

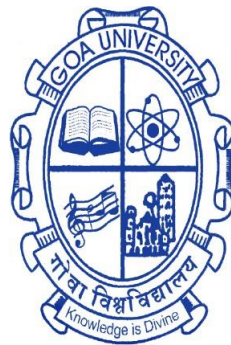
Intraseasonal and interannual physical and biogeochemical variabilities in Kongsfjorden, an Arctic fjord

A THESIS SUBMITTED IN PARTIAL FULFILLMENT
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By

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JUNE 2024

DECLARATION

I, Divya David T, hereby declare that this thesis represents work which has been carried out by me and that it has not been submitted, either in part or full, to any other University or Institution for the award of any research degree.

Place: Taleigao Plateau.

Date : 14 February 2025

Divya David T

CERTIFICATE

I hereby certify that the above Declaration of the candidate, Divya David T, is true and the work was carried out under my supervision.

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List of Abbreviations

ADCP	Acoustic Doppler Current Profiler
AO	Arctic Oscillation
ArW	Arctic water mass
Aw	Atlantic water
AW	Atlantic water mass
Chl <i>a</i>	chlorophyll <i>a</i>
CMEMS	Copernicus Marine Environmental Monitoring Service
CTD	Conductivity-Temperature-Depth
CWS	Continuous wavelet spectrum
DO	Dissolved Oxygen
ECMWF	European Centre for Medium-Range Weather Forecasts
ERA5	ECMWF Reanalysis v5
FWC	Freshwater content
FWV	Freshwater volume
HC	Heat content
HYCOM	Hybrid Coordinate Ocean Model
ICES	International Council for the Exploration of the Sea
ISO	Intraseasonal oscillations
IW	Intermediate water mass
LW	Local water mass
MDI	Mooring Dirigibile Italia
MLD	Mixed layer depth
NAO	North Atlantic Oscillation
NCODA	Navy Coupled Ocean Data Assimilation
NCPOR	National Centre for Polar and Ocean Research
NTU	Nephelometric turbidity unit
PAR	Photosynthetically active radiation
RCM	Recording current meters
ROMS	Regional Ocean Modeling System
SCM	Subsurface Chlorophyll Maximum
SPC	Spitsbergen Polar Current

SW	Surface water mass
TAW	Transformed Atlantic water mass
USGS	United States Geological Survey
WCW	Winter cooled water mass
WHCMSC	Woods Hole Coastal, and Marine Science Center
WSC	West Spitsbergen Current
WW	Winter water mass

Chapter 1

Introduction

1.1. Background

The Polar Regions have the greatest potential to affect global climate and are highly vulnerable to the current and projected climate change. On the northern side, warming in the Arctic has accelerated four times faster than in the rest of the world (Rantanen et al., 2022). The Arctic Ocean, surrounded by Russia, Canada, the United States, Norway, and Greenland, is particularly sensitive to climate change. During winter, the ocean's surface is mainly covered by sea ice, which expands and contracts with the changing seasons. The warming of the Arctic Ocean is not only due to atmospheric warming but also from internal oceanic heat sources. The Arctic Ocean is connected to the Atlantic Ocean through the Greenland and Norwegian Seas and the Pacific Ocean through the Bering Strait. These warm-saline waters from lower latitudes flow into the Arctic, influencing local conditions. However, the amplitude of the Arctic Ocean warming is not uniform. Certain areas, particularly the Atlantic-influenced eastern Arctic Ocean regions, have seen more significant temperature and salinity increases due to increased Atlantic water (Aw) inflows known as 'Atlantification' (Polyakov et al., 2017). The Aw from the Gulf Stream is transported to the northeast by the North Atlantic Current, which then becomes the Norwegian Atlantic Current in the Norwegian Sea, after which it bifurcates to two branches- one moving further north to the Arctic along the continental slope of west Svalbard known as the West Spitsbergen Current (WSC) and the other to the Barents Sea. Polyakov et al. (2017) reported that the Atlantification of the Eurasian basin ultimately leads to a new Arctic climate state. This includes melting sea ice with younger fragile ice replacing thicker ice, implicating the resilience and stability of the ice cover, more open waters and ice-albedo changes, increased storms, enhanced air-to-sea energy transfer and mixing, changes in ocean circulation, changes in

phytoplankton bloom timings and species composition ultimately affecting the food web and ecosystem, survival of species benefitting from a warming ocean, etc.

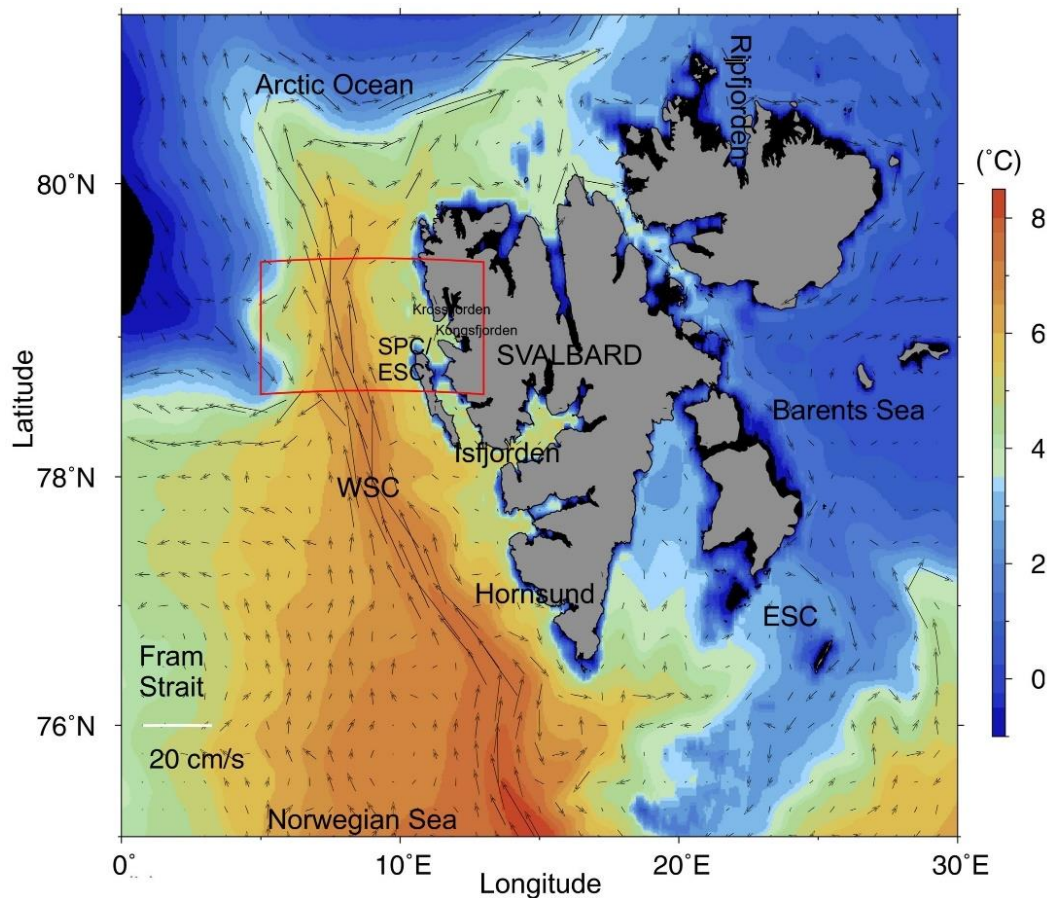


Figure 1.1: Map of Svalbard archipelago with 0–50 m depth-averaged ocean temperature (°C) in shades and currents (cm s⁻¹) in arrows for August 2002–2017. The red rectangle is the study region which is also the ROMS model domain used for the study. The temperature and current are from the GLORYS12V1 product, the CMEMS global ocean eddy-resolving reanalysis.

Within the eastern Arctic Ocean, the Svalbard archipelago is a group of islands about midway between continental Norway and the North Pole. Svalbard is surrounded at its left by the Fram Strait, Aw's largest opening and pathway to the Arctic Ocean, right by the Barents Sea, south by the Norwegian Sea, and north by the cold-fresh Arctic Ocean (Fig. 1. 1). Svalbard's climate is representative of the high Arctic, which makes it an ideal location for climate change research. The climate of Svalbard is influenced by changes in atmospheric circulation patterns (Hanssen-Bauer and Førland, 1998) and wintertime sea-ice extent (Divine and Dick, 2006; Day et al.,

2012). Sea ice is present in the east and north of Svalbard in winter and spring, while ice-free in the west (Stocker et al., 2020). The ice extent rarely reaches south of 76°N. The atmospheric circulation is influenced by cold, northeasterly polar winds from the Barents Sea (Hanssen-Bauer and Førland, 1998). These winds are frequently interrupted by the passage of low-pressure disturbances from the southwest. Being on the polar front, the region is dominated by precipitation in the form of snow and rain throughout the year (Rotschky et al., 2011). A recent study (Kleber et al., 2023) reported that the archipelago is warming two times faster than other areas in the Arctic and five to seven times faster than the rest of the world. Hence, this archipelago is an important indicator of climate change impacts in the Arctic, including changes in sea ice extent, glacier dynamics, permafrost thaw, and ecosystem shifts and their response to environmental changes, contributing to understanding broader Arctic and global climate trends.

An important feature of the Svalbard archipelago is its numerous fjords formed by glacial erosion. Fjords are glacially calved semi-enclosed basins, mostly with sills at their entrance that separate deep waters from adjacent coastal waters, thereby restricting circulation and oxygen renewal (Howe et al., 2010). Technically, they are highly stratified estuaries and are probable locations of dense water formation that form a crucial part of the global conveyor belt. Fjords in the polar, subpolar, and temperate climate regimes combined regulate Earth's climate by absorbing ~18 million tons of organic carbon yearly, equivalent to 11% of annual marine carbon burial globally, and the atmospheric carbon dioxide concentrations at times of varying ice sheet dynamics (Smith et al., 2015). Fjords in the Arctic are a critical pathway through which the sea can affect glacier stability (Luckman et al., 2015). Studies show the role of heat from warm oceanic waters transported to the fjords on the glacier terminus stability and increased melt rates (Rignot et al., 2010; Straneo et al., 2010; Luckman et al., 2015; Holmes et al., 2019). The contribution of marine-terminating glaciers to contemporary sea-level rise and associated processes (Moon et al., 2018; Halbach et al., 2019; Robel et al., 2019; Torsvik et al., 2019; Hopwood et al., 2020) is a critical response of the warming-induced changes. Glacial activity can transport sediments into the fjords, affecting water clarity and seabed conditions.

They are sites of locally enhanced sedimentation that provide high-resolution paleoceanographic records of late and postglacial changes. Hence, understanding the active physical processes in the fjord not only benefits the accurate understanding of biogeochemistry (Cottier et al., 2010) but also is vital to ascertain palaeoclimatic and palaeoceanographic conditions from sediment records (Majewski et al., 2009; Guruvayoorappan et al., 2019). Fjords are interesting research sites with their finite ocean basins and characteristic hydrography, biogeochemistry, fauna, and sedimentation dynamics (Syvitski and Shaw, 1995) and idealised locations to study these physical and biogeochemical processes (Wallace et al., 2010). Human activities and environmental changes make fjords vulnerable (Richerol et al., 2007). Moreover, fjords that link the atmospheric, relatively warm oceanic, and glacier-dominated terrestrial realms (Cottier et al., 2005) respond to and reflect whatever happens in the three realms. Fjords in Svalbard archipelago, especially on its western side, are highly responsive to variations in the realms and are, therefore, good indicators of the changes. Hence, they are suitable environments for studying ongoing climate changes in the region (Benestad et al., 2002).

