chemical converting treatment of DMSP, it is often measured on land.

2.6 Calculations of biogeochemical parameters

2.6.1 Apparent Oxygen Utilization (AOU)

AOU is the difference between the measured dissolved oxygen concentration and its equilibrium saturation concentration in water with the same physical and chemical properties. AOU is defined as:

$$\text{AOU} = [\text{O}_2]_{\text{sat}} - [\text{O}_2]_{\text{obs}}$$

$[\text{O}_2]_{\text{sat}}$: the value of oxygen in the water in equilibrium with the atmosphere at temperature and salinity of the water. It is called saturation; it implies that all water is in equilibrium with the atmosphere. $[\text{O}_2]_{\text{obs}}$: is dissolved oxygen observed in the water sample. Biological activities act to change the ambient of oxygen concentration; it leads to change in AOU. Primary production releases oxygen and increases its concentration, while respiration consumes and decrease its concentration. In shallow water, the water column is generally in contact with the atmosphere, so oxygen concentration is typically close to saturation, or AOU value is low. In deep water, water can be out of contact with the atmosphere for long periods of time and large AOU is possible (Garcia et al., 2006).

2.6.2 Fixed nitrogen (N^{**})

N^{**} is a modification of N^* , a parameter defined by Gruber and Sarmiento (1997) (Codispoti et al., 2005a):

$$\text{N}^{**} = [\text{NO}_3^-] + [\text{NO}_2^-] + [\text{NH}_4^+] - 16[\text{PO}_4^{3-}] + 2.9$$

This parameter estimates the impact of *in situ* fixed nitrogen additions and removals on water masses. N^{**} includes ammonium concentrations because of the high ammonium values (Codispoti et al., 2005a). N^{**} is a quasi-conservative tracer, the global mean ratio of 106:16:1 between carbon, nitrogen, and phosphorus (C:N:P). N^{**} is about zero if the Redfield ratio is retained, a negative N^{**} value indicates a deficit of dissolved inorganic nitrogen (due to denitrification), while a positive value indicates N fixation (Redfield, 1942; Gruber and Sarmiento, 1997).

Nishino et al. (2005) define for N^{**} that differs $0.1\mu\text{mol kg}^{-1}$. Pacific water enters the Arctic tend to have negative values due to the conversion of fixed-nitrogen to free nitrogen (N_2) that process enhanced in low oxygen water and sediments. Atlantic water has a positive value because of net nitrogen fixation (Codispoti et al., 2005b). Because of the influence of microbial processes in shelf sediments, it causes a reduction in N^{**} (Devol et al., 1997).

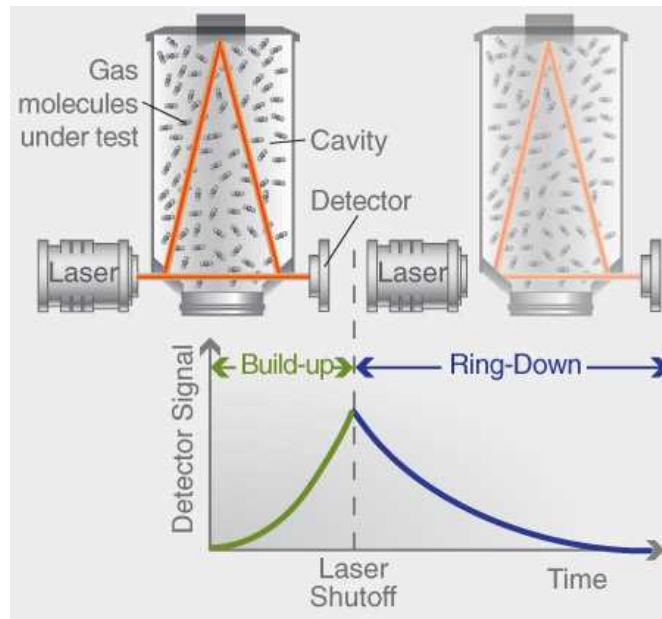


Fig. 2.1. Schematic of CRDS analyzer with a laser source, reflected mirrors and detector (upper) and profile of a ring down measurement (source: <https://www.picarro.com>).

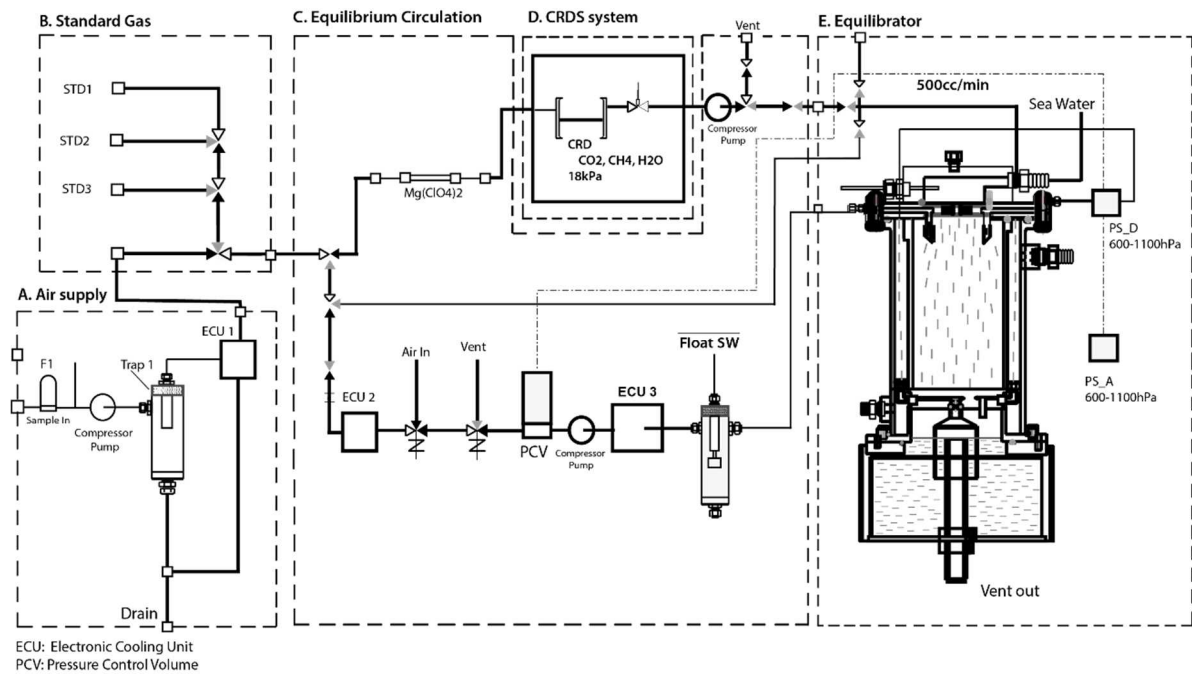


Fig. 2.2 Schematic diagram of underway system for measurements of CO₂ and CH₄, consisting of a CRDS analyzer and a shower-head-type equilibrator.

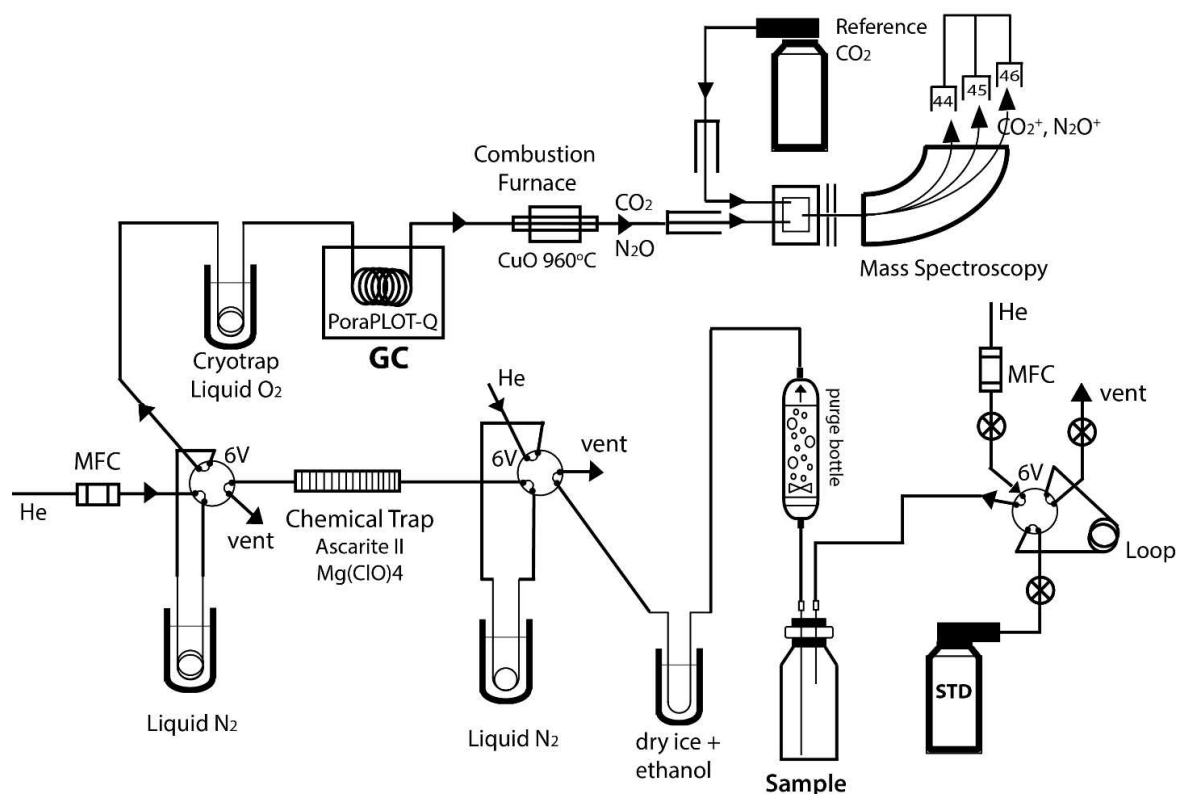


Fig. 2.3 Schematic diagram of a system using for CH₄ and N₂O analysis. 6V: six-port two-position Valco valve; STD: gas standard; GC: gas chromatography; MFC: mass flow controller.

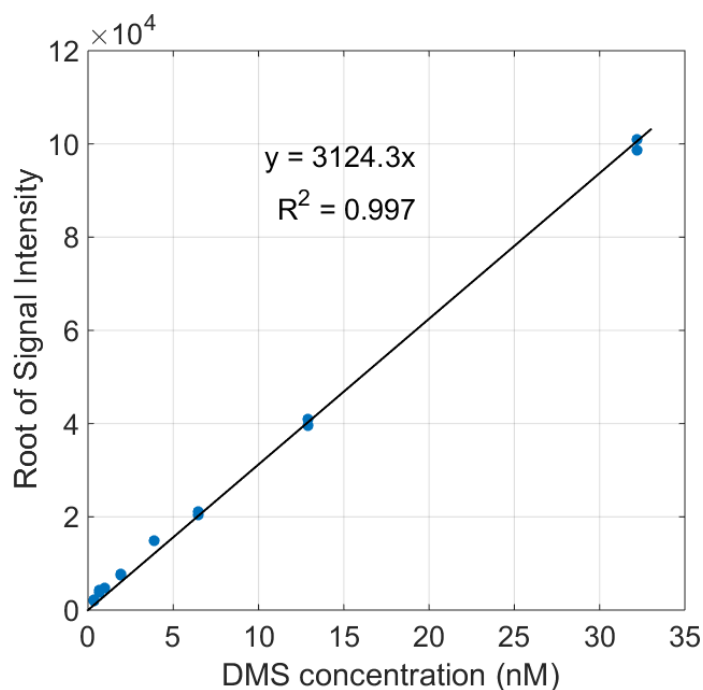


Fig. 2.4 The linear relationship between roots of signal intensity and DMS concentration of 30-mL DMSP standard was carried in the laboratory. Solid line shows the linear relationship between DMS concentration and signal intensity: $y = 3124.4x$ ($R^2 = 0.997$).

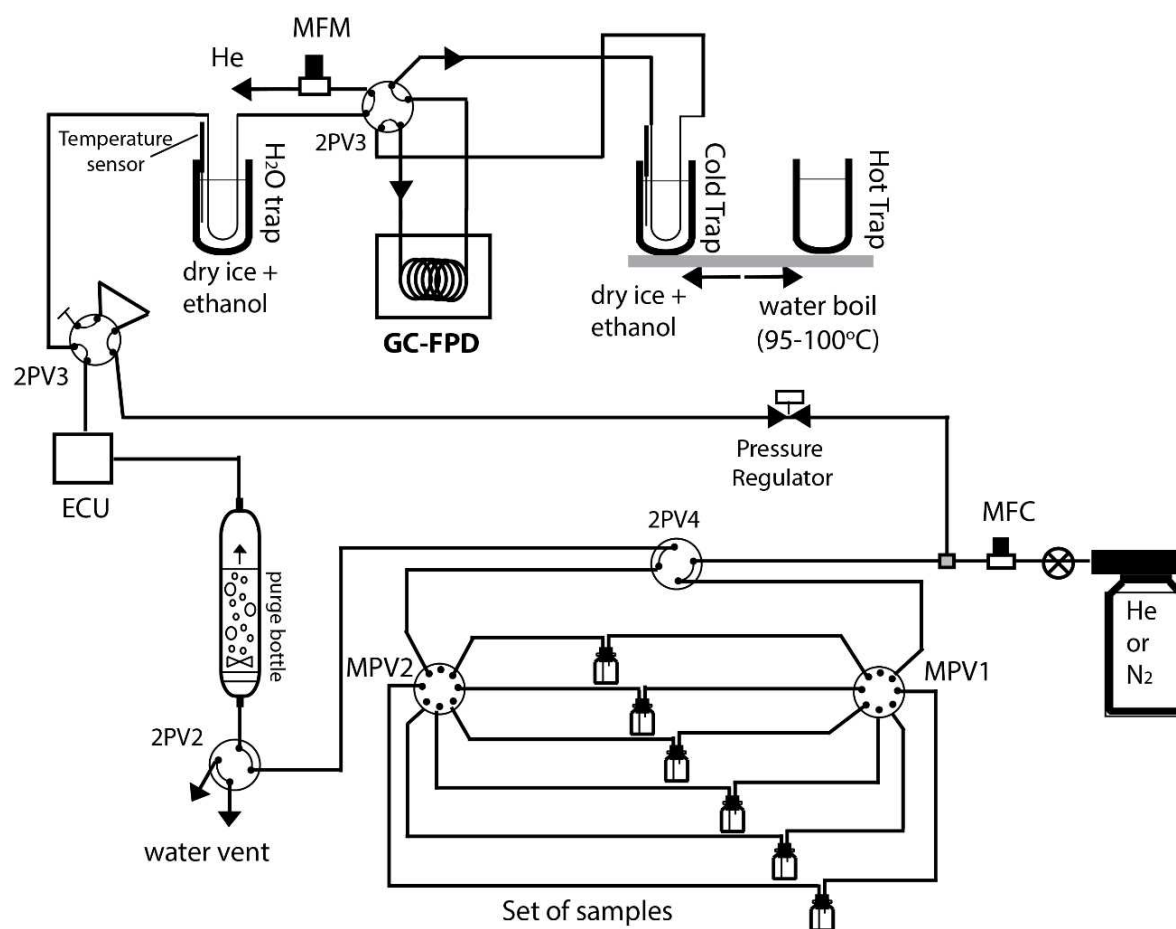


Fig. 2.5 Schematic diagram of the system using for DMSP analysis. MFC: mass flow controller; ECU: Electronics Cooling Unit; 2PV: two-port valve; MFM: mass flow meter; GC-FPD: gas chromatography equipped with Flame Photometric Detector.

CHAPTER 3:

HIGH-RESOLUTION METHANE EMISSIONS

IN THE SOUTHERN OCEAN

3.1 CH₄ distribution in the Southern Ocean

There are some works that have investigated CH₄ concentration in the Southern Ocean (Lagmontag et al., 1973; Bates et al., 1996; Tilbrook and Karl, 1994; Heeschen et al., 2004; Yoshida et al., 2011). Dissolved CH₄ concentrations are supersaturated but both supersaturation and undersaturation are known in the extratropics (Kelly and Jeffrey, 2002) and at high latitudes (Bates et al., 1996; Yoshida et al., 2011). The main factors that control CH₄ supersaturation in tropical and extratropical regions are equatorial upwelling and seasonal variations of sea surface temperature (SST), respectively (Bates et al., 1996). In the pelagic zone, bacterial CH₄ is released from the digestive tracts of zooplankton and from sinking organic particles; CH₄ is removed from ocean waters by microbial oxidation and by sea–air flux (Grunwald et al., 2009). In productive coastal areas, large amounts of CH₄ are produced by methanogenesis in buried sediment layers, which account in part for the high contribution of coastal areas and estuaries to the sea–air flux of CH₄ (Bange et al., 1994).

The CH₄ is produced by microbial methanogenesis in anaerobic environments and by bacteria in organic particles or in the gut of zooplankton (Karl and Tilbrook, 1994). CH₄ production from methyl phosphate decomposition in the aerobic environment has been reported (Karl, 2008). Recently, observation of a release CH₄ from sinking particles into the water column by measuring of dissolved CH₄ concentration and $\delta^{13}\text{C-CH}_4$ (Sasakawa et al., 2008). Methanogens have been shown to be zooplankton-species-specific (de Angelis and Lee, 1994), which contributes to the lack of a consistent correlation between CH₄ concentrations in seawater and commonly measured biological parameters. Burke et al. (1983) reported weak correlations between CH₄ concentrations and particulate and biological parameters in the eastern tropical North Pacific and that the CH₄ distribution there was largely controlled by physical oceanographic processes.

In the surface water in the Southern Ocean has been reported that undersaturated CH₄ with respecting to atmospheric CH₄ and a small sink to the atmosphere (Tilbrook and Karl, 1994). Undersaturation of CH₄ deep in the water column in the Arctic Ocean is a result of microbial oxidation (Rehder et al., 1999) and is consistent with the maximum seawater

temperature in the core of Warm Deep Water (Heeschen et al., 2004). In the Southern Ocean, low CH₄ in surface water is a consequence of entrainment by surface water of deep water depleted in CH₄ (Yoshida et al., 2011) because of its low content of organic particles (Fischer et al., 1988), upwelling of old deep water (Heeschen et al., 2004, Rehder et al., 1999), or both.

The Southern Ocean is one of the most biologically productive oceanic regions in the world; it is characterized by high biomasses of zooplankton, Antarctic krill, and salps (Knox, 2007), all of which produce large amounts of CH₄. Although some researchers have measured CH₄ concentrations in the Southern Ocean (Lagmontag et al., 1973; Bates et al., 1996; Tilbrook and Karl, 1994; Heeschen et al., 2004; Yoshida et al., 2011), accurate assessment of the sea–air flux of CH₄ has been limited by the sparsity of available data. This study aims to clarify the spatial distribution of the sea–air flux of CH₄ in the Southern Ocean and thus to reduce the uncertainties in estimations of the ocean as a source of atmospheric CH₄. Here, I present the distribution of the sea–air flux of oceanic CH₄ along the course of a scientific cruise in the Pacific and Indian sectors of the Southern Ocean during the 2012/13 austral summer; I use SR and AOU correlation to examine the climatological flux of CH₄ from the entire Southern Ocean south of 50°S during the austral summer.

3.2 Water masses in the Southern Ocean

In the Southern Ocean, major oceanic fronts separate water masses with different physical and biogeochemical properties (Sarmiento et al., 2004). The oceanic front is a boundary between two distinct water masses, moves around Antarctica. It was defined by SST and sea surface salinity (SSS). These are the Sub-Tropical Front (STF), Sub-Antarctic Front (SAF), Polar Front (PF), Southern Antarctic Circumpolar Current Front (SACCF), Southern limit of Upper Circumpolar Deep Water or Southern Boundary (SBDY), and Antarctic Slope Front (ASF) (Orsi et al., 1995; Rintoul et al., 1997; Rintoul and Bullister, 1999; Yaremchuk et al., 2001; Ainley and Jacobs, 1981). The Sub-Antarctic Zone (SAZ) locates between STF and SAF, the Polar Frontal Zone (PFZ) between SAF and PF, the Antarctic Zone (AZ) between PF and SACCF, the Southern Zone (SZ) between SACCF and SBDY, and the south of SBDY (SSBDY).

A major dynamic feature of the Southern Ocean is the Antarctic Circumpolar Current (ACC), which flows eastward around Antarctica and mixes with the various water masses along its path (Callahan, 1972; Georgi, 1981). The most voluminous water mass is Circumpolar Deep Water (CDW), which is carried around Antarctica by the ACC (Whitworth and Nowlin,

1987). Near Antarctica, CDW moves upward through the water column from north to south, approximately along the equal density surface, and mixes at the shelf break with shelf waters (Locarnini, 1994).

3.3 Regional Setting

Oceanic and atmospheric CH₄ and CO₂ measurements were made quasi-continuously onboard R/V *Mirai* (Japan Agency for Marine-Earth Science and Technology) during the Pacific and Indian sectors of scientific cruise MR12-05 (legs 2 and 3) in the Southern Ocean from late November 2012 to mid-February 2013 (Fig. 3.1). Fig. 3.1 depicts the distribution of oceanic fronts drawn on the basis of Orsi et al. (1995). Besides, I also compare CH₄ to another measurement of biochemical variables to investigate the controlling factors to CH₄ distribution. Along the cruise track, we determined the geographical position of each oceanic front on the basis of vertical profiles of temperature and salinity (Orsi et al., 1995; Rintoul et al., 1997; Rintoul and Bullister, 1999; Yaremchuk et al., 2001), SST, and SSS (Chaigneau and Morrow, 2002). The cruise included a zonal transect along the World Ocean Circulation Experiment Hydrographic Program line S4 and four meridional transects.

Along 140°E, the STF is the boundary between warm, salty subtropical water and cool, fresh sub-Antarctic water, which is located at 44°S. The SAF is recognized by a steep horizontal gradient of SST occurred around 54°S, which is the northern edge of the eastward-flowing Antarctic Circumpolar Current. The PF is characterized by a steep temperature gradient of SST and the northern limit of minimum-temperature water with temperatures of less than 2°C near a depth of 200 m, located at 59°S. The SACCF coincides with the southern limit of maximum temperature water over 1.8°C (Orsi et al., 1995), located around 62°S. SBDY is located at 63.5°S, where temperature maximum is 1.5°C (Orsi et al., 1995).

3.4 Results and Discussion

3.4.1 Spatial distribution of dissolved CH₄ in seawater

Calculated CH₄ concentration varies from 2.1–3.6 nmol kg⁻¹ with an averaged value of 3.1 nmol kg⁻¹ (Fig. 3.2) from CRDS system during the cruise. CH₄ concentrations tend to be increased toward the Antarctica Continent while dissolved CH₄ concentration is generally low in tropical regions. Concurrently, sea surface CH₄ concentrations by using GC measurements ranged 2.9–4.2 nmol kg⁻¹ (Fig. 3.3). The degree of saturation of dissolved CH₄ (SR) was

calculated from the observed CH₄ concentration (C_w) and the concentration of CH₄ in water equilibrated with ambient air (C_a , C_a is observed during the cruise). The overall SR in the sea surface ranged from 85-142% (Fig. 3.4), with an average CH₄ saturation was 104%. The estimated values were mostly within the ranges reported in the previous studies.

3.4.2 Comparison of C_w determined by the underway system and by GC-IRMS

C_w determined by the underway system ranged from 2.1 to 3.6 nmol kg⁻¹ (mean = 3.1 nmol kg⁻¹) during leg 2 of cruise MR12-05 (Fig. 3.1). I used reduced major axis (RMA) regression to compare C_w values determined by the underway system with those determined for discrete water samples by GC-IRMS. RMA analysis gave a regression line with a slope of 1.0 ± 0.1 ($r^2 = 0.84$, $n = 39$), which indicates that, within the precision of the analyses, the variations in C_w determined by the underway system were equivalent to those of the purge-and-trap technique we used for our discrete water samples (Reeburgh, 2007). A moderately large regression line intercept (-0.5 ± 0.2), indicates that the GC analyses gave C_w values that were ~ 0.5 nmol kg⁻¹ greater than those of the CRDS analyses. Similar differences were reported by Bates et al. (1996) who also used the underway system: they reported lower SR values for the underway system than those from comparable discrete water samples reported by Watanabe et al. (1995). Forster et al. (2009) also reported that equilibrated CH₄ concentrations measured underway were significantly lower than those obtained from rosette samples.

To allow further and more detailed discussion of the differences between the results of these two methods of analysis, I need to reconsider the need for collection of discrete seawater samples, the methods employed for preservation of such samples, GC-IRMS analysis of seawater samples, and equilibrium in the equilibrator (SCOR WG #143, 2014, <https://portal.geomar.de/web/scor-wg-143>). The mole-fraction scale of reference gases used and the natural variability of oceanic CH₄ must also be taken into account. Further discussion of the consistency of the results of the two methods is beyond the scope of this paper.

3.4.3 Mixing ratios of atmospheric CH₄

Temporal variation of atmospheric CH₄ in the Southern Hemisphere was examined to confirm that the atmospheric CH₄ mixing ratios measured by the CRDS analyzer were accurate (Fig. 3.6). During the cruise, the mean of atmospheric xCH_4^{air} was 1775.9 ± 5.4 ppb ($n = 33$) in late November 2012, 1767.2 ± 5.4 ppb ($n = 321$) in December 2012, 1755.7 ± 2.0 ppb ($n = 304$) in January 2013, and 1751.0 ± 1.5 ppb ($n = 150$) in early February 2013, and showed the same seasonal variations as reported for the Southern Hemisphere by Dlugokencky et al. (1994).

Mean values in December 2012 and January 2013 agreed well (within 5 ppb) with monthly means at Cape Grim (40.68°S, 144.69°E), Tasmania, and at the South Pole, on the basis of data from the NOAA/ESRL network (http://www.esrl.noaa.gov/gmd/ccgg/trends_ch4/).

3.4.4 Meridional distribution of CH₄ in surface seawater

The meridional distribution of pCH₄^{sw} between New Zealand and the SBDY (Fig. 3.7) showed a few pCH₄^{sw} peaks of up to 3.4 µatm and an SR range from 120% to 185% in the coastal area off New Zealand. These peaks can be attributed in part to large amounts of CH₄ that are produced by methanogenesis in buried sediment layers in productive coastal areas (Grunwald et al., 2009; Hovland et al., 1993; Scranton and McShane, 1991). CH₄ inputs from riverine and groundwater discharge can also contribute to CH₄ supersaturation in coastal areas (Bugna et al., 1996; Kim and Hwang, 2002; Reeburgh, 2007). In surface seawater off the New Zealand coast, strong pCH₄^{sw} peaks are accompanied by lower water density (lower SSS), which likely reflects the input of fresh water with high CH₄.

In the open ocean area from the STF to the SBDY (Fig. 3.7), pCH₄^{sw} decreased slightly and monotonously to the south and SR decreased from 119% to 100%; both of these results agree well with those of earlier studies (see Table 3.1; Bates et al., 1996; Kelly and Jeffrey, 2002). Atmospheric CH₄ has increased by 180 ppb (about 10%) between 1984 and 2012, although the rate of increase was slower between 1999 and 2006 (Dalsøren et al., 2016; McNorton et al., 2016). The fairly constant SR over these few decades suggests parallel increases of pCH₄^{sw} and pCH₄^{air}. South of the SBDY (63.5°S), pCH₄^{sw} decreased considerably along with decreases in SST and chlorophyll-a concentration and an increase in AOU. The lowest value of SR that we observed (85%) was in this region.

South of the STF, pCH₄^{sw} changed little (Fig. 3.7), but there were steep changes in SST and SSS at each of the other fronts. Because CH₄ solubility is a function of SST and SSS (Wiesenburg and Guinasso, 1979), the relatively constant pCH₄^{sw} indicates that there must have been changes in C_w at each front. According to Wiesenburg and Guinasso (1979; their equation 7), CH₄ solubility decreases by 2.3% for each unit increase in SST above 10 °C at constant SSS (34), and by 0.8% for each unit increase in SSS from 34 at constant SST (10 °C). The decreases of SST and SSS to the south caused large increases of C_w to the south. In the region between the PF and the SBDY, C_w tended to be high, suggesting enhancement of biogenic CH₄ production (Heeschen et al., 2004), lower rates of CH₄ oxidation and CH₄ flux between the sea and overlying air, or both.

The meridional variations of $p\text{CH}_4^{\text{sw}}$ during the sections of the cruise sailing toward and away from southern Tasmania, and during the northbound section sailing toward Fremantle (Figs. 3.1 and 3.7 to 3.10), were similar to those between the STF and the SBDY between New Zealand and Antarctica (Fig. 3.7). During the northbound sailing to Tasmania, $p\text{CH}_4^{\text{sw}}$ varied considerably between 1.8 and 2.9 μatm north of the SAF. Strong $p\text{CH}_4^{\text{sw}}$ peaks that continued over several tens of kilometers coincided with increases in chlorophyll-*a* concentration. However, about one week after the northbound sailing toward Tasmania we found no strong peaks of $p\text{CH}_4^{\text{sw}}$ at the same geographical positions (Fig. 3.9). During the northbound sailing to Fremantle, $p\text{CH}_4^{\text{sw}}$ north of the STF also varied considerably, from 1.8 to 2.7 μatm , but without correlation with chlorophyll-*a* concentration, SST, or SSS (Fig. 3.9). Thus, it appears that none of the other quantities I determined in this study can be reliably used to predict strong peaks of $p\text{CH}_4^{\text{sw}}$ north of the STF.

The minimum $p\text{CH}_4^{\text{sw}}$ we observed was over the Kerguelen Plateau, where it corresponded to local variations of other parameters: a decrease in SST, a slightly positive AOU (5 $\mu\text{mol kg}^{-1}$), and elevated chl-*a* concentration (Fig. 3.10). Gille et al. (2014) reported that cold SSTs in the Kerguelen region correlate with high wind speeds and that wind-mixing of the upper ocean there resulted in entrainment of cold water into the mixed layer and euphotic zone. Mashayek et al. (2017) also identified spatiotemporal changes of mixing of waters in a shallow region of the Southern Ocean by using trifluoromethyl sulfur pentafluoride (CF_3SF_5) as a passive chemical tracer. In the Southern Ocean, strong vertical mixing is induced by near-surface westerly winds that cause convergent flow at intermediate depths; such mixing partially offsets upwelling (Marshall and Speer, 2012). The fluxes of greenhouse gases between air and sea and those in the water column are affected not only by physical processes, but also by the availability of organic matter and oxygen (Codispoti et al., 2001). CH_4 is formed mainly by methanogens (Wuebbles and Hayboe, 2002) during anaerobic degradation of organic matter or by the transformation of methyl compounds by methylotrophs (Damm et al., 2008), whereas CH_4 consumption occurs as a result of aerobic methanotrophy (Hanson and Hanson, 1996). In this region, vertical mixing has been reported to support the development of chlorophyll-*a* blooms (Gille et al., 2014) and high AOU values. AOU is respiration in the ocean interior (without contact with the atmosphere), where lower concentrations of CH_4 would be expected because of its removal by oxidation. Positive AOU values in surface seawater suggest that air-sea exchange of O_2 and O_2 production due to photosynthesis by phytoplankton were too slow to establish equilibrium after or during vertical mixing. Because maximum CH_4 concentrations at high latitudes south of the PF are in surface waters and decrease with depth (therefore there

are no subsurface CH₄ maxima) (Yoshida et al., 2011), I would expect a negative relationship between pCH₄^{sw} and AOU.

3.4.5 Zonal distribution of CH₄ in surface seawater at high latitudes

The pCH₄^{air} at latitudes south of 60°S (Fig. 3.11) decreased westward, which was a response to seasonal variations in atmospheric CH₄ and decreases in barometric pressure. As noted above (see discussion of Fig. 3.6), pCH₄^{sw} east of 140°E decreased considerably, along with decreases in SST and chlorophyll-*a* and increases in AOU (Fig. 3.6). Holm-Hansen et al. (2005) reported that in a relatively shallow area of the Southern Ocean bounded by 145° to 170°E and 62° to 66°S, where the bottom topography is complex, and there are numerous islands, banks, and seamounts, upwelling of deep water creates regions of high-chlorophyll pelagic seawater. In very early December, which is before the onset of strong biological activity, observed positive AOU (41 μmol kg⁻¹) and low chlorophyll-*a* concentrations in this region.

Lowering of pCH₄^{sw} is to be expected with increasing AOU, as was evident over the Kerguelen Plateau (Fig. 3.10), which is consistent with higher AOU than surrounding (Aoki et al., 2007). Aoki et al. (2007) reported that cyclonic eddies of 100–150 km diameter appear every year in a particular area at around 140°E, and that the properties of the water in the eddies are similar to those of water over the continental slope. Large-scale spatial variability of pCH₄^{sw} might be caused by physical processes such as upwelling and vertical mixing and subsequent downstream biological production of CH₄.

3.4.6 CH₄ distribution in Coastal area

West of 140°E, surface seawater emitted CH₄ to the atmosphere, except at high latitudes along 38°E and 53.5°E (transects 1 (TS1) and 2 (TS2), respectively) where SST was lower than 0 °C and AOU was positive (Fig. 3.12). A broad low-amplitude peak of pCH₄^{sw} at around 110°E coincided with a region of considerable variability of SSS. The large decrease in SSS there suggests input of meltwater in response to sea-ice retreat, which reduces the density of surface seawater. The resultant shallow stratification can trap organisms in the sunlit (productive) layer. Although phytoplankton do not generate CH₄ directly, the high pCH₄^{sw} there is probably a result of Antarctic krill, zooplankton, or both feeding on phytoplankton and the subsequent microbial methanogenesis (Yoshida et al., 2011).

South of the ASF along 38°E and 53.5°E (at 67.5°S and 64.5°S along coastal TS1 and TS2, respectively), pCH₄^{sw} decreased toward the continent (Fig. 3.12; Fig. 3.13) within the range from 1.65 to 1.69 μatm. The ASF is a strong horizontal subsurface temperature front

defined by the 0 °C isotherm (Ainley and Jacobs, 1981) and clearly separates the cold and shelf waters from the modified CDW, which can extend all the way to onto the Antarctic Shelf (Williams et al., 2010). Along 53.5°E (TS2, Fig. 3.13), $p\text{CH}_4^{\text{sw}}$ also decreased southward, accompanied by an increase in AOU, which is the same pattern as that we observed east of 140°E and over the Kerguelen Plateau. The coastward decrease of $p\text{CH}_4^{\text{sw}}$ is the opposite trend to that observed in tropical and extratropical regions. In the Antarctica, the physical features are the same, either mixing or upwelling; but the upwelled water is distinct. Upwelling of old CH_4 -poor water characterized by former biological oxidation (Rehder et al., 1999) and/or by contact with a former atmosphere with low methane content off Antarctica (Heeschen et al., 2004), versus upwelled water in the tropics and subtropics fed by depleted oxygen subsurface water and potentially contact with organic rich sediments (Reeburgh, 2007).

3.5 Estimation of sea-air CH_4 fluxes in the entire of the Southern Ocean

The $p\text{CH}_4^{\text{sw}}$ and $p\text{CH}_4^{\text{air}}$ data we acquired along the cruise track allowed us to use Eq. (3) to evaluate the sea–air flux of CH_4 over a wide area of the Southern Ocean (Fig. 3.14). Calculated flux of CH_4 we determined ranged from -3.9 to $8.5 \mu\text{mol m}^{-2} \text{ day}^{-1}$ (average $0.19 \mu\text{mol m}^{-2} \text{ day}^{-1}$) and was greatest and most variable at lower latitudes, especially off New Zealand. In the pelagic zone between the SAF and SBDY, it ranged from -0.43 to $0.50 \mu\text{mol m}^{-2} \text{ d}^{-1}$ and there was CH_4 uptake by surface seawater in the area of the Kerguelen Plateau. I observed intermittent strong $p\text{CH}_4^{\text{sw}}$ peaks north of the SAF and STF (see section 3.4.4) that led to the higher sea-air flux of CH_4 there. It is important to examine the triggers of these strong $p\text{CH}_4^{\text{sw}}$ peaks at lower latitudes in our evaluation of oceanic CH_4 flux. In the area south of the SBDY and east of 140°E, our data (Fig. 3.14) show that surface seawater there acts as a net sink for atmospheric CH_4 .

The CH_4 flux data I have determined is limited to the course of cruise MR12-05. To gain a good understanding of the oceanic CH_4 cycle throughout the South Pacific and the Southern Ocean, I need a much greater geographic spread of data. This might be achieved by using variables such as SST and chlorophyll-*a* concentration, which are related to the distribution of SR (and hence to $p\text{CH}_4^{\text{sw}}$) and are measured remotely over wide areas. If such relationships exist between SR and other oceanographic processes, it would then be relatively simple to interpolate or extrapolate observed SR values across the Southern Ocean, which would allow a more precise evaluation of the sea–air flux of CH_4 . Biological activity (CH_4 production and oxidation), ocean dynamics (lateral flow, vertical mixing, and upwelling), CH_4 solubility, and the sea–air flux of CH_4 have a major effect on the spatiotemporal variation of SR. Based on the

assumption that these factors are directly or indirectly related to SR, we investigated possible relationships between SR at high latitudes (south of 50°S) and SST, chlorophyll-*a*, water density, and AOU.

First, we examined the SR–SST relationship south of 50°S (taken to be south of the PF). SR increased dramatically with only slight increases in SST (at around 0 °C) (Fig. 3.15). Although no clear relationship between *SR* and chlorophyll-*a* was found, such a relationship may exist within areas much smaller than our study area (Damm et al., 2008; Yoshida et al., 2011). Yoshida et al. (2011) reported a strong relationship between C_w and seawater density in the South Pacific – Southern Ocean region, but this relationship does not hold in regions of low SST. While cruising at high latitudes from December 2012 to February 2013, I noted a possible correlation between pCH_4^{sw} and AOU, which I investigated further (Fig. 3.16). The resultant analysis of data collected along the cruise track in the open ocean south of 50°S shows that the following linear relationship between SR and AOU applies, at least in the Southern Ocean south of 50°S:

$$SR = -0.294 \times AOU + 100.093 \quad (r^2 = 0.89, n = 1549) \quad (3.1).$$

I got a good correlation between AOU and SR (Eq. (3.1)). In the subsurface water which is brought upward by vertical mixing, is characterized by positive AOU and low pCH_4^{sw} (Heeschen et al., 2004). At the sea-air interface, the transfer velocity for gas exchange of oxygen and CH_4 is quite similar, it means that undersaturation due to on the same timescale. The correlation here is also a strong indication for lacking of importance of biological production.

To obtain a map of the distribution of the sea–air flux of CH_4 in the Southern Ocean south of 50°S, we first determined the distribution of *SR* by applying Eq. (6) to the long-term monthly mean AOU data ($1^\circ \times 1^\circ$ grid) for December, January, and February from the World Ocean Atlas (2013). Because monthly mean AOU data for each year are not available, so we estimated a *SR* distribution based on climatological AOU data. Then, we calculated the monthly mean sea–air flux of CH_4 and total CH_4 emission for the three months (Table 3.2 and Fig. 3.17). The sea–air flux of CH_4 is controlled mainly by $C_w - C_a$ (see Eq. (3)) and wind speed, and total CH_4 emission in the Southern Ocean is controlled by the sea–air flux of CH_4 and the amount of sea-ice coverage. Although the AOU data for each grid cell suggest strong uptake of CH_4 in seawater in December, the area of surface seawater in contact with the atmosphere is small at higher latitudes in December due to sea-ice coverage. Although the AOU data suggest that CH_4 uptake in the Weddell Sea and Ross Sea would be considerable, these areas are largely covered

by sea ice in December. From December to January, total CH₄ emission increased by 0.014 Tg yr⁻¹ owing to an increase in sea–air flux of CH₄ and the retreat of sea ice. Following the retreat of the sea ice, CH₄ flux tended to be higher slightly south of the sea ice and the coastal region showed a large negative sea–air flux of CH₄ (CH₄ uptake). From January to February total net CH₄ evasion decreased by 44% (0.02 Tg yr⁻¹) while both the area of surface in contact with the ambient air and the wind speed increased. In February, CH₄ flux was generally small in the area south of the edge of the sea ice. Surface seawater between 50° and 60°S, which is a region of high wind speed, was a source of atmospheric CH₄ from December to February. The CH₄ flux obtained from gridded AOU data along the cruise track (south of 50°S) was in the range from -2.4 to 2.1 μmol m⁻² d⁻¹ (mean 0.08 μmol m⁻² d⁻¹), which is lower than that obtained from our underway measurements (0.194 μmol m⁻² d⁻¹); we attribute the difference to the effect of strong winds during the cruise, especially near 50°S. In coastal regions, the higher CH₄ fluxes obtained from gridded AOU data might also reflect local variations of AOU and coastal dynamic mechanisms, which may have contributed to these higher fluxes. Thus, there are limitations to the use of Eq. (3.1) to extrapolate CH₄ fluxes across the entire Southern Ocean.

South of 50°S, the average CH₄ flux in austral summer is 0.024 Tg yr⁻¹, which accounts for a minor contribution of oceanic release CH₄ to the atmosphere (4–15 Tg yr⁻¹, IPCC, 2007). Considering that the area the Southern Ocean south of 50°S represents 13% of the global oceanic area, this efflux seems very small, but we attribute it to the high invasion of CH₄ at high latitudes. Although we have treated sea ice as an impermeable barrier to sea–air exchange of CH₄ in this study, recent studies suggest that gas exchange through sea ice does occur (Nomura et al., 2010, 2013, 2014). The estimates of CH₄ flux derived from the AOU–SR relationship is the main highlight of this study: this relationship can be extrapolated to apply at least to the entire Southern Ocean. To improve predictions of future changes of the global sea–air flux of CH₄, more observation data are required in areas where these fluxes are most variable, in particular, in the Southern Ocean in areas of seasonal sea ice coverage, in tropical and extratropical coastal regions, and in marginal seas.

3.6 Summary and Conclusion

A CRDS system coupled with a shower-head type equilibrator is used to obtain quasi-continuous underway measurements of the mixing ratio of oceanic CH₄ in seawater during legs 2 and 3 of cruise MR12-05 of R/V *Mirai* in the Southern Ocean from late November 2012 to mid-February 2013. Our pCH₄^{sw} data show that surface seawater was supersaturated with CH₄

in the tropics and extratropics, but both super- and undersaturated at high latitudes. Undersaturation occurred in areas where SST was low ($<0\text{ }^{\circ}\text{C}$), AOU was positive (up to $41\text{ }\mu\text{mol kg}^{-1}$), and chl-*a* concentrations were low ($<0.3\text{ }\mu\text{g L}^{-1}$), indicating that these areas were affected by upwelling and vertical mixing and limited biological production of CH_4 . The SR in our entire dataset ranged from 85% to 185% (average $104 \pm 4\%$). High SR in the north is biased due to coastal vicinity.

To attempt a regional evaluation of the sea–air flux of CH_4 over a wide area of Southern Ocean, beyond the limits imposed by using data only from the course of R/V *Mirai*, I examined the relationships of SR (hence pCH_4^{sw}) with several other oceanographic variables that are being measured remotely over wide areas. For the area south of the PF (50°S), we identified a strong linear relationship between SR and AOU that indicates that vertical mixing and upwelling are important factors that influence the spatial distribution of oceanic CH_4 . On the basis of this linear relationship, the regional climatological distribution of the sea–air flux of CH_4 is calculated in the Southern Ocean south of 50°S for the period from December to February. I estimated emissions of CH_4 from the Southern Ocean to be 0.03 Tg yr^{-1} in December, 0.04 Tg yr^{-1} in January, and 0.02 Tg yr^{-1} in February. Total CH_4 emissions increased from December to January in response to the combined effect of an increase in sea–air flux of CH_4 and retreat of seasonal sea ice. From January to February, total CH_4 emissions decreased because the sea–air flux of CH_4 decreased as a result of declines in both wind speed and the area of sea surface in contact with ambient air (increase in sea ice). Although the effect of CH_4 emissions from the Southern Ocean has limited global impact, our results clearly show climatological variations of the sea–air flux of CH_4 in the Southern Ocean from December to February.

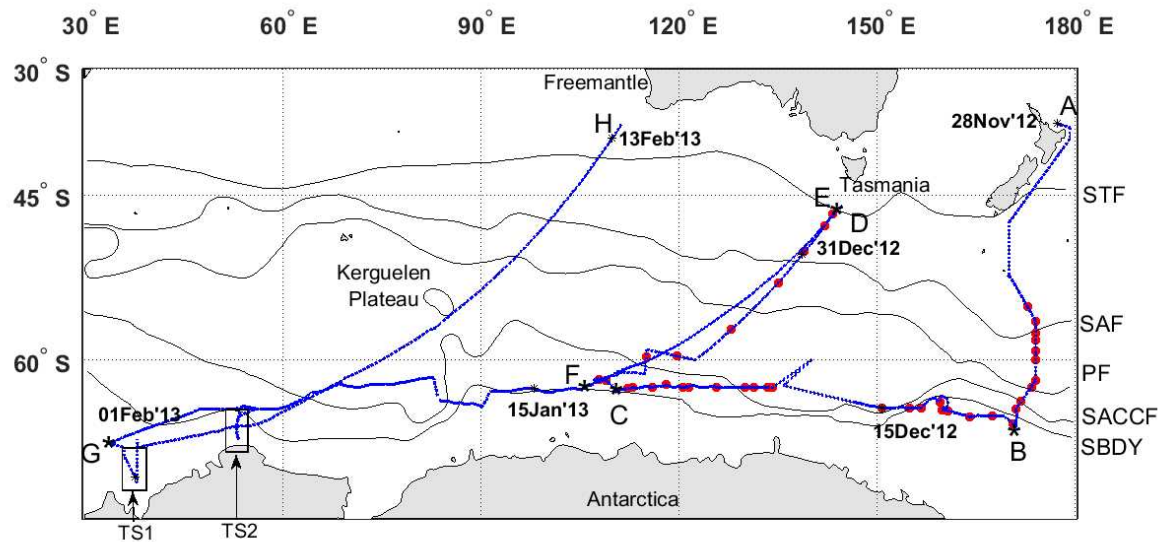


Fig. 3.1 Dash-dotted blue line shows the cruise track of R/V *Mirai* where underway measurements of oceanic CO₂ and CH₄ were made simultaneously. Red circles show hydrographic stations for measurements of CH₄ in discrete water samples. Black solid line shows the oceanic fronts from Orsi et al., (1995), and TS1 and TS2 coastal transects at 38°E and 53.5°E, respectively.

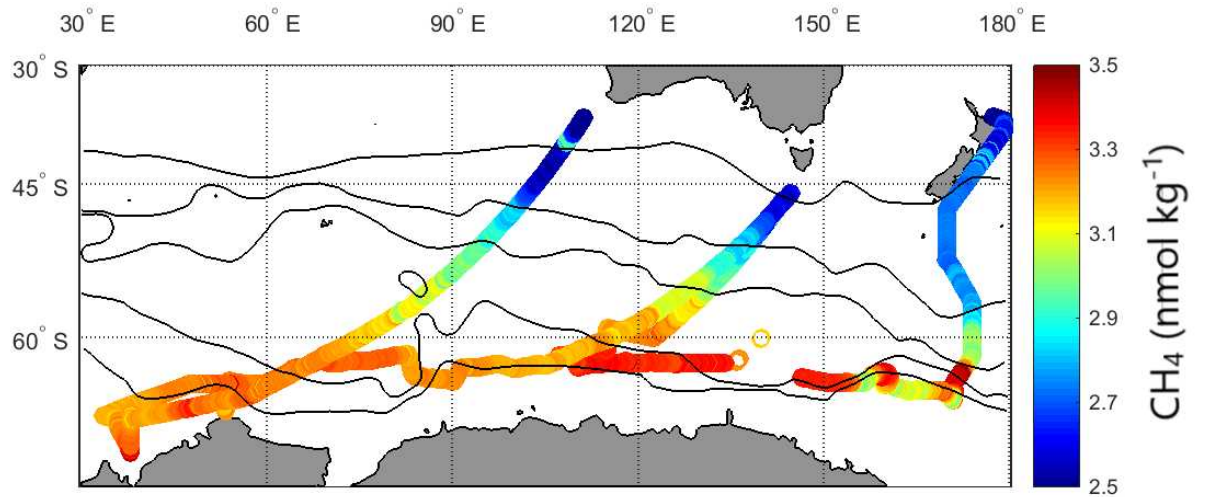


Fig. 3.2 Spatial distribution of methane concentration in surface seawater determined by using CRDS system

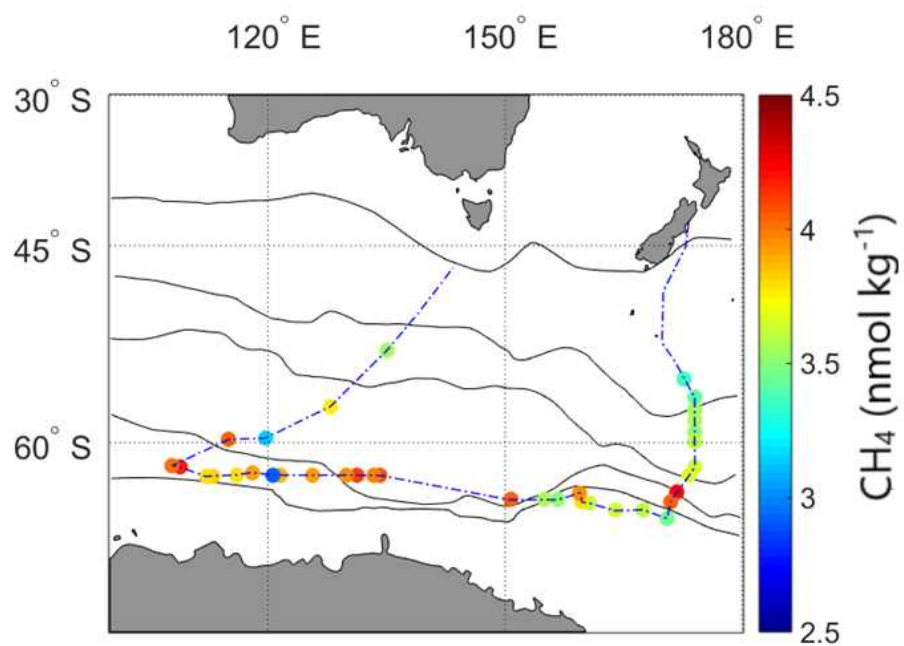


Fig. 3.3 Spatial distribution of methane concentration in surface seawater determined by using the GC-IRMS system

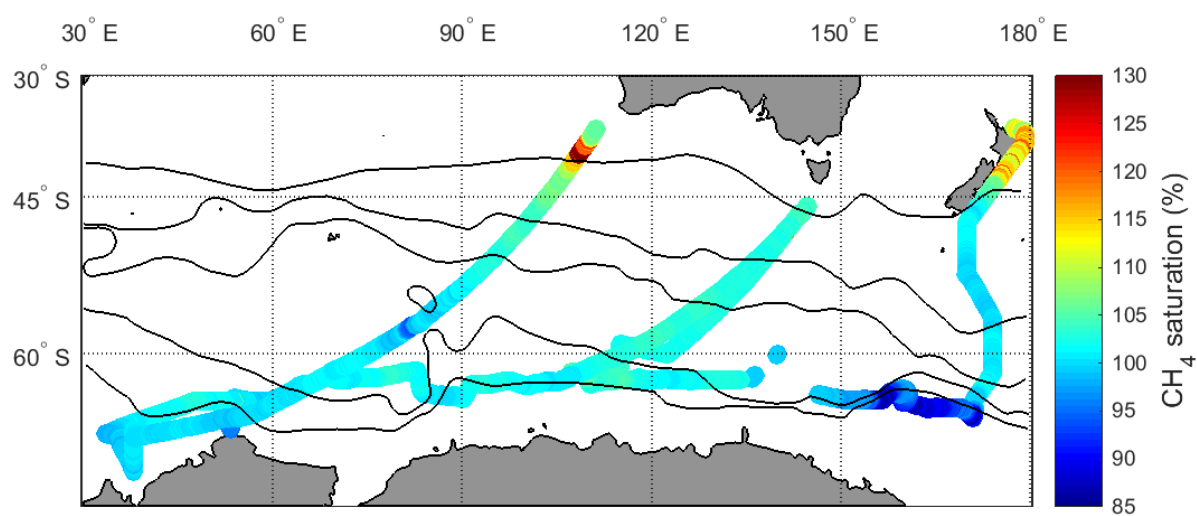


Fig. 3.4 Spatial distribution of methane saturation in surface seawater determined by using CRDS system

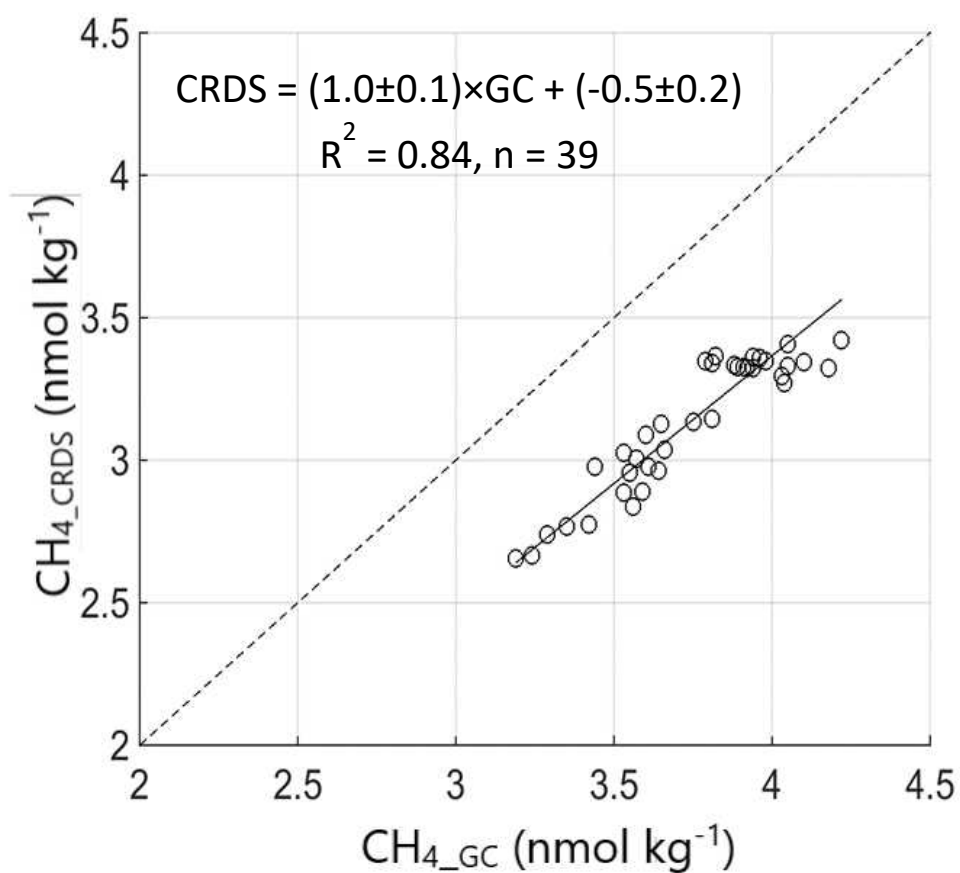


Fig. 3.5 Plot of CH_4 concentration between CRDS system versus GC system during the cruise track of R/V *Mirai* (MR15-03).

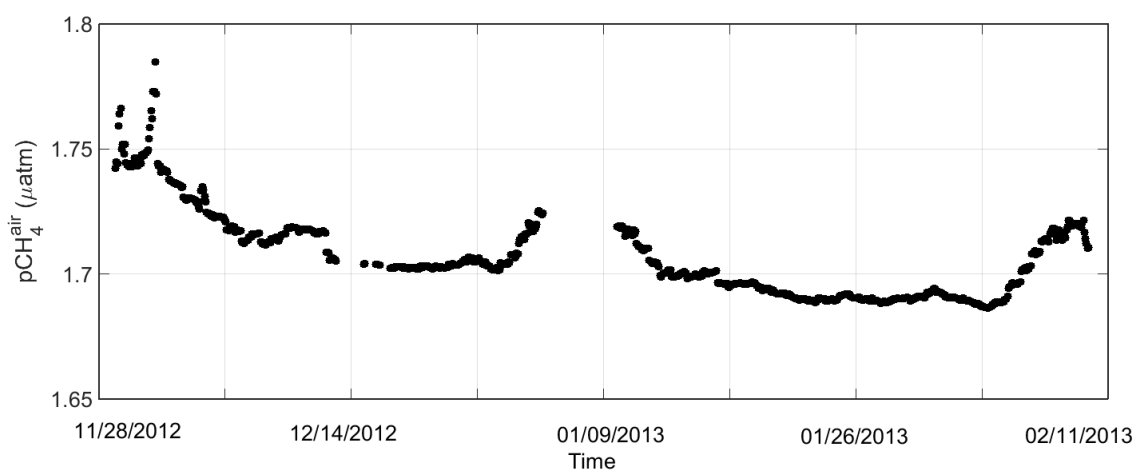


Fig. 3.6 Atmospheric CH_4 mixing ratios measured by CRDS analyzer during the cruise track of R/V *Mirai* (MR15-03).

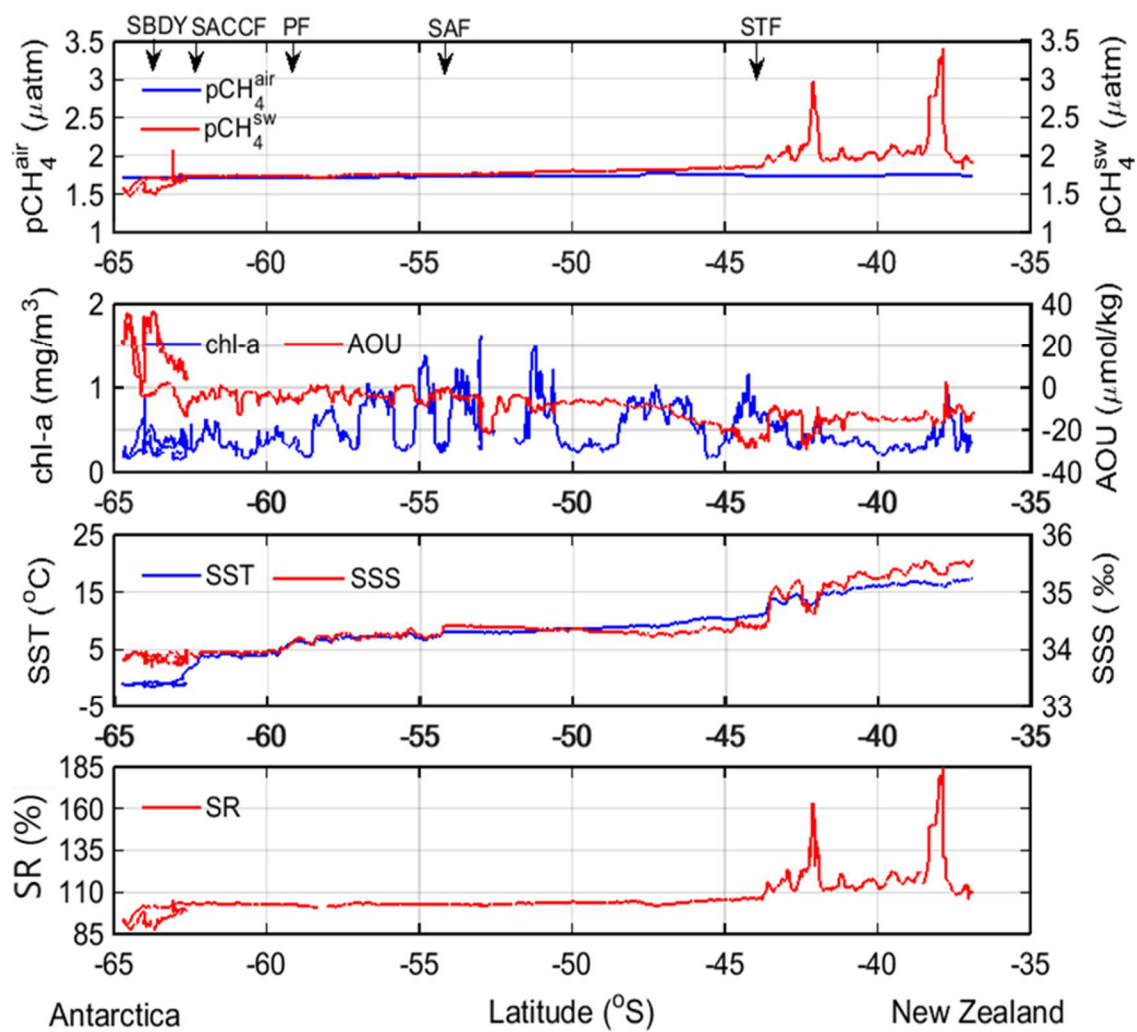


Fig. 3.7 Meridional distribution of pCH_4^{air} , pCH_4^{sw} , chl-*a*, AOU, SST, SSS, *SR* along transect from offshore New Zealand to Antarctica (A-B).

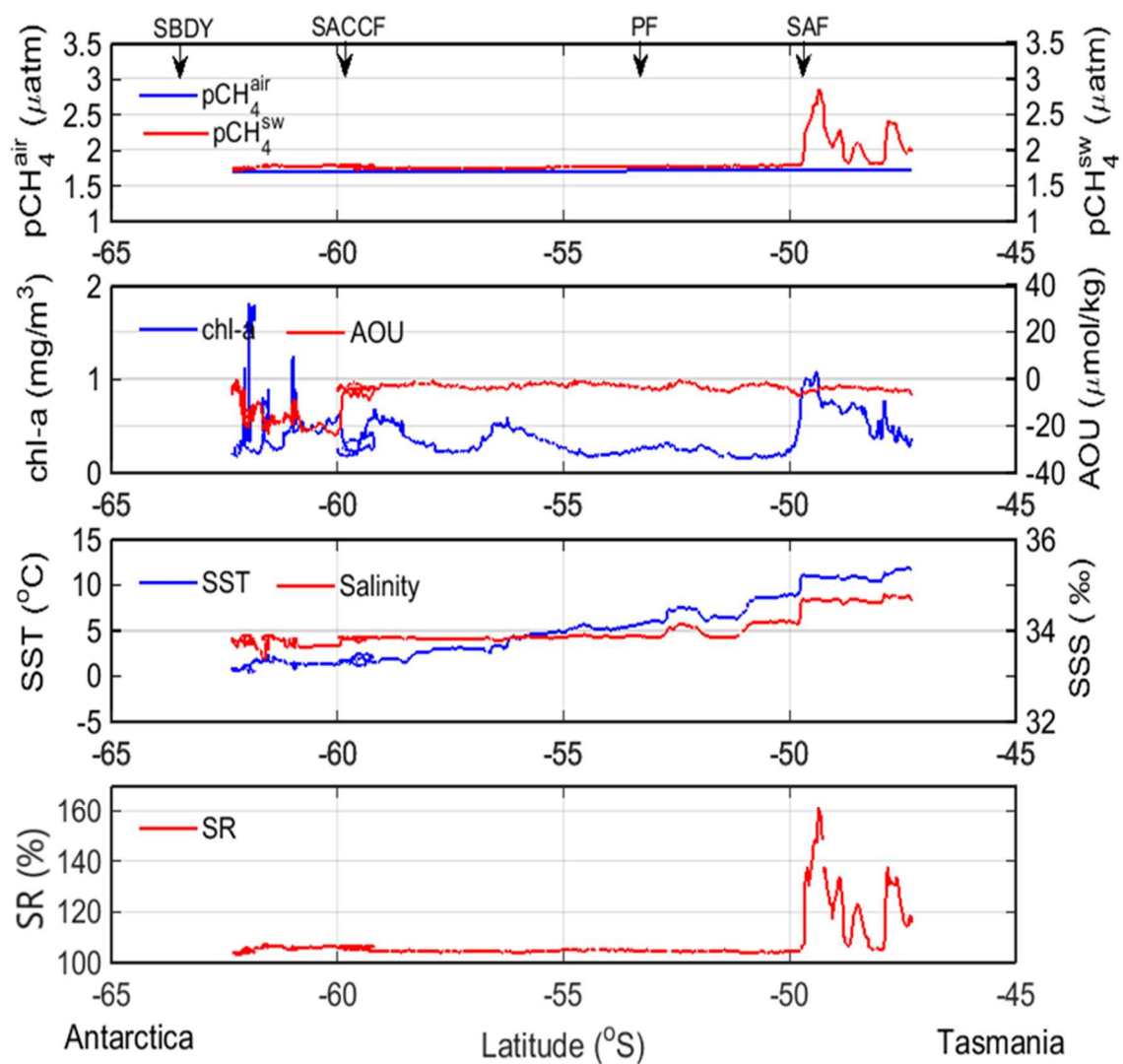


Fig. 3.8 Meridional distribution of pCH_4^{air} , pCH_4^{sw} , $chl-a$, AOU, SST, SSS, SR from SBDY to SAF during northbound sailing to Tasmania (Leg 2) (C-D).

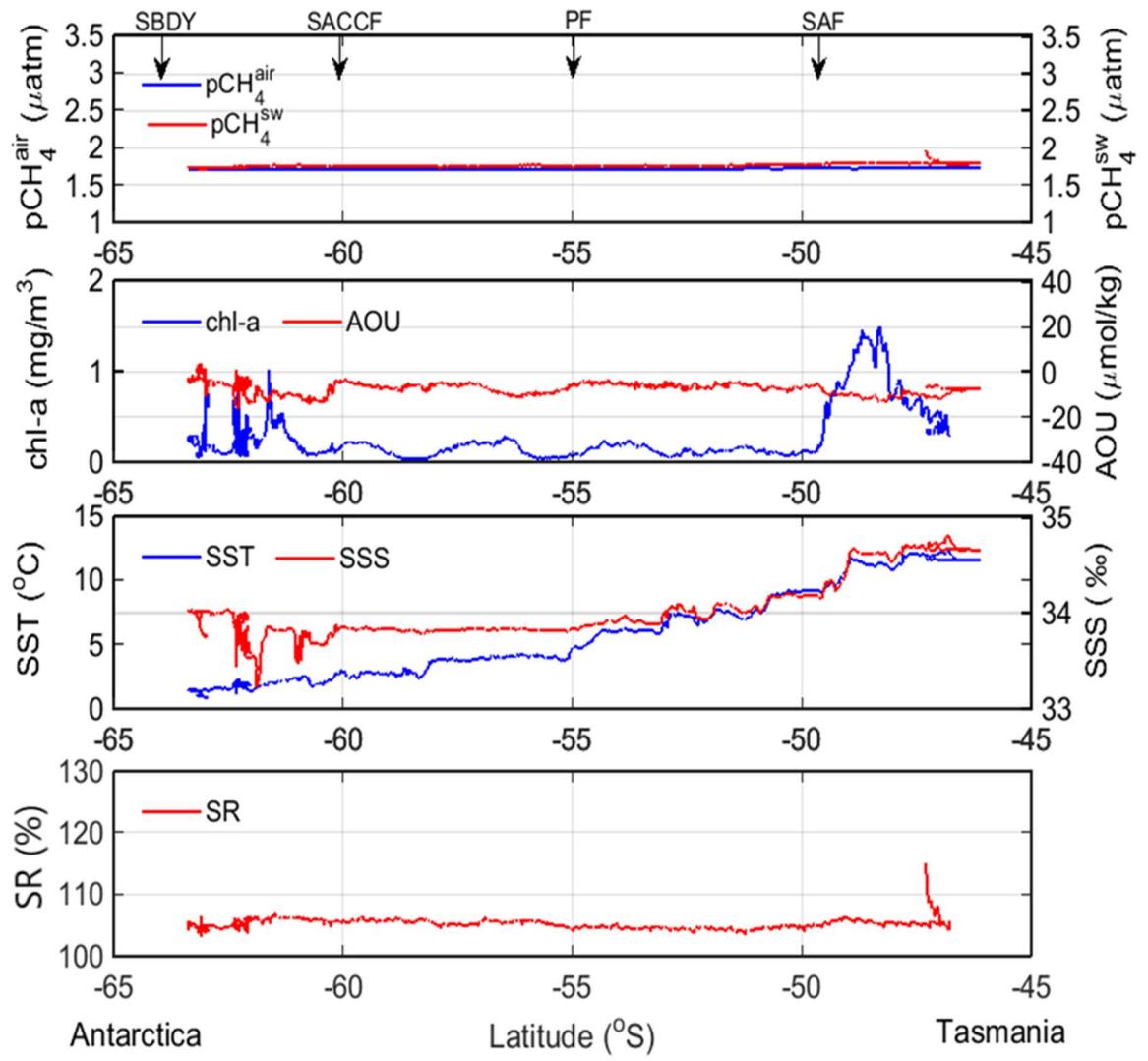


Fig. 3.9 Meridional distribution of pCH_4^{air} , pCH_4^{sw} , chl-*a*, AOU, SST, SSS, SR from STF to SBDY during southbound sailing (Leg 3) (E-F).

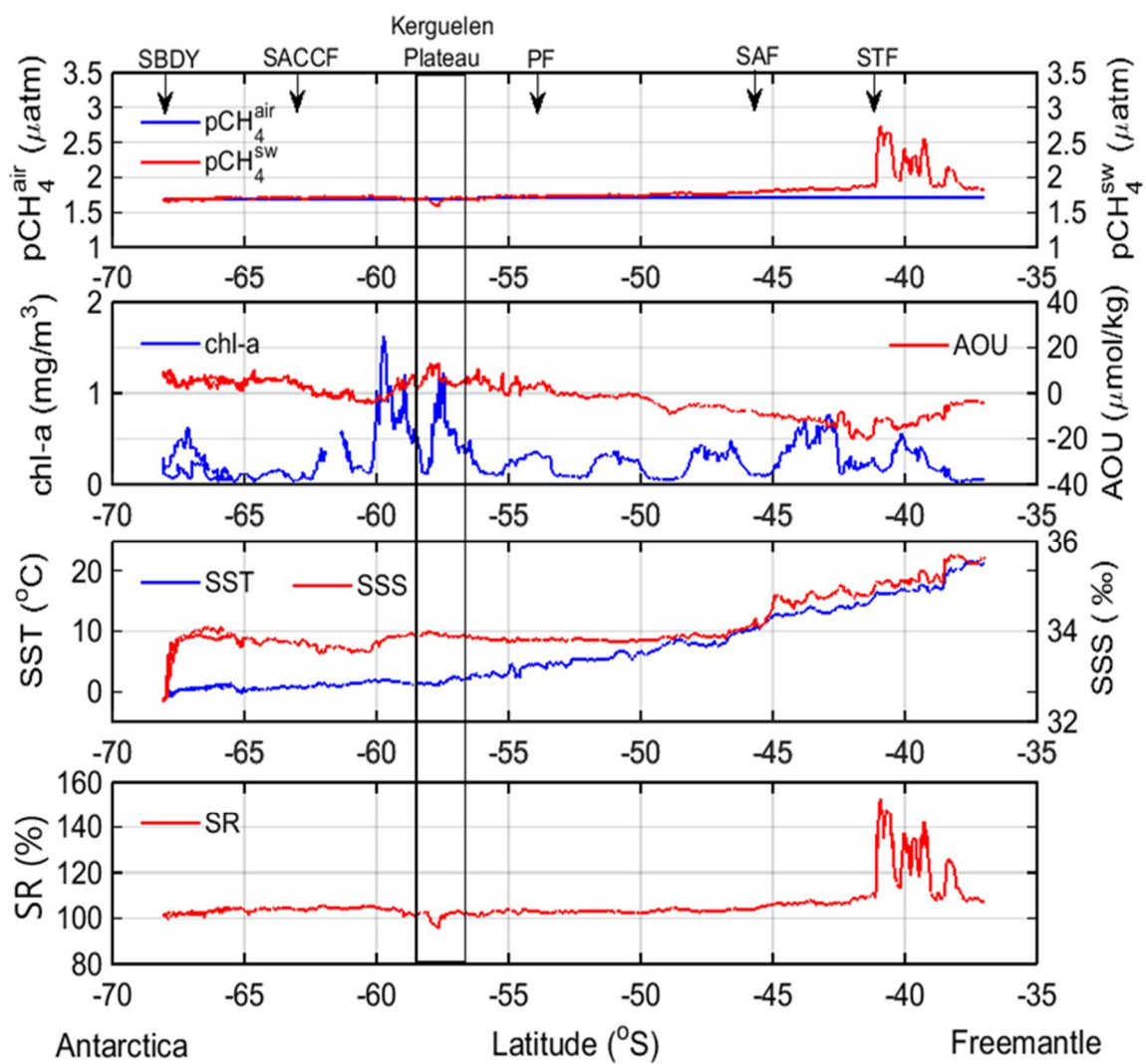


Fig. 3.10 Same as in Fig. 3.9 except from SBDY to STF during northbound sailing to Fremantle (Leg 3) (G-H).

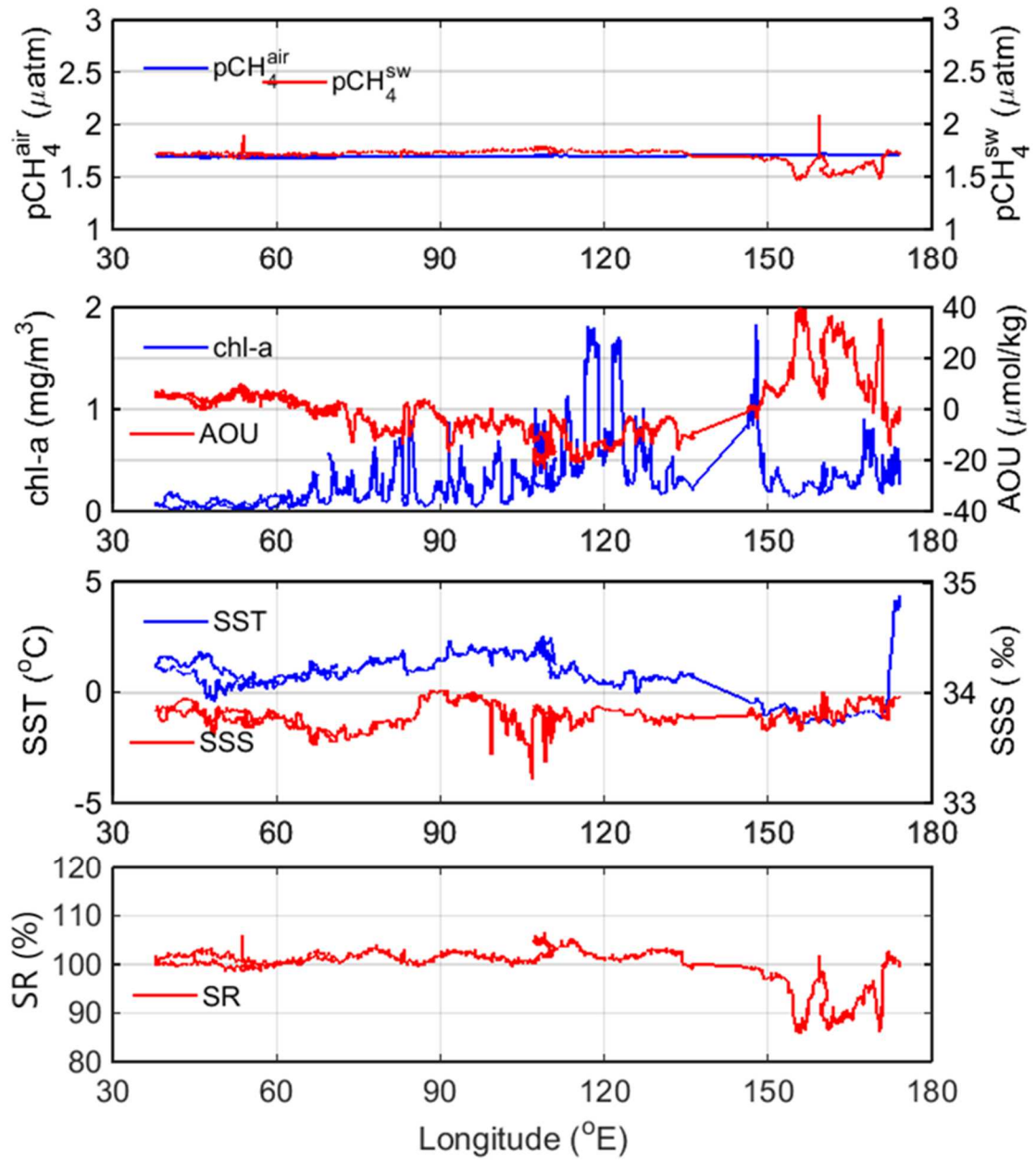


Fig. 3.11 Zonal distribution of pCH_4^{air} , pCH_4^{sw} , chl-*a*, AOU, SST, SSS, SR south of 60°S (B-C-G).