1.2. Kongsfjorden

1.2.1. Geographical and physical setting

Kongsfjorden, a fjord located northwest of Svalbard centered at 79 °N, and 12 °E, is oriented from southeast to northwest (Fig. 1. 2). It has a length of ~20 km, a varying width of ~4 km (inner part) to ~10 km (outer part) (Svendsen et al., 2002; Lindbäck et al., 2018), and a maximum depth of ~ 350 m in the mouth. The Coriolis effect is significant here as the fjord is wide concerning the 3-4 km Rossby radius of deformation. Kongsfjorden shares a common mouth with another fjord, Krossfjorden, to its north, oriented from south to north. Hence, Kongsfjorden-Krossfjorden is known as a twin fjord system. The geographical closeness and a common mouth are also reasons for the connectivity between the fjords with respect to the atmospheric and oceanic circulations, shelf-fjord exchanges of Atlantic water, and heat transport pathways. Hence, it is important to consider the hydrography and circulation of Krossfjorden while studying Kongsfjorden in detail. Kongsfjorden has a total volume of 29.4 km³ (Ito and Kudoh, 1997). It has a deep partial sill known as

Kongsfjordrenna, having a depth of ~ 250 m that extends as a deep canyon cutting across the West Spitsbergen shelf (Inall et al., 2015) to the open ocean. The fjord is strongly influenced

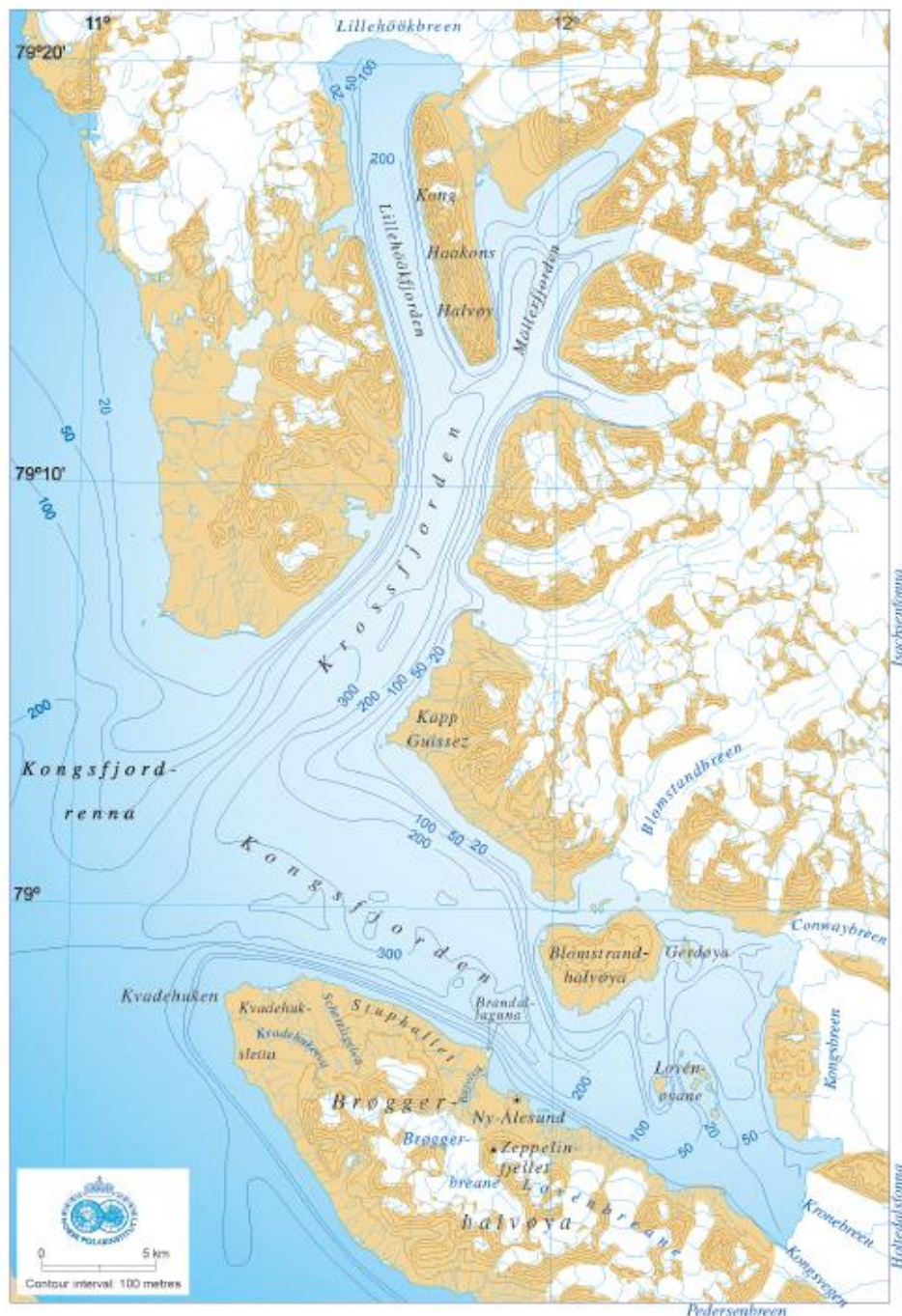


Figure 1.2: Kongsfjorden-Krossfjorden map with isobaths and glaciers [Svendsen et al., 2002].

by the tidewater glaciers of Kronebreen, Kongsvegen, and Kongsbreen at the head and Conwaybreen and Blomstrandbreen on its northern coast. A mean annual freshwater discharge of $\sim 1.4 \text{ km}^3$ with up to 30 % annual variation from the glacial input, icebergs, and precipitation was reported in Kongsfjorden (Svendsen et al. 2002). While the major input is from the glacial melting followed by seasonal precipitation (MacLachlan et al., 2007), the other freshwater sources are snow melt, run-off, and groundwater discharge (Svendsen et al., 2002; Cottier et al., 2005).

1.2.2. Climate

The climate in the region is affected by general atmospheric circulation, radiation conditions, ocean currents, and sea ice extent. Low pressure over Iceland and high pressure over Greenland and the Arctic Ocean drive the large-scale circulation inducing warm-humid air in the North Atlantic to Norwegian and Barents Seas along the cyclone tracks (Svendsen et al., 2002). The weather conditions, especially during winters, vary highly due to considerable differences in atmospheric temperature (Hansen-Bauer et al., 1990; Forland et al., 1997). The wind direction in Kongsfjorden is bi-modal, from inland to sea (downfjord), and the next most dominant direction is just opposite- upfjord (Svendsen et al., 2002). These winds are controlled by the orographic steering of large-scale wind fields and katabatic winds with cold air from the inland glaciers to the warmer fjord (D'Angelo et al., 2018). Seasonal and altitude variation modifies wind distribution (Beine et al., 2001). The wind direction in summer and winter is slightly different but remains between 100-160 degrees, clockwise from the north (Esau and Repina, 2012). The region's overall storm frequency and intensity have increased over the last few decades (Zhang et al., 2004; Sepp and Jaagus, 2010). Large-scale wind direction influences the local precipitation (Førland et al., 1997). Recently, warm winters altered the precipitation form in Kongsfjorden from snow to rain (Maturilli et al., 2015). The maximum precipitation was observed in January and the minimum in June, with the mountainous region receiving much precipitation and the inner fjord the least. The radiation climate of Kongsfjorden is characterised by significant seasonal fluctuations (Maturilli et al., 2015). The summer and spring seasons show large interannual variability in the monthly mean shortwave, longwave, and net radiations. The annual mean from 1992

to 2013 of observed shortwave radiation is $0.8 \pm 3.6 \text{ W/m}^2$, longwave radiation is $5.4 \pm 3.7 \text{ W/m}^2$, and the net radiation budget is $4.9 \pm 2.9 \text{ W/m}^2$ per decade. During the same period of 1992 to 2013, the annual mean surface air temperature increased by $+1.3 \pm 0.7 \text{ K}$ per decade while the summer and winter temperatures increased by $+0.7 \pm 0.5 \text{ K}$ and $+3.1 \pm 2.4 \text{ K}$ per decade, respectively.

Relatively cold-fresh Arctic waters brought by the Spitsbergen Polar Current (SPC, also known as the East Spitsbergen Current) originate from the east of Svalbard and are directed northward along the continental shelf to the west of the Kongsfjorden (Helland-Hansen and Nansen, 1909; Skogseth et al. 2020) while relatively warm-saline Aw brought by the WSC flows on the slope (Fig. 1. 2). The wind field tends to dominate over rotational effects in the Kongsfjorden's upper layer circulation (Svendsen et al., 2002). In summer, the upper layer circulation is confined to a shallow surface layer. The meltwater flows down the fjord at a speed ranging from 10 to 30 cm s^{-1} , and the total current velocity is a combination of tidal, freshwater, and wind-driven currents. Some brackish water flows through the opening between Blomstrandhalvøya and the glacier Blomstrandbreen (Ingvaldsen et al., 2001). A part of this branch flows along the northern Kongsfjorden coast, while another part recirculates around Blomstrandhalvøya, entering the inner fjord at the northernmost side. During rare and short-lived upfjord wind events, resultant upfjord surface currents stack up the brackish water at the fjord head, eventually turning the flow downfjord. In winter, the wind direction out of the partially ice-covered fjord removes surface water from the ice edge, creating a horizontal down-fjord pressure gradient, setting up a return flow that may supply Aw to the under-ice, slowing down the ice growth (Gerland and Winther, 2000). Part of the Aw reaching West Svalbard meet at the West Spitsbergen shelf, mix and exchange their properties with the fjords, and play an essential role in determining their water properties (Walczowski, 2013). According to Cottier et al. (2005), Aw on the shelf are prevented from entering the fjord in winter by a geostrophic control mechanism (Klinck et al., 1981) created by a significant lateral density gradient between the fjord and outer waters. Hence, AW is frequently observed in summer, when the freshwater influx in the fjord weakens the horizontal density gradient. However, strong winds on the shelf in winter can

potentially weaken the geostrophic control system, leading to AW incursions (Cottier et al., 2007). As a result, seasonal and synoptic scales—the occurrence of strong winds on the shelf and the collapse of the fjord's geostrophic control—are the most significant for AW intrusions. The inflows inside Kongsfjorden are aligned to the direction of the remotely generated Kelvin wave that enters along the southern coast (Svendsen et al., 2002). The pressure gradient across the fjord forces the wave to continue around the basin, forcing an outflow at the northern side with velocities not higher than 8 to 10 cm s^{-1} . Because of the strong pycnocline, deep-layer circulation is somewhat independent of the upper-layer flow.

Different ice types are found in Kongsfjorden; the dominant ice type is young ice in winter. Fast ice is observed in early spring. Once the melting season starts, a combination of fast ice and drift ice consisting of icebergs and ice pieces calved from glaciers are found in open water. Observations of sea ice cover between 1981 and 2019 (Mehlum, 1991; Lydersen and Gjertz, 1986; Gerland et al., 2020) point out interannual variation in the position of sea ice edge and the melting and breakup timing, even though the general breakup occurs in between April and July. Thus, the fjord hydrography is influenced by Aw, glacial melting, cold-fresh polar waters circulating on the shelf via SPC, and sea ice.

1.2.3. Water masses

Surface warming, glacial melt, mixing, entrainment, and AW intrusion result in the formation of different water masses inside Kongsfjorden (Svendsen et al., 2002; Cottier et al., 2005; Nilsen et al., 2008). Surface water mass (SW) is formed by the glacial melt and river run-off and is generally dominant in late spring and summer. Intermediate water mass (IW) is formed by vertical mixing of the SW with the water mass below it. Local water mass (LW) is formed by convection of the SW, which is cooled by losing heat to the atmosphere. Further cooling and brine release during sea-ice formation produce deeper convection, leading to the formation of the winter-cooled water mass (WCW). The LW and WCW found during winter and autumn are combined, called Winter water mass (WW) (Nansen, 1915; Nilsen et al., 2008), and play an important role in the shelf-fjord exchange mechanism.

The Atlantic water mass (AW) and Arctic water mass (ArW) originate outside the fjord in the WSC and SPC, respectively. Another external water mass observed inside Kongsfjorden is the Transformed Atlantic water mass (TAW) formed by mixing warm AW with cold ArW on its way to the fjord.

Table 1.1: Characteristic temperature-salinity ranges of water masses in Kongsfjorden based on Cottier et al. (2005) [Adapted from David and Krishnan, 2017].

Water masses	Temperature (°C)	Salinity (psu)
Internal:		
LW	$0.5 < T < 1.0$	$34.30 < S < 34.85$
WCW	$T < -0.5$	$34.40 < S < 35.00$
SW	$T > 1.0$	$S < 34.00$
Mixed:		
TAW	$1.0 < T < 3.0$	$S > 34.65$
IW	$T > 1.0$	$34.00 < S < 34.65$
External:		
AW	$T > 3.0$	$S > 34.65$
ArW	$-1.5 < T < 1.0$	$34.30 < S < 34.80$

Thus, the presence/absence of the type of water masses in the fjord can illuminate the changes it undergoes at temporal and spatial timescales and its shift from one season to another. More details on the physical characteristics of the water masses are given in Table 1.1.

1.3. Literature review on Kongsfjorden variabilities

Compared to other Arctic fjords (Svendsen, 1981; Farmer and Freeland, 1983), Kongsfjorden is highly vulnerable to ongoing warming due to its climatically and oceanographically critical location (Cottier et al., 2005). On the other hand, the fjord has higher Aw volume compared to other southern fjords in west Svalbard due to its relatively closer location to WSC, southeast-northwest orientation, and lack of a distinct sill at its entrance (Svendsen et al., 2002; Cottier et al., 2005; Nilsen et al., 2008). With the prominent increase in the air temperatures, winter 2005/06 had anomalously high Aw in Kongsfjorden that shifted the entire sea-ice environment (Cottier et al., 2007). Afterward, the fjord is mostly sea ice-free during winter with

an increasing influx of Aw. Conversely, the warming-induced increased glacial melting discharges freshwater into the fjord surface, mostly in summer (Svendsen et al., 2002). These cold fresh waters on the surface and warm saline waters below them change the vertical stratification, affecting its physical and biogeochemical processes. Beyond the overall change in trends, the changes considerably affect the seasonal and interannual variabilities in Kongsfjorden hydrography (Tverberg et al., 2019).

The summer hydrography data collection in Kongsfjorden started in 1994. The first review of the physical environment in Kongsfjorden (Svendsen et al., 2002) showed the interannual summertime and lateral variation in hydrography. Cottier et al. (2005) highlighted that the fjord adopts a “cold” or “warm” mode according to the amount of Aw occupation. Promińska et al. (2017) also attributed the summer variability in Kongsfjorden to an increased Aw intrusion. Meanwhile, David and Krishnan (2017) linked the increase in Aw in 2013 to the positive summer NAO index, increasing the Aw flow through the Nordic Seas and further into the fjord. Almost two decades after Svendsen et al. (2002), Tverberg et al. (2019) reviewed the physical studies from 1994 to 2014. This study reported seasonal and interannual variations in oceanographic conditions and interaction with shelf-slope water masses. While the annual cycle in Kongsfjorden hydrography from July 2014 to July 2015 was studied by Noufal et al. (2016) and Venkatesan et al. (2016), Cottier et al., 2021 reported an increasing trend in the interannual summer and winter averages of Kongsfjorden water temperature and salinity with robust interannual variability. Increased Aw advection from the WSC during the last decade, 2010 to 2020, increased temperature and salinity, and significantly influenced local conditions in the innermost fjord, were evidenced by De Rovere et al. (2022) using the Mooring Dirigibile Italia (MDI) just a few kilometres away from the tidewater glaciers Kongsvegen, Kronebreen, and Kongsbreen. The depth-averaged temperature and salinity in the inner Kongsfjorden rose by 0.26 °C/yr and 0.3 psu/yr, respectively, during the warmest months of the year, while they remained mostly steady in the colder months. The subsurface Aw advection that reaches the inner fjord adds to the melting of glaciers from below (Willem et al., 2016) and the calving of glaciers, collectively known as frontal ablation rates, increasing the freshwater discharge to the fjord and glacial instability

(Luckman et al., 2015; Holmes et al., 2019). This would further delay sea ice formation in the fjord during the cold season, resulting in positive feedback to the already ongoing warming and its implications.

Seasonal and interannual variabilities induced by the physical processes are observed in the fjord biogeochemistry. According to Piquet et al. (2014), glacier melt significantly impacts underwater light conditions and salinity since it adds to particulate matter and freshwater. The heterogeneity in the water column has been increased by mixing different water masses in the last two decades. The differences are evident in the surface, intermediate, and bottom physical and biogeochemical properties (Hop and Wiencke, 2019). The altered and interrelated properties of light, nutrients, temperature, mixing, respiration, and grazing pressure affect the growth of phytoplankton bloom in the fjord (Cabre et al., 2016; Rohr et al., 2017). The spring bloom, also known as the seasonal phytoplankton bloom, occurs yearly. The beginning and duration of this productive season are mainly defined by solar insolation and sea-ice cover (Oziel et al., 2019). Limited incoming solar radiation during polar night causes phytoplankton to be subjected to an extended period of darkness. The bloom begins when light returns in spring and causes a thermal stratification (Winder and Cloern, 2010). The stratification is strongest in summer; the spring bloom exhausts surface nutrients and dies (Piquet et al., 2010). A Subsurface Chlorophyll Maximum (SCM) forms after the bloom, below the mixed layer (Coupel et al., 2011; Ardyna et al., 2013). This phenomenon is caused by sinking biomass from the surface, and the presence of low-light acclimated algal cells with elevated chlorophyll *a* (chl *a*), and an active community of phytoplankton that maintains a balance between light and nutrients (Brown et al., 2015; Barbieux et al., 2019). According to Ardyna et al. (2013), this productivity in the subsurface is expected to impact primary production and biogeochemical cycles.

The changes in phytoplankton bloom phenology such as timing and characteristics are a sensitive indication of climate change (Ji et al., 2010). Strong blooms composed of warm water species are being observed in the Arctic (Hegseth and Sundfjord, 2008). Similar to the other areas in the Arctic, the bloom phenology in Kongsfjorden

is also changing. The bloom timing has been shifted between March and June depending on varying conditions of sea-ice and inflows of Atlantic water (Hegseth and Tverberg, 2013; Hodal et al., 2012; Piquet et al., 2014; van De Poll et al., 2016). The intrusion of Aw increases phytoplankton blooms by increasing euphotic depth caused by the reduction of sea-ice coverage. On the other hand, it decreases blooms by melting tidewater glaciers, enhancing sediment release that shortens the euphotic depth (Payne and Roesler, 2019). The effect of Atlantic water intrusion and glacial melt was also apparent in the dynamics of the phytoplankton community (Hegseth and Sundfjord, 2008; Piquet et al., 2014). Recent studies show a shift of phytoplankton species to a smaller size in spring and summer because of the rising water temperatures (Rasconi et al., 2015; Assmy et al., 2023). This change at the lower trophic level of phytoplankton will affect the higher trophic levels like zooplankton, fish, and seabirds and influence the prevailing food web and ecosystem dynamics in the region (Vernet et al., 2017).

Most physical and biogeochemical studies in Kongsfjorden are focussed on spring and summer periods. Even though these studies contributed vital information, the discrete datasets could not describe what happened during the missed seasons, which is crucial, especially when dealing with intraseasonal and interannual variabilities. Moreover, very few studies cover the annual cycle (Cottier et al., 2007; Iversen and Seuthe, 2011; Seuthe et al., 2011; Darnis et al., 2017; David and Krishnan, 2017; Tverberg et al., 2019; Cottier et al., 2021; D'Angelo et al., 2024). It is impossible to have an in-depth understanding of winter processes and the role of convective mixing in the year-to-year variability in biogeochemistry, including the phenology of the phytoplankton blooms (Mayot et al., 2017) with a discrete approach. Besides, a discrete approach also limits the understanding of phytoplankton responses to short-time scale variabilities like winds, heatwaves, and upwelling events (Winder and Cloern, 2010).

1.4. Objectives and thesis outline

As detailed above, the interannual variabilities in Kongsfjorden are being studied. However, this aspect needs a continuous update, considering the significance of the

ongoing and rapid variabilities in the Arctic Ocean. The interannual biogeochemical variability is less explored compared to seasonal and overall trends. The intraseasonal scale physical and biogeochemical variabilities in Kongsfjorden are also a less explored arena, considering their importance as building blocks of the interannual and long-term changes. Even though few studies investigated the role of the NAO in the interannual A_w variations and responses inside Kongsfjorden, no detailed studies have yet been carried out on the relation between the Kongsfjorden variabilities and the regional climate modes in the region. Addressing these knowledge gaps advances understanding of complex climate and oceanographic processes that will contribute to improved climate models and predictions, which are now widely used, especially in not-so-accessible Polar Regions, mitigate broader climate impacts, and benefit global climate science initiatives in the Arctic.

Kongsfjorden's unique geographical and physical characteristics, accessibility, and strategic location, which are climatically sensitive, make it an ideal region for investigating climate change, its impacts, and the processes involved. Studying the recent variabilities in Kongsfjorden is useful for a better understanding of the other Arctic Ocean realms with similar settings, and further help to assess the environmental and ecological consequences. Even though the Arctic Ocean is generally a data-sparse region, the fjord has a history of scientific research and long-term datasets. This will help to compare recent observations with historical data to identify trends and assess the rate and magnitude of variabilities. Due to the reasons mentioned above, India has implemented a long-term project for monitoring the hydrography of Kongsfjorden to understand the changing environment as well as to contribute to a broader understanding of the system to the global scientific community. The National Centre for Polar and Ocean Research, Goa, India is the nodal agency under the Ministry of Earth Sciences for monitoring the Kongsfjorden hydrography and biogeochemistry since 2008. The data collected continuously over the last decade, from 2011 to 2020, have been utilised to conduct the present research.

Based on the identified scientific gaps, as mentioned above, the present doctoral thesis addresses the following three objectives.

1. To decipher the physical variabilities in Kongsfjorden waters at intraseasonal and interannual timescales.
2. To investigate the role of physical processes in the annual cycle of chemical (nitrate) and biological (chlorophyll) variability in Kongsfjorden using time series mooring measurements.
3. To understand the role of climate modes, for example, the Arctic Oscillation and the North Atlantic Oscillation, in the observed variabilities in Kongsfjorden.

The present thesis consists of six chapters, including this chapter: 1.

Chapter 2 describes the hydrography and biogeochemistry data measured by NCPOR in Kongsfjorden over the last decade (2011-2020), which is the central part of the thesis. The data were collected from Kongsfjorden through Conductivity-Temperature-Depth (CTD) surveys at routine locations in the fjord during summers since 2011 and year-round time-series observations through the Indian Arctic subsurface mooring (IndARC) from July 2014 to December 2020. The details on the other in-situ, remote sensing, satellite, and reanalysis data used are also given. Besides the details on the Regional Ocean Modelling System for the study region and the simulations done, the chapter provides methodologies used for data processing, analysis, calculations, and illustrations.

Chapter 3 addresses the first objective of the research work and deciphers the physical variabilities in Kongsfjorden at interannual and intraseasonal timescales from 2011 to 2020. The variabilities and trends in the Kongsfjorden water column hydrography as a whole and from the perspective of the two major water masses that reflect the two crucial ongoing processes in the overall Arctic Ocean - (1) SW (glacial melt freshwaters at the surface) and (2) AW (warm-saline subsurface to deep Atlantic water intrusions) also are given. Further, the factors and processes responsible for the surface and deeper fjord variabilities are determined and comprehended. The chapter details the intraseasonal variabilities in the Kongsfjorden water column hydrography

from the time series mooring observations from 2014 to 2020. A detailed study on intraseasonal oscillations and responsible factors is done for the first time in Kongsfjorden.

Chapter 4 addresses the second objective. The sensor-based time-series subsurface biogeochemistry and hydrography observations in the upper 50 m of Kongsfjorden from the IndARC mooring during 2015-2018 were used to discuss the annual, interannual, and intraseasonal variations in the fjord biogeochemistry and its coupling to the oceanic-atmospheric variations. A special focus is given to the annual cycle of biological (chlorophyll) and chemical (nitrate) variability over the years. The chapter discusses the interannual variation in spring and secondary blooms with their evolution and connection to distinct environmental conditions each year. Besides, the chapter also discusses the role of Atlantic water in the nutrient supply and variations in Kongsfjorden.

Chapter 5 addresses the third objective and evaluates the influence of regional climate modes of NAO and Arctic Oscillation (AO) in Kongsfjorden temperature, salinity, and AW volume. The chapter further focuses on the influence of the NAO on the fjord variabilities. Thus, the chapter provides a detailed picture of how the subpolar Atlantic influences the polar Kongsfjorden.

Chapter 6 summarises the key findings of the doctoral research.

Chapter 2

Data and Methodology

2.1. Definition of seasons

For the seasonal interpretations in the study, June to September is considered as summer, October to November as fall, December to March as winter, and April to May as spring (Svendsen et al., 2002).

2.2. In-situ Data

2.2.1 Summer Conductivity-Temperature-Depth (CTD) survey

The water temperature and salinity data collected using SBE 19plus V2 CTD profiler from July to September in Kongsfjorden (Fig. 2. 1) from 2011 to 2020 were used to study the summer hydrography variations. The accuracy of temperature and conductivity was ± 0.005 °C and ± 0.0005 S/m, respectively. Data processing followed the Seabird recommended methodology used in David and Krishnan (2017), which involved data conversion, wild edit, filter, align CTD, cell thermal mass, loop edit, derive, and bin average in the order of execution. Using the scan count of data acceptability obtained from the pressure versus scan count plots, the downcast and upcast data of pressure, temperature, and conductivity were selected. After the correction for the sensor time constant for pressure and temperature ‘wild edit’ omitted the spikes in the profiles. The wild edit estimated true standard deviation of the data with its first pass and then replaced the fitting data with bad flag in its second pass. This was followed by a data smoothening through forward and backward run low pass filter to remove the hastily changing data. Next, the data was aligned in time relative to pressure to confirm that the same parcel would be used to derive salinity, density, and DO. At this step, a 5 s advance time was given to the oxygen sensor relative to pressure. Further, the thermal mass effects of the conductivity cell were removed from the measured conductivity using the thermal anomaly amplitude of 0.03 and thermal anomaly time constant of 7. The oscillations by ship heave and initial surface soak of the CTD was further cleaned. After these steps, depth, salinity, and density were derived from the profiled pressure, temperature, and conductivity.