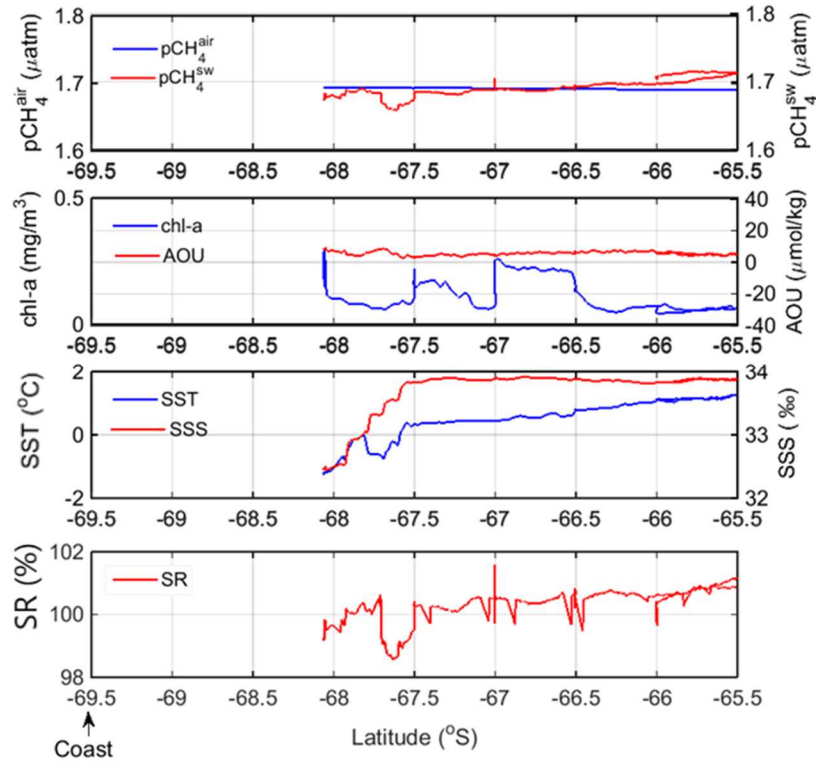


Fig. 3.12 Meridional distribution of pCH_4^{air} , pCH_4^{sw} , chl-*a*, AOU, SST, SSS, *SR* at transect 1 (38°E). An arrow at the bottom indicates the position of the coast.

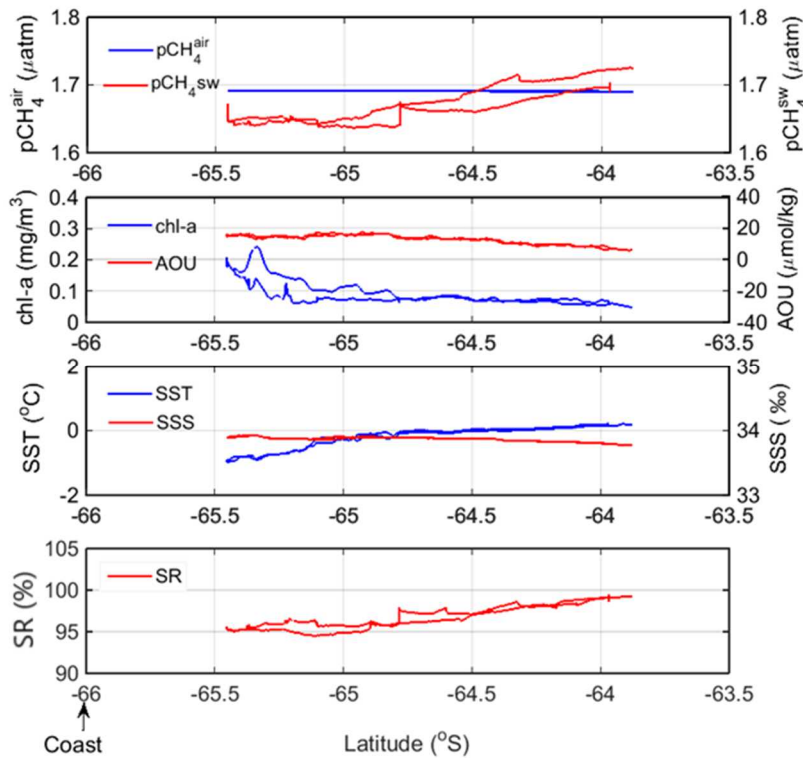


Fig. 3.13 Meridional distribution of pCH_4^{air} , pCH_4^{sw} , chl-*a*, AOU, SST, SSS, *SR* at transect 2 (53.5°E). An arrow at the bottom indicates the position of the coast.

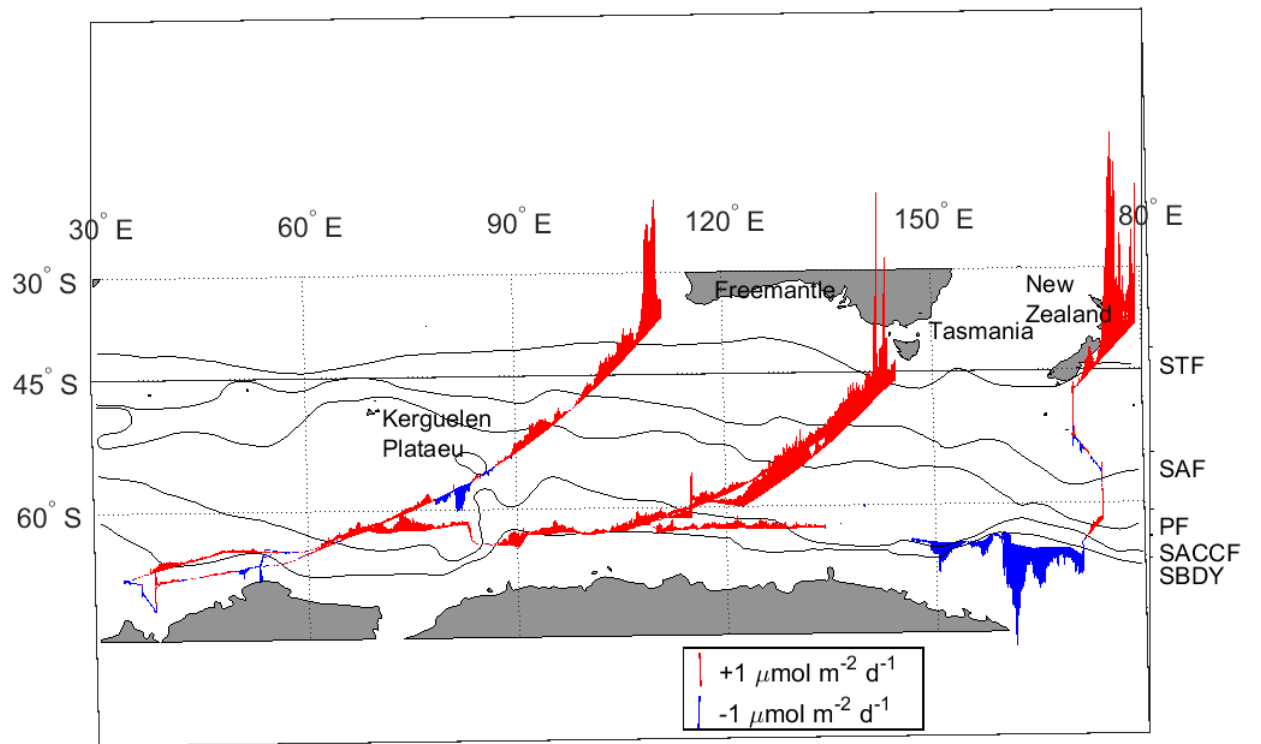


Fig. 3.14 The CH₄ flux along the cruise track (red color indicates CH₄ release, and blue color indicates CH₄ uptake in seawater)

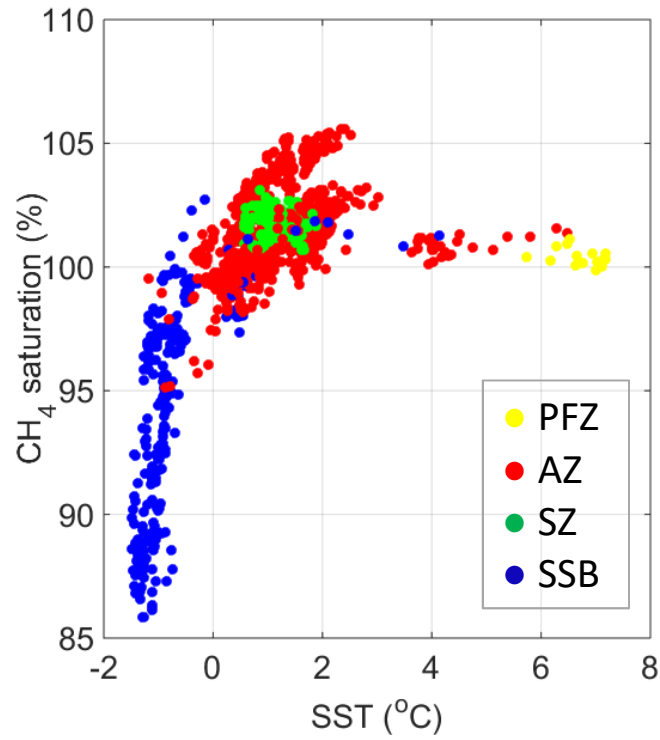


Fig. 3.15 The relationship between CH₄ saturation (SR) and SST south of 50°S in the Southern Ocean ($n = 1549$). Colorful dots indicate for data in water masses in the Southern Ocean.

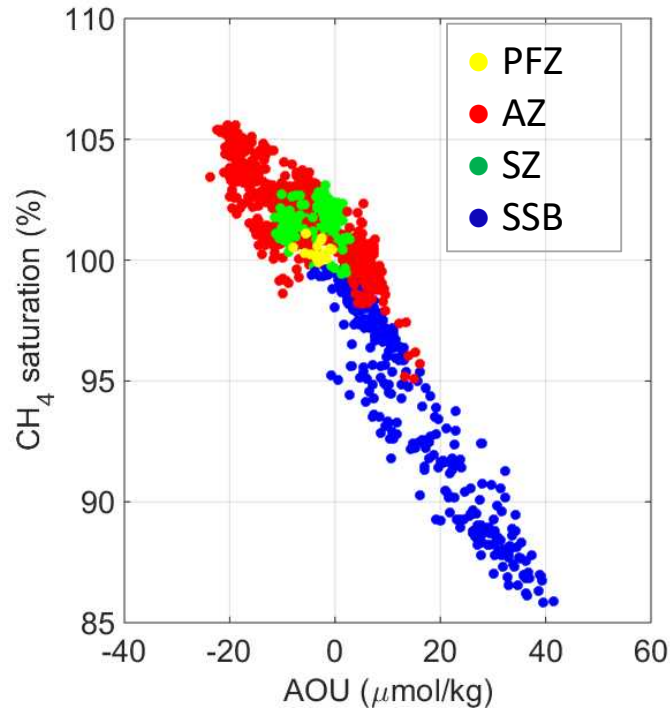


Fig. 3.16 The relationship between SR and AOU south of 50°S in the Southern Ocean. Solid line shows the linear relationship between SR and AOU: $SR = -0.294 \times AOU + 100.093$ ($r^2 = 0.89$, $n = 1549$). Colorful dots indicate for data in water masses in the Southern Ocean.

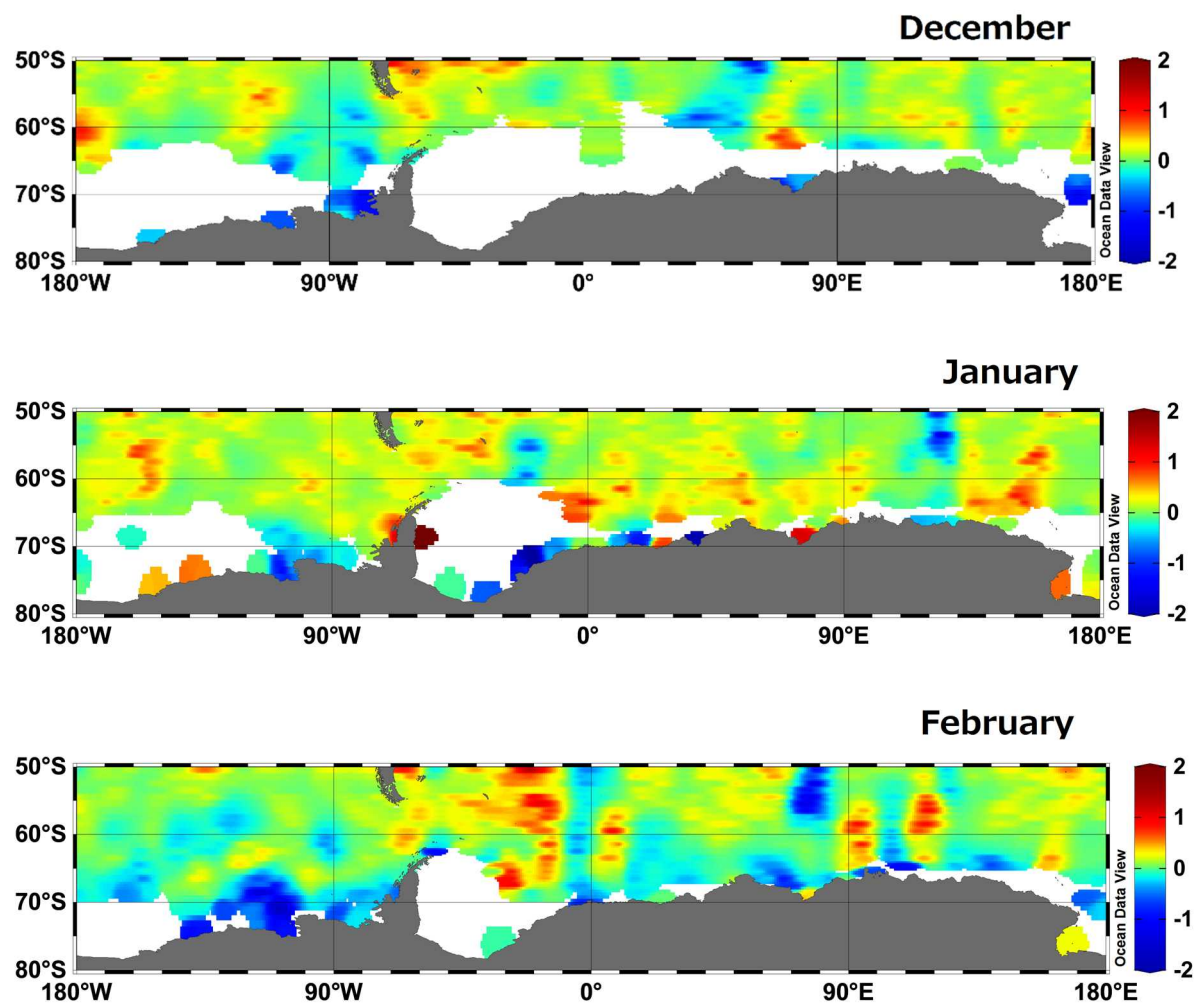


Fig. 3.17 Monthly mean sea-air CH₄ flux in the Southern Ocean from December to February (from upper panel to bottom panel). White areas are sea ice coverage.

Table 3.1 Comparison of CH₄ saturation (SR) to the previous studies

High latitudes (> 60°S)	Mid-latitudes (30°S – 60°S)	References
65°S, 150 – 170°E Under: 85-100%	Super- 100-142%	This study
60°S – 66°S, 140°E Jan: super- 100-210% Feb: under- 87-100%		Yoshida et al., 2011
60°N – 70°S, 140°W Under-: 97-100% super-: 100-102%	97–109%	Bates et al., 1996
5°N – 42°S, 75-80°W Under- 90-100%	Super- 100 – 110%	Kelly et al., 2002

Table 3.2 Monthly mean AOU, wind speed, the sea-air CH₄ flux in the Southern Ocean (south of 50°S)

	Dec	Jan	Feb	Average
Mean AOU ($\mu\text{mol kg}^{-1}$)	-2.5	-3.2	-0.1	-1.9
CH ₄ saturation ratios (SR, %)	100.84	101.04	100.13	100.67
Mean wind speed (U, m s^{-1}) (avg 30 yrs)	6.9	6.78	7.44	7.03
CH ₄ flux ($\mu\text{mol m}^{-2} \text{ day}^{-1}$)	0.09	0.14	0.06	0.08
CH ₄ emission (Tg yr^{-1})	0.027	0.04	0.019	0.024

CHAPTER 4:

INFLUENCE OF A WARM-CORE EDDY ON CH₄ DISTRIBUTION IN THE ARCTIC OCEAN

4.1 Dissolved CH₄ distribution in the Arctic Ocean

Large sources of CH₄ are stored in the Arctic (Shakhova et al., 2014). Large sources of CH₄ in the Arctic shelves contribute to the distribution of CH₄ in the water column (Damm et al., 2015b; Matveeva et al., 2015) and atmosphere (Myhre et al., 2016). In the Arctic Ocean, the principal sources of CH₄ are anaerobic methanogenesis in decaying organic matter in thawing permafrost, degradation of gas hydrates, and thermogenic CH₄ from cold seeps from hydrocarbon deposits (Reeburgh, 2007).

Release of unstable CH₄ hydrates is thought to be a large source of CH₄ emissions. This release occurs in shallow (Phrampus et al., 2014) and deep water (Coffin et al., 2013; Ruppel, 2015) in the Beaufort Sea, Eastern Siberian Arctic Shelf and Eurasian shelves (Shakhova et al., 2010a; 2014), which is generally considered the most sensitive to climate change effects (Kvenvolden, 1993; Ruppel, 2011). Previous studies show elevated CH₄ fluxes into seawater can potentially lead to increase CH₄ emission from the ocean into the atmosphere. In addition, extensive gas bubbling from the sea floor has been observed in the shallow water along the West Svalbard continental margin (Westbrook et al., 2009; Graves et al., 2015; Myher et al., 2016). Expelled CH₄ from sea floor is released into the water column and thus does not reach the atmosphere (James et al., 2016). Conversely, the recent publications have shown that high level of atmospheric CH₄ has considered as a release of oceanic CH₄ in the shallow areas (Myhre et al., 2016).

Some observations of CH₄ have been investigated in the Arctic Ocean (Shakhova et al., 2010; Damm et al., 2010, 2015; Lorenson et al., 2016; Fenwick et al., 2017; Li et al., 2017). Results of CH₄ observations have been reported as an evidence for high CH₄ emission in many areas of the Arctic shelves. High CH₄ in surface water (mean value of 18 nmol L⁻¹) has been attributed to the thawing of subsea permafrost on the Eurasian shelf (Shakhova et al., 2010a, Kosmach et al., 2015) while less CH₄ (average of 8.1 nmol L⁻¹) has been previously found in the permafrost-free Chukchi Sea (Kosmach et al., 2015). Significant CH₄ is also observed in the Beaufort Sea (up to 13.8 nmol L⁻¹) (Lorenson et al., 2016), however, amount of CH₄ is smaller than that of in the East Siberian Arctic shelves (ESAS) with large amount of CH₄

escape from seabed into the water column (flux of $\sim 70\text{--}450 \text{ nmol m}^{-2} \text{ s}^{-1}$ or 17 Tg yr^{-1}) (Shakhova et al., 2014).

Amount of CH_4 stored within gas hydrates or from marine sediments releases into the water column and then reaches to the atmosphere is still unclear. Emitted CH_4 in the atmosphere could cause positive feedback and could accelerate climate warming (Ferré et al., 2012). Almost the previous studies have focused on biological processes that directly affect biological CH_4 production and consumption in the water column. Notwithstanding rare studies present combination of physical and biological factors affect the transportation of dissolved CH_4 or CH_4 distribution. In this research, I investigate the effects of eddies to CH_4 distribution in the water column.

4.2 Characteristics of water masses and eddies in the Arctic Ocean

Research area in Chapter 4 covers the Bering and Chukchi Seas and a part of Canada Basin. Thus, a brief introduction for its characteristics of water masses and eddies is described below.

4.2.1 Characteristics of water masses in the Arctic

Water masses in the Arctic Ocean have specific physical and biological features because of the interaction of freezing and melting processes in the water column, river run-off into the continental shelves and advection into the Arctic Ocean (Rudels et al., 1996). The Bering and Chukchi Seas are dominated by northward flowing Pacific water that comprises several distinct water masses such as Anadyr Water (AnW), Bering Sea Water (BSW), Pacific Winter Water (PWW), Pacific Summer Water (PSW) and Alaskan Coastal Water (ACW). The inflow water flows across the Chukchi Shelf including two or three distinct branches which governed by the bathymetry of the shoals and canyons (Paquette and Bourke, 1981; Weingartner et al., 1998; Woodgate et al., 2005). The western side of the Bering Strait follows Herald Canyon (e.g., AnW) composing of lowest temperature, upwelled, saline, nutrient-rich deep water. A central inflow follows the Central Channel with intermediate salinity is unknown as BSW. In the eastern side, the inflow is dominated in late-spring to early-fall by the warm, relatively fresh ($S < 31.8$), nutrient-poor ($< 1.5 \mu\text{mol L}^{-1} \text{ NO}_3^-$) ACW (Paquette and Bourke, 1974) as a part of the PSW (Shimada et al., 2001). Pacific water flows into Arctic Ocean with composing of PWW and PSW but the difference between PWW and PSW results from the period of seasonal sea-ice growth. In the winter, convective overturning caused by brine rejection during sea-ice formation, this process helps to stir up the sediments and add regenerated sedimentary nutrients

to PWW (Pickart et al., 2016). PSW ($S = 31-32$, $T > -1^{\circ}\text{C}$, $D = 25-26\sigma_t$) is found between 40-100m while PWW ($S = 32-33.1$, $T < -1^{\circ}\text{C}$, $D = 26-26.5\sigma_t$) lies between $\sim 100-200$ m or deeper water characterizing by low temperature and high salinity feature (Woodgate et al., 2005). When Pacific water transit over the shallow Bering and Chukchi shelves into the Arctic Basin and are modified by biological and physical processes (Yamamoto-Kawai et al., 2006) from both of the surface water and above sediment layer (Cooper et al., 1997). During summer the fresher, warmer and northward flowing PSW gradually displaces the PWW even though the appearance of remanent PWW exists on the northern Chukchi shelf (Pickart et al., 2016).

The Canada Basin comprises deep water (~ 4000 m), which is bordered by shallow seas and continental shelf. Surface water of the Canada Basin is characterized by a Polar Mixed Layer (PML) of about 40 m depth, which is relatively fresh ($S \sim 26-31.2$) and is nutrient-depleted (Macdonald et al., 2004; Carmack et al., 2006). Dense PWW sinks to the deepest area of the canyon (below 35 m depth) and is overlain by PSW and ACW (5-35 m) as well as a surface melt-water layer (0-5m) (Gong and Pickart, 2015). When Pacific water enters the Canada Basin from the Chukchi Sea, it subducts below the PML into the Upper Halocline Layer (UHL). Below the UHL, the halocline layer occupies a depth range of $\sim 200-400$ m and derives from Atlantic origin water while Canada Basin Intermediate Water (CBIW) occupies at a depth range of 400-2000 m. Canada Basin Deep Water (CBDW) is found below 1500 m depth, which is relatively homogeneous and has been isolated from the atmosphere for about 450 years (Timmermans et al., 2003).

4.2.2 Introduction to eddy in the Arctic Ocean

Chukchi Sea, Beaufort Sea and Canada Basin are populated with a large number of sub-mesoscale eddies (Muenchow et al., 2000). Three types of eddies have been observed in the southern Canada Basin. Firstly, warm-core surface-intensified eddies (0-50 m) contains Alaskan Coastal water, which belongs to the season (late-summer/early-fall). Other two types of eddies which occur in subsurface water, are either a cold- or a warm-core. The former is spawned during spring and early summer when the current is advected dense and cold water. The latter are formed later in fall when the boundary current has been replaced with PSW, which is warmer in summertime Chukchi Sea shelf water.

The primary difference between both eddies is characteristics of itself water-core including physical, chemical and biological parameters. The most commonly observed type of eddy in the southern Canada Basin is the sub-surface cold-core anticyclone ($S = 33.1$; $T = -1.6^{\circ}\text{C}$) (Pickart and Stossmerster, 2008) with a diameter of 10-20 km (Manley and Hunkins,

1985) and thickness of 75 m (Mathis et al., 2007). In contrast, warm-core eddies are rarely detected by oceanographers with a small fraction of the total eddies (3%). This water-core entirely is warmer than ambient environment, coming subsurface water of PSW. The warm-core eddies also transport heat and nutrients from the Chukchi Sea shelf to the Canada Basin like cold-core eddies (Mathis et al., 2007; Nishino et al., 2011; Kawaguchi et al., 2012; Watanabe et al., 2014) but the warm-core water may control nutrient distribution and affect primary production in the euphotic zone (Nishino et al., 2011). Enhancement of the magnitude of the biological pump attributed to arising of eddies, driven transport of shelf-origin nutrients and biological materials towards the Canada Basin (Watanabe et al., 2014). Even though the frequency of the occurrence of warm-core eddies is rare (accounting for 3% of the total number), its role in transporting material is undeniable. Few studies of warm-core eddies has examined their influences on the nutrient distribution and biological activities but the effect on CH₄ distribution that has not been investigated yet until now. Therefore, this study focuses on the impact of warm-core eddies on CH₄ distribution in the Arctic Ocean.

4.3 Data and Methods

4.3.1 Field sampling

Over the period from August to October 2015, seawater samples were collected on R/V *Mirai* (MR15-03 Leg.1, Fig. 4.1a) from surface water to the bottom water. A warm-core eddy was observed in the shelf-slope of the Chukchi Sea during the cruise. Majority of sampling was conducted over Chukchi Sea shelf-slope and the southwestern Canada Basin (see Fig. 4.1b), to identify the vertical distribution of CH₄ concentration inside/outside of the eddy zone. Vertical seawater samples were taken using 12-L Niskin bottles mounted on a CTD rosette for analyses of dissolved CH₄. Approximately 170 seawater samples were collected at 19 stations for the CH₄ analysis. A Sea-Bird CTD system (SBE9plus, Sea-Bird Electronics Inc., WA, USA) equipped with ancillary sensors was used to measure salinity, temperature, and other hydrographic parameters. These CH₄ samples were measured at Nagoya University by GC, and description of GC system is also mentioned in Chapter 2. The cruise crossed some distinct water masses such as PSW, PWW, ACW, MW, and Atlantic Water (AW); and crossed the warm-core water.

Another measurement system is CRDS system coupling with an equilibrator which is described in detail in Chapter 2. It was deployed to obtain mixing ratios of CH₄ at that time during the cruise.

The atmospheric concentration of CH₄ was taken as 1850 ppb, based on the average mixing ratios in Barrow, Alaska in 2015 (Data provided by NOAA ESRL Global Monitoring Division, Boulder, Colorado, USA).

4.3.3 Chemical and Biological Oceanography parameters

4.3.2.1 Dissolved Oxygen (DO)

Seawater samples were collected with Niskin bottle attached to the CTD-system and transferred from bottles to a calibrated flask of 100 cm³ with overflowing of 3 times volume of the flask. Then two reagent solutions (Reagent I and II) of 0.5 cm³ were added immediately into the sample flask and the stopper was sealed carefully. The sample flask was shaken at least twice to mix the content and dissipate the precipitate. At least 2 hours after shaking, 1 cm³ sulfuric acid solution and a magnetic stirrer bar were added to the sample flask and stirred. These samples were measured by using a burette for sodium thiosulfate and potassium iodate APB-510/APB-620 (Kyoto Electronic Co. Ltd, Japan)

AOU is the difference between the measured dissolved oxygen concentration and its equilibrium saturation concentration in water. It is defined in section 2.6 Chapter 2.

4.3.2.2 Nutrients

Nutrients data included ammonium (NH₄⁺), nitrate (NO₃⁻), nitrite (NO₂⁻), phosphate (PO₄³⁻) and silicate (SiO₂) were analyzed by using a QuAatro 2-HR system. Fixed nitrogen (N^{**}) values were calculated as a measure of the relative excess or deficit of fixed nitrogen relative to phosphate (see section 2.6 Chapter 2; Codispoti et al., 2005a). These biological parameters (e.g., nutrients) allow us to constrain the biogeochemical processes that influence on trace gas cycling.

4.4 Existence of a warm-core eddy during the cruise

During the cruise, a warm-core eddy was observed in the Chukchi Sea shelf slope and near Barrow Canyon. The position of the warm-core eddy is pointed out in Fig. 4.2. In general, the warm-core eddy appeared near the surface or at a depth range was shallower than the depth of cold-core eddy (Pickart and Stossmerster, 2008). This eddy appeared near the surface at depths of 20–70 m (Fig. 4.2b), and the core of the eddy was characterized by potential temperatures of 4–6°C at a depth of 50 m. The characteristics of the water mass indicated that the warm water had been derived from Pacific Summer Water (PSW) (salinity = 31–32;

temperature $> -1^{\circ}\text{C}$) (D'Asaro, 1988). This warm-water core was distributed laterally along an isopycnal plane with a potential density (σ_{θ}) of $\sim 24\text{--}25$. The patch of warm water created a sharp thermohaline front against the surrounding water to the north (72°N or at 100 and 150 km) that resulted in a horizontal velocity maximum shear with a squared magnitude of $(1\text{--}2) \times 10^{-4} \text{ s}^{-2}$ (Kawaguchi et al., in preparation). The interior of the warm core was characterized by a low Brunt-Vaisala frequency (N) of about 3 cycles per hour (Fig. 4.2b) Kawaguchi et al. (in preparation) have pointed out that the magnitude of the maximum tangential velocity within the warm-core radius was about 0.55 cm s^{-1} . The colder and denser water below the warm water, which was characterized by a salinity (S) of $32\text{--}33$ and σ_{θ} of $25\text{--}26$, was determined to be Pacific Winter Water (PWW). The location of the warm water mass in a subsurface layer suggested that it was a warm-core eddy. The eddy was presumed to have formed over the Chukchi Sea shelf and to have been advected laterally into the cold water above the shelf slope. The eddy was thin, about $30\text{--}50 \text{ m}$ thick, and its diameter was $30\text{--}50 \text{ km}$. Eddies can transport warm water (PSW, ACW) to a depth of 70 m or more but still lie above the upper boundary of PWW.

4.5 Results and discussion

4.5.1 Spatial distribution of dissolved CH₄ distribution in high latitudes

4.5.1.1 Spatial distribution of dissolved CH₄ from CRDS measurement

Continuous dissolved CH₄ concentration range was 2.6-17.6 nmol kg⁻¹ in the surface seawater with an averaged value of 4.6 nmol kg⁻¹ (Fig. 4.3

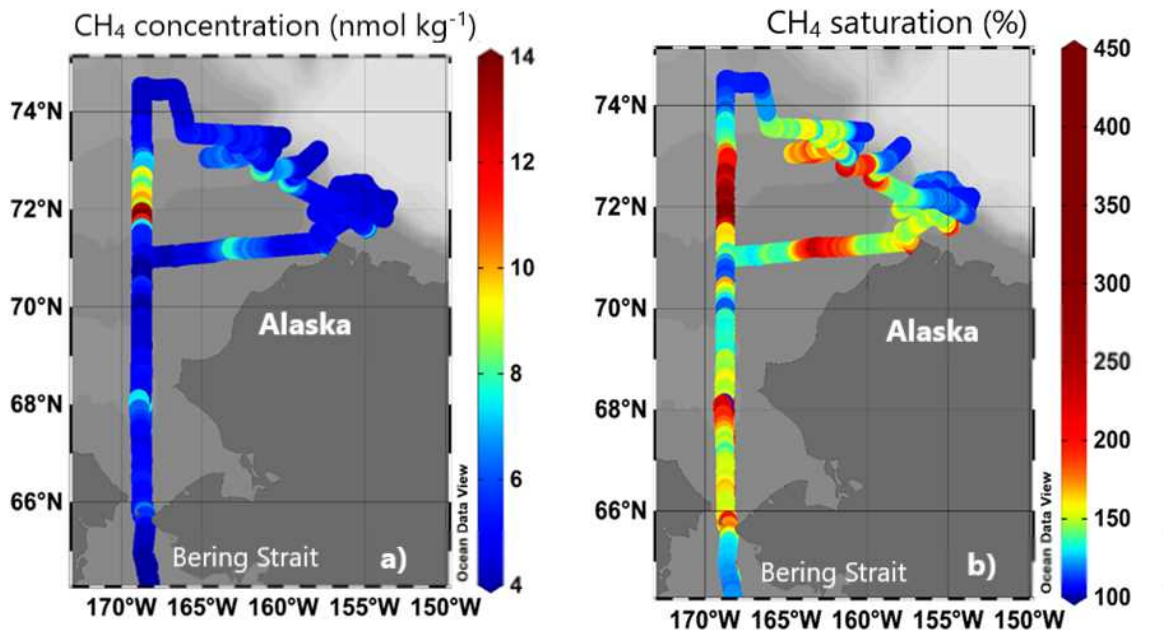


Fig. 4.3) over the whole region. Dissolved CH₄ concentration in the surface was supersaturated (103-514%), averaging 137.9% in the surface water. CH₄ concentration increased considerably from south Bering Strait (ca. 3 nmol kg⁻¹) to the northern Chukchi Sea (17.6 nmol kg⁻¹). High CH₄ concentration was double near the Point Hope with a value of 7.7 nmol kg⁻¹ in the Bering Sea (SR = 235%; $F_{CH_4} = 2.25 \mu\text{mol m}^{-2} \text{d}^{-1}$). In addition, high CH₄ concentration the Chukchi sea shelf slope (72°N) was 17.6 nmol kg⁻¹, which was also observed by Li et al. (2017). The CH₄ maxima at the shelf break (72°N) may be resulted of diffusion from bottom water into the water column (Li et al., 2017).

4.5.1.2 Spatial distribution of dissolved CH₄ from GC measurement

Over whole region, dissolved CH₄ concentration range was 4.8-14.3 nmol kg⁻¹ and the average was 6.4 nmol kg⁻¹ in near-surface seawater (10 m depth) (Fig.4.3a). Dissolved CH₄ was obviously supersaturated (range: 141–416%; average 185%) in the surface seawater. In comparison, the CH₄ concentration in near surface seawater in this study was generally lower

than that reported in the previous reviews (10.08 nmol kg⁻¹, Li et al., 2017; 11.2 nmol kg⁻¹, Savvichev et al., 2007).

Near the coast, CH₄ concentration was as high as up to 14.3 nmol kg⁻¹ (SR = 416%), it was 2.5 times higher than surrounding surface water (~5 nmol kg⁻¹; SR ~150%). At the stations with shallow water depth (~50 m), dissolved CH₄ concentration increased with depth to the bottom (Fig. 4.4 b,c). At a few stations with a water depth of ~200 m (Fig. 4.5), the CH₄ concentration increased with depth up to 50 m but decreased with depth at greater depths. The lowest CH₄ concentration at 200 m depth was 1.7 nmol kg⁻¹ (SR = 48.8%) in the deeper water. In this study, the observed CH₄ concentration was lower in Atlantic-derived water than in Pacific-derived water (PSW, PWW and ACW) (Fig. 4.6). The highest CH₄ supersaturation appeared in bottom water above the sediment. In that water the SR of 947% (35.81 nmol kg⁻¹) probably resulted from high *in situ* methanogenesis of organic matter in the sediments (Reeburgh, 2007) and diffusion of the produced CH₄ into the water column. High CH₄ production at the Chukchi Sea shelf may be laterally transported into the Canada Basin by ACW (Matveeva et al., 2015). The ACW captured higher CH₄ from riverine before it came to Chukchi Sea, which result from the methanogenesis of organic matters in degradation of permafrost (Striegl et al., 2012). In contrast, Ruppel (2015) reported that there is no influence of subsea permafrost or gas hydrate on the elevated CH₄ in water column, but it may results from methanogenesis in the sediments. In general, the degree of CH₄ supersaturation in the Chukchi Sea was smaller than that of the East Siberian Sea (ESS), had been reported by Shakhova et al. (2010).

The Chukchi sea is a productive regions because of its high primary productivity, large quantity of particulate organic matter settles to the sea floor (Grebmeier et al., 2006; 2015; Lepore et al., 2007). It may support for high rates of microbial processes for CH₄ formation in the upper sediments (Savvichev et a., 2007). At the shelf break, CH₄ concentration exceeded the equilibrium concentration with the atmosphere by 3 orders of mangitude. Maxima CH₄ concentrations at the bottom would have been released upward and diluted. A few CH₄ hotspot within the Chukchi Sea has been revealed (Savvichev et al., 2007).