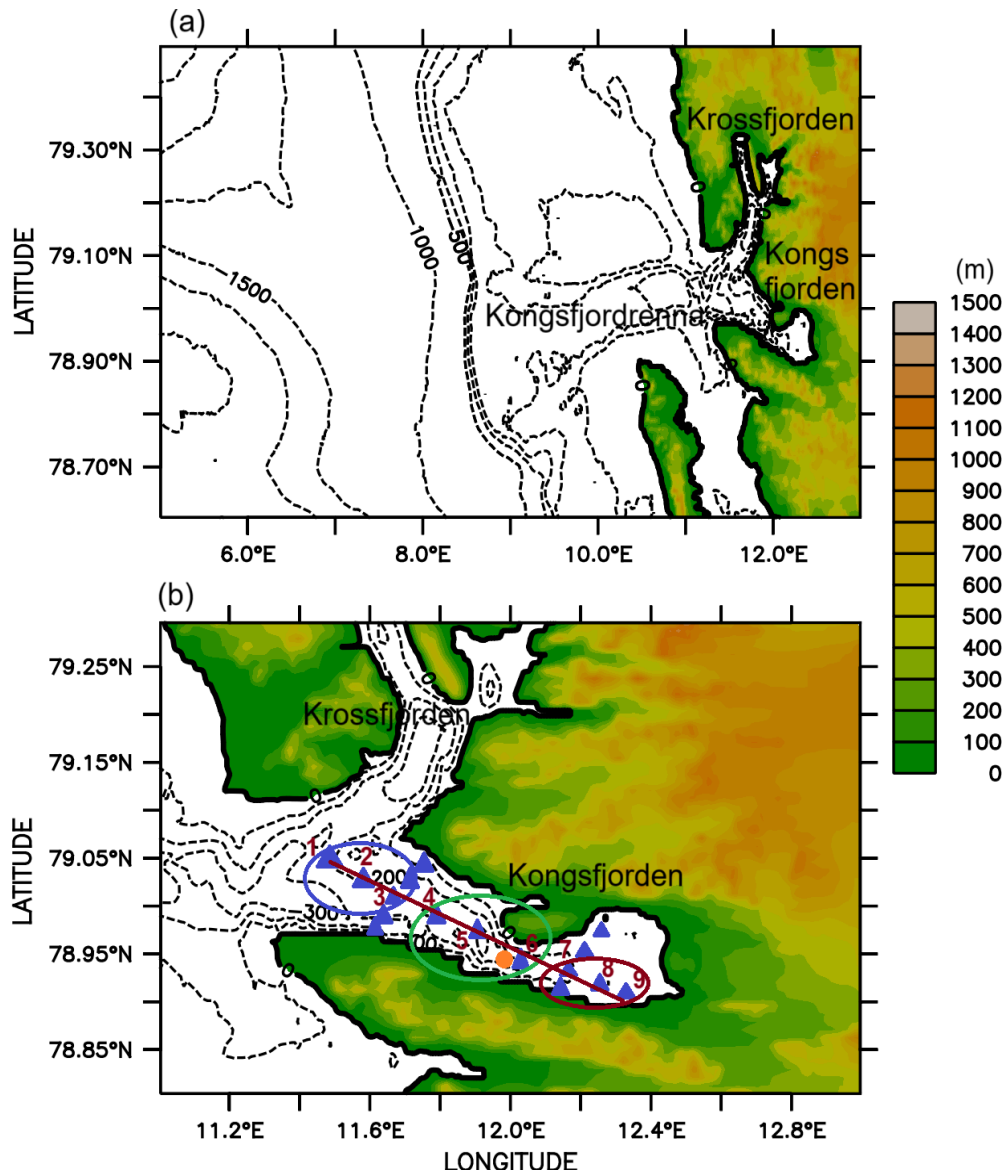


Figure 2. 1: (a) The ROMS model domain including Kongsfjorden-Krossfjorden and the western continental shelf-slope region. (b) Enlarged map of Kongsfjorden with the CTD locations (blue triangles). The red, green, and blue ellipses represent the inner, central, and outer zones of Kongsfjorden, respectively while the orange dot is the IndARC mooring location. The topography in shades and bathymetry in contours in (a) and (b) are in metres. The CTD locations from 1 to 9 in increasing order are located from the open ocean to the glacier side along the Kongsfjorden transect shown in red line. The CTD locations- 3.1, 3.2, 3.3, and 3.4 are across the Kongsfjorden mouth from its southern to northern shore with the first two locations on the left and the last two locations on the right of the CTD location 3. The CTD locations 7.1, 7.2, and 7.3 are across the inner fjord basin from its southern to northern shore with the first location on the left and the last two locations on the right of the CTD location 7.

The derived salinity was compared with the salinity analysed in the lab using Autosol from water samples collected at discrete depths and discrete times to confirm the data quality for the studies. Finally, the downcast data was averaged to 1 m depth bins, despiked again based on the standard deviations, and then used. Further, σ_t is calculated by subtracting 1000 from the density in kg m^{-3} (Pond and Pickard, 1983). Sixteen CTD locations over Kongsfjorden (Fig. 2. 1, Table 2.1) whose depth varied between < 50 m at the head of the fjord and ~ 350 m at the fjord mouth were used.

Table 2.1: Geographical coordinates of the CTD profiles in Kongsfjorden surveyed in summer-fall from 2011 to 2020.

Kongsfjorden		
CTD Locations	Latitude ($^{\circ}\text{N}$)	Longitude ($^{\circ}\text{E}$)
1.0	79.0354	11.2836
2.0	79.0099	11.4245
3.0	78.9931	11.5547
3.1	78.9840	11.5090
3.2	79.0100	11.6080
3.3	79.0250	11.6540
3.4	79.0410	11.7020
4.0	78.9755	11.6915
5.0	78.9587	11.8224
6.0	78.9408	11.9576
7.0	78.9228	12.0937
7.1	78.9170	12.0740
7.2	78.9400	12.1410
7.3	78.9540	12.1750
8.0	78.9050	12.2278
9.0	78.8951	12.3201

The CTD survey has been carried out roughly at a biweekly interval in summer every year. The number of CTD profiles each summer from 2011 to 2020 is shown in figure 2.2. The temperature and salinity profiles from the CTD were then averaged to monthly and summer means according to the objective of the present study.

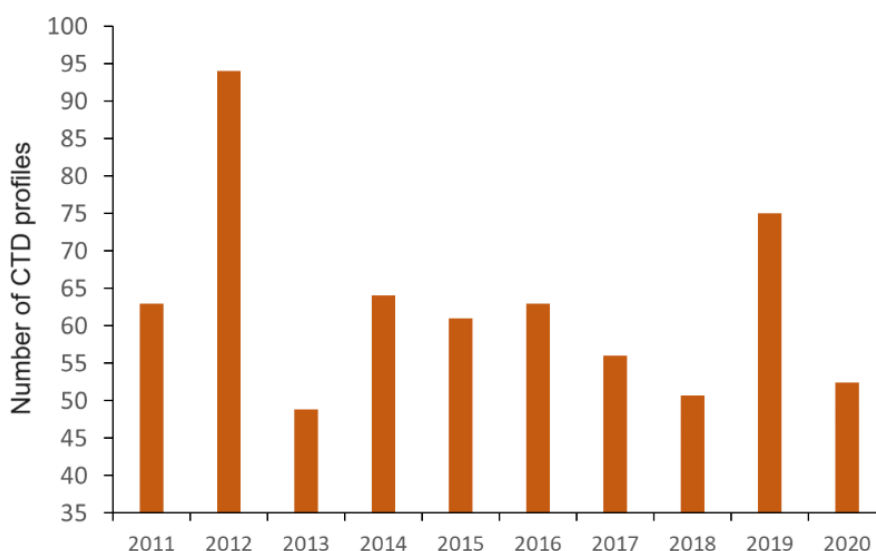


Figure 2.2: Number of CTD profiles collected during July-September from 2011 to 2020.

2.2.2 International Council for the Exploration of the Sea data

A few studies conducted earlier (Cottier et al., 2005, 2007; Nilsen et al., 2008; Promińska et al., 2017) for comparing the Kongsfjorden hydrography with various other west Svalbard fjords, reported that Kongsfjorden undergoes with more Aw intrusions and associated implications. Even though Kongsfjorden and Krossfjorden are so close to each other geographically and share a common mouth on the open ocean side, not many surveys are being carried out, except some biological and geological studies, for comparing the hydrography between them, especially during the present study period. Hence, utilising the CTD data during the same time from both the fjords a comparative study was conducted between Kongsfjorden and Krossfjorden to corroborate the earlier findings as well as to understand the factors or processes responsible and the pathways of circulation and heat transport causing the variability in Kongsfjorden. This comparison proved Kongsfjorden as a hotspot, compared to Krossfjorden and the other west Svalbard fjords, to conduct future studies for a comprehensive understanding of the study region.

For the comparison, the peak summer month of August from 2013 to 2018 has been considered as the data was available on the nearby dates in this month every year. Historical CTD data from the ICES database [<http://ocean.ices.dk/HydChem/>] also has been used to get the data for the nearby dates in August 2013 and 2014 from the

Kongsfjorden and Krossfjorden as the NCPOR data during these two periods are with larger time intervals. The ICES database is a repository of CTD data from expeditions conducted by different international research groups and also through the ships of opportunity. The details of the dates and the total number of profiles used are given in Table 2. 2. As the data is very specifically used for the comparison in the thesis, more details on the geographical locations of the profiles are given in Chapter 3 for easier connectivity between the results and discussions.

Table 2.2: The dates of the NCPOR and ICES CTD data availability with the total number of profiles in August from 2013 to 2018 in Kongsfjorden and Krossfjorden were used for the comparative study between the fjords.

	Kongsfjorden			Krossfjorden		
Year	Dates of CTD observation		Number of profiles	Dates of CTD observation		Number of profiles
	Indian	ICES		Indian	ICES	
2013		29, 31	17		21, 30, 31	22
2014		29, 30	10		29, 30	21
2015	17, 18		15	3, 17		17
2017	29		16	31		9
2018	6		6	6, 10, 17		19

2.2.3 Mooring Data

The IndARC, the first Indian Arctic mooring initiative, is a single-point subsurface ocean mooring deployed in the central Kongsfjorden at 78.94 °N and 12.01 °E (Fig. 2. 1) in July 2014. The IndARC has been designed as a subsurface mooring leaving a safe upper water depth of ~20 m after considering the interannual winter ice thickness and calved floating iceberg draft observed during summer in Kongsfjorden from the discussions with the researchers of the Scottish Association of Marine Sciences, Scotland, who were maintaining another mooring towards the open ocean side in Kongsfjorden since 2002. The mooring line consists of sensors collecting physical (temperature, salinity, and currents), chemical (DO, nitrate), and biological (fluorescence, PAR) parameters at discrete depths. A general schematic of the IndARC mooring line with the sensors to measure the above-mentioned variables and other components is given in Figure 2.3. The data details of the five deployments with

the sensors and their depths from 24 July 2014 to 31 December 2020 used in the present study are given in Table 2.3.

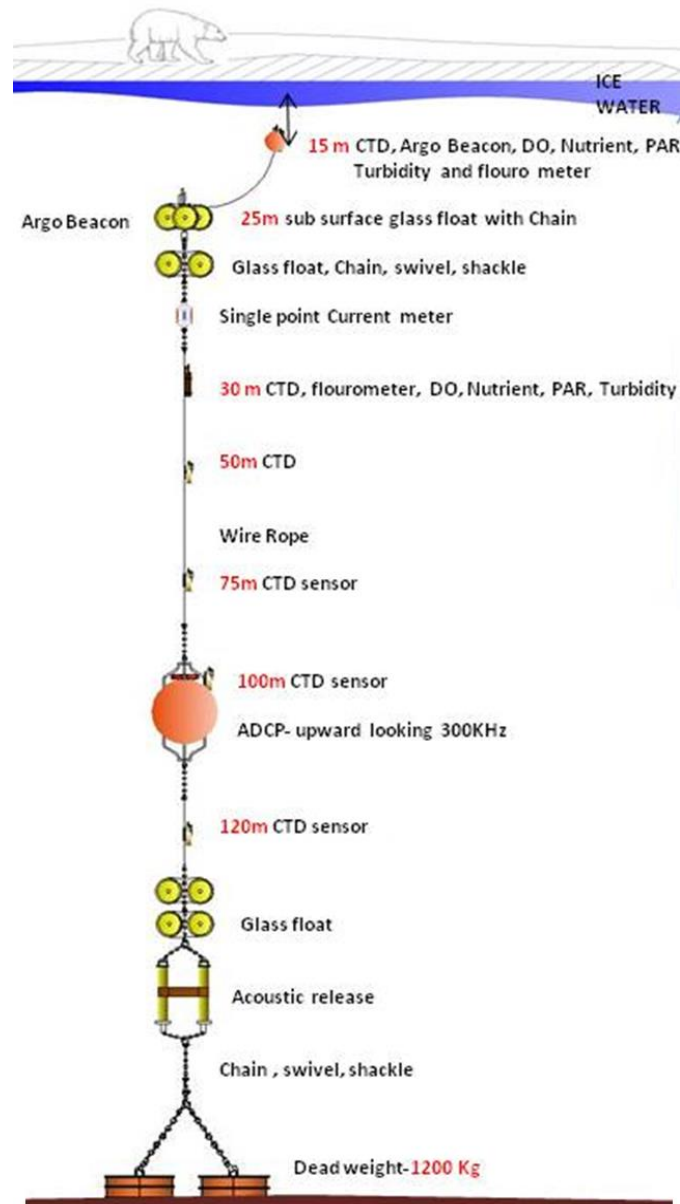


Figure 2.3: Schematic of the IndARC mooring with the sensors and other components at defined nominal depths. The depths varied slightly over the deployment years.

The deployment-retrieval gap was 4-6 days over the entire observational period. During the deployment in July 2019, NCPOR used an easily detachable link assembly unit, technically called a ‘weak link assembly’, attached from the upper glass floats as can be seen in Fig. 2.3 to collect data above 20 m. However, the weak link assembly

along with the sensors on it was detached in July 2021 when hit by a calved iceberg. This weak link assembly with sensors was then retrieved using a boat while the main mooring was retrieved later in August 2021. More technical details on the IndARC mooring configuration are given in Venkatesan et al. (2016).

Table 2.3: The location depth and the nominal depths of CT sensors, ADCP, and RCM during each deployment-retrieval period of the IndARC mooring in Kongsfjorden. The date in 2020 with a star denotes the time of the last data that is used in the present study and not the retrieval date.

Deployment-Recovery period	Location depth (m)	Nominal depths of CT sensors (m)	Nominal depth of ADCP (m) with frequency (KHz)	Depth of RCM (m)
24 July 2014 – 14 July 2015	192	22, 30, 55, 80, 105, 140	175 (150)	
20 July 2015 – 23 July 2016	190	27, 40, 62, 88, 114, 150, 170	175 (300)	
27 July 2016 – 18 July 2017	198	31, 41, 68, 96, 121, 161, 181	89 (300)	
22 July 2017 – 14 July 2019	200	25, 38, 60, 91, 115, 150, 170	84 (300)	
17 July 2019 – 31 December 2020*	198	17, 50, 100, 150, 182		28, 80

The hourly temperature and salinity were recorded using SBE 37-IMP microcat Conductivity-Temperature-Pressure (CT) sensors. The temperature accuracy is 0.002 °C for a range of -5 to 35 °C, and that of conductivity is 0.0003 S m⁻¹. The temperature and salinity data were collected at an hourly resolution. The depths of the upper and lowermost CT sensors were slightly different between the five deployments. However, the depth from 31 m to 140 m was found to be the common depth range over the entire observational period. The temperature and salinity data close to 31 m over the entire period was considered for the analysis done at the upper layer depth. The data was first vertically spline interpolated and from the last deployment period values closer to 31 m were considered with the other periods to denote the upper layer depth. The temperature and salinity data close to 140 m over the entire period was considered for the analysis done at the lower layer depth. Hence, hereafter, the upper and lower layer common depths are mentioned as 31 m and 140 m for ease of understanding.

Current velocities were recorded by the Workhorse model upward-looking Acoustic Doppler Current Profiler (ADCP) from Teledyne RD Instruments in the first four deployment periods. In the last deployment period, two SeaGuard Recording current meters (RCM) were used; the details of the depth of current sensors in the mooring line each year are given in Table 2.3. The current data resolution was 30 min and 4 m depth bins in the first two deployments, while it was one hour and 5 m depth bins for the latter years. The data were processed following the U.S. Geological Survey (USGS), Woods Hole Coastal, and Marine Science Center (WHCMSC) procedures (Côté et al., 2011). Only data with a percentage good limit of 80 and above were considered. The uppermost accurate data was found at 22 m depth from the surface. Hence, the ADCP could not give us an idea of the currents in the surface layer. The RCM's current resolution was also hourly. The hourly mooring temperature, salinity, and currents were then converted to daily averages for the present study. Since the biogeochemical variables are discussed only in chapter 2, their data details are given in that chapter to create a better connection in the fact line.

2.2.4 Weather Station Data

The meteorological parameters such as daily air temperature at 2 m above the ground, wind speed and direction at 10 m above the ground, and total precipitation were obtained from the Norwegian Meteorological Institute weather station at Ny-Ålesund [<http://sharki.oslo.dnmi.no>].

2.3 Reanalysis and other data

The daily net shortwave radiation (J m^{-2}), net longwave radiation (J m^{-2}), sensible heat (J m^{-2}), and latent heat (J m^{-2}) having a $0.75^\circ \times 0.75^\circ$ spatial resolution at 79.50°N and 12°E were extracted from ERA-Interim database (<http://apps.ecmwf.int/datasets>). From these, the daily net surface heat flux, F_{net} (W m^{-2}), was calculated. We use a sign convention where a positive surface heat flux corresponds to a net heat gain by the fjord and vice versa. The surface ocean temperature and ocean currents from the GLORYS12V1 product, the CMEMS global ocean eddy-resolving ($1/12^\circ$ horizontal resolution, 50 vertical levels) reanalysis [https://resources.marine.copernicus.eu/?option=com_csw&view=details&product_id=GLOBAL_REANALYSIS_PHY_001_030] and the bathymetry and topography from the SRTM30_PLUS estimated

topography, 30 s, global, v11 data [https://topex.ucsd.edu/WWW_html/srtm30_plus.html] were used to generate plots. The monthly NAO and AO indices are obtained from the National Centers for Environmental Information, the National Oceanic and Atmospheric Administration [<http://www.ncdc.noaa.gov/teleconnections/>]. The topography data used is the global ice sheet surface ETOPO1 with a spatial resolution of $0.0166667^\circ \times 0.0166667^\circ$ [<https://coastwatch.pfeg.noaa.gov/erddap/griddap/etopo180.html>].

2.4 Regional Ocean Modeling System (ROMS)

The Kongsfjorden-Krossfjorden hydrography and circulation were simulated using the three-dimensional numerical model- ROMS version 3.6 (Shchepetkin and McWilliams, 2003, 2005, 2009). The domain of the model comprises of the two fjords and their next-to-shelf and slope regions (Fig. 1.1, chapter 1). The western open boundary was stretched to the deep ocean including the WSC. The horizontal resolution of the model was set to ~ 500 m. The grid of the model was prepared using the ‘romspre-processing toolbox’ gridding package developed by ROMS-AGRIF (ftp://ftp.legos.obs-mip.fr/pub/romsagrif/DATA_ROMS/user_manual/doc_roms_agrif_v2.1_19_07_2010.pdf). The topography in the model was obtained from the Shuttle Radar Topography Mission (SRTM) 30+ bathymetry data. This bathymetry was developed by merging satellite-derived and ship-acoustic-sounding bathymetry (Becker et al., 2009). SRTM data incorporated the Arctic bathymetry from the International Bathymetric Chart of the Ocean (IBCAO). The model used forty vertical sigma layers with high resolution in the upper water column ($\sim O(10$ m)) and low resolution towards the deeper ocean. Hourly wind and atmospheric flux at the surface from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA5 reanalysis data [<https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels?tab=overview>] were used to force the model. The boundary and initial conditions were from the high-resolution ($1/12^\circ$) Hybrid Coordinate Ocean Model (HYCOM) and Navy Coupled Ocean Data Assimilation (NCODA) global $1/12^\circ$ reanalysis product (<https://hycom.org/data/glb0pt08/expt-19pt1>). The daily sea level, temperature, salinity, and velocity components were given at three open boundaries at the north, west, and south. Glacier discharge from the five tidewater glaciers in the Kongsfjorden was given from the subsurface snow model output (Pramanik et al.,

2018) as a point source in the model. Since the snow model output is available only for the tidewater glaciers around Kongsfjorden, the discharge was set to zero in Krossfjorden. The nonlocal K-Profile Parameterization mixing scheme (Large and Gent, 1999) was used to parameterize the sub-grid scale vertical mixing with background eddy viscosity of $10^{-5} \text{ m}^2 \text{ s}^{-1}$ and diffusivity of $10^{-6} \text{ m}^2 \text{ s}^{-1}$. Horizontal mixing in the model was parameterized by Smagorinsky-type viscosity and diffusion schemes (Muschinski, 1996; Griffies et al., 2000). The background eddy viscosity and diffusivity were set as $3 \text{ m}^2 \text{ s}^{-1}$ and $10 \text{ m}^2 \text{ s}^{-1}$, respectively. The model run was carried out from 2002 to 2018 due to the non-availability of freshwater discharge data in Kongsfjorden after that.

2.5 Calculations

2.5.1 Mixed Layer Depth

The density gradient, dp/dz where ρ is the ocean density and z is the depth, was computed from the temperature, salinity, and depth profiles from the temperature and salinity data in the fjords. The mixed layer depth (MLD) was then estimated using the criterion of depth at which the maximum density gradient was found (Meire et al., 2016).

2.5.2 Freshwater content and volume

The FWC in the fjords was calculated following Dølven (2015).

$$\text{FWC} = \int_0^z (S_{\text{ref}} - S/S_{\text{ref}}) dz$$

where S is the measured salinity, S_{ref} is the reference salinity, and z is the depth in meters of the water column.

Here, the S_{ref} was taken as 34.65, which demarcated the external and internal fjord water masses. Further, the total freshwater volume (FWV) was estimated for the Kongsfjorden by multiplying the surface area of the fjord, 267.81 km^2 (Promińska et al., 2017) with the FWC. The FWV in the Krossfjorden was estimated by multiplying the surface area of the fjord, 135 km^2 (length of 30 km multiplied by average width of 4.5 km), adapted from Svendsen et al. (2002) with the FWC.

2.5.3 Stability of water column

The stability of the water column can be learned from the Brunt–Väisälä frequency, which is a stratification measure. The Brunt–Väisälä frequency,

$$N^2 = -g/\rho \times (d\rho/dz),$$

where g is the acceleration of gravity, ρ is the potential density.

2.5.4 Fjord heat content

The vertically integrated fjord heat content (HC) in Jm^{-2} at the mooring position was calculated (Skogseth et al., 2003) over a common depth range of 31 to 140 m and 60 to 140 m for the mooring period using equation (1).

$$\text{HC} = C_p \rho_{\text{ref}} \int (T - T_{\text{ref}}) . dz \dots \dots \dots (1)$$

where C_p is the specific heat capacity of seawater taken as $3850 \text{ J/kg } ^\circ\text{C}$,

T_{ref} is the reference temperature, which was $-1.93 ^\circ\text{C}$, the freezing point of seawater was calculated for the salinity maximum value of 35.12 psu (S_{ref}) during the observation period in the fjord, and ρ_{ref} is the reference density in kg m^{-3} estimated from T_{ref} and S_{ref} .

2.5.5 Wind stress

The daily wind stress in N m^{-2} was calculated as $\rho_{\text{air}} C_D U^2$ (Stewart, 2007) where ρ_{air} is the density of air (1.22 kg m^{-3}) and ' C_D ' is the dimensionless drag coefficient (1.2×10^{-3}), and U is the daily wind speed in m s^{-1} .