4.5.2 Sea-air CH₄ flux

There was a minor source of CH₄ with sea-air flux averaging 1.02 μmol m⁻² d⁻¹ (ranged from 0.001 to 19.45 μmol m⁻² d⁻¹) from CRDS system was compared to 2.4 μmol m⁻² d⁻¹ (Fig. 4.7) measuring from GC.

Sea-air fluxes of CH₄ in this study from southwestern Canada Basin were approximated the observed CH₄ emission of Fenwick et al. (2017) in the Bering Sea (0.2 $\mu\text{mol m}^{-2} \text{d}^{-1}$), in the Canada Basin (1.3 $\mu\text{mol m}^{-2} \text{d}^{-1}$) and in the Chukchi Sea (1.9 $\mu\text{mol m}^{-2} \text{d}^{-1}$).

4.5.3 Comparison of dissolved CH₄ distribution inside and outside of the warm-core eddy

Comparison of vertical profiles of dissolved CH₄ concentration within and outside the warm-core eddy is shown in Fig. 4.8. CH₄ concentration peaked at approximately 50 m and 100 m depth outside and inside the eddy, respectively. I hypothesize that the greater thickness of subsurface CH₄ peaks inside than that outside the warm-core reflected the characteristics of the eddy. Which factors may have been responsible for this phenomenon?

Does *in situ* CH₄ production in the water column affect the subsurface CH₄ peaks? We could not detect spreading of dissolved CH₄ from either the sea surface or from the bottom water (Fig. 4.8). Is there a contribution from *in situ* production in the water column? If a large amount of CH₄ production occurred in the water column, a large source of particulate organic matter would be needed. However, there was no clear correlation between chlorophyll-*a* concentration and CH₄ at the subsurface water (0-50 m depth). In addition, CH₄ concentrations were correlated with the concentrations of nutrients such as ammonium, phosphate, and silicate (Fig. 4.10). The similarity of the characteristics of the subsurface nutrient peaks and the dissolved CH₄ concentrations implies that there was *in situ* source of CH₄ in the subsurface water. The CH₄ maxima at 100-200m depth might be transported laterally along the broaden shelf water or by Alaskan Coastal Current (Hioki et al., 2014; Kudo et al., 2018). Besides, another mechanism, such as ocean physics, needs to be considered.

Do physical factors (e.g., seawater temperature) affect the distribution of CH₄? At the shelf stations, CH₄ supersaturation was detected where the seawater temperature was high (up to 6°C) and the salinity low. Inside the warm core, the water temperature was as warm as 5–6°C, whereas below the warm core the water was relatively cold (–1.46°C) and dense. Salinities inside and outside the warm core were the same (30.9), but the temperatures differed, 4.39°C and 1.18°C, respectively. The corresponding methane solubilities were 3.25 and 3.55 nmol kg^{-1} , respectively. The difference in CH₄ solubilities was 2.9%, whereas the difference in CH₄ saturation was 13.5%. Physical processes such as seawater temperature therefore did not strongly affect the spreading of subsurface CH₄ concentration.

The spreading of CH₄ that occurs in a warm-core eddy may explain the entrainment and lateral transport of warm-core water from the shelf water. A maximum tangential velocity of

about 0.55 cm s^{-1} in magnitude was observed within the warm core, as we mentioned above. In addition, remarkable maxima of vertical shear were found along the top and bottom of isopycnals corresponding to $\sigma_\theta = 24.5\text{--}25.0$ and $26.0\text{--}27.0 \text{ kg m}^{-3}$ (corresponding to depths of 50, 100, and 150 m, respectively). A broad extension of sub-surface CH_4 peaks was therefore present inside the radius of the eddy where the shear velocity was a maximum. The spreading that occurred inside of warm-core eddy can therefore be explained by entrainment of CH_4 from shelf water associated with nutrients's tracer.

4.5.4 Eddies transport CH_4 from shelf-slope into the basin

The primary mechanism of water that is transported off shelf water into the interior basin in the Arctic by ocean currents remains largely unclear. One possible mechanism is formation of eddies at or near the continental shelf-break in the Chukchi/Beaufort Seas, migrates the water into the central basin including nutrients (Nishino et al., 2011), implying the eddy can transport dissolved CH_4 at the same time. I would like to discuss whether the eddy has the meaningful role in transporting of CH_4 into the southwestern Canada Basin.

By tracking sub-sections, I could see that warm-core eddies apparently helped to transport dissolved CH_4 from the shelf slope ($35.81 \text{ nmol kg}^{-1}$ at sta.47) into the basin and transported the CH_4 into deeper water (32.3 and $18.7 \text{ nmol kg}^{-1}$ at sta. 37 and 35, respectively) (see Fig. 4.9). Possible mechanism contributing to CH_4 transport in this area was lateral transport, which would have carried dissolved CH_4 from the shelf-slope into the Barrow Canyon and Canada Basin. A CH_4 “hotspot” at a distance of 150 km from the coast appeared at the deeper water. This hotspot may have resulted from downward transport in the warm-core eddy further northward.

In this study, I detected the influence of warm-core eddies on the transport and distribution of dissolved CH_4 in the water column even though it accounted for only about 3% of the total number of eddies in the western Canada Basin and Chukchi Plateau (Plueddemann and Krishfield, 2007). However, approximately 125 cold-core eddies per year that occur in Arctic water may affect material transports at the shelf water if these eddies have an average radius of 10 km and a thickness of 75 m (volume of 25 km^3) (Pickart et al., 2005). Cold-core (anticyclonic) and warm-core (cyclonic) eddies probably transport dissolved CH_4 in similar ways.

A basic difference, however, between anticyclonic and cyclonic eddies is the mechanism that pulls down or lifts up subsurface CH_4 peaks, respectively. This difference between the eddies would affect the timing of CH_4 release and removal mechanism from the

shelf. A cyclonic warm-core eddy may lift up sub-surface CH₄ peaks until the eddy dissipates. The CH₄ is diluted and eventually released into the atmosphere because the CH₄ is volatile. A warm-core eddy contributes directly to advection of CH₄ at the shelf break and in the upper halocline. This eddy transport was directly responsible for maintaining CH₄ maxima at 100 m depth at eddy stations in the southwestern Canada Basin in our study. In a cold-core eddy, this process may affect the release of CH₄ into the atmosphere from the shelf water. Mathis et al. (2007) estimated the amount of nutrients that eddies remove from the shelf and concluded that cold-core eddies are the dominant removal mechanism in the western Arctic Ocean. Even though we assume that cold-core eddies contribute to CH₄ effluxes compared to warm-core eddies, we have not yet observed CH₄ transport in a cold-core eddy. Future field observations including of cold-core eddies will be needed to obtain a comprehensive understanding of how eddies affect CH₄ distributions.

4.5.5 Correlation between CH₄ and Nutrients

High concentrations of nutrients (e.g., ammonium, phosphate, silicate) are characterized for large amount of organic matters link to CH₄ methanogenesis. High ammonium concentration during summer indicates for decomposing of organic matter which is deposited on the Chukchi Sea shelf (Cooper et al., 1997; Nishino et al., 2005). Ammonium concentration increased rapidly up to 3.4 $\mu\text{mol kg}^{-1}$ at a depth of 40-50 m compared to the continental shelf. A ammonium plume at a distance of 150 km from the coast was found at 70-150 m depth with concentration of 2.6 $\mu\text{mol kg}^{-1}$ (Fig. 4.10a), which was at the same depth to the CH₄ hotspot. Nevertheless, at the northern of the continental shelf slope, ammonium concentration was low and almost close to zero. Plumes of silicate and phosphate were also observed at the same depth of ammonium (Fig. 4.10b, c). In contrast, nitrate concentration was increased rapidly at a depth of 70-80 m at the center of the transect (Fig. 4.10d). At 150-km distance from the coast, the warm-core eddy contained high ammonium, high-silicate and high-phosphate contents compared to the surrounding water. It is completely different from Nishino et al. (2011) reported that the warm-core eddy packed the low-silicate and low-phosphate water. The warm-core eddy could carry high concentration nutrients trapped on the shelf water during formation to Canada Basin.

Negative values of fixed nitrogen (N^{**}) indicate the occurrence of significant denitrification within the shelf sediments (Codispoti et al., 2005a; Gruber and Sarmiento, 2009). Minimum values of N^{**} was observed at the shelf water, and lied on isoline of 32-33 or 25-26 σ_θ . In particular, minima N^{**} (-12.3 $\mu\text{mol kg}^{-1}$) correlated to maxima nutrients (NO_3^- : 14.71

$\mu\text{mol kg}^{-1}$, PO_4^{3-} : $1.765 \mu\text{mol kg}^{-1}$) (Fig. 4.10) at the same layer in the UHL ($< 200 \text{ m}$) because N^{**} derived from the nitrogen and phosphate concentration. Low nitrate concentration is probably attributable to low N^{**} values in the Bering shelf water. The similarity of minimum N^{**} has been reported at the same depth by Hioki et al. (2014). During denitrification process, nitrate is consumed instead of oxygen during the bacterial decomposition of organic matter in low-oxygen water in the sediment pores. This process causes a reduction in N^{**} which is found in the Bering/Chukchi Seas (Devol et al.; 1997; Yoshikawa et al., 2006) and indicates a net loss of nitrate in the ocean interior (Devol et al., 1997).

Strong correlations at the coast were found between CH_4 and NH_4^+ ($\text{CH}_4 = 9.11 \times \text{NH}_4^+ + 4.52$, $R^2 = 0.87$, $n = 20$); silicate ($\text{CH}_4 = 0.62 \times \text{SiO}_2^{2-} + 3.35$, $R^2 = 0.77$, $n = 20$); and phosphate ($\text{CH}_4 = 25.07 \times \text{PO}_4^{3-} - 9.28$, $R^2 = 0.77$, $n = 20$); fixed nitrogen (N^{**}) was: ($\text{CH}_4 = -5.49 \times \text{N}^{**} - 28.37$, $R^2 = 0.8$, $n = 20$). This finding implies that CH_4 is produced at the shelf stations where response for high primary productivity and large quantity of particulate organic matters settling to the seafloor (Grebmeier et al., 2006, 2015; Lepore et al., 2007) with a high rate of microbial CH_4 processes in the upper sediments (Savvichev et al., 2007). This result is consistent with previous research that obtained high CH_4 concentration upper the sedimentary in the Chukchi Sea shelf (Matveeva et al., 2015; Li et al., 2017).

Apparent oxygen utilization (AOU) was high ($> 60 \mu\text{mol kg}^{-1}$) (Fig. 4.10) near the bottom layer at the coastal stations indicated low dissolved oxygen (DO) concentration even though it still maintained aerobic biological activities. Above the sedimentary layer, CH_4 concentrations was high as well as AOU concentration. A positive correlation between AOU and CH_4 at coastal station 47 ($\text{CH}_4 = 0.276 \times \text{AOU} + 17.64$, or $\text{SR} = 6.90 \times \text{AOU} + 489.56$, $R^2 = 0.6$) means CH_4 -rich water or existence of CH_4 production at the positive AOU. At the other stations without effect of the eddy, a negative correlation between AOU and CH_4 ($\text{CH}_4 = -0.05 \times \text{AOU} + 7.18$ or $\text{SR} = -1.617 \times \text{AOU} + 206.42$, $R^2 = 0.62$) was found.

In this study, a distinct CH_4 maxima occurred at which corresponds to the near isoline of 32 that is similar to the publication of Li et al. (2017). Profiles of nutrients, N^{**} and CH_4 against to potential density (σ_θ) showed a well-developed halocline in the range of $24\text{-}27.5 \sigma_\theta$ (Fig.4.11). In the shelf water, maximum CH_4 was observed in the bottom water ($\sigma_\theta = \sim 26$). The fact that the maximum concentration of nutrients and N^{**} occurred at $\sigma_\theta = \sim 26$ in the shelf, slope and further region; it strongly suggests the lateral transport of these nutrients and CH_4 from cold and dense bottom water ($T \leq 1^\circ\text{C}$, $S \sim 32$, $\sigma_\theta = \sim 26$) of the shelf to the halocline in the shelf slope region. Amount of organic matters is delivering to in the upper halocline water, it would be contributed to oxygen consumption in shelf slope of the Canada Basin. We suggest

that the subsurface CH₄ maxima originates from the shelf water as a result of northward spreading water by oceanic currents and propagation of the eddy.

4.5.6 Correlations between nutrients and CH₄ inside/outside the warm-core eddy

The relations of CH₄ and nutrients at the coastal station have been found in the previous section. These nutrients and CH₄ could be simultaneously trapped and maintained within the warm-core during transporting from shelf water to the southwestern Canada Basin. As if the relations between them are the same within and outside the eddy, the relations between CH₄ and nutrients between within and outside of the eddy are considered as follows.

For ammonium at outside the eddy, the correlation between them was agreed well ($\text{CH}_{4\text{out}} = 9.6 \times \text{NH}_4 + 5.3$, $R^2 = 0.81$) (Fig. 4.12). Within the eddy, it showed weaker correlation ($\text{CH}_{4\text{in}} = 7.97 \times \text{NH}_4 + 5.15$, $R^2 = 0.64$) comparing to the outside. Lower concentration of CH₄ inside the eddy is discussed in the previous section. Lesser concentration ammonium inside the eddy may be caused by physical factors (e.g., later transport, mixing or dilution) or by biological factors (ammonium is taken up by phytoplankton's existence) (Nishino et al., 2011).

For nitrate, phosphate, and N^{**}: Fig. 4.12 showed lesser correlation within the eddy than at the outside eddy. Because they are non-conservative and are easily consumed by nitrification and denitrification (Codispoti et al., 2001; Nishino et al., 2005). However, both of them have the same distributional structure between within and outside the eddy with CH₄. I suggest the relation of CH₄ and each nutrient is a similar feature between within and outside the warm-core eddy.

Phytoplankton may be affected by the warm-core eddy. The chl-*a* concentration was high on the shelf area (sta. 47 and sta.15) and the shelf slope, for example, at the sta. 23 and sta. 30 (Fig. 4.13). However, integrated chl-*a* concentration over the warm-core eddy (sta. 37) was lower than that of the water outside the eddy in the top of 50 m layer (approximately the euphotic zone) is shown in the Fig. 4.13. The chl-*a* concentration of large phytoplankton (> 10 μm) drastically decreased between the shelf area comparing to the eddy area (green bars). Chl-*a* concentration of 2-10 μm was slightly higher in the shelf area than within the eddy and shelf-slope (red bars). There is no different from chl-*a* concentration of picoplankton (< 2 μm) (blue bars) between stations. In the shelf area, large phytoplankton is dominated such as diatoms because of high concentration of nutrients in the euphotic zone. Diatoms can uptake and store nutrients at a more rapid rate than picoplankton can (Smetacek, 1998). Over the warm-core eddy (sta. 37), composition of size-fractionated chl-*a* was similar to at the shelf water (sta. 47

and sta. 15), but chl-*a* concentration itself was low. It might be caused by the maintenance of large phytoplankton inside of the eddy.

4.5.7 Correlation between CH₄ saturation (SR) and AOU at isoline of 32 and N minimum layer**

Lateral transport supplied negative N**, CH₄ and nutrients from the shelf water into the basin interior as well as AOU. Maxima CH₄ linked to negative N** values and positive AOU at the coastal station. Observed CH₄ maxima was discovered on the isoline of 32 (Li et al., 2017) in the Chukchi shelf and was also detected on the isopycnal of $\sim 26\sigma_\theta$ in the UHL (Fig. 4.10e) (Hioki et al., 2014). AOU indicates for oxygen utilization of organisms, CH₄ production could be detected with high AOU concentration. Thus, correlation between CH₄ (or SR) and AOU is expected to find (Fig. 4.14). In general, SR and AOU have a negative relation excepting for some scattered points in two ellipsoids (Fig. 4.14). The relation shows high CH₄ agreed to low AOU concentration in the aerobic water, which could be derived from rich-CH₄ water from the bottom water by mixing processes. A negative correlation outside the warm core indicates that CH₄ is generally oxidised in the aerobic environment (Fig. 4.15a) except for coastal stations.

On the contrary, inversed relation between SR and AOU appears for coastal station and inside the warm-core water, are shown as scattered points in the upper trend arrow. The positive correlation points out that it is completely different from general case (Fig. 4.14) showing in two ellipsoid shapes. CH₄ production requires strictly anaerobic conditions by microbes in hypoxic microenvironments (sinking organic particles and within guts of zooplankton). However, high CH₄ is consistent with positive AOU even though low dissolved oxygen indicates for the aerobic environment reflected CH₄ supply from low dissolved oxygen upper sedimentary. For scattered blue circles, it is obtained from the shelf water which is discussed in the previous section. For scattered yellow circles is shown in Fig. 4.14 the positive correlation derives from the warm-water core (Fig. 4.15b, $SR = -4.172 \times AOU + 503.92$; $R^2 = 0.87$). Two correlations between SR and AOU has a similar pattern to at the shelf water but lower CH₄ concentration was observed. Within the warm-core, water-core is isolatable to contact with surrounding water to exchange oxygen-rich water. Low dissolved oxygen within the warm-core was caused by oxygen consumption to maintain biological production via remineralization (Mathis et al., 2007) and/or biological activities.

The high CH₄ concentrations observed inside the warm-core radius were trapped as the vortex formed and propagated. As the eddy spread, local *in situ* CH₄ production and oxidation may have occurred. CH₄ can be produced in small anaerobic environments in organic particles

(Karl et al., 2008) or in the digestive tracts of organisms (de Angelis and Lee, 1994). Methane oxidation rates of $0.0038 \text{ nmol hr}^{-1}$ ($= 0.0912 \text{ nmol day}^{-1}$) and $0.03\text{--}0.09 \text{ nmol day}^{-1}$ have been reported in the Beaufort Sea (Kitidis et al., 2010) and above the Chukchi Sea shelf (Savvichev et al., 2007), respectively. Here I assumed that CH_4 oxidation occurred during the week when the eddy was propagating. The average calculated CH_4 concentration inside the warm core after 7 days was calculated to be $18.91 \text{ nmol kg}^{-1}$ and $23.52 \text{ nmol kg}^{-1}$, respectively, with initial CH_4 concentration of $35.81 \text{ nmol kg}^{-1}$. However, the observed CH_4 concentration inside the warm core was $32.3 \text{ nmol kg}^{-1}$, higher than the calculated CH_4 concentration (Table 4.3). Matveeva et al. (2015) have indicated that the loss of CH_4 from the Chukchi Sea shelf are is caused by ocean currents, mixing, and diffusion. Thus, within the warm-core eddy, CH_4 was produced, or methane oxidation rates were lower than $0.06 \text{ nmol day}^{-1}$. I therefore speculate that both CH_4 losses caused by currents, mixing, and diffusion and CH_4 production contributed to the CH_4 distribution inside the warm core during propagating of the eddy.

4.5.8 Nitrous oxide (N_2O)

Nitrous oxide concentration in the southwestern Canada Basin ranged from 9.0 to $24.4 \text{ nmol kg}^{-1}$ (saturation range: 71.5-161.6%), with a difference in a distinct water masses. In the surface freshwater layer, low concentration of N_2O was observed in river-input or melt-water (MW) and in the ACW whereas higher N_2O extent in the PSW (Fig. 4.17). The N_2O concentration was almost supersaturated respecting to atmospheric N_2O , with maximum concentration up to $16.7 \text{ nmol kg}^{-1}$ ($\text{SR} = 111\%$). In contrast, high concentrations of N_2O were observed in the PWW due to mixing in the shallow water in early spring and summer. In shallow Pacific water masses (PSW, ACW, MW) ($< 25.5\sigma_t$) in the southwestern Canada Basin, N_2O concentrations that were low close to atmospheric N_2O equilibrium. It implies that processes in this region likely a combination of sea-air exchange and dilution with low N_2O water masses, it leads to decrease in N_2O concentrations. N_2O maxima was also observed on the $26\sigma_\theta$ isopycnal line ($S \approx 32$) like CH_4 that is transported from the shelf water off-shore into the basin interior, was discussed in the previous section of CH_4 distribution. In deeper water, N_2O concentrations tend to decrease with increasing density, low N_2O concentration was observed with a mean value of $\sim 15 \text{ nmol kg}^{-1}$ and 97% saturation. This observation is relatively consistent with the recent observation of Zhang et al. (2015) and Fenwick et al. (2017).

A negative correlation between $\Delta\text{N}_2\text{O}$ and N^{**} is examined that providing an important evidence of N_2O source. The correlation implies the denitrification is a primary source of N_2O from sediment into the water column like Zhang et al. (2015) and Fenwick et al. (2017) strongly

suggested that the N₂O production from the sediment of the Bering Sea and Chukchi Sea are a primary source in the Arctic Ocean. However, no relation between N^{**} and N₂O has been found because of its significant scatter ($R^2 = 0.067$). This could be attributed by another process impacting on the N₂O production. Recently, processes of denitrification-nitrification coupling have played a significant process in the nitrogen cycle on the sediments of the Bering and Chukchi shelves (Horak et al., 2013; Brown et al., 2015) and N₂O production (Naqvi et al., 1998). However, the trend of N^{**} and ΔN_2O shows that denitrification may work as a N₂O source in this region.

4.5.9 DMSP as a precursor of methane

CH₄ could be produced from methylated compounds primarily and from a metabolic by-product of algae, which are considered to be a source for methanogenesis (Scranton and Brewer, 1977; Scranton et al., 1995). One of the methylated compounds produced by marine phytoplankton is DMSP (Keller et al., 1989). Primary production or phytoplankton abundance in the Arctic Ocean is regulated by nutrients in seawater while DMSP accounts only for less than 10% of total phytoplankton carbon (Kiene et al., 2000). In this study, DMSPt correlated with chl-*a* concentration with $DMSPt = 0.25 \times chl-a + 0.63$ ($R^2 = 0.64$, $n = 22$). *In situ* excess CH₄ is produced during phytoplankton bloom, and DMSP and its degradation product in oxygen-rich polar water (Damm et al., 2008; 2010) has been reported, but no correlation between them did obtain in this study. CH₄ is not agreed well to DMSPt because of the effect of high production of CH₄ from marine sediment release.

4.6 Summary and Conclusion

During the cruise of the R/V *Mirai* (MR15-03 Leg.1) in summer 2015, CH₄ observation was examined to investigate vertical profiles of CH₄ in the Bering Sea to the Chukchi Sea and southwestern Canada Basin. Seawater samples were collected within and outside the warm-core eddy. A CRDS coupled with a shower head type equilibrator was also used to obtain continuous underway measurements of the mixing ratios of CH₄ in the surface water. In the surface seawater, CH₄ was slightly supersaturated overall region from 103-514% from CRDS system measurement. In additional, both undersaturated and supersaturated CH₄ ranged 49-947% in the southwestern Canada Basin from surface seawater to the bottom with extremely high CH₄ concentration (947% SR) reflected decomposing of organic matters.

The warm-core eddy was observed in the southwestern Canada Basin. The vertical profiles of CH₄ were examined within and outside the eddy. I found high CH₄ concentrations

in bottom water on the shelf, which located outside the warm-core eddy. I also found that subsurface CH₄ peaks within the radius of the maximum velocity of the eddy were broader at depth than that outside the eddy. Lateral transports of ocean currents and propagating of the eddy were likely the controlling factors responsible for subsurface peaks within the warm-core water. This mechanism could carry high CH₄ at the shelf water and transport to the shelf-slope water and Canada basin interior. Furthermore, nutrients (e.g., ammonium, phosphate, and silicate, fixed nitrogen) were also conveyed as well as CH₄. Two patterns of the correlation between SR and AOU were found. Negative correlation obtaining outside the warm core indicates that CH₄ is generally oxidised in the aerobic environment. In contrast, positive correlation at the coastal stations indicates high CH₄ in the aerobic water which could be derived from rich-CH₄ water or CH₄ diluted upward from upper sedimentary release with positive AOU into the water column. Within the warm-core water, positive correlation also has been found, CH₄ production and high CH₄-rich water, which was trapped when eddy formed. The positive AOU values here may be caused by biological activities, CH₄ oxidation and remineralization of organic matters. The present study indicates that eddies play a crucial role in controlling CH₄ distributions and in transporting CH₄ laterally and nutrients as well.

CH₄ and N₂O concentrations in each water mass was examined. The highest CH₄ and N₂O concentrations were simultaneously detected in the bottom water (PWW) of the shallow Chukchi Sea where active microbial metabolism under low sedimentary oxygen condition leads to high rates of CH₄ and N₂O production. Low CH₄ and N₂O in Atlantic-derived water (AW) and subsurface Pacific-derived water (e.g., ACW and MW). Similar to CH₄, high N₂O concentrations on the shelf water are transported through the Canada Basin. N₂O concentration is used as supported data for CH₄ analysis. Both super-saturation and under-saturation of N₂O were detected due to capacity of its sources and denitrification process in oxygenated water.

Sea-air fluxes of CH₄ were 1.02 (μmol m⁻² d⁻¹) and 2.4 (μmol m⁻² d⁻¹) that has been measured by CRDS and GC measurement, respectively. They approximated to CH₄ fluxes of the previous studies in the Chukchi Sea and southwestern Canada Basin. However, sea-air CH₄ flux is generally lower than that of in the East Siberian Sea or Barent Sea. I could not make a comparison of sea-air CH₄ fluxes between within and outside the eddy because high CH₄ emission from the Chukchi Sea shelf to the atmosphere compared to the further region in the southwestern Canada basin.

DMSP and chl-*a* were considered as precursors of *in situ* CH₄ production during DMSP degradation in oxygen-rich polar water and phytoplankton bloom, respectively. However, I could not find the correlation between CH₄ and its precursors in this area. Therefore, CH₄ did

not produce from marine phytoplankton and methylated compounds (DMSP). High CH₄ production in the Chukchi Sea and southwestern Canada basin reflected by sedimentary methanogenesis release.

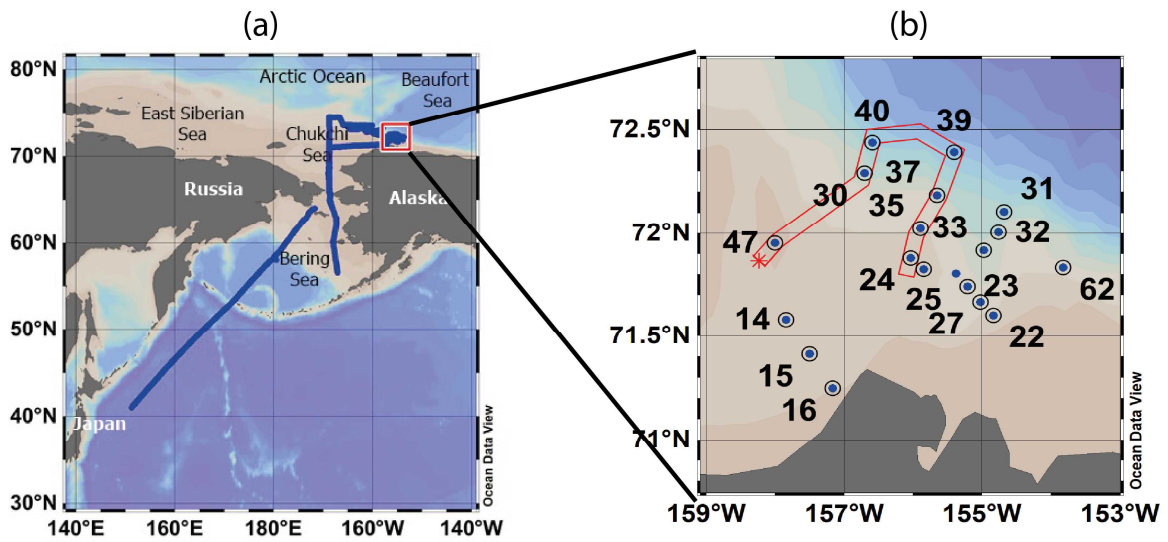


Fig. 4.1 (a) Blue line shows the cruise track of R/V *Mirai* cruise (MR15-03 Leg. 1) from August to October 2015. Red box shows the area where the discrete CH₄ samples were taken. (b) Map of sampling stations of discrete seawater. Red line shows the transect crossing a warm-core eddy during the cruise.

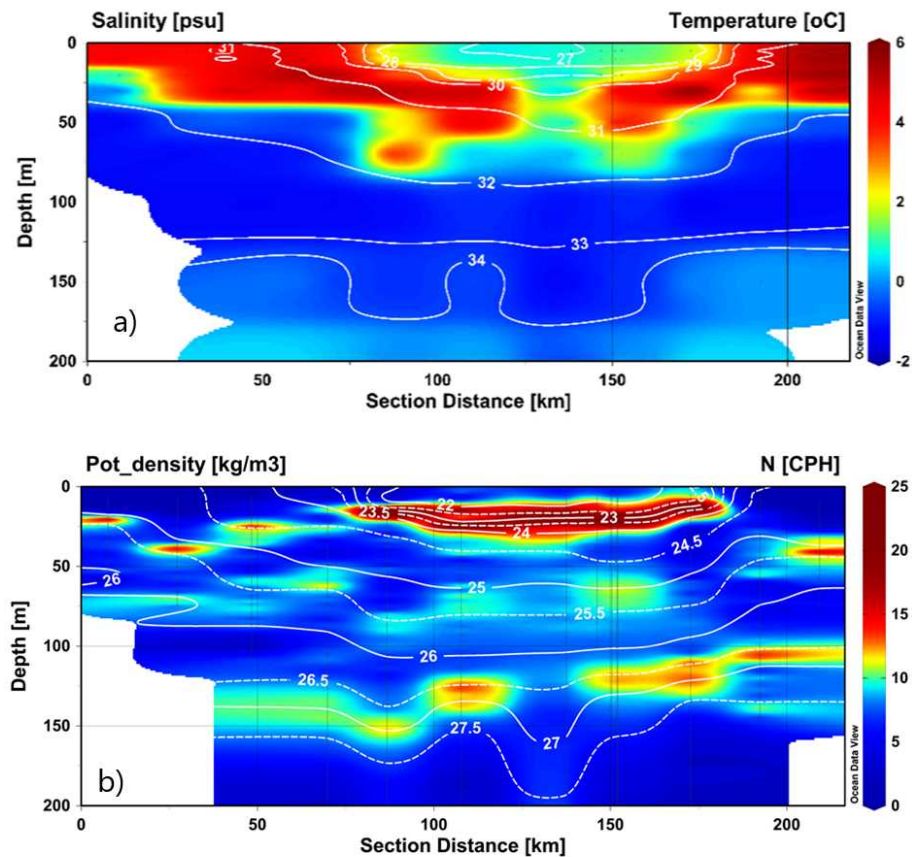


Fig. 4.2 Vertical sections of potential temperature (°C) (color) and salinity (contours) (a); Brunt-Vaisala frequency (cph) (color) and potential density (contours) (b) from surface water to 200 m depth.

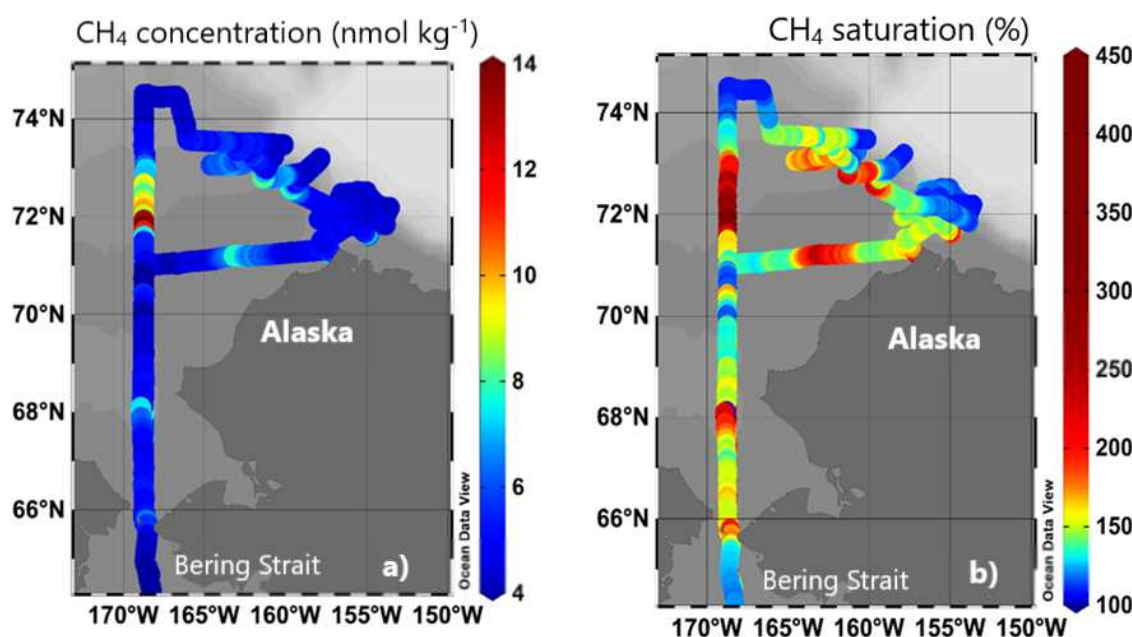


Fig. 4.3 The spatial distribution of CH₄ concentration (a-left) and CH₄ saturation (b-right) were measured by CRDS system in high latitudes in the Arctic Ocean.

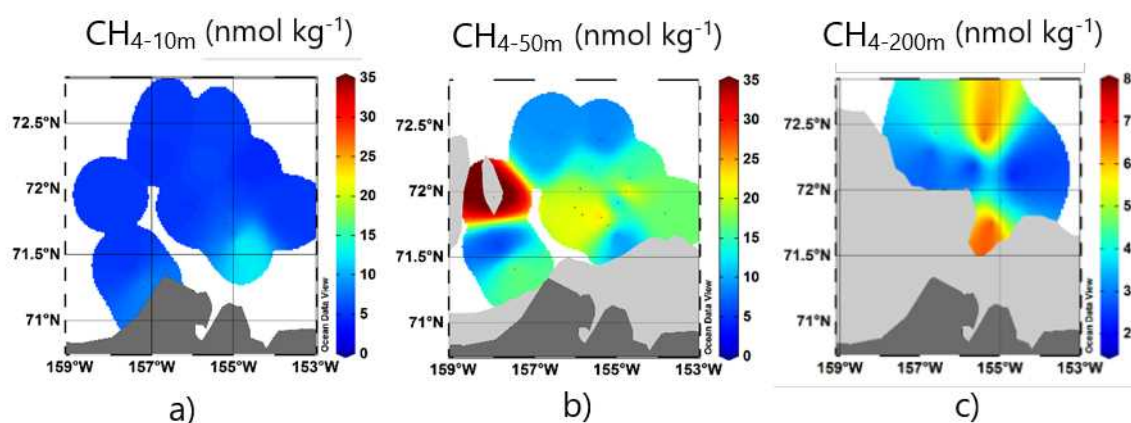


Fig. 4.4 The spatial distribution of CH₄ concentrations at depths of 10 m (a), 50 m (b), and 200 m (c) corresponding to the water near the surface, within, and below the warm-core eddy at all stations during the cruise.

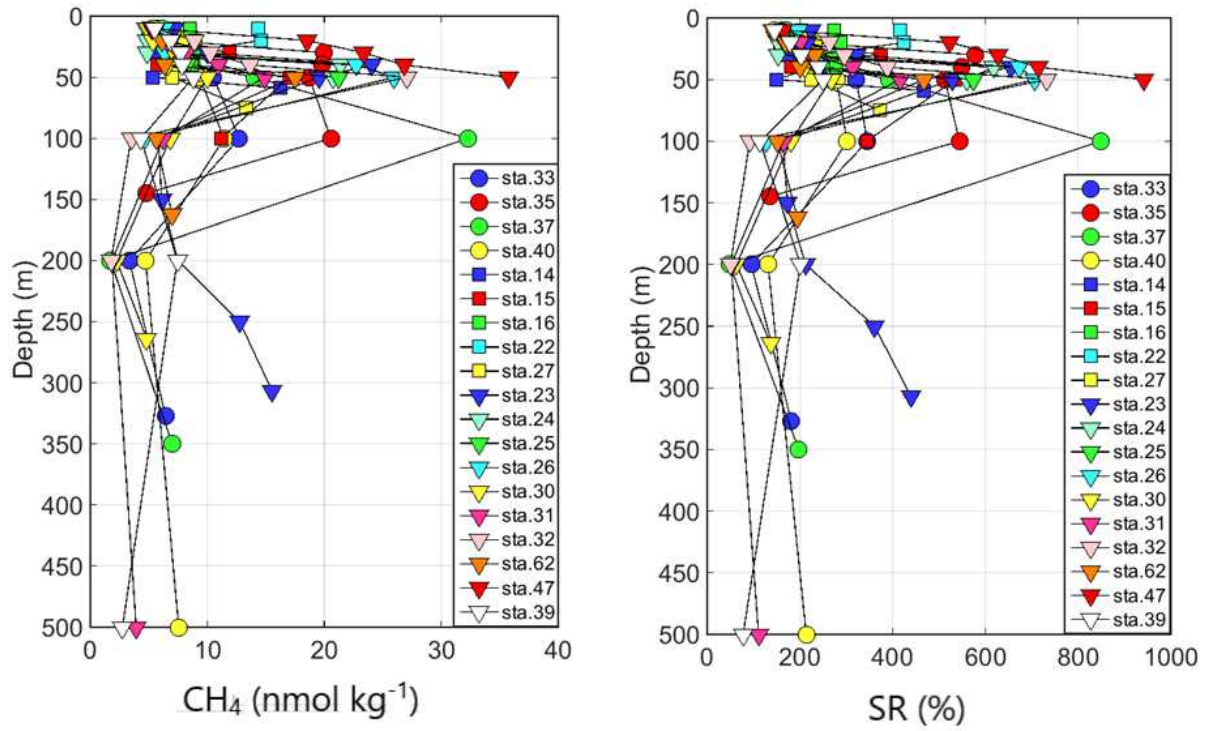


Fig. 4.5 The vertical profiles of CH₄ concentration (left) and SR (right) eddy at whole stations during the cruise.