2.6. Spectral Analysis

The chi-square distribution method (Emery and Thompson, 1997) was applied at a 95% confidence level to identify significant peaks in the time series data. This analysis compares spectral power against random noise to confirm meaningful frequencies. Another spectral analysis used was the continuous wavelet transformation (Grinsted et al., 2004) to understand variations in power across the time and frequency domains and thereby to detect periodicities and assess how they vary over time from time series data. Further, to examine the relationship between causative factors and fjord hydrography, wavelet coherence analysis (Grinsted et al., 2004) was conducted. The

wavelet coherence analysis reveals both short- and long-term oscillations and indicates the strength and phase of relationships. In these diagrams of wavelet coherence analysis, right-pointing arrows denote an in-phase relationship, while left-pointing arrows indicate the opposite. Both the continuous wavelet transformation and wavelet coherence spectrum analysis used the Morlet wavelet as the mother wavelet, as it balances time and frequency resolutions.

2.7. Graphical packages

The Flexible Extensible Research, Rendering, and Exploration Tool known as FERRET developed by NOAA's Pacific Marine Environmental Laboratory (PMEL) widely used for geophysical research is used for most of the visualisations in this thesis. The spectral analysis and plotting were done using the chi-squared and wavelet spectral toolboxes in MATLAB while the other line graphs were plotted using Origin2018.

Chapter 3

Physical variabilities in Kongsfjorden waters at interannual and intraseasonal timescales

3.1. Background

The chapter deciphers the physical variabilities in Kongsfjorden at interannual and intraseasonal timescales from 2011 to 2020 using data from summer CTD surveys and mooring. Interannual and seasonal studies have been done systematically in Kongsfjorden over the decades (Svendsen et al., 2002; Tverberg et al., 2019). The present study is focussed on the latest decade, a continuation and an addition to the existing knowledge base. This becomes important under the scenario of ongoing rapid changes in the region. On the other hand, studies on subinertial coastally trapped waves with a 2-day periodicity (Inall et al., 2015), tides (Svendsen et al., 2002; Tverberg et al., 2019), near-inertial and internal waves (Subeesh et al., 2022), etc., that play a significant role in the Atlantic water intrusions to the fjord are gradually increasing with the new state of decreasing sea ice and more open waters in the region. Very few studies have (De Rovere et al., 2022) investigated the spectra between the seasonal and lower periodicity variabilities. These periodicities, which are the intraseasonal oscillations (ISO), which are also crucial building blocks to the larger period variabilities, have not yet been done in detail in the fjord, which forms the second part of this chapter.

The chapter is organised as follows:

Sections 3.2 and 3.3 deal with the interannual variabilities observed in Kongsfjorden from 2011 to 2020. Section 3.2 particularly is a comparative study that shows Kongsfjorden as a hotspot that responds more to changes in the Arctic than Krossfjorden and the other fjords in West Svalbard. The section proves Kongsfjorden as a perfect miniature system to study further the recent changes in this region of the Arctic Ocean. This is followed by Section 3.3 on the interannual summer variations in Kongsfjorden from 2011 to 2020, the observed trends, and the factors responsible for

the variations. Section 3.4 describes the interannual variabilities of other seasons using the time series data from the Indian mooring in Kongsfjorden since 2014 while section 3.5 reveals the intraseasonal variabilities in Kongsfjorden hydrography and its triggering mechanisms using the time series mooring data. Finally, section 3.6 summarises the findings.

3.2. Kongsfjorden versus Krossfjorden, its twin fjord, and other west Svalbard fjords

Kongsfjorden has been known to respond more than the other fjords in west Svalbard to any variabilities that this part of the eastern Arctic Ocean undergoes. Even though Kongsfjorden hydrography surveys have been done since 2011 summer, the observations in Krossfjorden were started a few years later. Hence, in this section, hydrography data collected from Kongsfjorden and its twin fjord- Krossfjorden, with which it shares a common mouth, in the peak summer month of August from 2013 to 2018 were used, based on the common data availability for both the fjords.

Out of the Kongsfjorden-Krossfjorden system situated between the latitudes of 78° and $13^{\circ} 40' N$ and $79^{\circ} 30' N$ and longitudes of $11^{\circ} 03' E$ and $13^{\circ} 06' E$ (Fig 1.2), Krossfjorden which is further north to Kongsfjorden is 30 km long and 3–6 km in width, with a volume of 25 km^3 (Svendsen et al., 2002). Krossfjorden is aligned from the north to south direction, whereas Kongsfjorden is aligned from the northwest to southeast, as mentioned in Chapter 1. Both the fjord channels intersect at Kongsfjordrenna, the deep glacial basin. The depth (MacLachlan et al., 2007) of the outer part of Krossfjorden is $\sim 374 \text{ m}$. Krossfjorden bifurcates into two minor fjords towards the north, with water depths $\sim 250\text{--}100 \text{ m}$ (Sexton et al., 1992): Lilliehookfjorden and Mollerfjorden by Kong Haakons Halvøya. The glacier discharge from Lilliehookbreen (a tidewater glacier), at the head of Krossfjorden strongly influences the fjord system, while Kongsfjorden has five tidewater glaciers. The fjords are important climatically and oceanographically and are located east of the WSC (Cottier et al., 2005). The overall variability and increase in the Atlantic water inflows to the eastern Arctic can be identified as a change in the intruded Atlantic water volume inside the fjords, and this is primarily due to their location (Svendsen et al., 2002; David and Krishnan, 2017; Tverberg et al., 2019). As in the case of

Kongsfjorden, freshwater sources in Krossfjorden are meltwater, snowmelt, river runoff, groundwater discharge, and precipitation (Cottier et al., 2005). The hydrographic properties of these fjords, affected by the Atlantic and glacier meltwaters, result in strong physical gradients (Svendsen et al., 2002; Sundfjord et al., 2017).

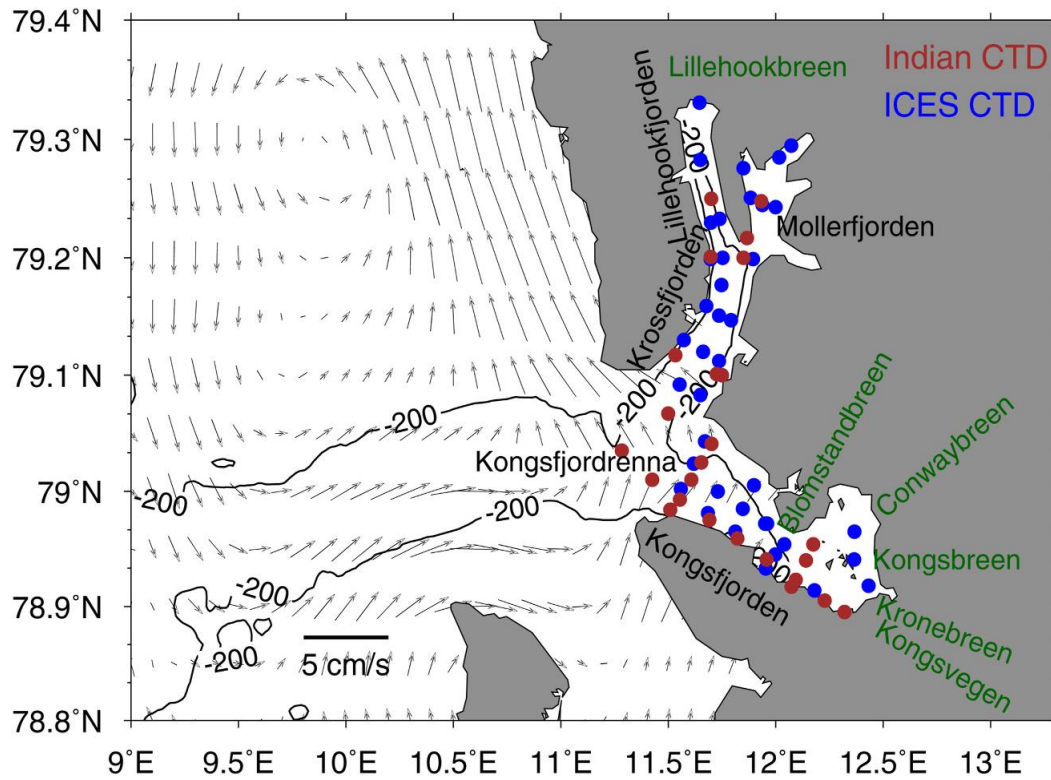


Figure 3.1: The CTD locations used in Kongsfjorden- Krossfjorden are marked with blue (ICES) and red (NCPOR) dots. Isobath in the diagram denotes 200 m depth. The ocean current in vectors are from the GLORYS12V1 product, the CMEMS global ocean eddy-resolving (1/12° horizontal resolution, 50 vertical levels) reanalysis [https://resources.marine.copernicus.eu/?option=com_csw&view=details&product_id=GLOBAL_REANALYSIS_PHY_001_030]. The bathymetry and topography are from the SRTM30_PLUS estimated topography, 30 s, global, v11 data [https://topex.ucsd.edu/WWW_html/srtm30_plus.html].

There are comprehensive reviews (Svendsen et al., 2002) focused on modern oceanographic and physical conditions (Tverberg et al., 2019) as well as marine biology (Hop et al., 2002, 2006) and ecosystems in the Kongsfjorden–Krossfjorden system (Hop and Wiencke, 2019). However, with Kongsfjorden being one of the most

extensively studied fjords in west Svalbard, studies reported an increase in sea surface temperature by 0.6 °C in Krossfjorden during August over 60 years from 1953 to 2014 (Guruvayoorappan et al., 2019). The study reasoned it to increase in CO₂ forcing and air-sea interaction rather than by the Atlantic water. Hence, the recent variabilities in Kongsfjorden and Krossfjorden are compared in this study to confirm the direction of the research focus in Kongsfjorden. In addition to the hydrography data collected by NCPOR in these fjords, the ICES historical CTD data from 2013 to 2018 are also used (Table 3. 2) to determine the continuation and maximum profiles to be included in the study. The spatial distribution of these CTD locations in the fjords is shown in Fig 3.1, and the geographical locations of the Kongsfjorden and Krossfjorden stations are given in Tables 2.1 and 3.1, respectively. The CTD survey, which was conducted with the total number of profiles in August each year, is presented in Table 3.2. The data for the same period in both the fjords was available in all the years from 2013 to 2018 except 2016. The temperature and salinity profiles in August were averaged to obtain the monthly mean.

Table 3.1: Geographical locations of the Indian CTD survey in Krossfjorden.

Krossfjorden		
CTD Locations	Latitude (°N)	Longitude (°E)
I1	79.0667	11.5000
I2	79.1167	11.5333
I3	79.1013	11.7245
I4	79.1000	11.7500
I5	79.2009	11.6968
I6	79.2000	11.8500
I7	79.2167	11.8667
I8	79.2500	11.7000
I9	79.2476	11.9334

3.2.1. Hydrography

Generally, in summer (June–September), the surface temperature of Kongsfjorden–Krossfjorden increases from June, peaks in July–August, and then decreases (Cottier et al., 2005; Hegseth and Tverberg, 2013; Sundfjord et al., 2017). Fig. 3.2 shows the water temperature, salinity, and density averaged over each fjord in August. The mean water column temperature (vertical) of the fjord varied between 0 and 7 °C. The mean vertical water temperature in Kongsfjorden was warmer by ~0.1–1 °C compared to

Krossfjorden except the upper 30 m in 2015 and 2018, which was ~ 1.5 °C and 1 °C warmer, respectively, in Krossfjorden.

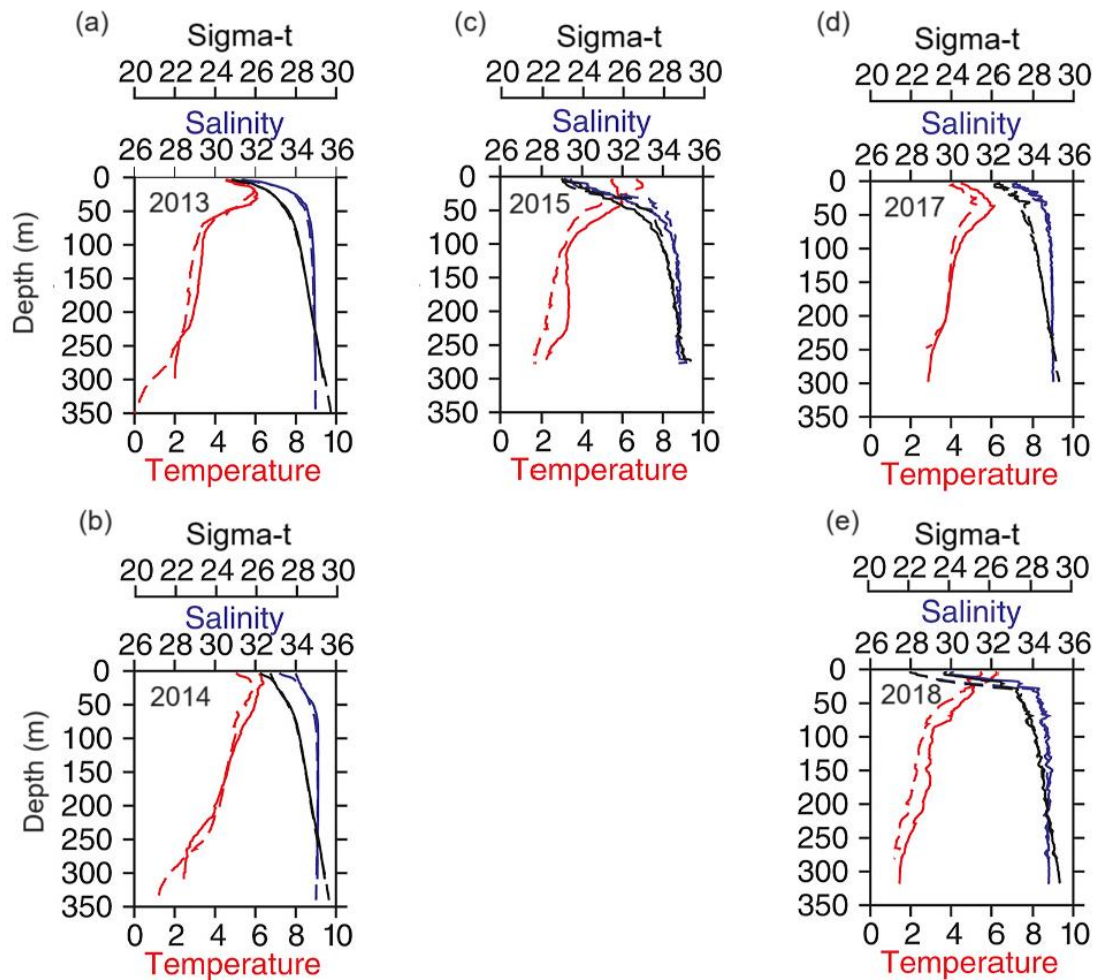


Figure 3.2: The fjord averaged vertical profile of water temperature in °C (red), salinity in psu (blue), and sigma-t (black) in Kongsfjorden and Krossfjorden for August in (a) 2013, (b) 2014, (c) 2015, (d) 2017, and (e) 2018. The continuous lines represent Kongsfjorden profiles, and the dashed lines represent Krossfjorden profiles.

With the arrival of summer, the fjord surface water temperature rises, and the lower water column is gradually heated from above (Sundfjord et al., 2017), resulting in a vertical two-layer system of warmer, less-saline waters over colder, high-saline waters. During the peak summer months (August–September), the freshwater discharge in Kongsfjorden-Krossfjorden would be maximum (Svendsen et al., 2002) from the snow melting, surface and submarine melting of glaciers, calving, river runoff, groundwater discharge, and precipitation (Cottier et al., 2005) forming an

uppermost thin, cold, less-saline layer over the existing two layers. Thus, in August, the fjord transforms into a structure with warm, high-saline waters sandwiched between the cold, less-saline surface, high-saline deep waters, balancing the density stratification, thus transforming into a three-layer system. This formation of warmer water temperatures between the colder waters up and down is the temperature inversion layer in the fjord. Both Kongsfjorden and Krossfjorden showed temperature inversions. The inversion in Kongsfjorden was often deeper than that in Krossfjorden (Fig. 3.2). The inversion layer depth was 40 m in Kongsfjorden and 30 m in Krossfjorden in August 2017, which is the deepest and strongest inversion during the observational period. However, the rate of temperature rise within the inversion layer in both the fjords was $0.04\text{ }^{\circ}\text{C/m}$ in 2017. The surface salinity varied between 28.0 and 33.3 psu, while it varied between 28.0 and 35.0 psu vertically, with the minimum at the surface. Salinity exhibited minimal variability in the lowermost depths compared to the surface. The salinity of the fjords did not significantly differ, especially below 100 m. But in Kongsfjorden, the upper 15–20 m in August 2014 and 2018 was more saline (1–2 psu). The density differences between the fjords closely followed the salinity variations.

Generally, the mixed layer depth (MLD) in Kongsfjorden and Krossfjorden is deep in winter. It then shallows in summer with strong upper water column stratification created by the surface meltwater discharge into the fjords (Svendsen et al., 2002). The MLD (subsection 2.4.1, chapter 2) and FWV (subsection 2.4.2, chapter 2) in the fjords were estimated from the August profiles in Kongsfjorden and Krossfjorden and then averaged for each of the fjords. The depth of 34 isohaline from the surface in the profiles was considered as the thickness of SW while the $T_{\max}$ and $S_{\min}$ within this depth were determined from each of the profiles before averaging. The thickness of the AW layer was calculated by subtracting the upper depth from the lower depth, both determined upon satisfying the characteristic temperature and salinity values of AW (Table 1.1) while the $T_{\max}$ and $S_{\max}$ between these depths were determined from each of the profiles in the fjords in August which were then averaged. These averages for August from 2013 to 2018 are given in Table 3.2.

Table 3.2: The MLD, Surface water mass (SW) characteristics (thickness, maximum temperature ($T_{\max}$), and minimum salinity ($S_{\min}$)), Atlantic water mass (AW) characteristics (thickness, $T_{\max}$, and maximum salinity ($S_{\max}$)), and freshwater volume (FWV) in August from 2013 to 2018 in Kongsfjorden and Krossfjorden.

Year	Kongsfjorden							
	MLD (m)	Thickness (m) of SW	$T_{\max}$ of SW ($^{\circ}\text{C}$)	$S_{\min}$ of SW	Thickness (m) of AW	$T_{\max}$ of AW ($^{\circ}\text{C}$)	$S_{\max}$ of AW	FWV (km^3)
2013	6	50	6.9	30.2	100	7.5	35.0	0.3
2014	50	25	6.3	33.6	200	6.5	35.2	0.11
2015	10	100	7.0	26.5	50	3.9	35.0	1.45
2017	11	30	6.0	31.2	230	6.0	35.9	0.21
2018	18	60	7.0	28.4	35	4.8	35.4	0.64

Year	Krossfjorden							
	MLD (m)	Thickness (m) of SW	$T_{\max}$ of SW ($^{\circ}\text{C}$)	$S_{\min}$ of SW	Thickness (m) of AW	$T_{\max}$ of AW ($^{\circ}\text{C}$)	$S_{\max}$ of AW	FWV (km^3)
2013	8	60	6.8	29.6	75	7.2	34.8	0.19
2014	9	25	6.2	32.0	200	6.5	35.2	0.08
2015	33	90	7.6	26.7	20	3.8	35.0	0.71
2017	16	30	4.8	31.4	230	6.0	35.8	0.09
2018	29	50	8.0	26.1	20	4.4	35.1	0.64

The MLD in both the fjords showed large interannual variability. Except in 2014, when it was 50 m, the Kongsfjorden MLD in August varied between 6 and 18 m. Out of this, the shallowest MLD was in August 2013 in Kongsfjorden. The MLD in Krossfjorden varied between 8 and 33 m. The deepest MLD was in 2015, and the shallowest was in 2013. Thus, except in 2014, the MLD in Kongsfjorden was shallower than in Krossfjorden indicating more stratified waters due to more freshwater discharge in Kongsfjorden. The difference in the MLD between these fjords varied between 2–23 m, except in 2014 when the Kongsfjorden MLD was deeper than that in Krossfjorden.

3.2.2. Circulation and heat flow

The spatial variation in the hydrography was analysed to understand the circulation and heat flow in the region. With bifurcation of Krossfjorden towards the glacier side, two parallel transects along the western and eastern coasts of Krossfjorden and a

transect along Kongsfjorden along which the fjord hydrography shows strong lateral gradient are present (Svendsen et al., 2002).

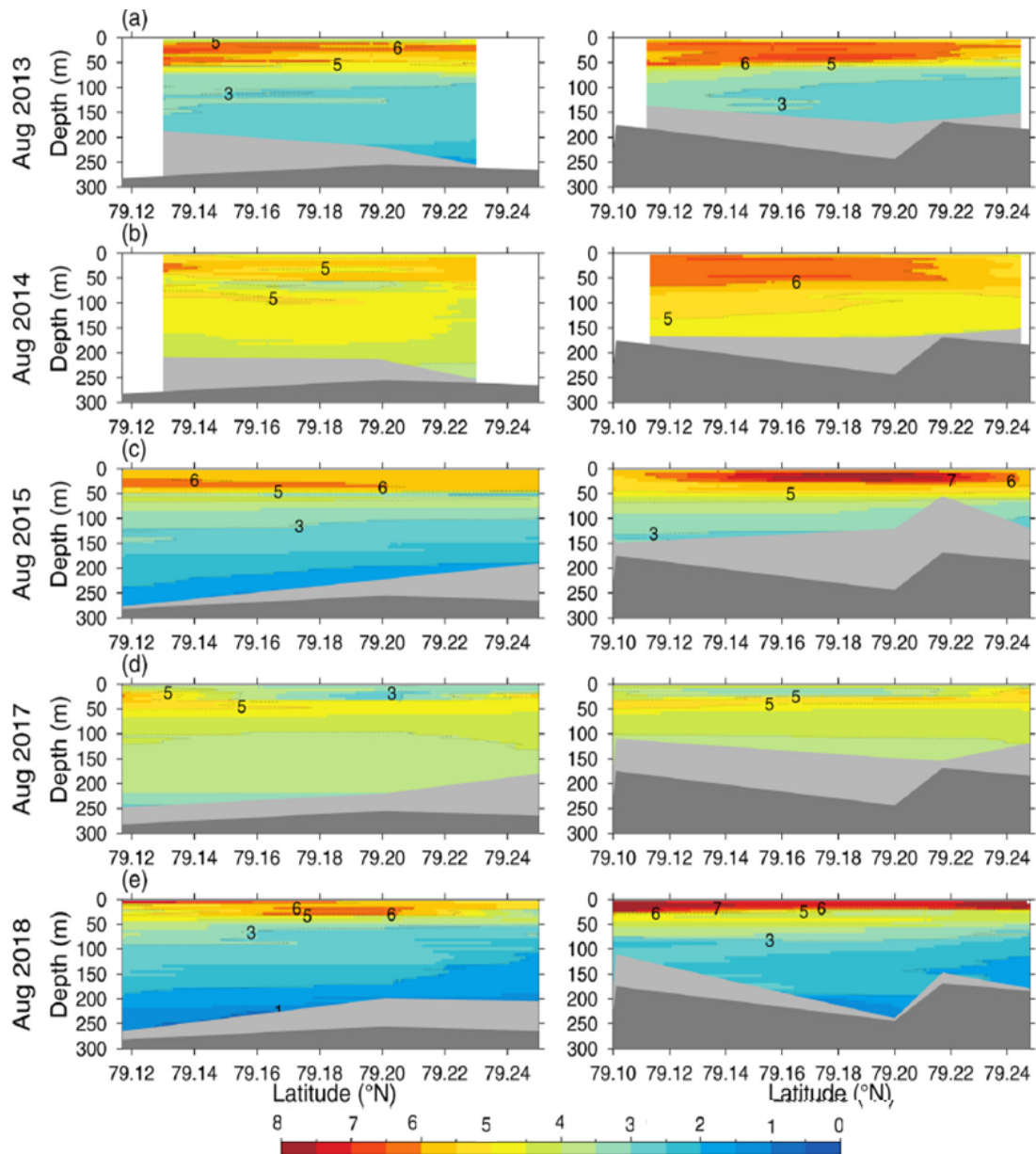


Figure 3.3: Spatial distribution of water column temperature ($^{\circ}\text{C}$) along the western (left) and eastern (right) Krossfjorden in August (a) 2013, (b) 2014, (c) 2015, (d) 2017, and (e) 2018. Lower to higher latitudes on the x-axis correspond to outer to inner Krossfjorden. The dark grey shade in the plots indicates the depth of CTD locations while the light grey shade represents no data region.