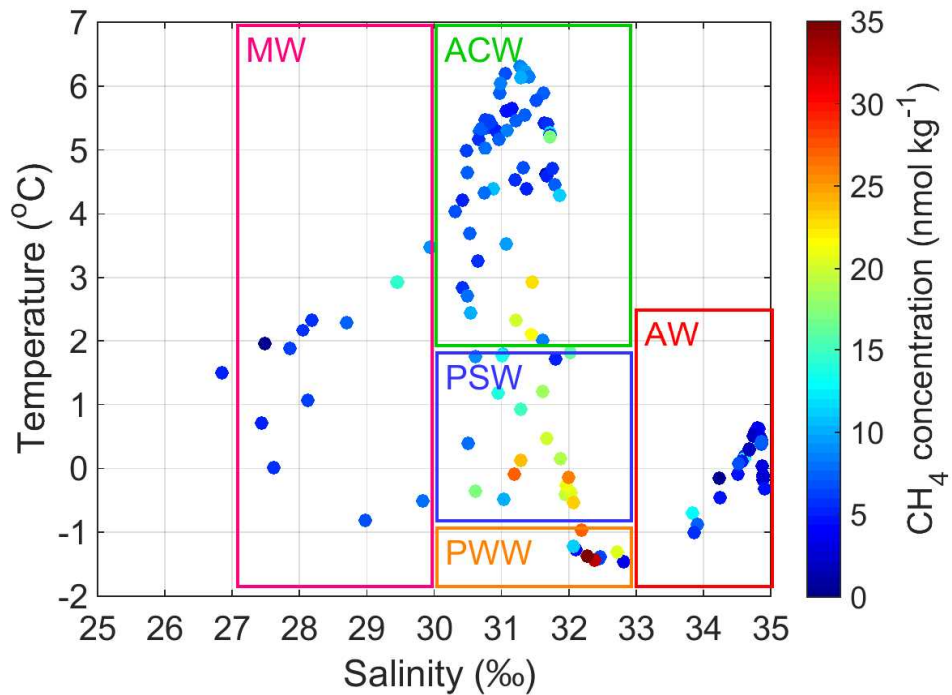


Fig. 4.6 Salinity vs. temperature plot showing the different water masses observed in the southwestern Canada Basin. Circles indicate for dissolved CH₄.

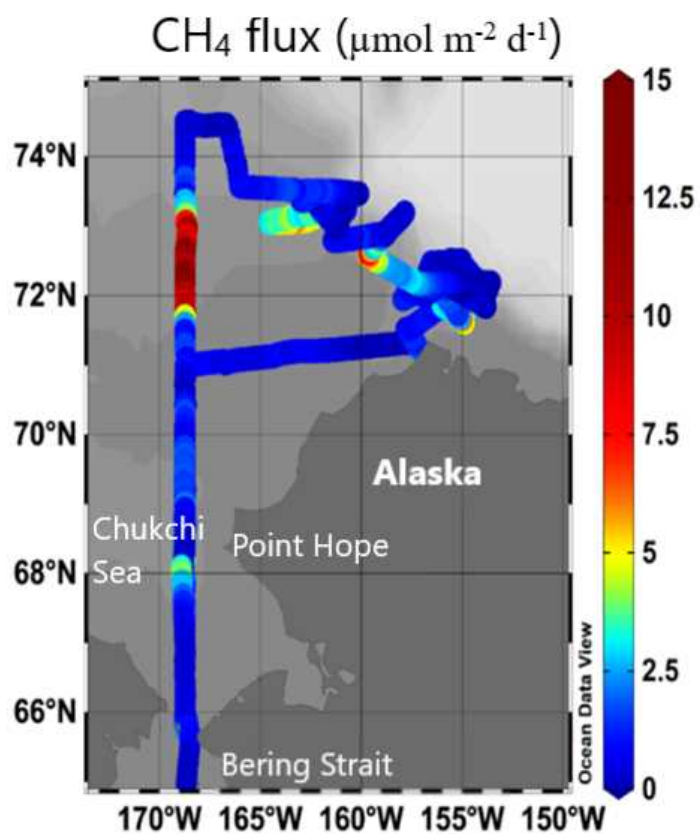


Fig. 4.7 The CH₄ flux in high latitudes in the Arctic Ocean from CRDS measurement during the cruise.

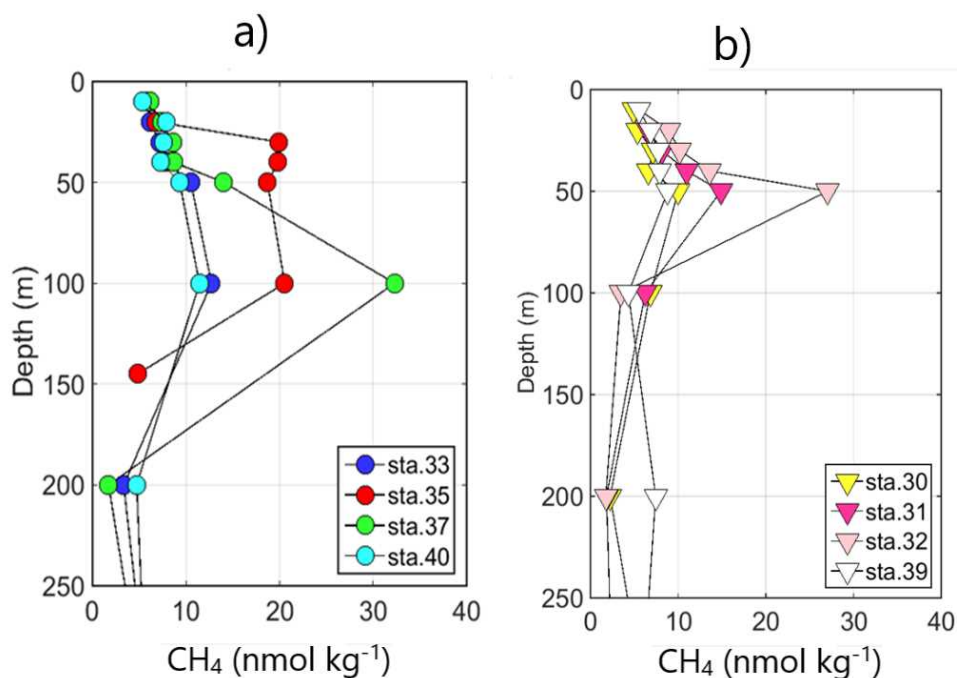


Fig. 4.8 Comparison of the vertical profiles of dissolved CH₄ concentrations in the southwestern Canada Basin inside (a) and outside of the warm-core eddy (b) from GC measurement.

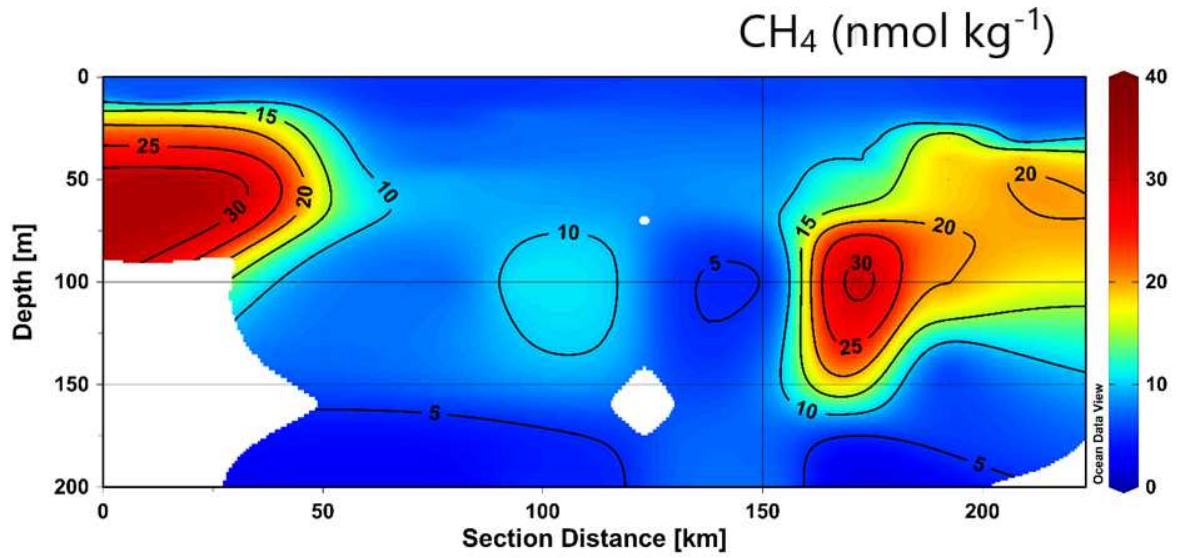


Fig. 4.9 Vertical transect distribution of CH_4 concentration was measured by GC following the red line of Fig. 4.1b. White background in both sides shows shelf area.

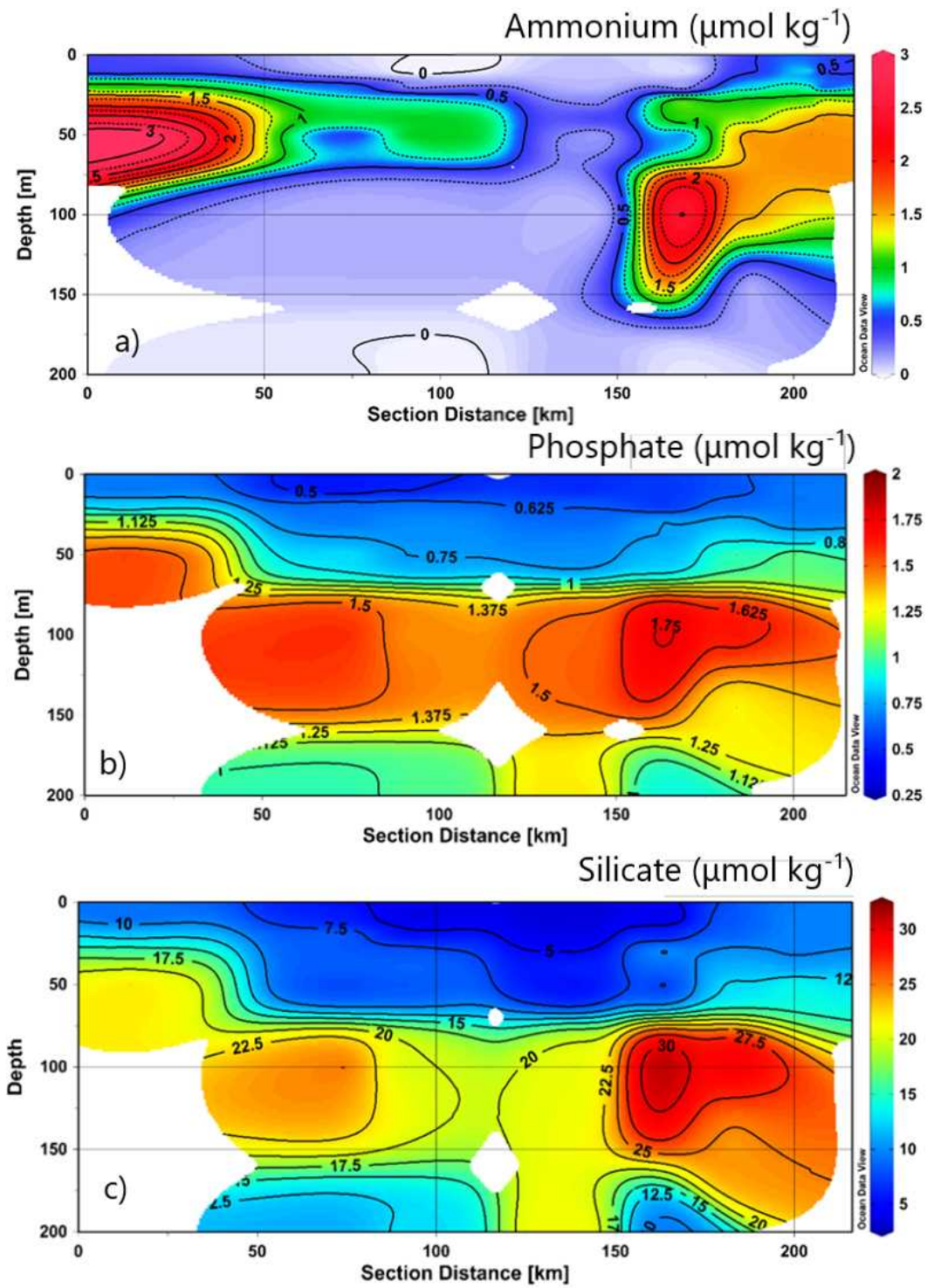


Fig. 4.10a Vertical transect distribution of ammonium (a), phosphate (b), silicate (c) concentration following the red line of Fig. 4.1b. White background in both sides shows shelf area.

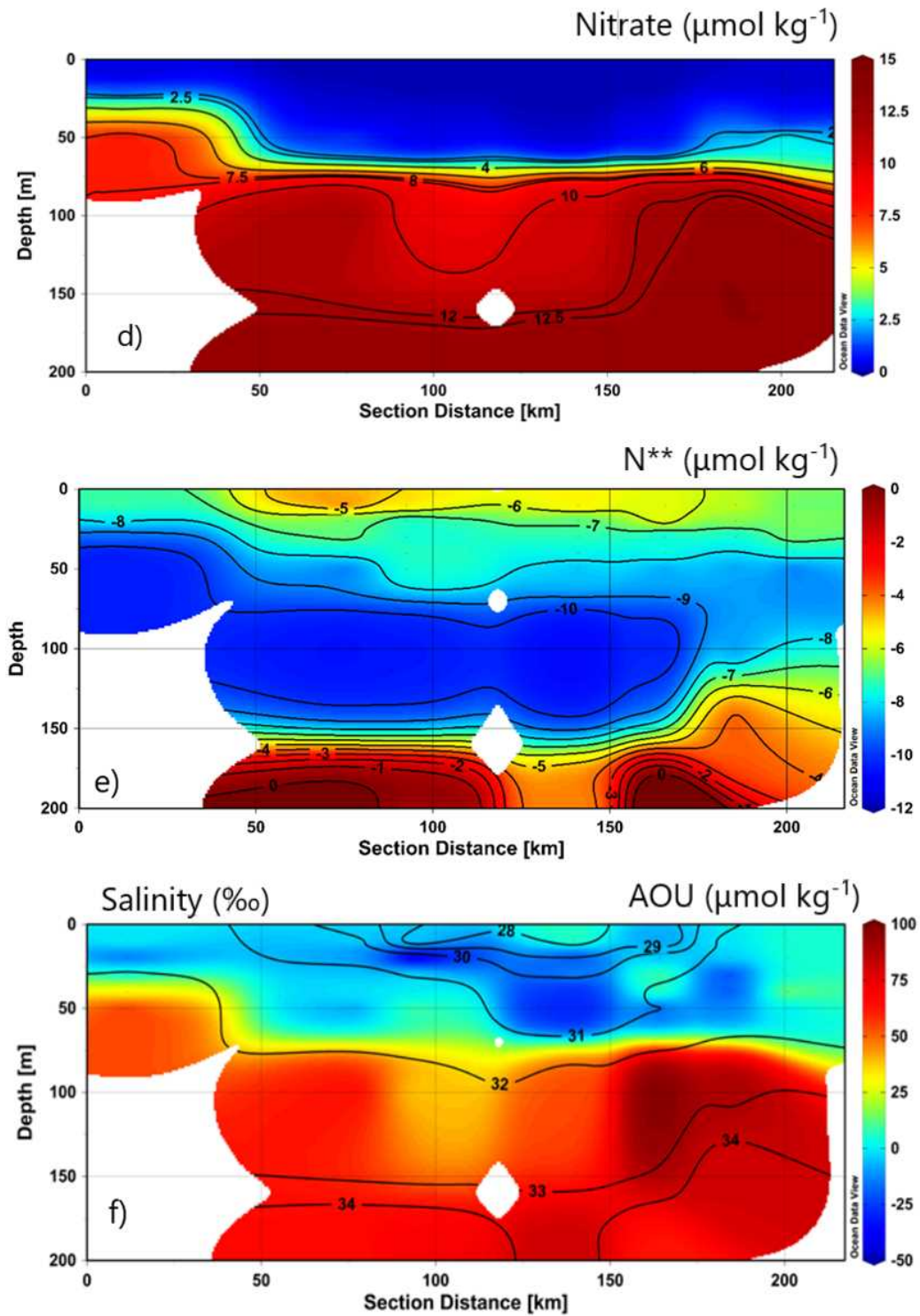


Fig. 4.10b Vertical transect distribution of nitrate (d), fixed nitrogen (e) concentration and AOU (f) following the red line of Fig. 4.1b. White background in both sides shows shelf area.

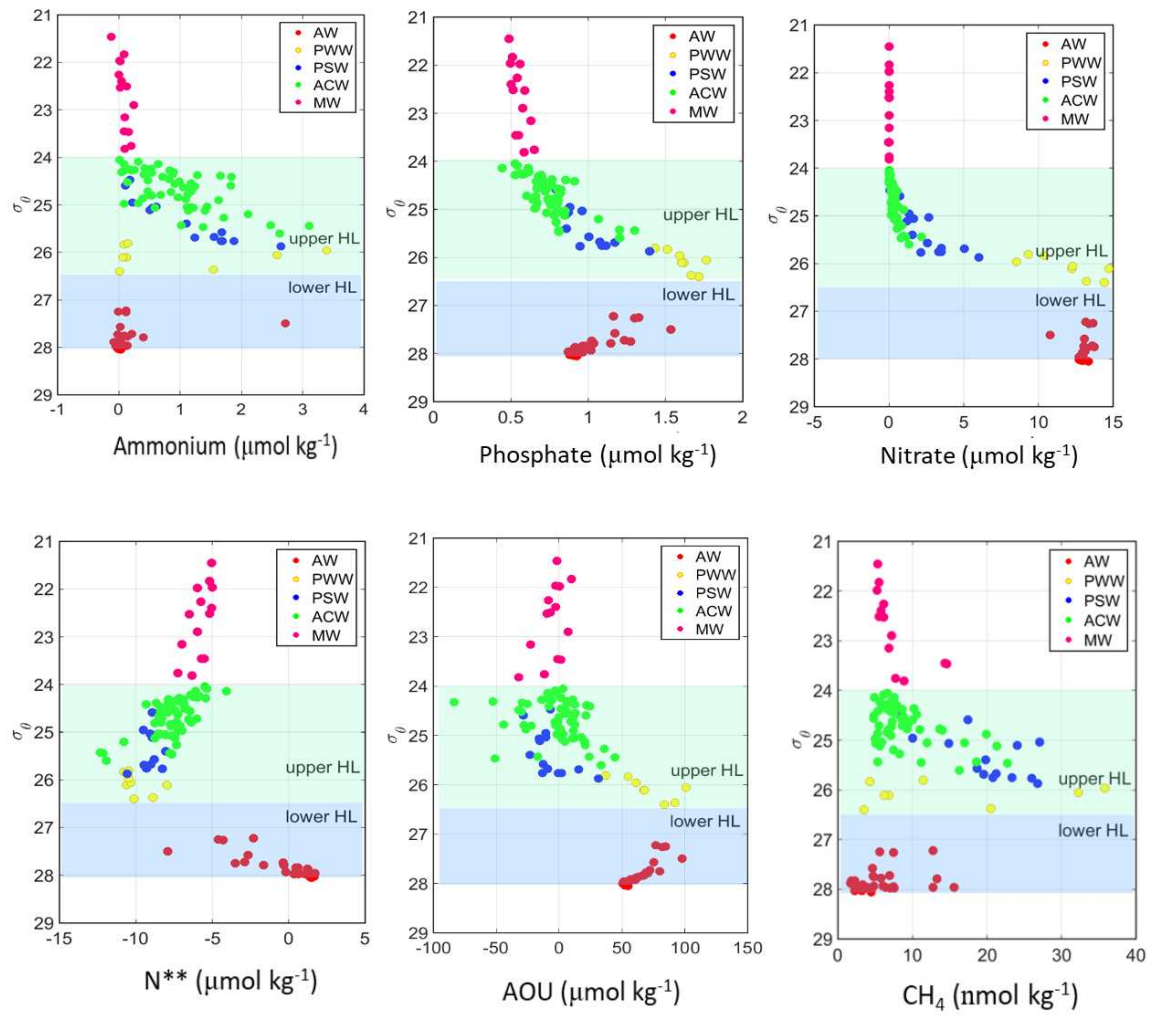


Fig. 4.11 Plot of ammonium, phosphate, nitrate, N^{**} , AOU and CH_4 against potential density (σ_θ). Colorful dots indicate for data in each water mass.

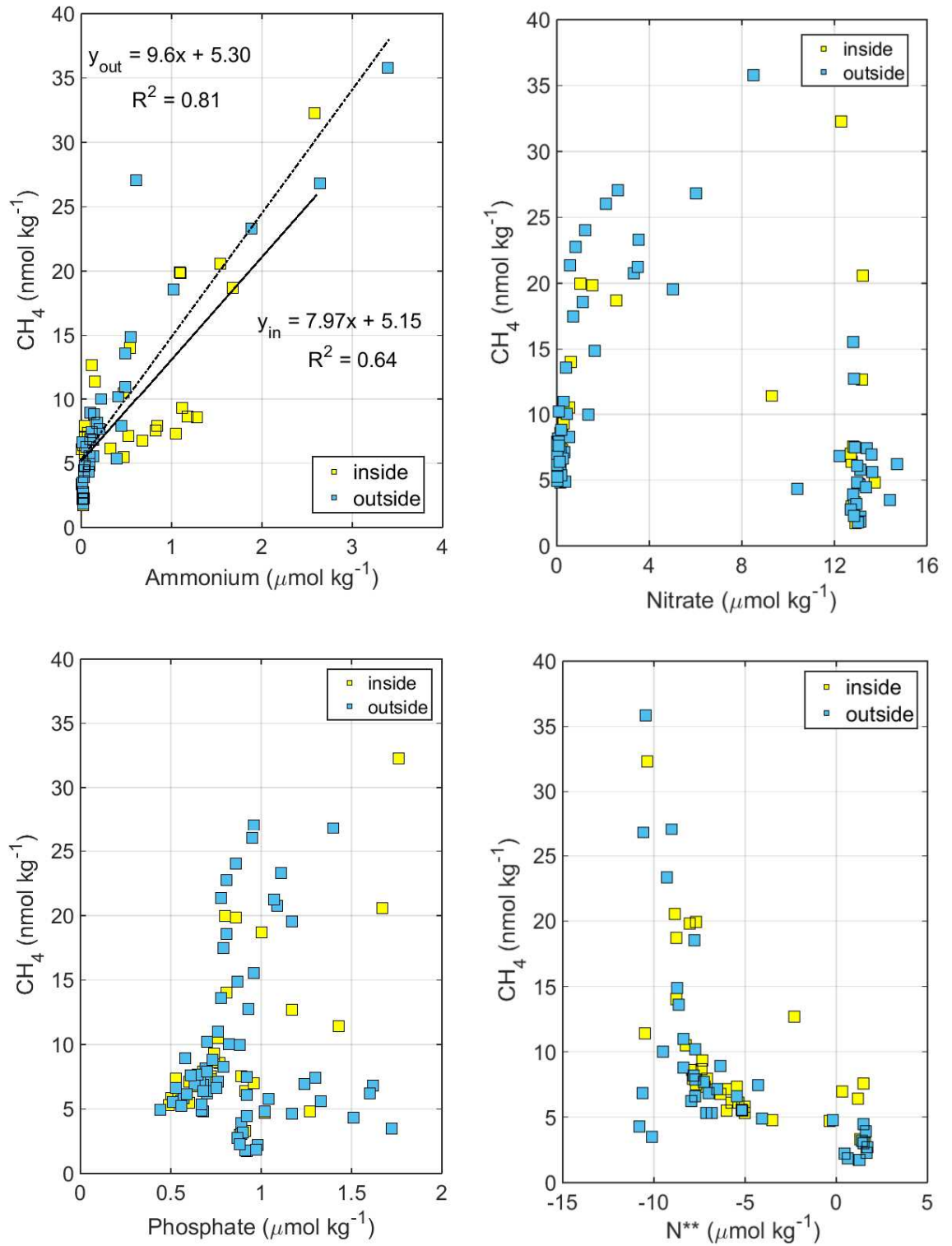


Fig. 4.12 Comparison of the relation between CH_4 and ammonium, nitrate, phosphate and fixed nitrogen concentration within (yellow squares) and outside the warm-core eddy (blue squares).

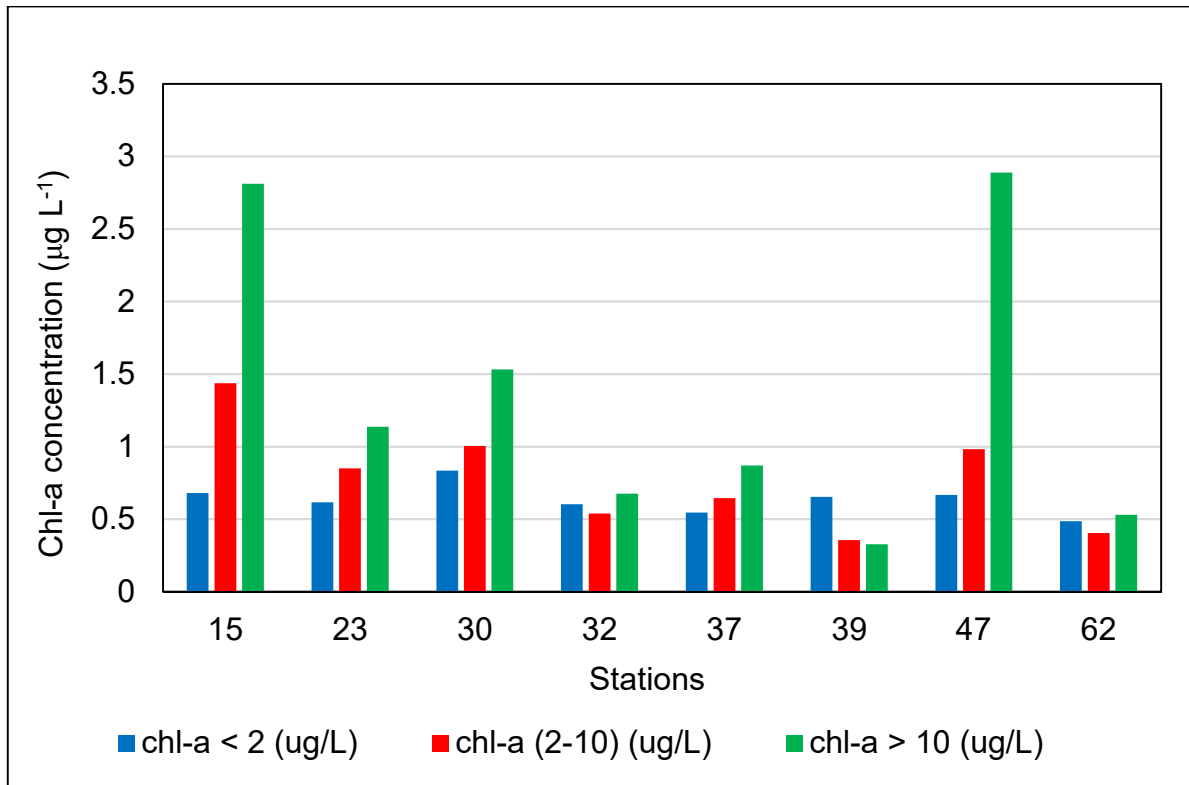


Fig. 4.13 Chl-*a* concentration ($\mu\text{g L}^{-1}$) integrated over the water column from the surface to a depth of 50 m (approximately the euphotic zone) for chlorophyll-*a* in cells $< 2 \mu\text{m}$ (blue bars), cells $2\text{-}10 \mu\text{m}$ (red bars), and cells $> 10 \mu\text{m}$ (green bars). The coastal area was at the station 15, 30 and station 47; the warm-core eddy was at the station 37. Shelf-slope was at the station 23 and station 62.

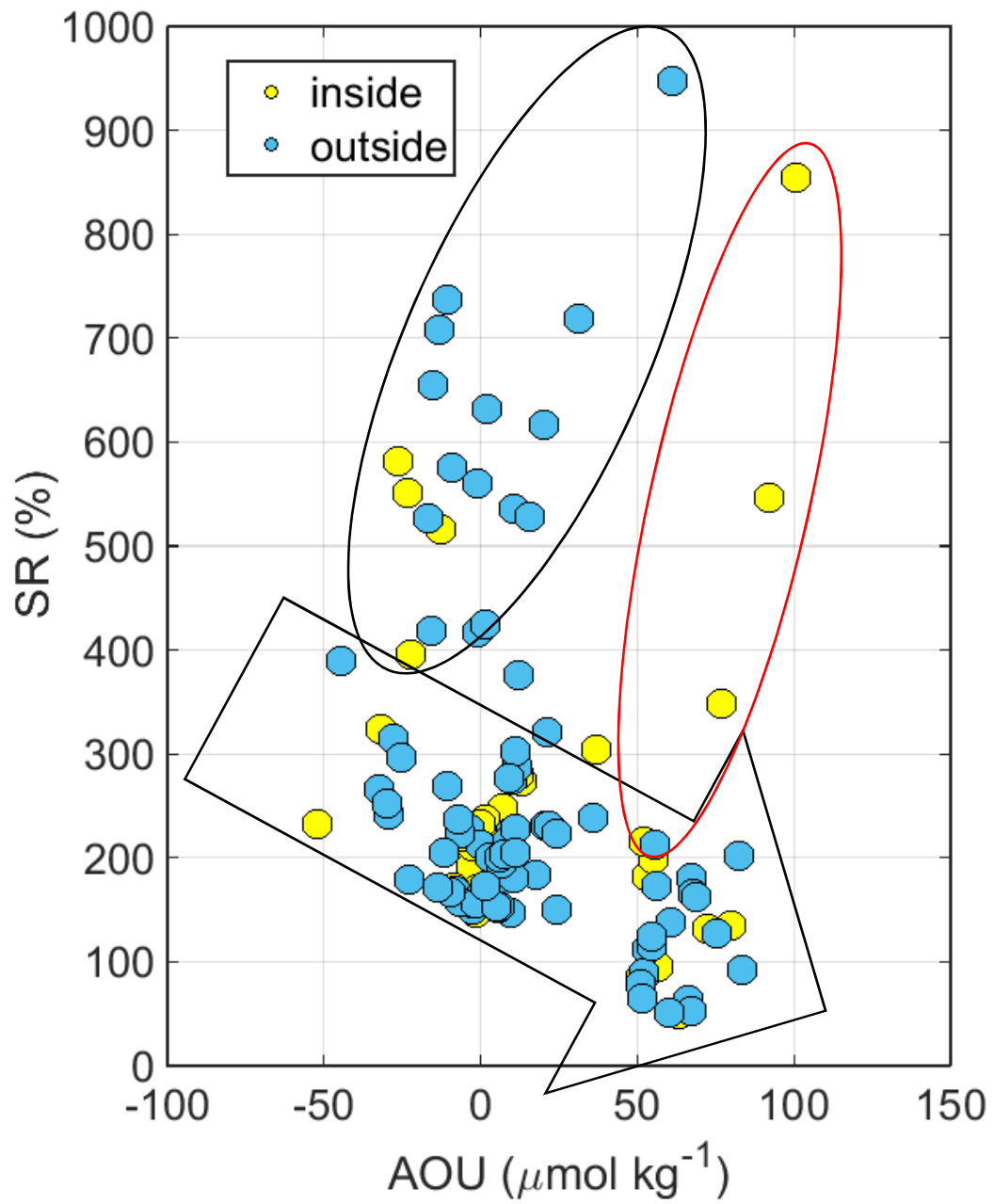


Fig. 4.14 Comparison of the relation between methane and AOU between within (yellow) and outside (blue) the warm-core eddy. Black and red ellipsoid point out for the coastal stations and at the water-core eddy (100 m depth only), respectively.

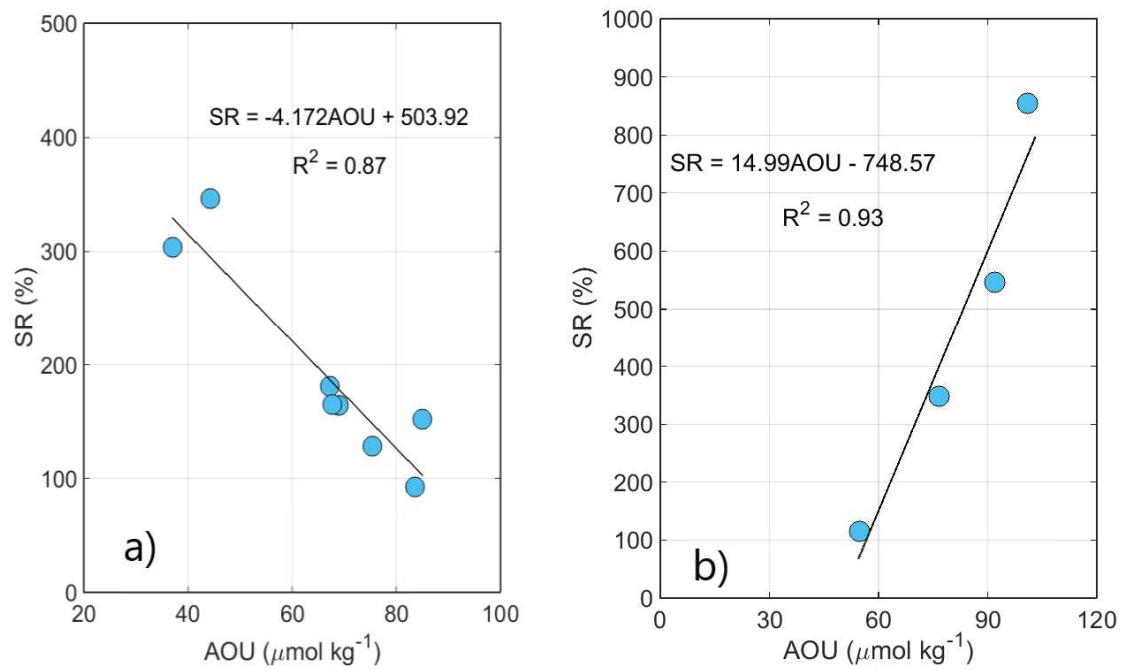


Fig. 4.15 Comparison of the correlation between CH_4 and AOU (a) outside and (b) inside the eddy at isoline of 32 ($\sim 26\sigma_\theta$) and N^{**} minima.

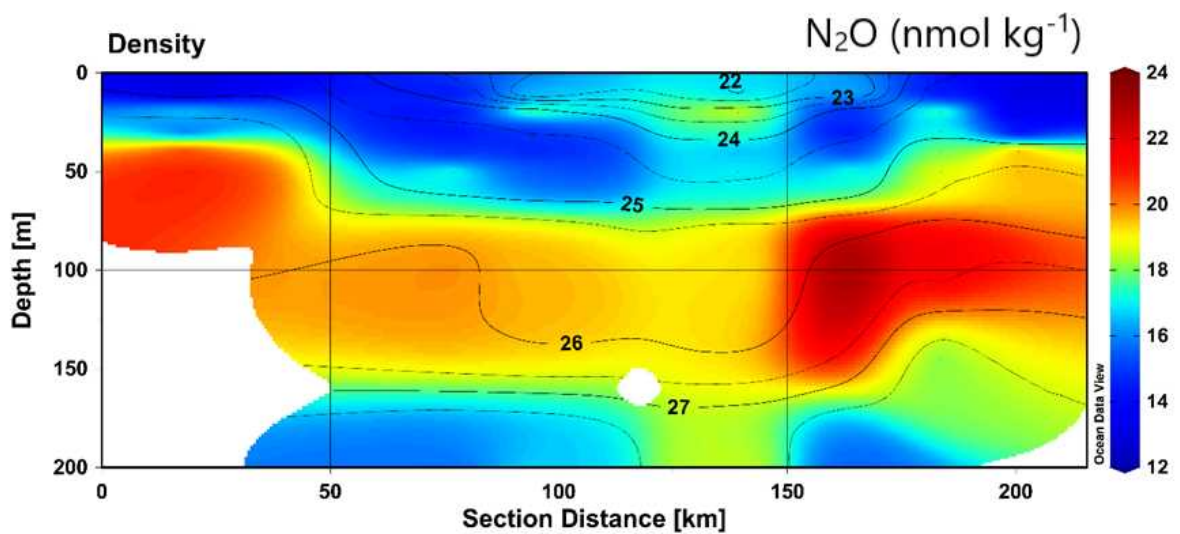


Fig. 4.16 Vertical sections of N_2O concentration (nmol kg^{-1}) (color) and potential density (contours). White background in both sides shows shelf area.

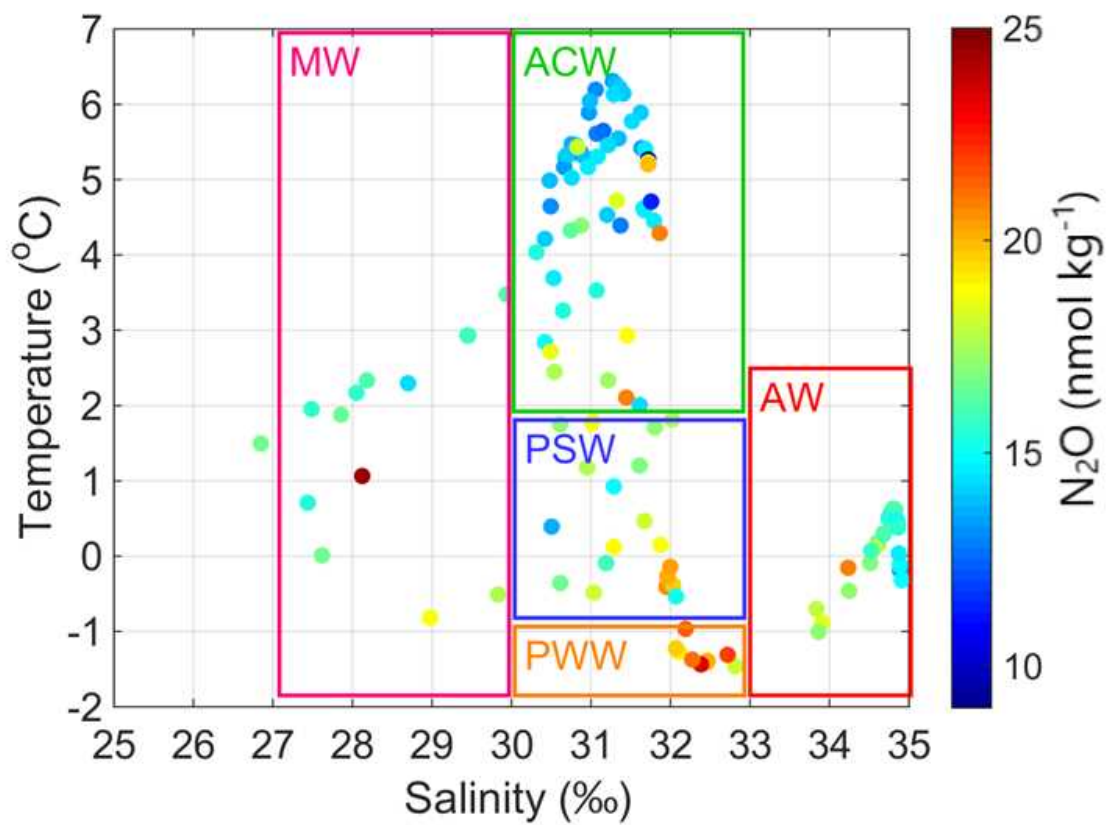


Fig. 4.17 Salinity vs. temperature plot showing the different water masses observed in the southwestern Canada Basin during the cruise. Colorful circles indicate for dissolved N₂O concentration.

Table 4.1 CH₄ concentrations and sea-air CH₄ flux in the Arctic Ocean

Region	CH ₄ concentration/Saturation	Flux ($\mu\text{mol m}^{-2} \text{d}^{-1}$)	Reference
East Siberian Arctic shelves (ESAS)		70-450 $\text{nmol m}^{-2} \text{s}^{-1}$	
ESAS	17.6 nmol L^{-1} ; 5000%		Kosmach et al., 2015
ESS		229.375	Shakhova et al., 2010; Sapart et al., 2017
Barents and Kara Seas	10.5 nmol L^{-1} ; 2200-2500%		Kosmach et al., 2015
Chukchi Sea & Bering Sea	8.1 nmol L^{-1} ; 1000%		Kosmach et al., 2015
Chukchi Sea		10.08	Li et al., 2017
Beaufort Sea shelf		8.75 – 26.87	Lorenson et al., 2016
Bering & Eastern Chukchi Sea	4.8 nmol L^{-1}	1.9 ± 1.4	Fenwick et al., 2017
Central Arctic Ocean (ice-covered)		35	He et al., 2013

Table 4.2 Observation of sea-air flux of CH₄ in the Arctic Ocean from the previous studies

Region	Flux ($\mu\text{mol m}^{-2} \text{d}^{-1}$)	References
East Siberian Arctic shelves (ESAS)	70-450 $\text{nmol m}^{-2} \text{s}^{-1}$	
ESAS		Kosmach et al., 2015
ESS	229.375	Shakhova et al., 2010; Sapart et al., 2017
Barents and Kara Seas		Kosmach et al., 2015
Chukchi Sea & Bering Sea		Kosmach et al., 2015
Chukchi Sea	10.08	Li et al., 2017
Beaufort Sea shelf	8.75 – 26.87	Lorenson et al., 2016
Bering & Eastern Chukchi Sea	1.9 ± 1.4	Fenwick et al., 2017
Central Arctic Ocean (ice-covered)	35	He et al., 2013

Table 4.3 Estimation of methane oxidation within the warm-core eddy with different oxidation rates.

Oxidation rate (nM day^{-1})	Initial CH ₄ concentration (nmol kg^{-1})	Expected CH ₄ concentration over time (7 days)	References
0.0912	35.8	18.91	Kitidis et al., 2010
0.03-0.09	35.8	19.06 - 29.01	Savvichev et al., 2007
	35.8 (at station 47)	32.30	This study

CHAPTER 5:

CH₄ DISTRIBUTION IN SEA-ICE COVERED AREA IN THE ARCTIC OCEAN IN SUMMERTIME

5.1 Dissolved CH₄ under sea ice cover in the Arctic Ocean

Arctic region is the most vulnerable area to global warming (IPCC, 2007). A rate of ice loss in magnitude increased from 0.35×10^6 km² per decade during the period 1979-1993 (Comiso, 2006) to 0.9×10^6 km² per decade between 1993-2007 (Deser and Teng, 2008). According to Stroeve et al. (2012), the extent of sea-ice loss during summer has become remarkable recently, so the area of open water has increased in summertime. Decline in the Arctic sea-ice extent may affect changes in large-scale atmospheric circulation (Overland et al., 2008) and its chemical composition (Comiso, 2010). Sea-ice reduction accelerates light penetration into the ocean water, so it may enhance phytoplankton photosynthesis and supply foods for higher trophic level organisms. In the western Canada Basin, diatom productivity was greater in 2004 than in 1994; it implied that the biological pump was enhanced during 1994-2004 due to the sea ice retreat (Nishino et al., 2009). Sea ice acts as a physical barrier, and removal of this barrier directly affects the air-sea exchange of greenhouse gases as well as the exchange of energy and water (Bates et al., 2006; Loose et al., 2014). Sea ice completely prevent air-sea exchange in winter, and a closed sea ice cover may trap bubbles reaching the surface water. Characteristics of physical oceanography such as water masses, geography and biogeochemistry which influence on CH₄ distribution in the Arctic Ocean have been described in Chapter 4. Thus, in this chapter, additional information of CH₄ distribution especially in sea-ice cover region is added for further understanding.

Some previous reviews have compared CH₄ flux between ice-free and ice-covered regions. On the East Siberian Arctic shelf, CH₄ fluxes beneath the sea-ice cover (multi-year ice) are 2-10 times higher than in ice-free region (Shakhova et al., 2010; Kitidis et al., 2010). CH₄ fluxes in both ice-free region and ice-covered region range from 0.02 Tg CH₄ yr⁻¹ and 0.30 CH₄ yr⁻¹, respectively (Lorenson et al., 2016) in the Beaufort Sea. It suggested that accumulation of CH₄ under ice could enhance the annual sea-air flux due to CH₄ release after melting in seasonally ice-covered regions (Lammers et al., 1995). Increase of atmospheric CH₄ concentration has been detected over the open leads and regions with fractional sea-ice cover (Kort et al., 2012). During period of sea ice formation, gases are expelled from the sea-ice and

transported to the deeper layer or reached to the atmosphere while during summer melting ice dilutes the gas content of seawater (Anderson et al., 2004; Kitidis et al., 2010; Damm et al., 2015a).

Arctic continental shelves are well-known as “Arctic carbon hyper pool” refer to the huge content of organic carbon within the sedimentary basin (Semiletov et al., 2012). High level of CH₄ emission have been observed on the East Siberian Sea (Shakhova et al., 2010) and in the upper continental slopes (Shakhova et al., 2010; Ruppel and Kessler, 2017). The produced CH₄ from sediments could be transported by the ocean currents and diluted into the oceanic water. In the central Arctic, super-saturation of CH₄ in the oceanic water was also observed and is supposed to aerobic CH₄ production in the water column via degradation of DMSP (Damm et al., 2010). Sea ice is a DMSP-rich water where abundance of precursors of CH₄ in the polar water. Sea ice processes have been identified as potential CH₄ sources associated with microbial activities in anoxic brine channels (Damm et al., 2015a). The sea ice zone could biologically produce CH₄ with depleted nitrate and available phosphate (Karl et al., 2008; Damm et al., 2010). A low N:P ratio enhances the ability of bacteria to compete for phosphate while the phytoplankton metabolite DMSP is utilized as a C source. During this process, methyl-phosphonate is converted into CH₄ without requiring anoxia (Karl et al., 2008). Hence, there are potentially important CH₄ cycle processes in the sea ice zone, however, relation of CH₄ with sea ice is not clear at this stage.

Most of studies on CH₄ emission in Arctic Ocean focused on shelf areas whereas the central Arctic Ocean has rarely been investigated (Damm et al., 2010; 2015) as a result of long-term ice cover. The effect of reduced sea-ice cover may affect the amount of CH₄ emission into the atmosphere. There are not as much as investigations on the CH₄ distribution in the sea ice covered area. The role of CH₄ distribution under sea ice cover has not been understood completely yet, thus comprehension of this role is needed to investigate to understand the gas exchange between sea-ice and atmosphere in the Arctic Ocean. In this study, I investigated CH₄ flux from the southern Bering Sea crossing the Chukchi Sea shelf to central Arctic Ocean under two conditions: beneath sea-ice cover and sea ice-free regions. This allowed us to gain some insights into the role of sea ice on CH₄ emission and budget of CH₄.

5.2 Experimental Setting

During the period of August 2016, seawater samples were collected from IBRV *ARAON* cruise (ARAON-07B) throughout the water column. The majority of sampling was conducted

over the Bering Sea, Chukchi Sea and Arctic Ocean (Fig 5.1) during the cruise in order to investigate the vertical profiles of dissolved CH₄ in the Arctic Ocean under ice-free region and sea-ice cover region. Vertical seawater samples were taken using 24 Niskin bottles (12L) mounted on a rosette, were used to make observation at 31 stations. At each station, a Sea-Bird CTD system (SBE 911+) equipped with ancillary sensors was used to measure salinity, temperature, and other hydrographic parameters. Over 130 seawater samples were collected for CH₄ analysis at 21 stations with various depths (from surface seawater to the bottom water). CH₄ samples have been measured by GC-IRMS and description of GC system has been mentioned in Chapter 2. Approximately 190 seawater samples for each DMS, DMSPp, and DMSPt were also collected at 31 stations from surface to about 300 m depth. DMS samples was immediately measured on board as soon as possible after sampling. DMSP samples were treated by adding acid solution (see Chapter 2) and stored in a dark place at room temperature. These samples were measured by GC-FPD on land at Hokkaido University, and description of GC-FPD measurement has also been described in Chapter 2.

This research aims to examine the vertical profiles of CH₄ and DMSP distribution in Bering and Chukchi Sea (coastal and ice-free region) and the central Arctic Ocean (beneath sea-ice). The cruise crossed main distinct water masses influenced by a range of physical and biological processes that impact salinity, temperature, and nutrients concentrations, such as PSW, PWW, ACW, MW, and AW (these water masses are mentioned in Chapter 4). An available CH₄ emission above seafloor sediments and comparison of CH₄ distribution between ice-free region and sea-ice cover region were investigated.

The atmospheric concentration of CH₄ was assumed as 1850 ppb, based on the average mixing ratios in Barrow, Alaska in 2015 (Data provided by NOAA ESRL Global Monitoring Division, Boulder, Colorado, USA). Weekly sea-ice concentration was downloaded from National Oceanic & Atmospheric Administration (NOAA), which was produced on 1°x1° grid (<https://www.esrl.noaa.gov/>).

5.3 Results and Discussion

5.3.1 Dissolved CH₄ distribution in the water column

Over the region, the dissolved CH₄ concentration ranged from 3.5-23.1 nmol kg⁻¹ and the average was 7.5 nmol kg⁻¹ in the surface seawater (at 0 m depth) (Fig 5.2a). Dissolved CH₄ was ubiquitously supersaturated (range: 124.7-604.6%; average: 216%) in the surface seawater

(Fig 5.4b) respecting to atmospheric CH₄ concentration. The highest CH₄ oversaturation were 605% (23.1 nmol kg⁻¹) in surface water at Chukchi sea shelf (sta #12).

At the Bering Sea and Chukchi Sea shelf, CH₄ concentration increased with depth to the bottom from station #1 to station #13. At the shelf break, CH₄ concentration was as high as 38.94 nmol kg⁻¹ (SR = 1019%), approximately 7 times higher than surrounding surface water (~5 nmol kg⁻¹; SR ~150%) (Fig 5.4). Away shelf water, at a few stations with depth greater than ~200 m, the CH₄ concentration increased with depth up to 50-10 m depth but decreased with greater depths. The lowest CH₄ concentration at 200 m was 0.96 nmol kg⁻¹ (SR = 27%).

High CH₄ supersaturation appeared in bottom water above sediment. Water contains SR of 1019% (38.94 nmol kg⁻¹) probably resulted from high *in situ* methanogenesis with organic matter in the sediment layer (Reeburgh, 2007) and then diffused the produced CH₄ into the water column. High CH₄ concentration at the shelf slope (station #12) was also observed (Li et al., 2017) with high organic matter settling over bottom layer (Nishino et al., 2005). Maximum CH₄ (35.8 nmol kg⁻¹) concentration upper the bottom water (sta. #12) in our observation was higher than that of Kudo et al. (2017) (20 nmol kg⁻¹) at the same position. Methanogenesis rate in the sediment of the Chukchi Sea was up to 67 μmol m⁻² d⁻¹ (Savvichev et al., 2007), implying that it may be a major source of CH₄ releasing into the water column.

Further the shelf slope, CH₄ concentration increased up to SCM layer (~40 m) and decreased with depth until the bottom. The lowest CH₄ value of 0.96 nmol kg⁻¹ (SR = 27%) at the deeper water. High CH₄ concentrations were above 100 m depth at a distance of 3000 km of transect may be transported to the northern Arctic by northward flowing of ocean currents of (e.g., ACW (Watanabe et al., 2014) or Anadyr water (AnW) with CH₄ concentration of 10.11 nmol kg⁻¹ (SR = 277.5%). At a distance of 1500 km of the transect, a plume of CH₄ was observed, which might be derived from East Siberian Coastal Water where high CH₄ concentration was observed (Shakhova et al., 2005).

5.3.2 Dissolved CH₄ in distinct water masses

Dissolved CH₄ concentration varied in different water masses. In this study, observed CH₄ concentrations were lower in Atlantic-derived water (AW) than in Pacific-derived water (Pdw, including PSW, PWW, ACW) (Fig 5.3b). The ACW contained high inputs of CH₄ from riverine water before it enters the Chukchi Sea and the Arctic. High CH₄ in the river water reflects the methanogenesis of organic matter in degrading permafrost (Striegl et al., 2012). On the other hand, Ruppel (2015) reported that subsea permafrost or gas hydrate did not affect the

elevated CH₄ here. This observed CH₄ concentration is consistent with the observation of Fenwick et al. (2017) and Li et al. (2017). In MW, CH₄ concentration was relatively medium and was less than 10 nmol kg⁻¹ excepting for a few points. CH₄ concentration is high in PWW (cold and saline) that is suggested by a brine rejection process in the winter. This process causes convective overturning and mixes up from the sediments, releasing regenerated nutrients and gases into the PWW (Pickart et al., 2016). Therefore, methanogenesis in the sediments may contribute to the high CH₄ concentration in the PWW and ACW.

Fig 5.5 shows CH₄ concentration was negatively correlated with sea water temperature, especially in the ACW ($\text{CH}_4 = -1.585 \times \text{Temp} + 19.92$, $R^2 = 0.51$, $n = 36$). This correlation explains high-CH₄ water in warmer water where derived river water containing high CH₄ production. Near the bottom-colder water or in the PWW (yellow circles), CH₄ concentration focused on at the cold water temperature, which derived from mixing of CH₄-bottom water on the shallow shelf water. In the cold water, temperature did not limit the production and emission of CH₄ was suggested in early research. In contrast, MW including meltwater and river water contained relatively high CH₄ concentration (5-10 nmol kg⁻¹) but there was no correlation between them.

No correlation between CH₄ concentration and salinity in whole water column has been found (Fig 5.6) compared to its correlation in each water mass. However, CH₄ looks like positively correlated to salinity in the MW; it will be discussed later on.

5.3.3 Sea surface CH₄ distribution in open water and sea ice cover region

Sea-ice presents as a barrier for restriction at sea-air interface and variation in the sea-air flux of CH₄ (Loose et al., 2014). In summer when sea ice melting, CH₄ from temperature destabilization of gas hydrates on the continental shelf (Westbrook et al., 2009) and trapped CH₄ gas beneath sea-ice can release into the atmosphere.

Basing on distribution of sea ice coverage, I divided research area into two sub-regions: shelf water and sea ice cover region. Shelf water covered from station #1 to station #11 and station #16. Especially, CH₄ plume on the shelf water includes the stations from #3-8. Sea ice cover region was the rest of stations, including CH₄ plume at station #12. The averaged CH₄ in the shelf water was lower than in sea ice cover region with concentration of 6.06 nmol kg⁻¹ (SR = 199%) and 7.38 nmol kg⁻¹ (SR = 231%) (excluding station #12), respectively (Fig 5.7). In this study, two assumptions of CH₄ supersaturation in the sea ice cover is supposed as follows:

(1) CH₄ plume originating from nearby sediments was advected or mixed up into the water column and then the CH₄ was accumulated under sea-ice (Kitidis et al., 2010; Lorenson et al., 2016) or

(2) Due to mineralization of organic matters derived from sea ice algae (Rysgaard and Glud, 2004) which can act as a substrate for methanogenesis. Methanogenesis is detected during zooplankton grazing (DeAngelis and Lee, 1994) and phytoplankton blooms (Damm et al., 2008).

Not only those, high CH₄ was also observed under a multi-year sea-ice or in the marginal ice zones (MIZ) (Kitidis et al., 2010) implying that sea ice cover could influence the exchange of CH₄ between seawater and air. Appearance of sea ice directly influences on seawater temperature and salinity. Due to high CH₄ concentration in the sea ice cover, the correlation between CH₄ and SST at surface seawater is considered. No clear correlation between them was found at the shelf water (blue circles) but CH₄ concentrated at low sea water temperature in the sea ice cover (red squares) (see Fig 5.8).

The difference in salinity responses to CH₄ concentration results in different local sources of CH₄ (Yoshida et al., 2011; Fenwick et al., 2017) in surface seawater. In this study, CH₄ concentrations was positively agreed to sea surface salinity (SSS) (Fig 5.9) with two distinct correlations. In the shelf water, the positive correlation is found ($\text{CH}_4 = 2.92 \times \text{SSS} - 86.02$, $R^2 = 0.6$, $n = 10$) at SSS was higher than 30 (30-33‰) which contains the Pacific-derived water and CH₄ sources is suggested from methanogenesis of organic matter. The produced CH₄ upper sediments is stirred up with higher salinity by vertical mixing. Then it was dispersed and diluted in the water column. In the sea ice cover region, I found the positive correlation between CH₄ and SSS ($\text{CH}_4 = 1.14 \times \text{SSS} - 25.45$, $R^2 = 0.68$, $n = 9$) at SSS was lower than 30 (‰), meaning fresh and low density surface water. CH₄ was generally supersaturated in this water. In this region, low salinity water may result from rivers or ice melting. Rivers input could also be supplied CH₄ from methanogenesis in sediments Chukchi sea shelf at higher salinity whereas at lower salinity CH₄ could be supplied from ice melting. Because sea ice contains anaerobic brine channels, which can contain high concentration of CH₄ due to decomposition of organic matter (Damm et al., 2015). CH₄ from open water region is transported and then is accumulated beneath sea-ice is immediately released when sea ice cracked (Damm et al., 2015b). In recent, sea ice contains rich-DMSP water (Asher et al., 2011; Stefels et al., 2018) especially sea ice melting directly impacts DMS(P) distribution. As the sea ice begins to melt, this brine extracts into the surface layer of the ocean. Observed CH₄ belongs to MW water mass including river

water and melting water (Fig 5.9). The slope is greater at the shelf water than in the sea-ice cover, implying that CH₄ production is faster at higher salinity than at low salinity. However, Fenwick et al. (2017) have been observed inverse correlation between CH₄ and salinity at the Canadian Arctic Archipelago. The supersaturated CH₄ observed in the surface water could be produced in local melting of sea ice as Damm et al. (2015a) suggested methane excess in Arctic surface water enhanced by sea ice formation and melting. This study is generally consistent with Damm et al. (2015b) that sea ice acts as a CH₄ source and a semi-permeable barrier to sea-air.

5.3.4 Correlation of CH₄ and nutrients in seawater

The Arctic Ocean water comprises both low salinity, nutrient-rich water and relatively nutrient-poor, more saline water (Pabi et al., 2008). In the Chukchi Sea shelf, high ammonium concentration is produced during summer by decomposing of organic matter at the place where high organic matter is deposited at the bottom (Nishino et al., 2005). The correlation between CH₄ and nutrients is collected at 0-50 m depth.

In summer, chl-*a* concentration increased approximately at the Bering and northern Chukchi sea shelf with the maximum greater than 7.0 mg m⁻³. The correlation between CH₄ and chl-*a* in the shelf water was not found ($\text{CH}_4 = 11.19 \times \text{chl}a + 6.41$, $R^2 = 0.48$) so strongly as in the sea ice cover region ($\text{CH}_4 = 4.82 \times \text{chl}a + 6.10$, $R^2 = 0.48$).

Over the shelf-water, high ammonium concentration increased at the bottom (5.8 μmol kg⁻¹) (Fig 5.11) while ammonium concentration was low (~0 μmol kg⁻¹) in high latitudes. This implied that high ammonium concentration could be expected high organic matter and produced CH₄ through methanogenesis by decomposing of organic matter in anoxic sediments. Thus, the correlation between ammonium and CH₄ was found ($\text{CH}_4 = 3.09 \times \text{NH}_4 + 9.52$, $R^2 = 0.65$) at the shelf water (Fig 5.12) and in the sea-ice cover region ($\text{CH}_4 = 12.38 \times \text{NH}_4 + 13.90$, $R^2 = 0.55$).

Bering and Chukchi water derive water of Pacific which is continuously depleted in nitrate relative to phosphate (Yamamoto-Kawai et al., 2006). Nitrate + nitrite concentrations increased with depth. Unfortunately, there was not data of nitrate and nitrite separately. Total concentration were maximum at the shelf slope (1000 km distance of the transect) approximately 18 μmol kg⁻¹ (Fig 5.11). The correlations between CH₄ and total nitrate + nitrite concentration were found due to the existence of nitrite in the shelf water and ($\text{CH}_4 = 9.63 \times (\text{NO}_2 + \text{NO}_3) + 9.33$, $R^2 = 0.65$) and in the sea ice cover region ($\text{CH}_4 = 1.36 \times (\text{NO}_2 +$

NO_3) + 8.11, $R^2 = 0.87$). Nitrate + nitrite concentration in the sea ice cover was generally lower than in the shelf water. For phosphate, at the shelf water phosphate concentration was as high as $\sim 2.5 \mu\text{mol kg}^{-1}$ upper the sedimentary layer and further offshore phosphate was high at around 100 m depth (Fig 5.11). The positive relationships between CH_4 and phosphate were found at the shelf water ($\text{CH}_4 = 24.19 \times \text{PO}_4 - 2.35$, $R^2 = 0.81$) and in the sea ice cover region ($\text{CH}_4 = 14.89 \times \text{PO}_4 - 1.5$, $R^2 = 0.87$) (Fig 5.12). Tight correlations between CH_4 and phosphate in both regions mean that excess phosphate is available for bacterial biomass production produced DMSP and possible source to produce CH_4 biologically. Phosphate availability may influence the utilization of DMSP and produce CH_4 as a precursor (Damm et al., 2008; 2010). Relation between CH_4 and DMSP concentration should be considered.

In the surface layer, low nitrate concentration is probably attributable to the inflow of Bering shelf water with relatively low N^{**} values and the biological utilization during the inflow of the Pacific water into the Arctic Ocean through Bering Strait. A broad continental shelf of Bering and Chukchi Sea is denitrification sites, and minimum values of N^{**} was observed (Codispoti et al., 2005a; Gruber and Sarmiento, 2009). In particular, minima N^{**} ($-17.9 \mu\text{mol kg}^{-1}$) correlated to maxima nutrients of PO_4^{3-} : $2.51 \mu\text{mol kg}^{-1}$ (Fig 5.13) at the shelf slope where CH_4 concentration was maxima. The similarity of minimum N^{**} results in the Bering, Chukchi sea shelf water, and further Arctic has been reported at the same depth (Devol et al., 1997; Yoshikawa et al., 2006; Hirota et al., 2009; Hioki et al., 2014).

Apparent oxygen utilization (AOU) was high ($> 151 \mu\text{mol kg}^{-1}$) (Fig 5.11) near the bottom layer at the shelf slope indicated low dissolved oxygen (DO) concentration even though it still maintained aerobic biological activities. High CH_4 was obtained at high AOU that indicates anaerobic environments where CH_4 could be released from the sediment into the water column. There were no clear correlations between AOU and CH_4 at the shelf water and in the sea ice cover region in the whole region. However, in the CH_4 plume on the shelf water near Point Hope, CH_4 relatively correlated to AOU ($\text{CH}_4 = 0.096 \times \text{AOU} + 14.59$, $R^2 = 0.6$) at the station #2-7 whereas it scattered at the Chukchi shelf slope station #12. In contrast, at the station #18 and #19 this water contains high CH_4 and negative AOU. At the negative AOU, it was not a condition for producing CH_4 , so the CH_4 may be resulted in water mixing derived from the East Siberian Sea before arriving here. In short, CH_4 relatively correlated with nutrients, especially with phosphate within Pacific-derived water. High concentrations of CH_4 at the shelf water were consistent with high organic matter settling above the sea bed.

5.3.5 Distribution of Nitrous oxide (N₂O)

Nitrous oxide concentration in the Arctic Ocean ranged from 8.5 to 78.9 nmol kg⁻¹ (Fig 5.14). In surface water, N₂O concentration was almost supersaturated (range: 51.6-633.7%, averaged saturation: 132%) respecting to atmospheric N₂O, with maximum concentration up to 21.4 nmol kg⁻¹ (Fig. 5.5). High N₂O concentrations were at the Chukchi Sea shelf averaging of 140-150% saturation whereas higher N₂O concentration (~400%) (or N₂O “hotspot”) from East Siberian Sea might be transported into this region by lateral transport observed at the stations #18 and #19 due to no existence of high N₂O at the bottom water. At deeper water, N₂O decreased with depth (> 200 m) and was under-saturated (~50% saturation) comparing to shallow water. This observation is relatively higher than the recent observation of Zhan et al. (2015) and Fenwick et al. (2017), who reported an average dissolved N₂O concentration of 12.5 and 13 nmol kg⁻¹ in the same region, respectively. N₂O concentrations were similar to the previous observations of Fenwick et al. (2017) and Kitidis et al. (2010). N₂O saturation at the shelf slope (123%) was higher in the sea ice cover (112%).

Δ N₂O is determined by difference between observed N₂O and equilibrium N₂O concentrations. In this study, positive Δ N₂O located on the negative N^{**} implied the N₂O productions through denitrification in this region. Zhang et al. (2015) and Fenwick et al. (2017) strongly suggested that the N₂O production from the sediment of the Bering Sea and Chukchi Sea are a primary source in the Arctic Ocean. However, the positive Δ N₂O concentration was at the positive N^{**} shows not only denitrification work as a N₂O source in Atlantic origin water.

5.3.6 DMSP and Precursor of CH₄ in the Arctic

5.3.6.1 Distribution of total DMSP (DMSPt), DMSPd and DMS concentrations

During the cruise in summer (August 2016), high DMSPt concentrations were observed in the Chukchi Sea ranged from 0 - 90.8 nmol L⁻¹ in the whole region with an averaging of 8.2 nmol L⁻¹. At the depth of about 250-300 m depth, DMSPt concentration is very low and almost zero. DMS concentration ranged from 0 to 8.87 nmol L⁻¹, averaging of 0.71 nmol L⁻¹ from surface seawater to about 300 m depth. In the surface water, DMS concentration ranged from 0.195 to 7.99 nmol L⁻¹ with an averaged value of 1.27 nmol L⁻¹. In this observation, high DMS concentration in the Chukchi Sea shelf was up to 8.87 nmol L⁻¹. DMS concentration from our observation is lower than that of Sharma et al. (1999) in off Alaskan Coast and western Canada Basin (range: 2-23 nmol L⁻¹, average: 12.1 nmol L⁻¹). High DMS concentration over there is caused by high productive shelf region (Gosselin et al., 1999) as well as high organic matter

settling (Nishino et al., 2005). DMSP may convert into DMS in the water column via photochemical and bacterial processes (Kiene et al., 2000; Stefels et al., 2007) and produce another product as dimethylsulphoxide (DMSO).

In the water mass in the Arctic, high DMS concentration lied in the warmer and less saline water ($T > 0^{\circ}\text{C}$, $S = 31\text{-}33 \text{ ‰}$) or Pacific-derived water (Fig 5.17; Fig 5.18; Fig 5.19) whereas low DMS concentration located on the cold water such as AW, PWW, and MW. In addition, sea surface DMSPt generally increased to sea ice concentration, but DMS varied in the sea ice cover region (Fig 5.20). High concentration of DMSPt at the sea ice cover has been reported by Asher et al. (2011) and Brabant et al. (2011).

5.3.6.2 Precursors of DMSP

DMSP is one of the methylated compounds produced by variety of marine phytoplankton (Keller et al., 1989). DMSP plays a significant role in marine surface water (Kiene et al., 2000) because of DMSP accounting for 1-10% of total phytoplankton carbon. Thus, the correlation between DMSP and chl-*a* concentration is needed to examine in the polar mixed water (<50 m depth). High DMSPt concentrations were consistent with degradation of chl-*a* concentration over shelf water whereas DMSPt concentration increased with enhancement of chl-*a* concentration in the sea ice cover region (Fig 5.21; Fig 5.16). Weak negative correlation between DMSPt and chl-*a* concentration was found in the shelf water ($R^2 = 0.28$), while tighter positive correlation was stronger in the sea ice cover ($R^2 = 0.65$) even though those concentration is smaller 3-4 times than that of in the open water. Observed chl-*a* may result from ice melting release when trapped in ice formation in winter. Phytoplankton growth is restricted to the short melting period in spring and summer in a seasonal ice cover. At the ice edge zone (IEZ), region of seasonal melting is known as an area of intensive biological activity (Smith and Nelson, 1985).

DMS in the surface mixed layer is exposed to removal processes includes photochemical oxidation, micro-biological degradation, ventilation to the atmosphere, advection and vertical mixing. DMS in surface water appears to be primarily controlled by the biological cycle of DMS and DMSP (Kiene and Bates, 1990; Leck et al., 1990). Global DMS estimation from the ocean range from 250-1600 Gmol S yr⁻¹ (Spiro et al., 1992), which accounts for about 50% from ocean, plants, and soils (Bates et al., 1992).

5.3.6.3 Precursor of CH₄

Over the shelf water, high DMSP, chl-*a* (>2 mg m⁻³) and CH₄ concentrations were observed near the CH₄ hotspot. Phytoplankton is believed to be a main source of DMSP and DMS (Challenger and Simpson, 1948) and CH₄ (Damm et al., 2008) in the ocean as well. Kudo et al. (2018) suggested DMSP as a source of CH₄ in the Arctic Ocean. In the shelf water, DMSPt degradation correlated to CH₄ production except for the stations high CH₄ release upper the sedimentary layer. It is shown in the negative correlation between CH₄ and DMSPt ($\text{CH}_4 = -0.52 \times \text{DMSPt} + 25.9$, $R^2 = 0.58$) (Fig 5.22a). The finding equation implies that CH₄ production increased when DMSPt degraded. *In situ* excess here CH₄ could produced during the phytoplankton bloom while DMSP and its degradation products acts as potential source for CH₄. The degree of oligotrophic water in the Arctic and the decreasing DMSPt concentration reflect an increasing utilization of DMSP as a carbon source (Damm et al., 2010). In contrast, in the sea-ice cover region a positive correlation between CH₄ and DMSPt ($\text{CH}_4 = 0.44 \times \text{DMSPt} + 5.38$, $R^2 = 0.67$) indicates that increase in CH₄ production correlated to increase in DMSPt production (Fig 5.22b). The two correlations between and CH₄ and DMSPt concentration reflect two different potential sources of CH₄ over Chukchi Sea shelf water containing high nutrient concentrations and organic matters and in the sea ice cover. The excess phosphate water and the presence of sea ice ability (see section 5.3.3) influence on DMSPt production and CH₄ production as well. Thus, in the sea ice cover region, both sea ice and DMSPt work as *in situ* potential CH₄ sources.

5.4 Summary and Conclusion

CH₄ distribution in the Arctic Ocean was investigated under two conditions: in the shelf water and sea ice cover region in order to evaluate the influence of sea ice coverage on CH₄ contribution and CH₄ emission. During the cruise of R/V *ARAON* (ARAON-07B) in the Arctic Ocean in summer 2016, seawater samples of CH₄ from the surface seawater to the bottom were collected and analyzed. CH₄ was obviously supersaturated in the surface water over the region from 124-604%. High CH₄ concentration (1019%) was observed in the Chukchi shelf water close to CH₄ hotspot release from the anaerobic environment in sediments into the water column, which is linked to the positive AOU upper. The correlations between CH₄ and ammonium, nitrite+nitrate, phosphate and N^{**} concentration reflect the high CH₄ production co-existence to high organic matter. Beneath the sea-ice cover region at deeper water (> 200 m depth), low CH₄ concentration in deep water was controlled by CH₄ oxidation in aerobic

environmental water. CH₄ characteristic is clearly different in each water mass (ACW, PSW, PWW, MW, AW).

In the sea ice cover, surface seawater CH₄ concentrations in the sea ice were lower than that of over the shelf water. High CH₄ at surface seawater in the sea ice cover may be trapped by northward flowing into the Arctic Ocean advected and accumulated under sea ice cover. On the other hand, mineralisation of organic matter in sea ice brine channel can supply for producing CH₄ inside sea ice and expel into seawater or the atmosphere when ice melting. However, CH₄ generally increases with sea ice concentration suggests CH₄ can be produced in local melting or formation of sea ice. This study clarifies sea ice act as the semi-permeable barrier to sea-air exchange.

DMSP and CH₄ production is directly affected by phytoplankton bloom, which is presented by chl-*a* concentration, was considered as their sources. DMSP is examined as the precursor of CH₄ with the two distinct correlations under two conditions was found. I supposed that DMSP is considered to be a precursor of CH₄ in the Arctic Ocean and directly affect to amount of CH₄ emission into the atmosphere, especially in the sea ice cover region.

High N₂O near the bottom water in the Chukchi shelf water and existence of N₂O “hotspots” in the sub-surface water and off-shore water originated with lateral transport from East Siberian Sea or a branch of Anadyr Water with remarkably high N₂O concentration because “hotspots” of N₂O did not link to the bottom water. The N₂O data is used to trace out the mechanism of CH₄ sources and transport.

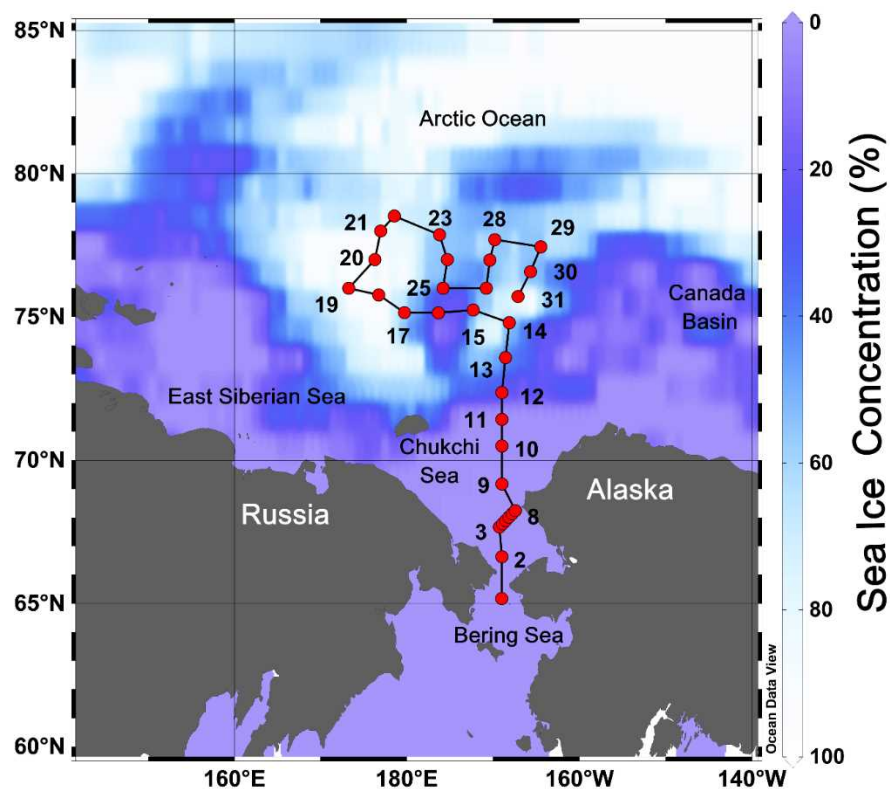


Fig 5.1 Sampling stations of cruise track of IBRV *ARAON* (ARAON-07B) on August 2016 in the Arctic Ocean. Discrete seawater samples is shown in red circles. Sea ice concentration (%) is shown in white. This figure is plotted by Ocean Data View.

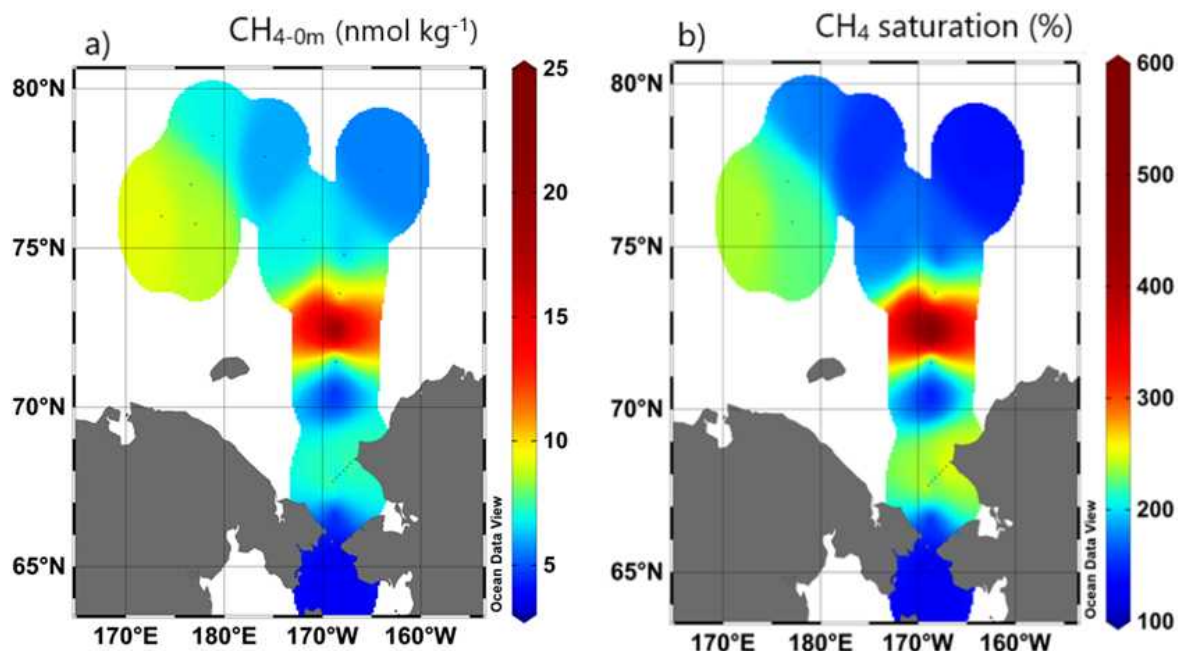


Fig 5.2 The spatial distribution of CH_4 concentration (nmol kg^{-1}) (a) and CH_4 saturation (%) (b) in surface water in the Arctic Ocean during the cruise of IBRV *ARAON* (ARAON07B) August, 2016. The figure is plotted by Ocean Data View.

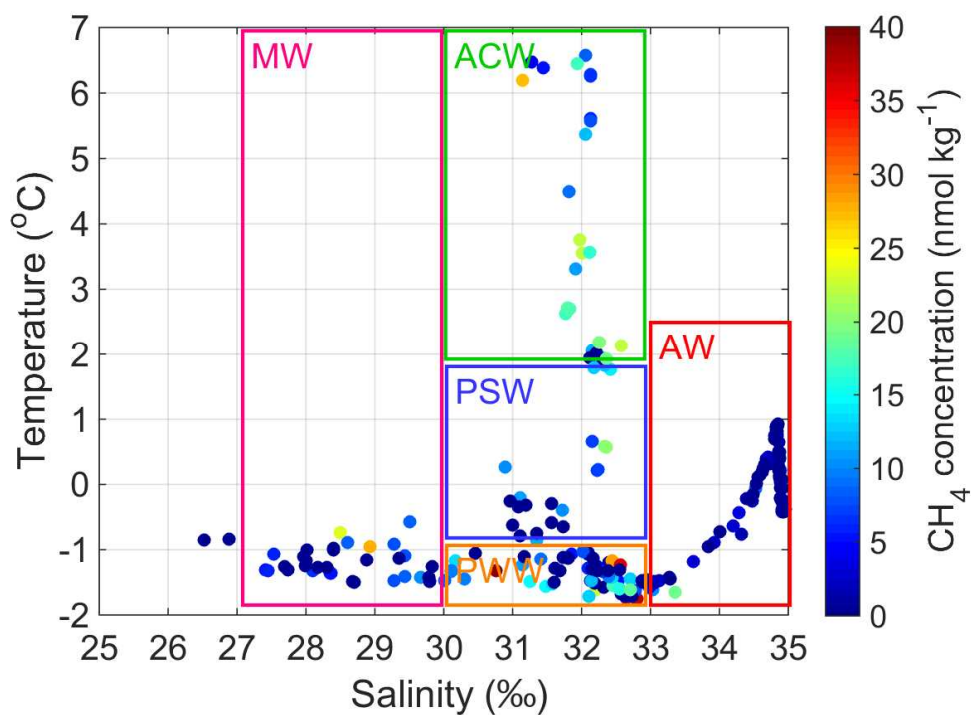
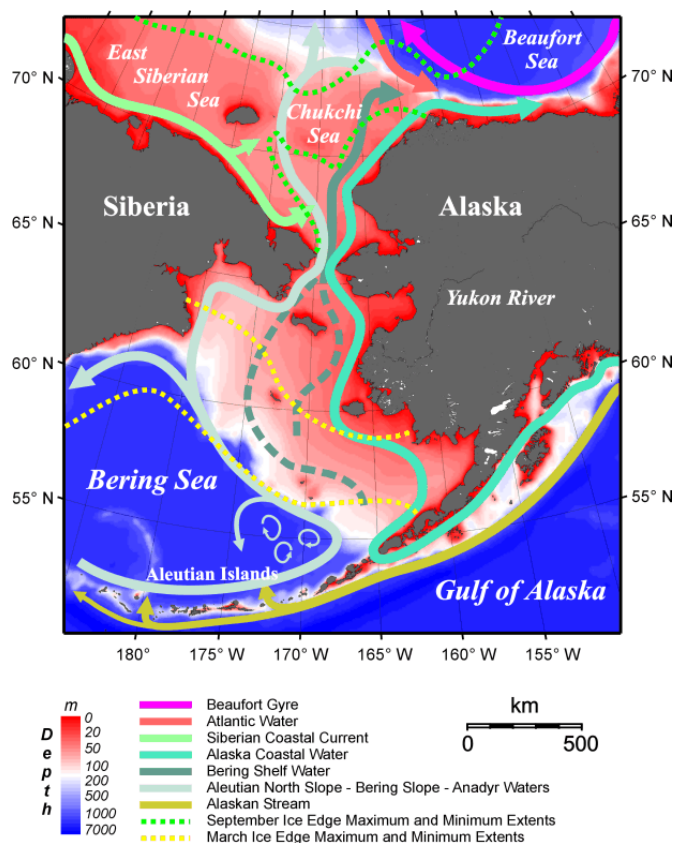


Fig 5.3 (a) Oceanic currents in the Bering, Chukchi and Arctic Ocean (Source: <http://research.cfos.uaf.edu/nprbdrifters/>). (b) Salinity vs. potential temperature plot showing the different water masses observed in the Arctic Ocean. Figure shows CH_4 concentration. This figure is plotted by MATLAB.

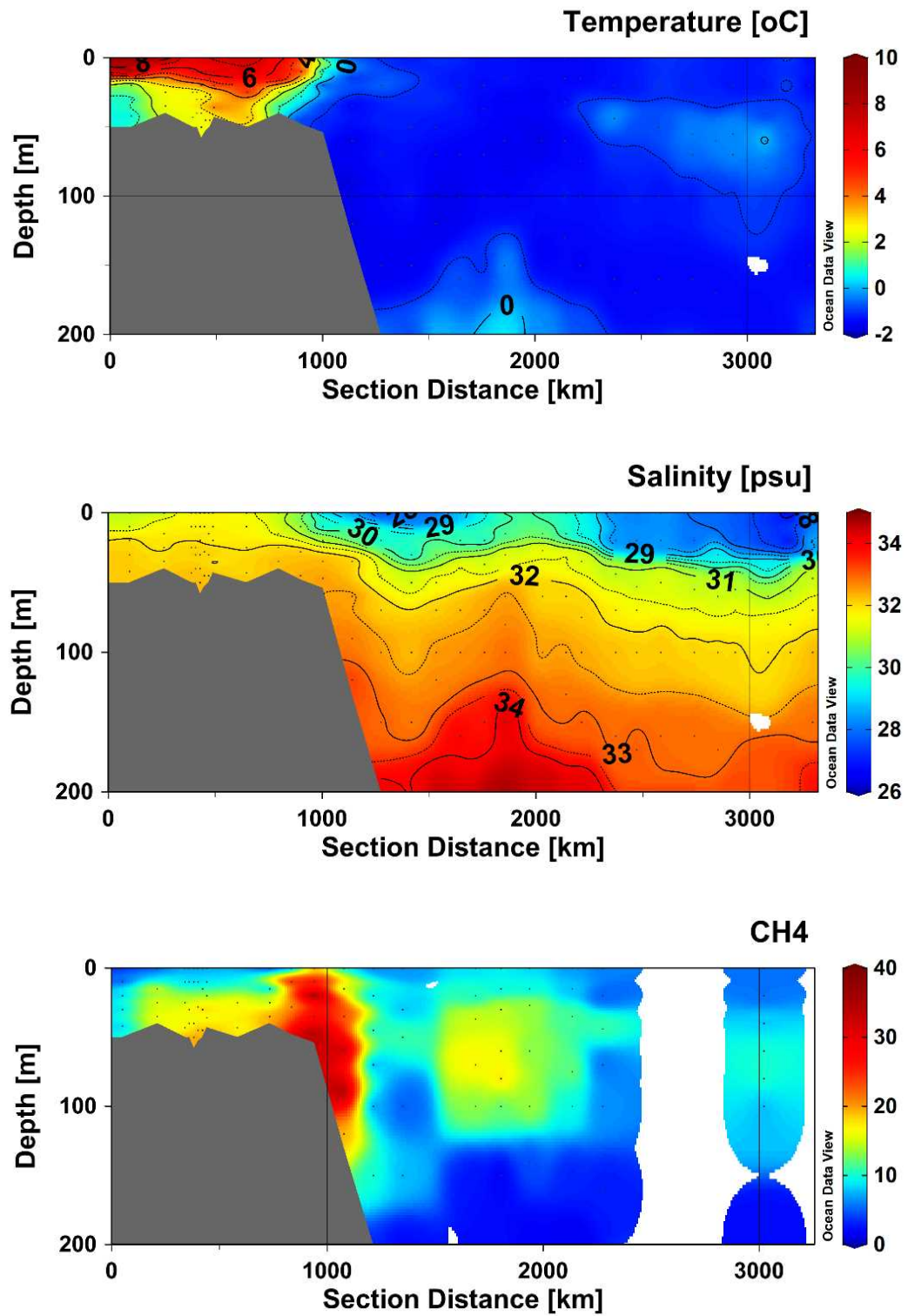


Fig 5.4 Vertical transect distribution of temperature (°C), salinity (‰) and dissolved CH₄ concentration following in order stations (1-31).

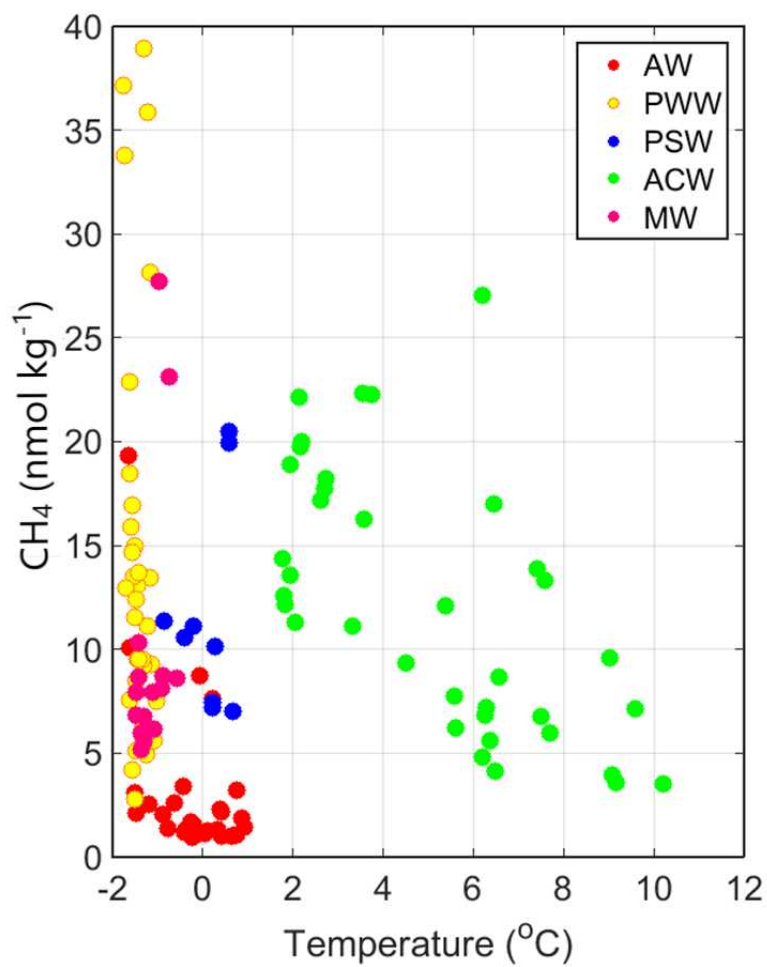


Fig 5.5 Plot of CH₄ concentration versus temperature within each water masses in the Arctic Ocean (AW, PWW, PSW, ACW, MW).

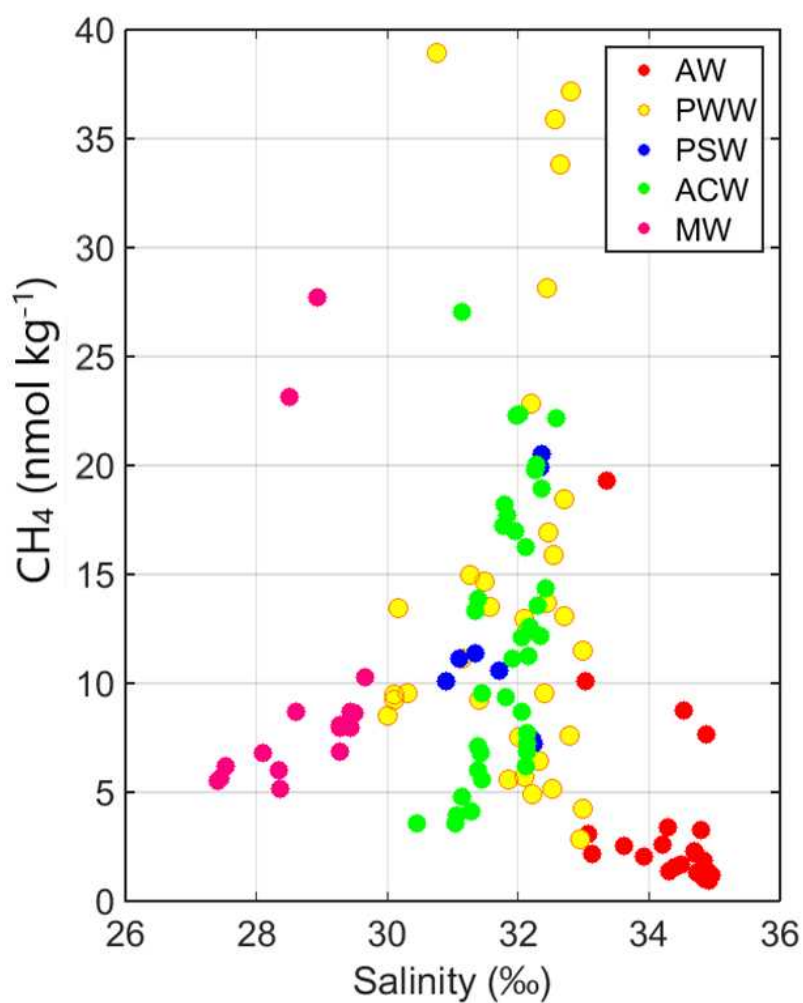


Fig 5.6 Plot of CH₄ concentration versus salinity within each water masses in the Arctic Ocean (AW, PWW, PSW, ACW, MW).

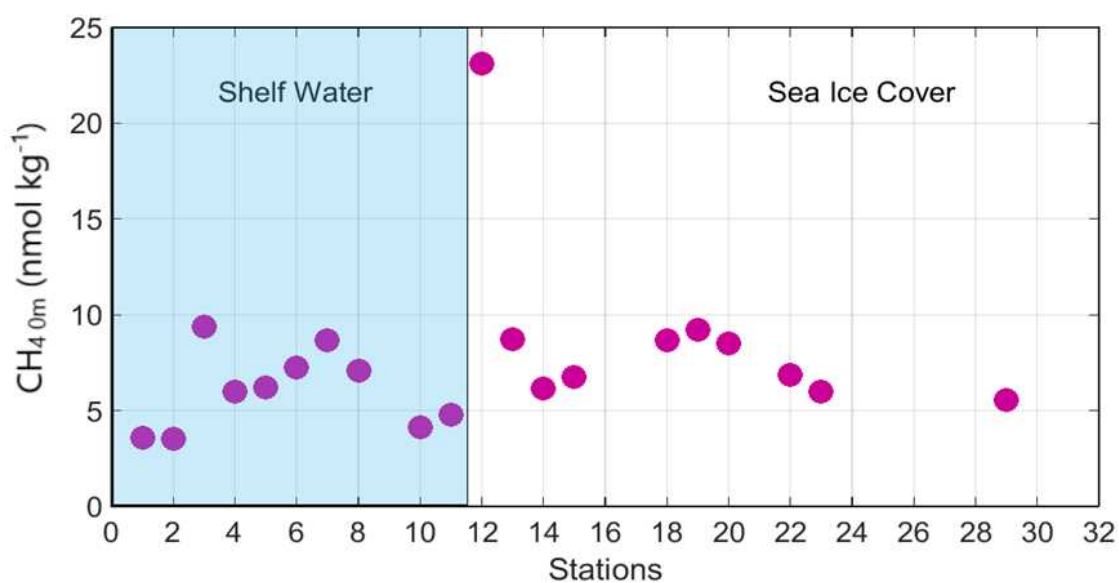


Fig 5.7 The distribution of CH₄ concentrations in the shelf water and in the sea ice cover region

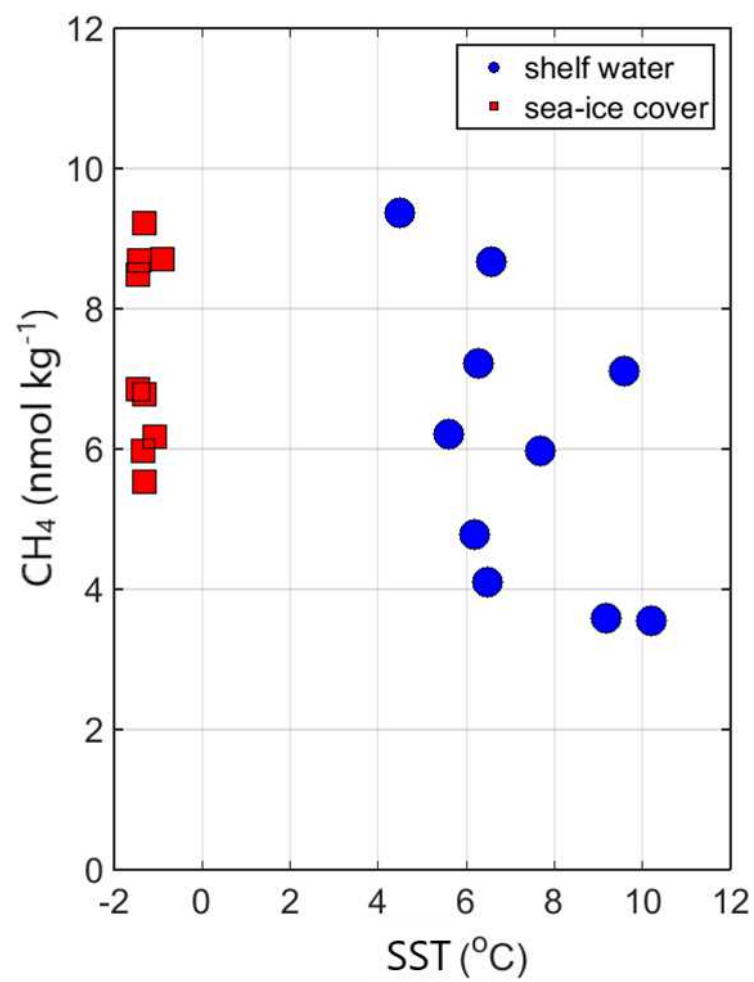


Fig 5.8 Plot of CH_4 concentration versus SST in the shelf water (blue circles) and sea-ice cover region (red-squares).

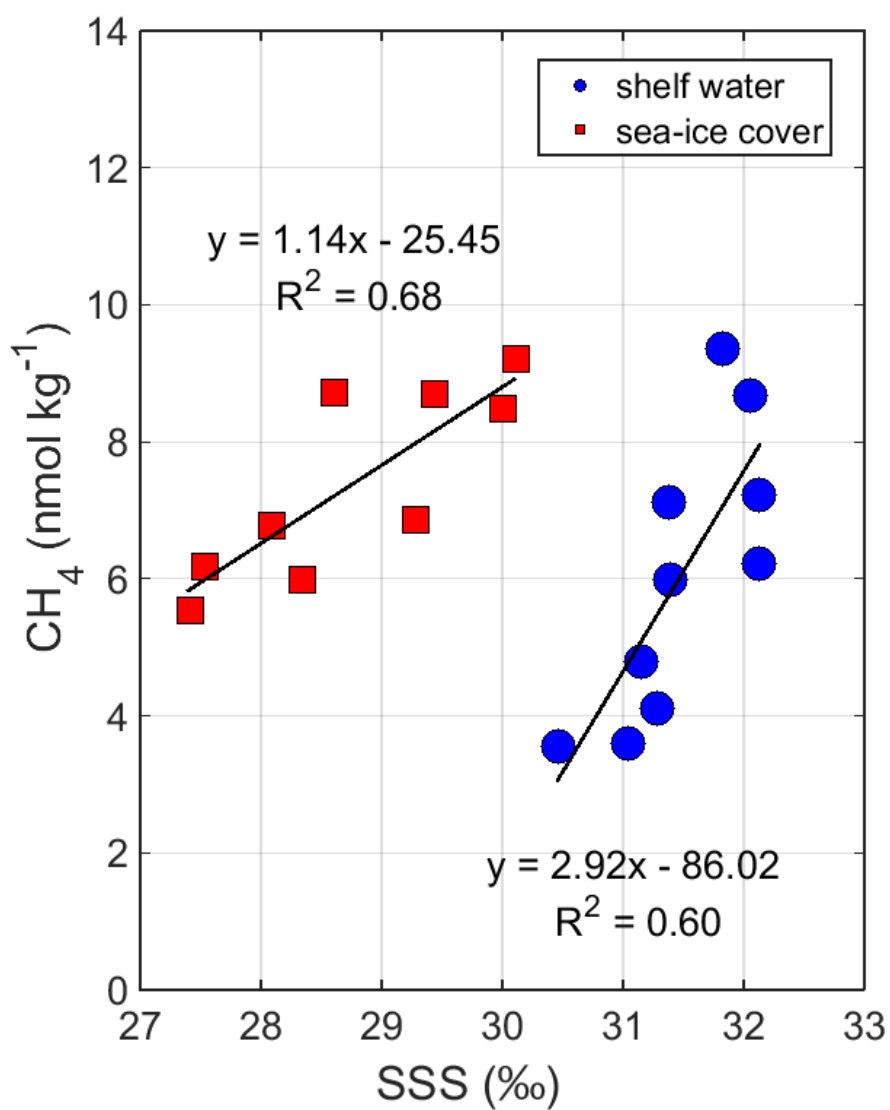


Fig 5.9 Plot of CH₄ concentration versus salinity in the shelf water (blue circles) and sea-ice cover (red-squares).

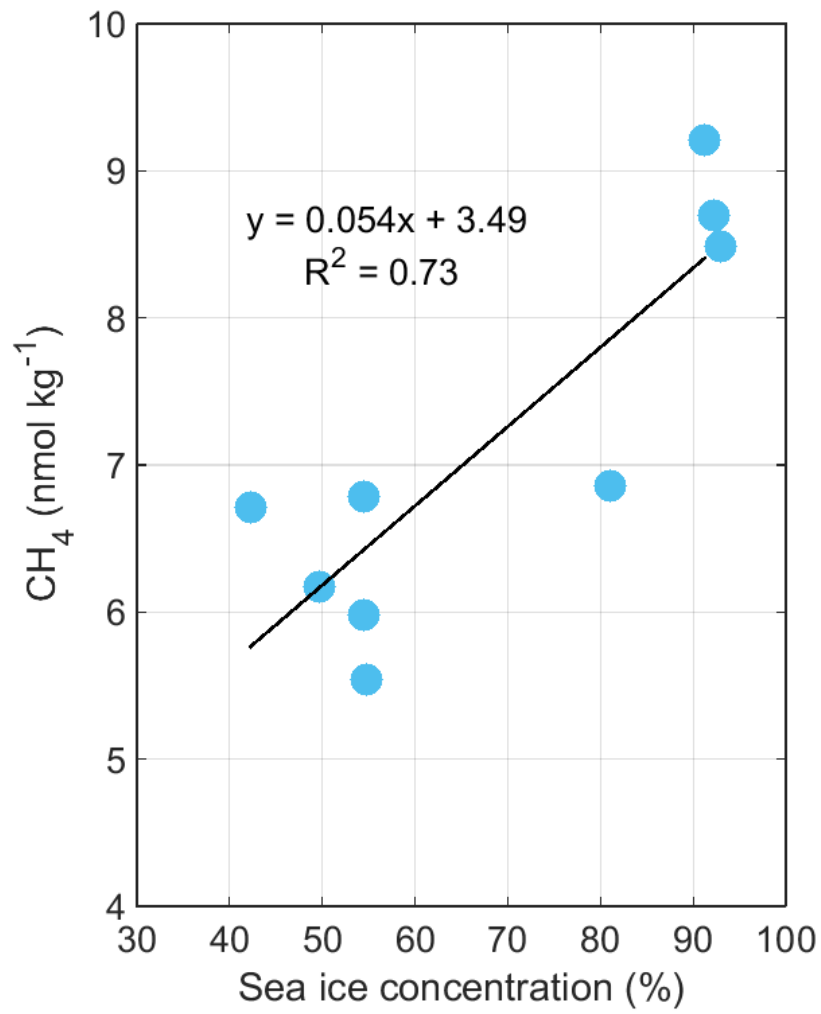


Fig 5.10 Correlation between CH₄ and sea ice concentration in the sea ice cover region.

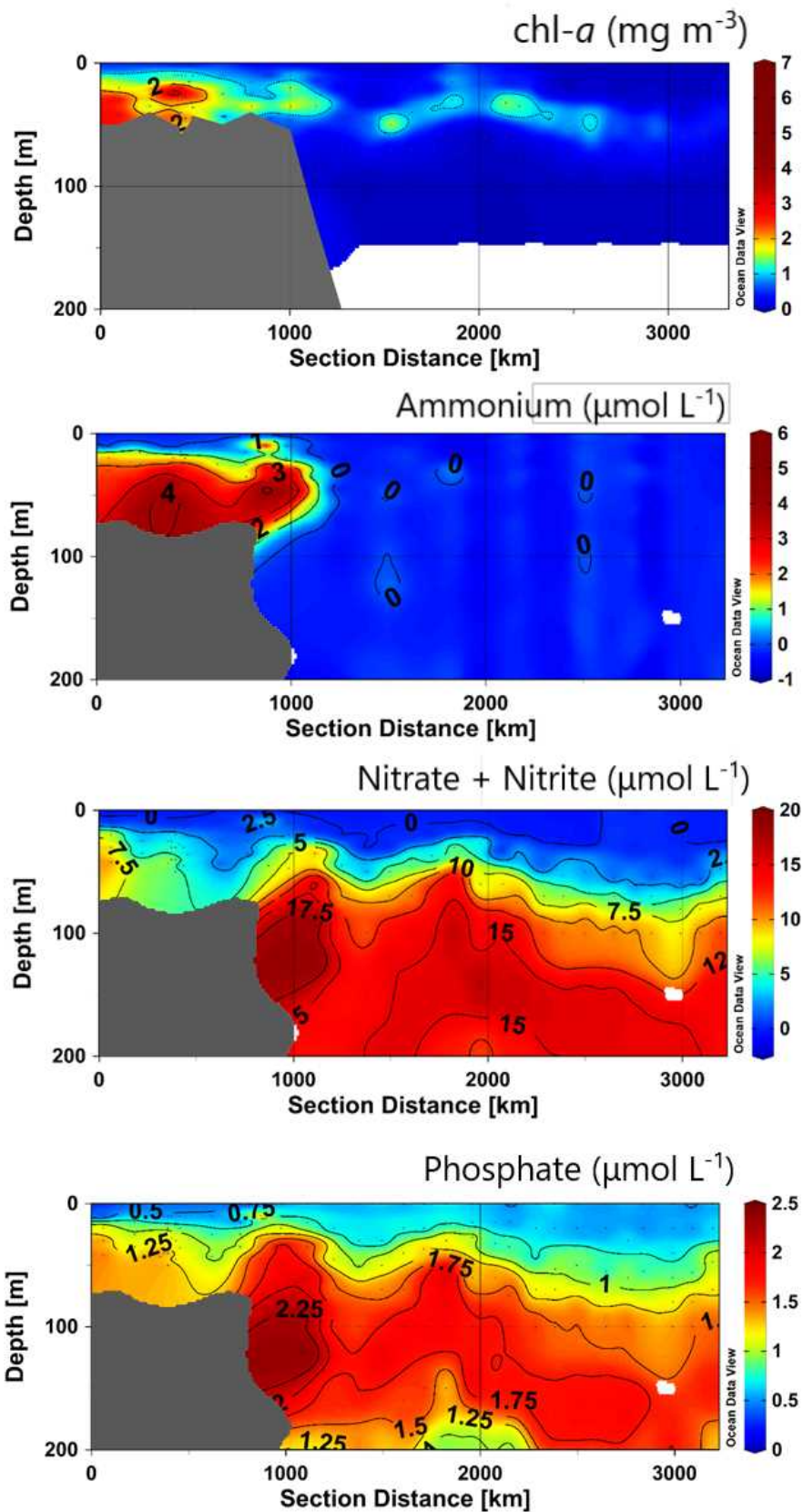


Fig 5.11a Vertical transects of distribution of (a) chl-*a*; (b) Ammonium; (c) nitrate and nitrite; (d) phosphate; (e) silicate; (f) fixed nitrogen (N^{**}); (g) Dissolved Oxygen (DO); (h) AOU concentrations following in order stations (1-31).

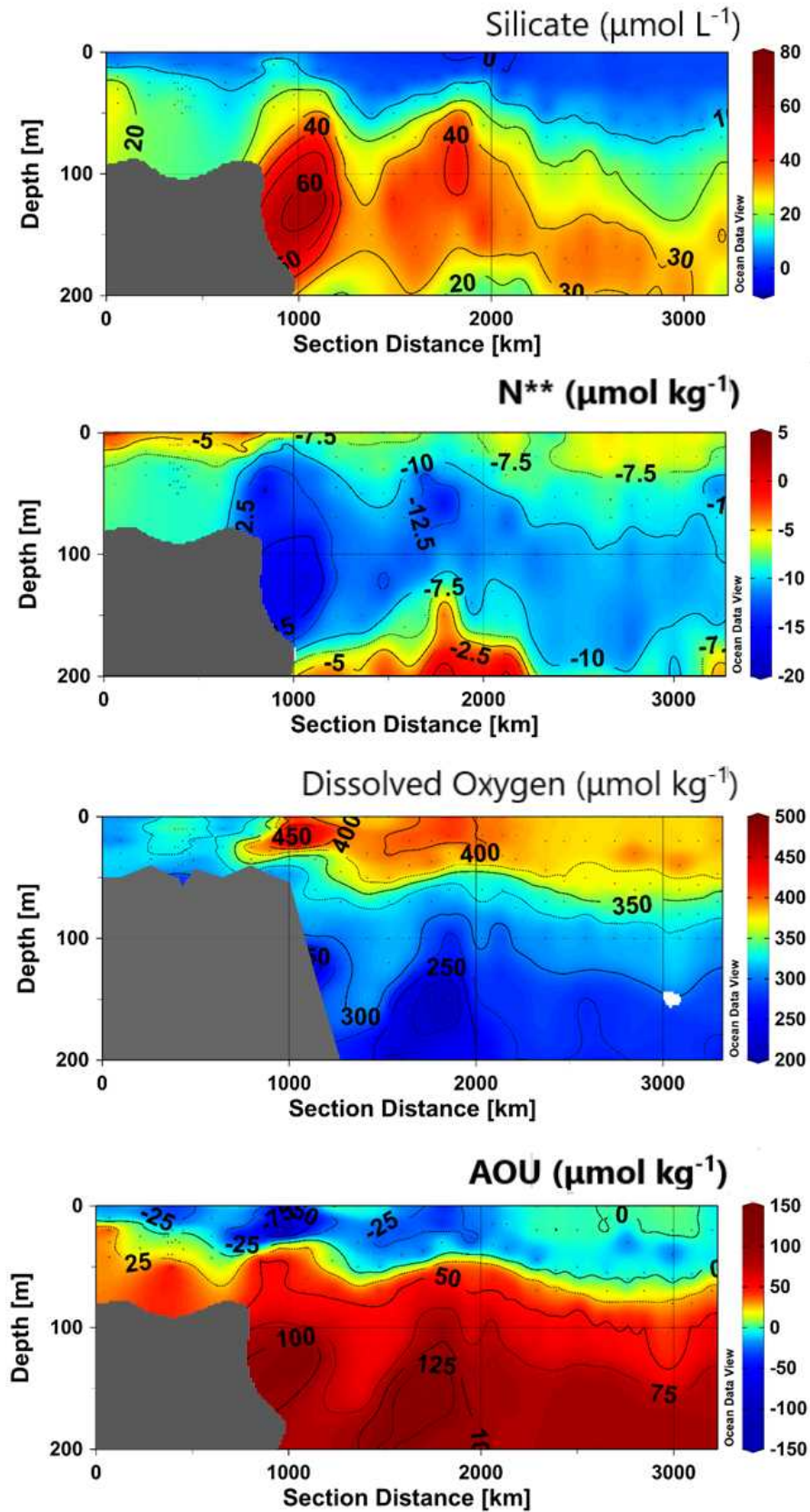


Fig. 5.11b Vertical transects of distribution of (a) chlorophyll-a; (b) Ammonium; (c) nitrate and nitrite; (d) phosphate; (e) silicate; (f) fixed nitrogen (N^{**}); (g) Dissolved Oxygen (DO); (h) AOU concentrations (to be continued) following in order stations (1-31).

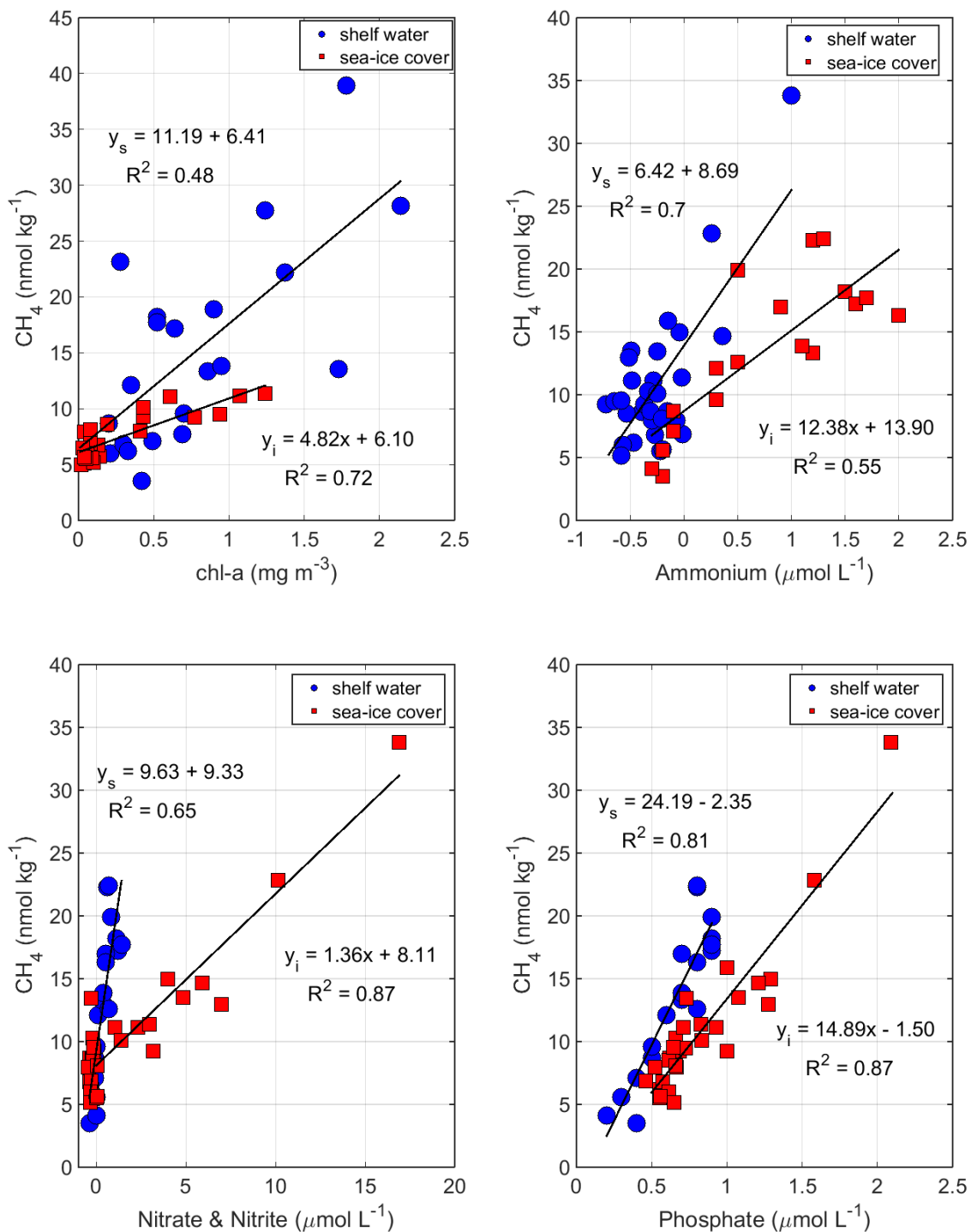


Fig 5.12 The correlation between CH_4 and chl-a, ammonium, total dissolved (nitrite + nitrate) and phosphate concentration over the shelf water region and sea ice cover region.

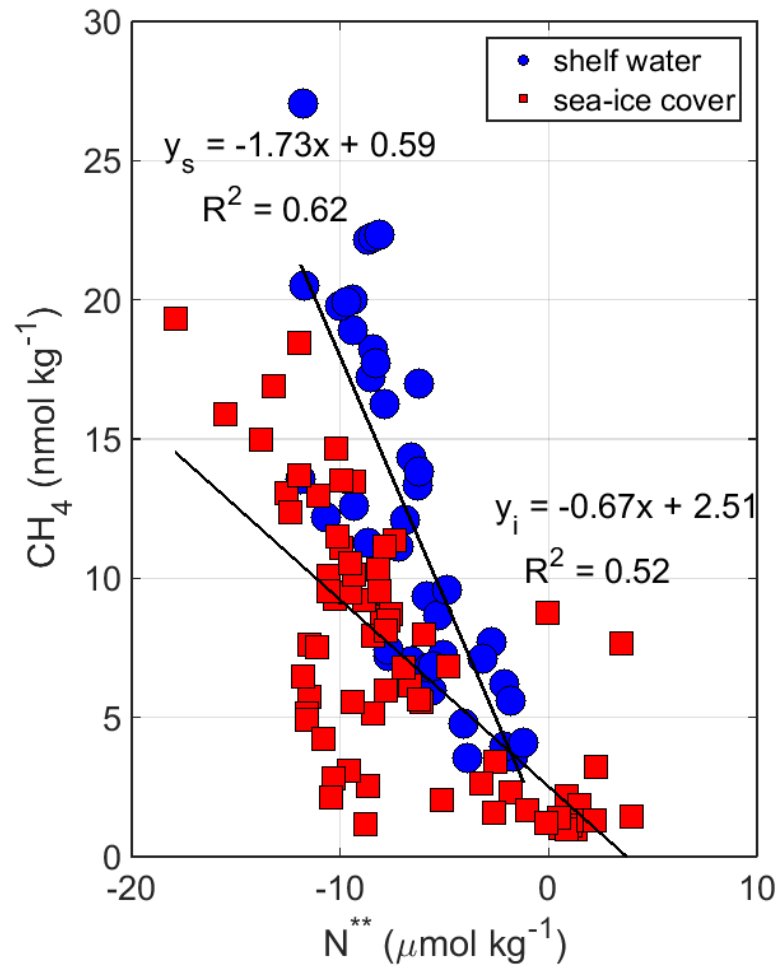


Fig 5.13 Correlations between CH_4 and N^{**} over the shelf water region and sea ice cover region.

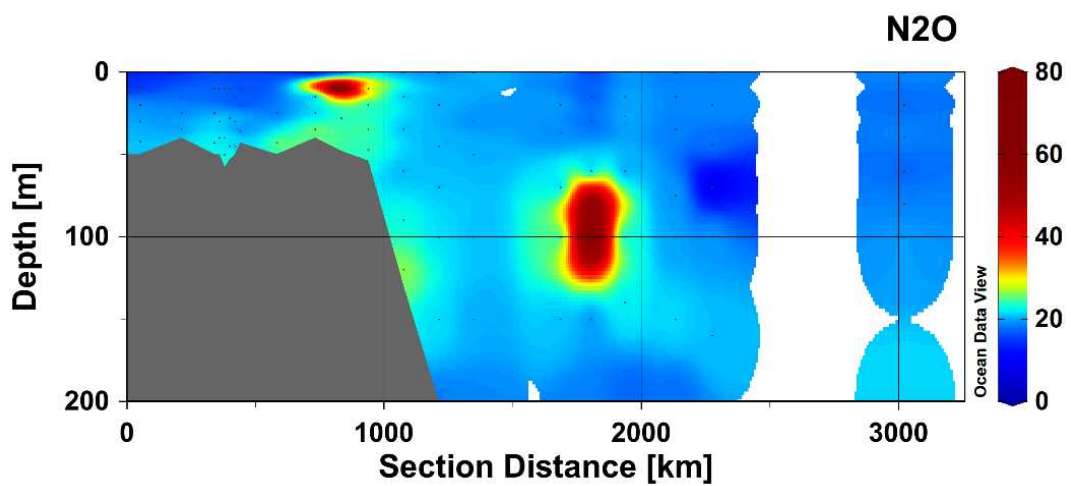


Fig 5.14 Vertical distribution of nitrous oxide (N_2O) during the cruise following in order stations (1-31).

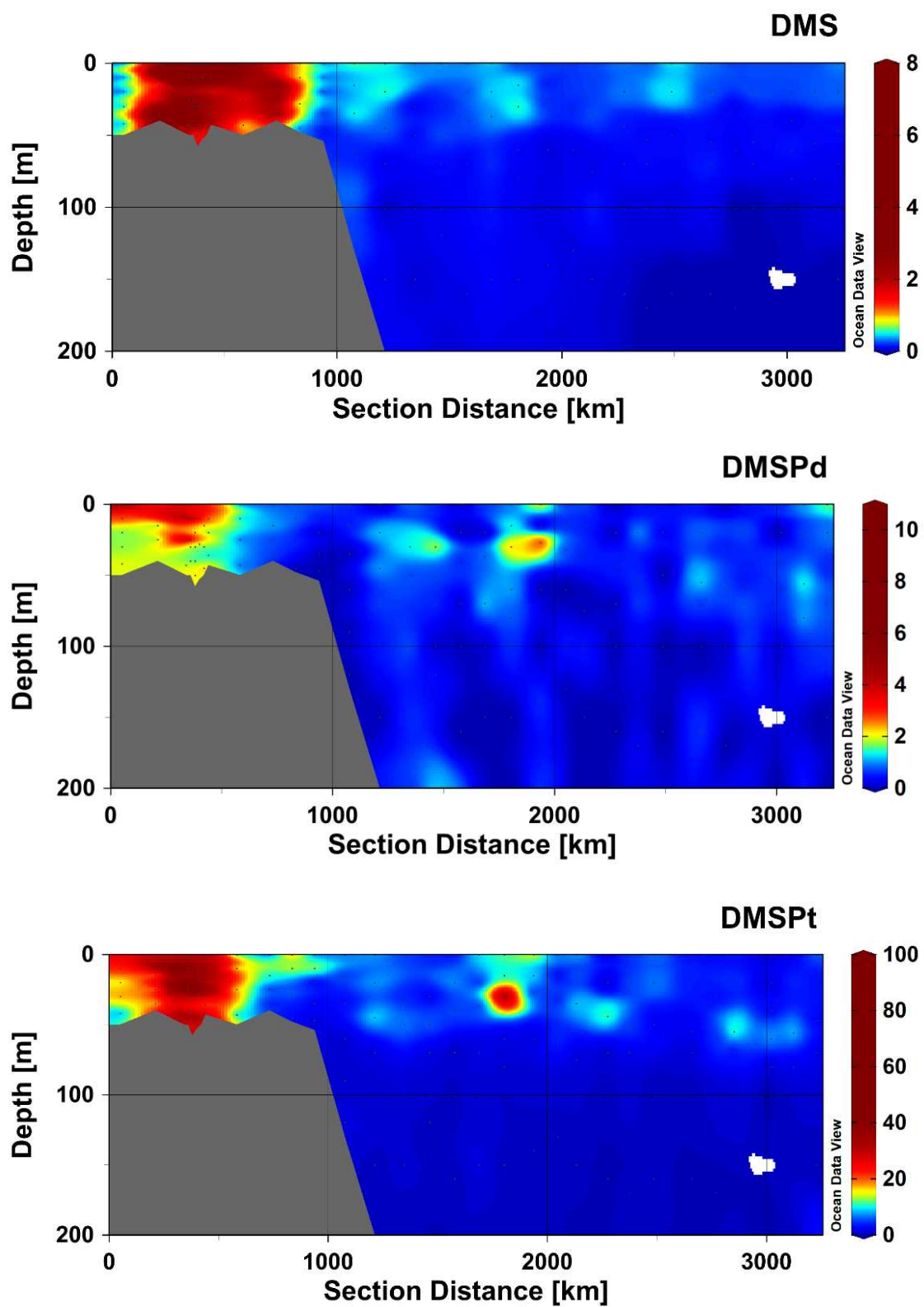


Fig 5.15 Vertical transects of distribution of (a) dimethyl sulphide (DMS); (b) dissolved dimethyl sulphide (DMSPd); (c) total dimethyl sulphide (DMSPt) during the cruise.

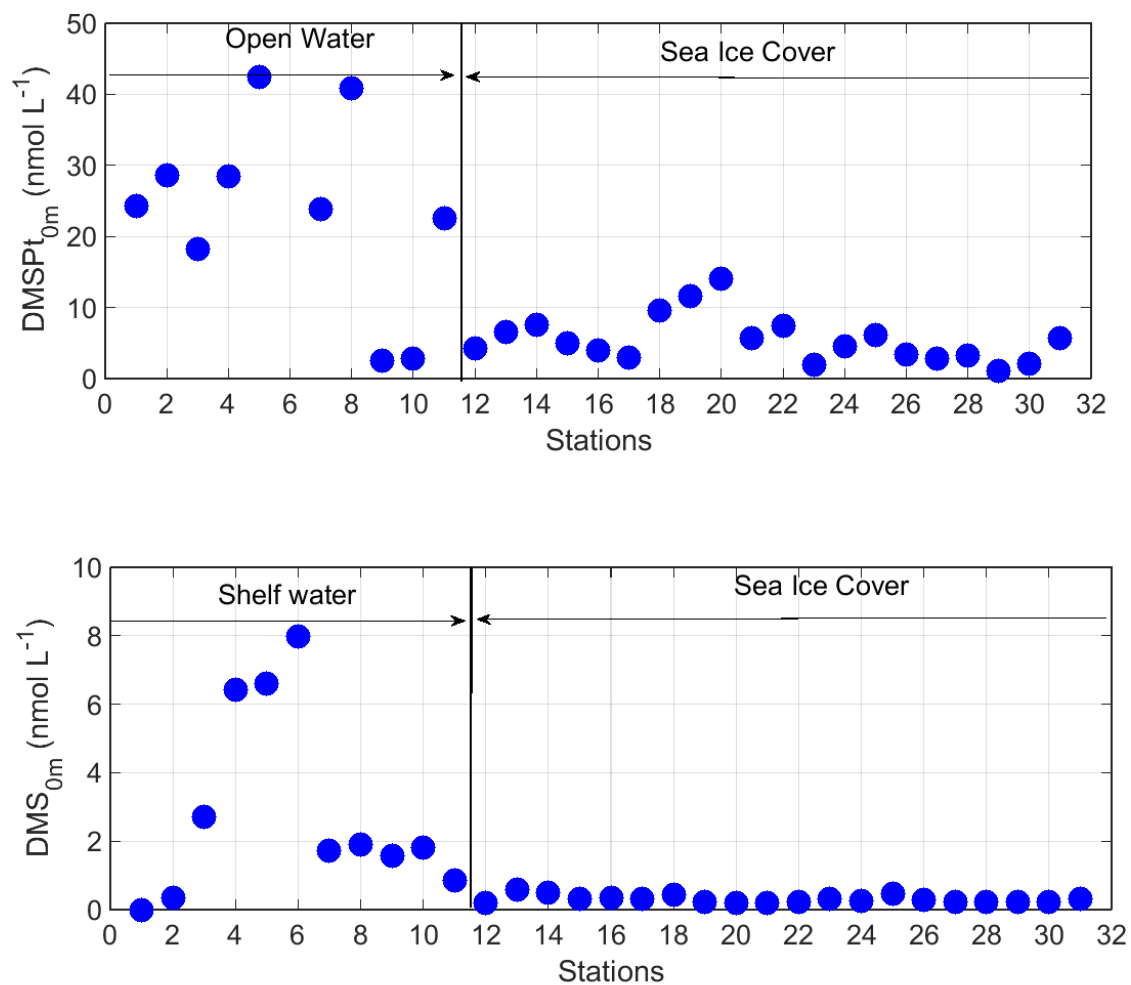


Fig 5.16 DMSPt (upper panel) and DMS (bottom panel) concentration in surface water in order of stations in the Arctic Ocean during the cruise.

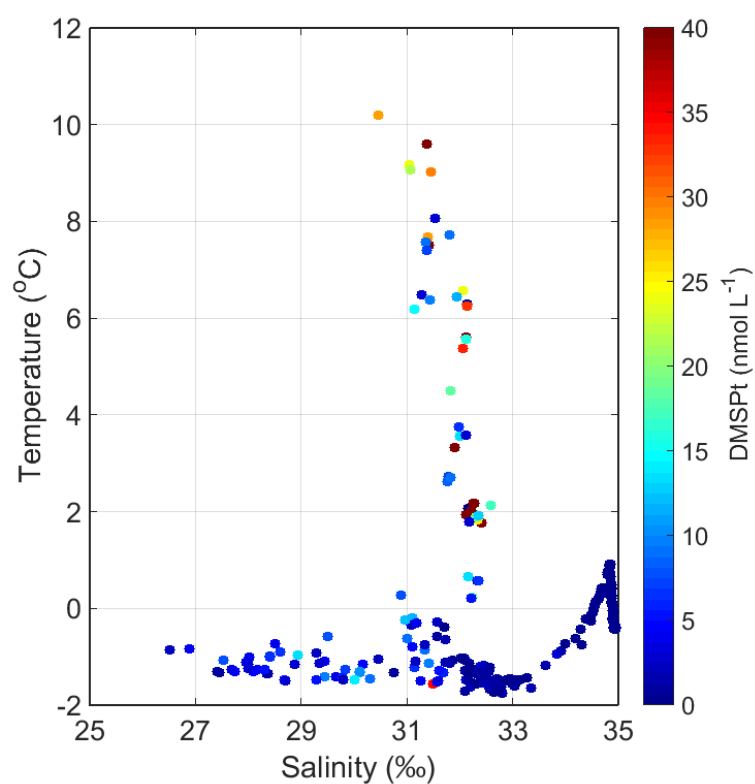
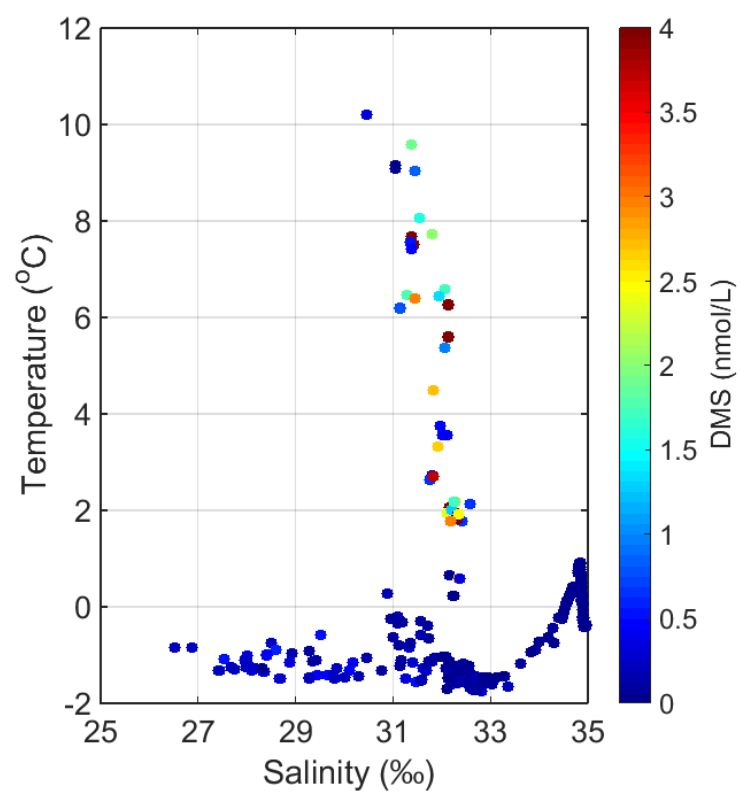


Fig 5.17 Salinity vs. potential temperature plot showing the different water masses observed in the Arctic Ocean. Figure shows DMS and DMSPt concentrations. This figure is plotted by MATLAB.

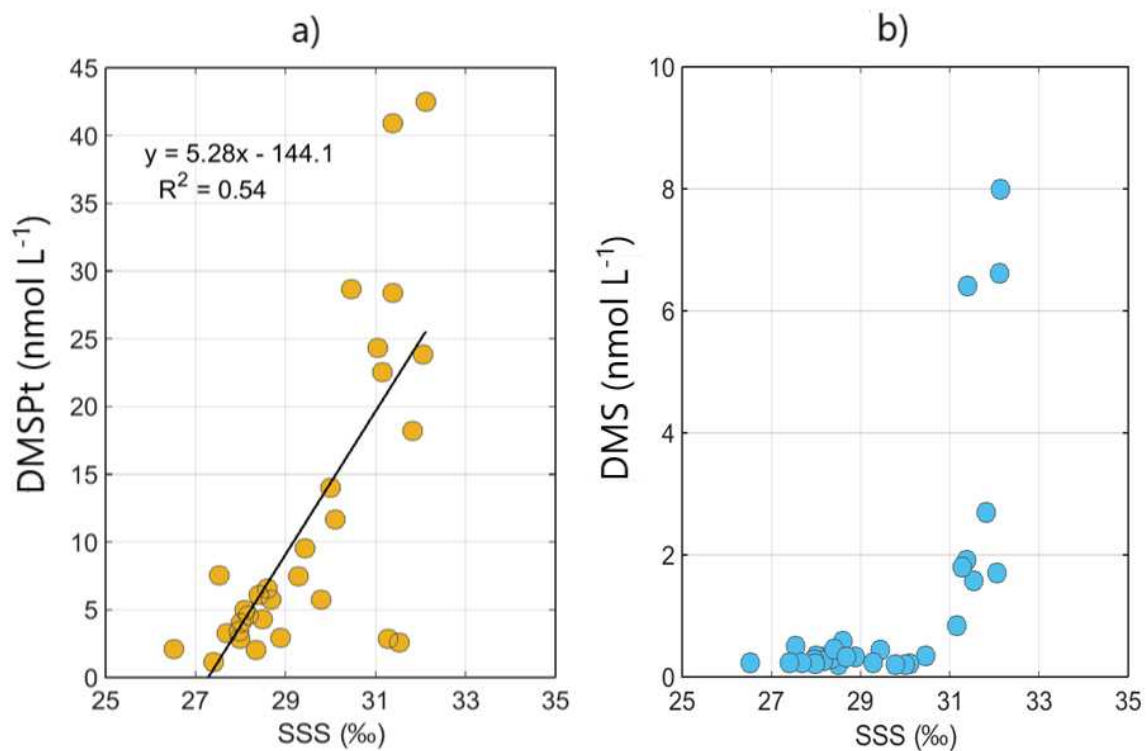


Fig 5.18 Plot of DMSPt and DMS versus seasurface salinity (SSS)

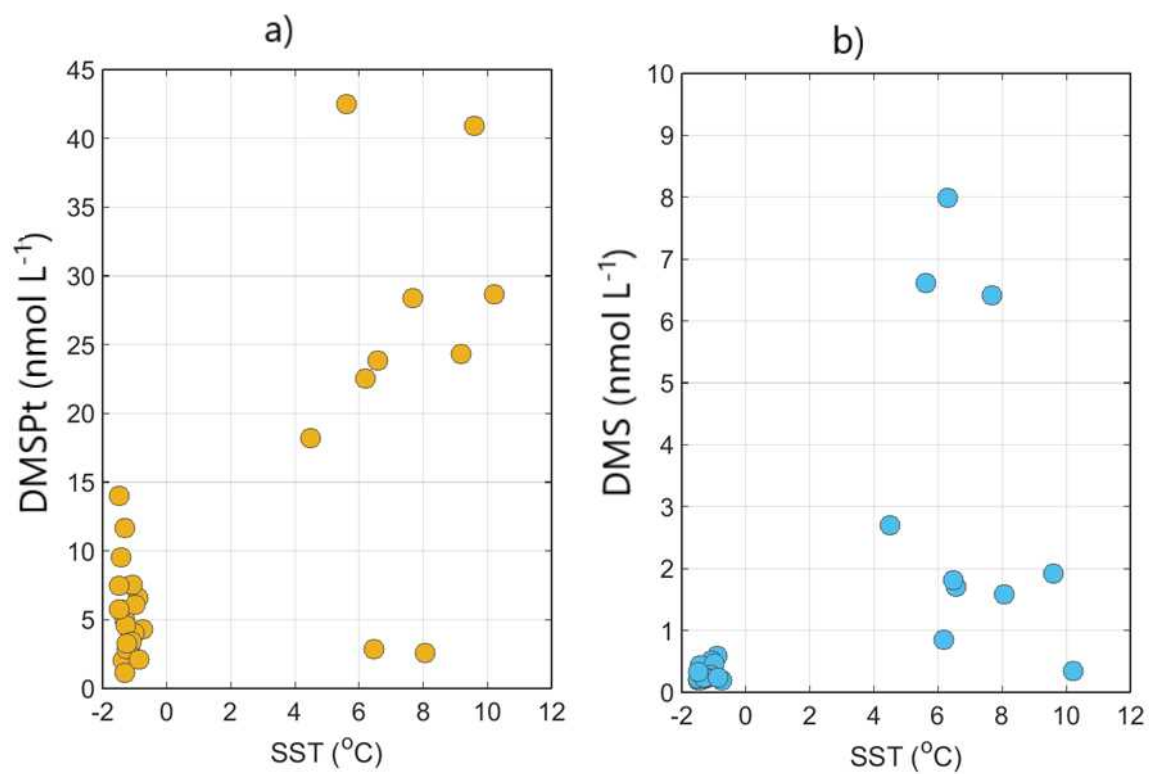


Fig 5.19 Plot of DMSPt and DMS versus seasurface temperature (SSS)

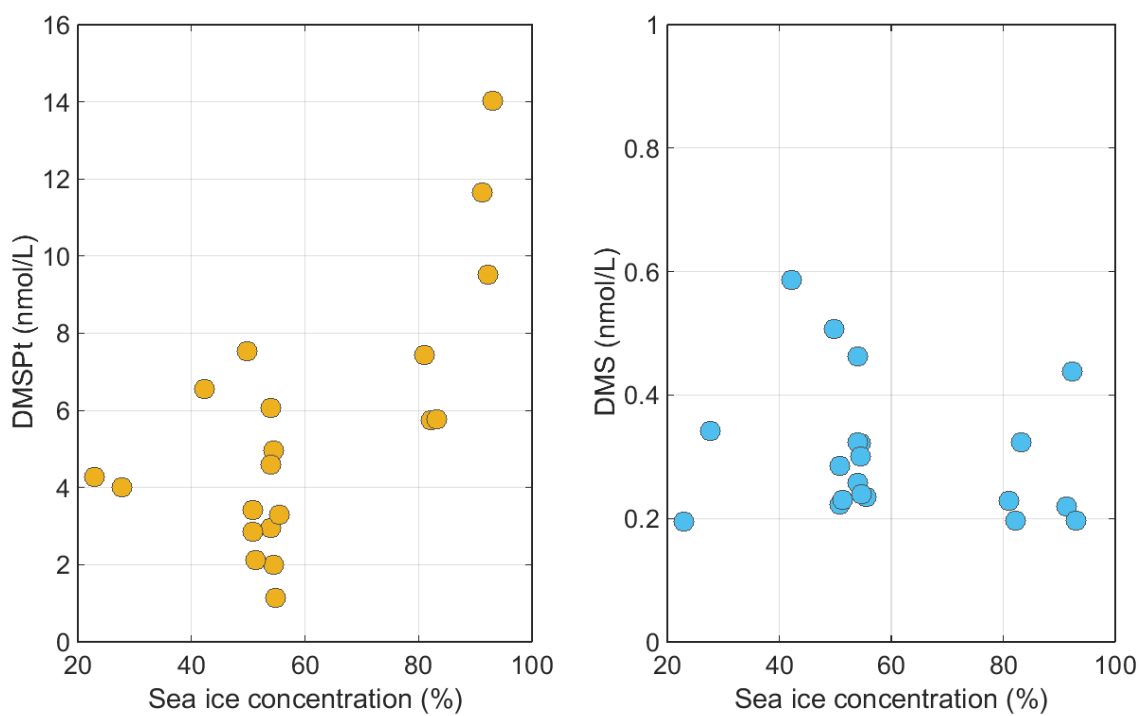


Fig 5.20 Plot of DMSPt and DMS versus sea ice concentration

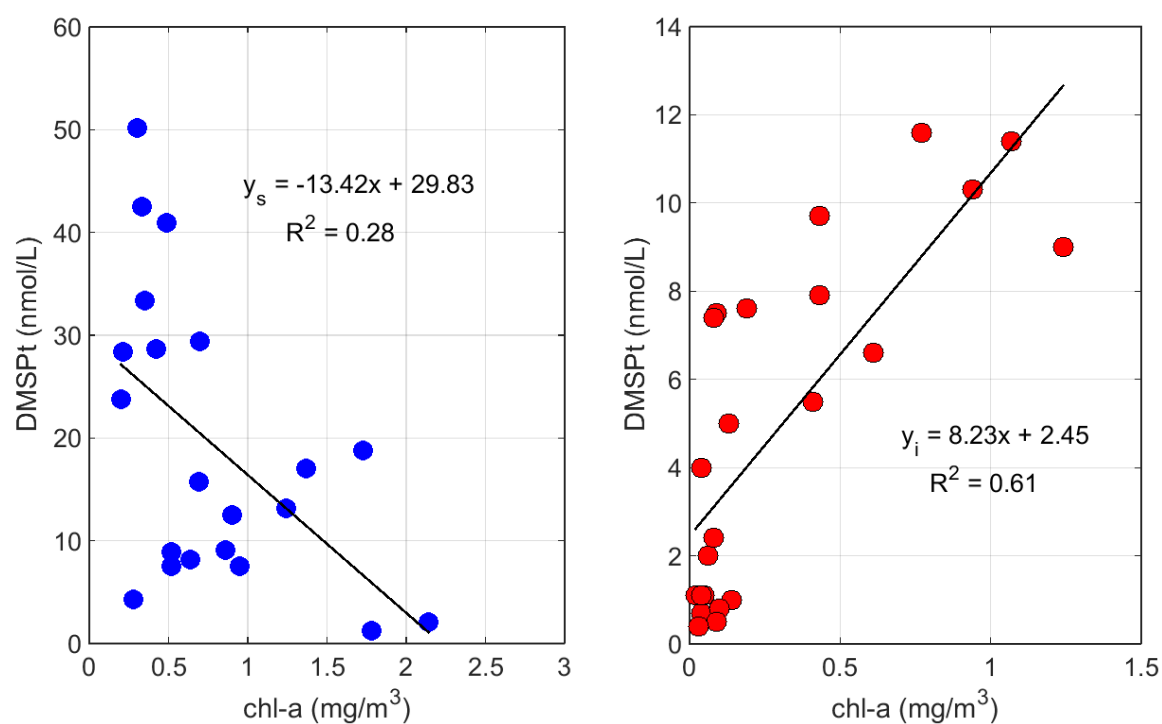


Fig 5.21 Plot of DMSPt versus chl-*a* concentration in the shelf water (left panel) and in the sea-ice cover region (right panel). Black line shows regression line between DMSPt concentration and chlorophyll *a* concentration.

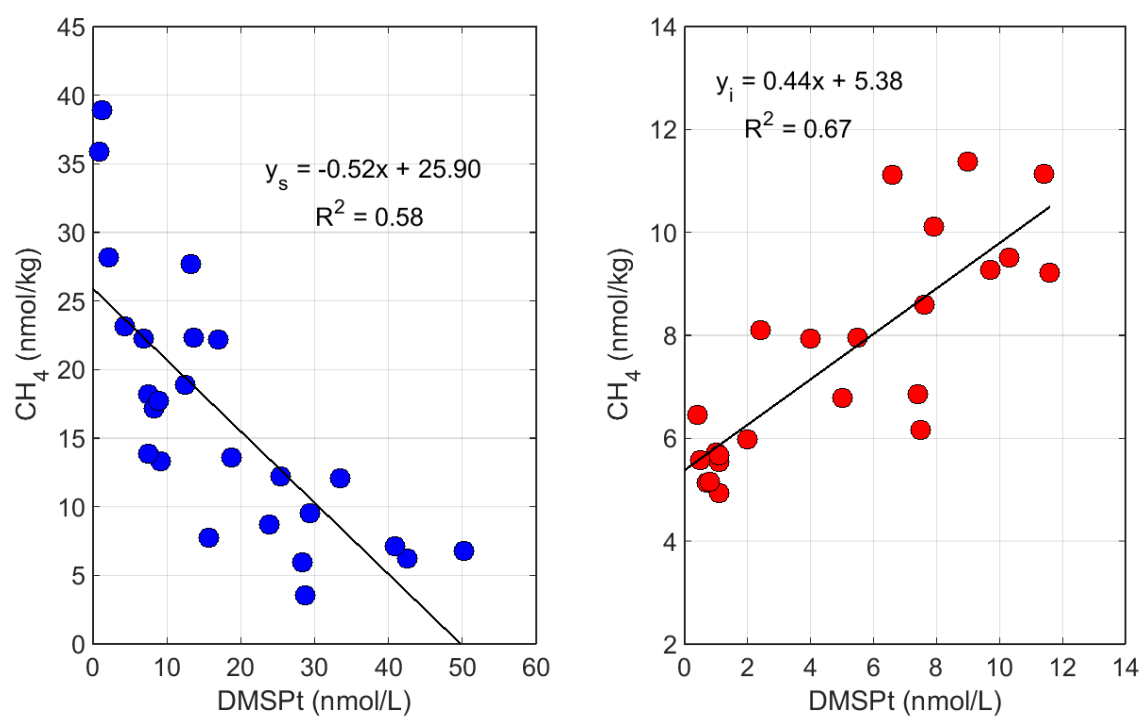


Fig 5.22 CH₄ and DMSPt concentration in the shelf water (left panel) and in sea-ice cover region (right panel). Black line shows regression line between methane concentration and chlorophyll a concentration.

CHAPTER 6: GENERAL CONCLUSION

6.1 Conclusions of this study

In this study, I examined the CH₄ distribution in seawater in the polar regions and investigated the controlling factors including physical and biological factors affect CH₄ distribution. The continuous mixing ratios of CH₄ was measured by new continuous measurement system (CRDS) coupling with an equilibrator. Discrete dissolved CH₄ was measured by GC – IRMS. This observations have provided the dissolved CH₄ distributions under the effects of controlling factors and amount of CH₄ efflux from the ocean to the atmosphere.

Obtained high-resolution CH₄ data in the polar waters (Chapter 3, 4) aims to reduce less uncertainties in estimation of CH₄ flux and supports for research of CH₄ (Chapter 3, 4, and 5). Controlling factors is considered including physical factors and biological factors. Physical factors include the vertical mixing (Chapter 3), eddies transport and lateral transport (Chapter 4), and physical barrier of sea ice cover (Chapter 5). The finding negative relationship between SR and AOU points out the vertical mixing and upwelling in shallow areas supplies poor-CH₄ water from deeper water to surface water in the Southern Ocean and influences on the spatial distribution of CH₄ (Chapter 3). Lateral transport and warm-core eddies play a role in transporting CH₄ from the shelf water into the Arctic/Canada basin. Eddies plays a key phenomenon in transporting and contributing to dissolved CH₄ both in horizontally and vertically (Chapter 4). Achieved results have shown combination of biological and physical factors influence on dissolved CH₄ distribution. However, further research would be required to elucidate the difference in lateral transport and transport of warm-core eddies.

In addition, biological parameters was mentioned such as components of inorganic nutrients, chl-*a* indicating for abundance of organic matter settling upper the sediments producing CH₄ through methanogenesis in anaerobic environment (Chapter 4, 5). As a result, the negative correlation between SR and AOU is examined in the Arctic Ocean shows CH₄ is generally oxidized in the aerobic water column except for producing CH₄ showing in the positive correlation. The results indicate that high CH₄ production were at negative AOU while CH₄ oxidized at the positive AOU both in the Southern Ocean and in the Arctic Ocean (Chapter 3, 4). In fact, it is hard to determine the limitation of AOU co-existence with CH₄ production. Moreover, significant correlation between SR and AOU was observed in the Southern Ocean suggesting that it would be useful for estimating CH₄ flux from AOU data directly.

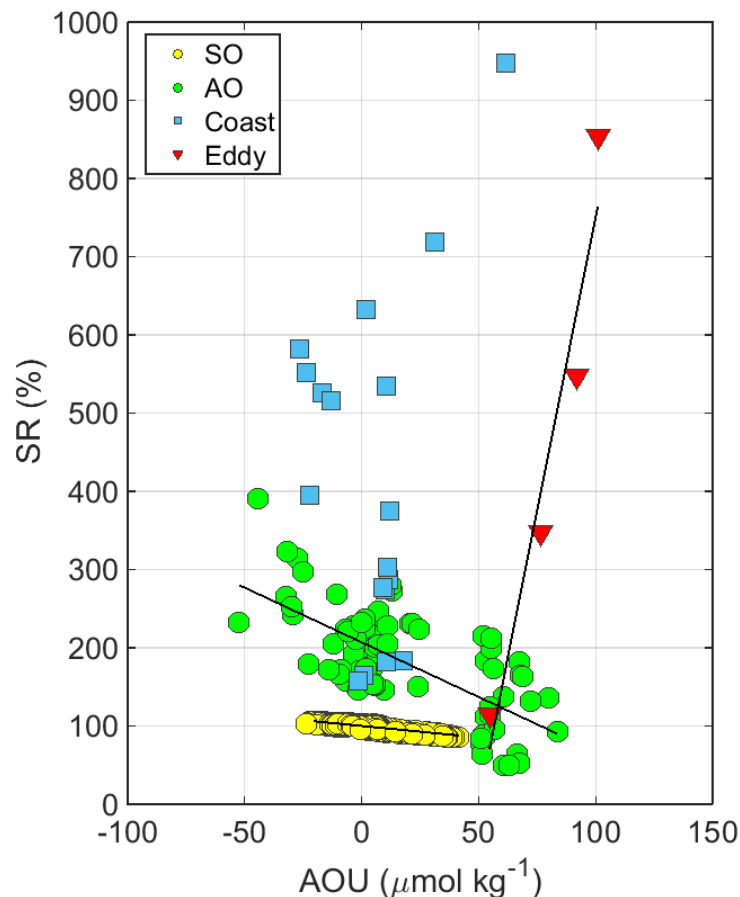


Fig. 6.1 Integral correlation between SR and AOU concentration in the two polar regions. Data were conducted in the Southern Ocean (yellow circles), in open water in the Arctic Ocean (green circles), in the coastal Arctic Ocean (blue squares) and within the warm-core eddy in the Arctic. Data in the Arctic Ocean were conducted from R/V *Mirai* (Chapter 4). The black lines show the regression lines.

As if sea ice coverage prevent sea-air exchange, observed high CH_4 when sea ice cracking may accumulated of northward flowing water beneath sea ice. Mineralisation of “regenerated” organic matter in brine channel can supply for methanogenesis and then release into seawater or the atmosphere when ice melting. In contrast, appearance of sea ice could be enhanced CH_4 exchange at sea-air interface that is expressed via the positive correlation between CH_4 and sea ice concentration. Salinity could be a key factor controlling CH_4 distribution because CH_4 can be produced in local sea ice formation or ice melting. Sea ice acts as semi-permeable barrier to CH_4 sea-air exchange (Chapter 5).

These results obtained from this study elucidate distinct factors impact the spatial distribution of CH_4 between the Southern Ocean and Arctic Ocean as well as in the shelf water

and in the sea ice cover region. It would contribute to the future observation of CH₄ emission from the ocean to the atmosphere.

6.2 Possible feedback to climate change

The present study demonstrated that the controlling factors influence on the dissolved CH₄ distribution in the polar regions. Climate change can affect ocean biogeochemical cycles, ocean water masses, sea ice cover, CH₄ emission from the ocean surface and create a feedbacks to climate. Upwelled high AOU level in recent would raise up the poor CH₄ water and then prediction of the contribution of CH₄ emission from surface seawater in the Southern Ocean would reduce. In contrast, the Arctic is the most sensitive to climate change presenting increase in seawater temperature and melting of sea ice. Reducing in sea ice cover on the permafrost regions that would expel the produced CH₄ under the permafrost and/or CH₄ hydrates into the water column and lead to rapid increase in the atmospheric CH₄ and trigger global warming. Increases in CH₄ levels in the ocean and in the atmosphere constitutes a positive feedback for global warming. However, it should be noted that biogeochemical cycles, characteristics of ocean water masses, frequencies and magnitudes of the occurrence of eddies would differ among the ocean areas depending on biological conditions, hydrographic conditions, and other environmental conditions.

6.3 Further understanding of CH₄

Further, I would examine the vertical profiles of CH₄ in the Southern Ocean where the vertical mixing helps to upwell the poor-CH₄ water in deeper water to the surface water. The relation between SR and AOU and other factors may impact CH₄ distribution in vertically need to examine. Consideration of the influence of the warm-core eddy on CH₄ distribution even though it accounts a small fraction of the total number of eddies (3%). From this study, it clearly shows that the effect of the warm-core directly contribute to the vertical distribution of dissolved CH₄ in the water column. Cold-core eddies occupy approximately 125 (or 97% of the total number of eddies) in the Arctic Ocean so that cold-core eddies would transport CH₄ like the warm-core eddies did. Further field observation of CH₄ crossing a cold-core eddy is a chance to understanding CH₄ distribution within and outside the water core. Furthermore, the effects of climate change impact physical, biogeochemical ocean characteristics as well as CH₄ distribution. Changing in CH₄ distributions and estimations need to do further research.

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