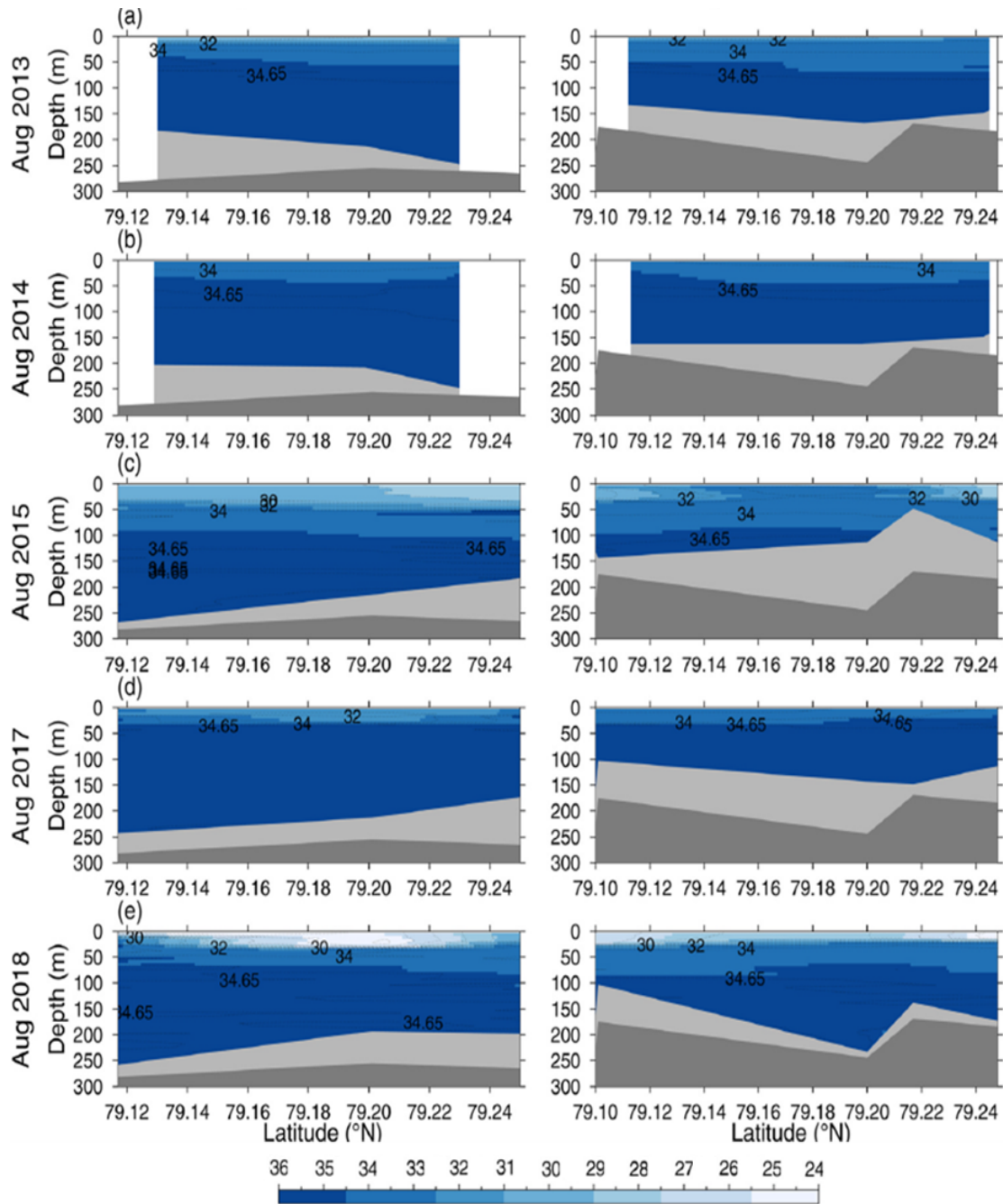


Figure 3.4: Spatial distribution of water column salinity (psu) along the western (left) and eastern (right) Krossfjorden in August (a) 2013, (b) 2014, (c) 2015, (d) 2017, and (e) 2018. Lower to higher latitudes in the x-axis correspond to outer to inner Krossfjorden. The dark grey shade in the plots indicates the depth of CTD locations while the light grey shade represents no data region.

The western side of Krossfjorden, which has large calving (Den Norske Los, 1992) and freshwater discharge (Ingvaldsen et al., 2001), is colder than the eastern side of

Krossfjorden. This temperature difference between the western and eastern sides is evident in the vertical temperature distribution (Fig. 3.3) and salinity (Fig. 3.4) from Krossfjorden. The upper 60 m eastern Krossfjorden was 0.5–1.0 °C warmer than the western side in August 2013 (Fig. 3.3 (a)). However, the salinity (Fig. 3.4 (a)) did not differ much. The temperature difference in the entire water column between the eastern and western Krossfjorden was again observed in 2014 (Fig. 3.3 (b)) and 2015 (Fig. 3.3 (c)). In 2015, the eastern fjord was 1–1.5 °C warmer than its western side (5–5.5 °C).

In 2017, even though the upper 25 m water column was colder than the earlier observations, the temperature contrast between both sides was still present (Fig. 3.3 (d)). The increase in the upper water temperature from the north (head) to the south (mouth) along the western transect was also evident in 2017. The upper 50 m salinity (Fig. 3.4 (d)) showed a prominent freshening in the western Krossfjorden. In 2018, the core of the upper layer warming was up to 15 m on the western side and 50 m on the eastern side (Fig. 3.3 (e)). Besides, the eastern and western Krossfjorden had colder deep waters than the previous years. Thus, as observed, the east Krossfjorden was 0.5–1.5 °C warmer than the west Krossfjorden during the peak summer.

Fig. 3.5 shows the spatial distribution of water column temperature and salinity in Kongsfjorden. The CTD profiles were not available at the outer and innermost locations in August 2013 (Fig. 3.5 (a), (b)) and 2014 (Fig. 3.5(c), (d)) while it was not available in the outermost location (Fig. 3.5 (e), (f)) in 2015. In August 2013, a core of maximum water column temperature was found around 30–40 m (Fig. 3.5 (a)) with cooler waters above and below it, as was in Krossfjorden (Fig. 3.3 (a)), which is the temperature inversion. The ~30 m shallower isohaline of 34.5 (Fig. 3.5 (b)) than that in Krossfjorden (Fig. 3.4 (a)) indicates more saline deep waters in Kongsfjorden. The water column, especially below 50 m, was warmer in 2014 (Fig. 3.5 (c)) than in 2013. On the other hand, the upper 50 m was fresher (Fig. 3.5 (d)). The water temperature (Fig. 3.5 (c)) and salinity distribution in 2014 (Fig. 3.5 (d)) was very similar to that found in the eastern Krossfjorden (Figs. 3.3 (b), 4 (b)). But, the temperatures in the upper Kongsfjorden were 0.5–1 °C higher than that in the Krossfjorden (Fig. 3.3 (b)).

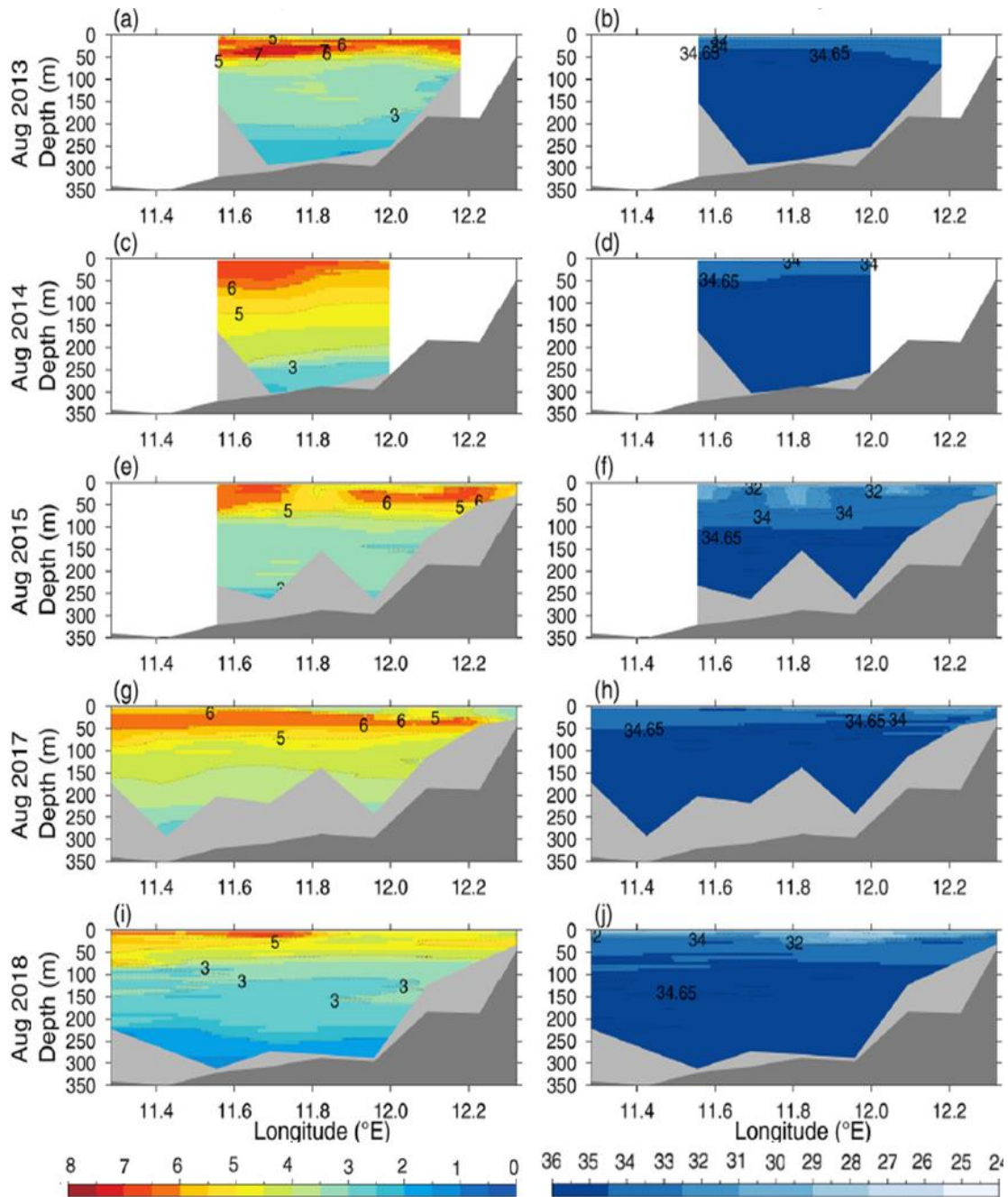


Figure 3.5: The water column temperature in °C (left) and salinity in psu (right) from the CTD profiles in Kongsfjorden in August (a, b) 2013, (c, d) 2014, (e, f) 2015, (g, h) 2017, and (i, j) 2018. Lower to higher longitudes on the x-axis correspond to the outer to the inner fjord. The dark grey shade in the plots indicates the depth of the CTD locations while the light grey shade represents no data region.

The salinity distribution in Kongsfjorden (Fig. 3.5 (d)) and Krossfjorden (Fig. 3.4 (b)) were similar. Kongsfjorden showed a temperature inversion in 2015 (Fig. 3.5 (e)) centred around 30 m (6.5–7 °C), similar to that found in the eastern Krossfjorden (Fig.

3.3 (c)). The upper layer salinity was lower in 2015 (Fig. 3.5 (f)) for both the fjords than in 2013 and 2014. The head-to-mouth lateral gradient in the temperature (Fig. 3.5 (g)) and salinity (Fig. 3.5 (h)) is seen in the upper 50 m Kongsfjorden in 2017, as was seen in the Krossfjorden (Fig. 3.3 (d)). The water column hydrography distribution appeared similar to that in 2014. As discussed earlier, August 2018 had colder waters than the previous years except in 2015, especially below 50 m, in both the fjord systems (Figs. 3.5 (i), 3 (e)). The spatial distributions in August 2018 (Fig. 3.5 (i), (j)) appeared similar to 2015 (Fig. 3.5 (e), (f)). In short, the Kongsfjorden transect behaved closer to the eastern Krossfjorden but was warmer and saline than Krossfjorden. In addition, the eastern Krossfjorden was warmer and more saline than the western part of it. This spatial variability was examined in detail to understand more about the circulation and heat flow patterns in the fjords using ROMS model simulations (details in Chapter 2) due to the limited in-situ current data available spatially. Since summer is the time of maximum Atlantic water intrusions (Svendsen et al., 2002; Tverberg et al., 2019) and also as the freshwater discharge data was available for Kongsfjorden for the period, the ROMS model simulations were done for 2013–2018. Further, the model output for August is considered for studying the circulation and heat flow during summer in the fjords. The simulated temperature, circulation pattern, and the Atlantic water inflows closely follow the earlier model simulations (Ingvaldsen et al., 2001; Svendsen et al., 2002; Cottier et al., 2005; Arild et al., 2017).

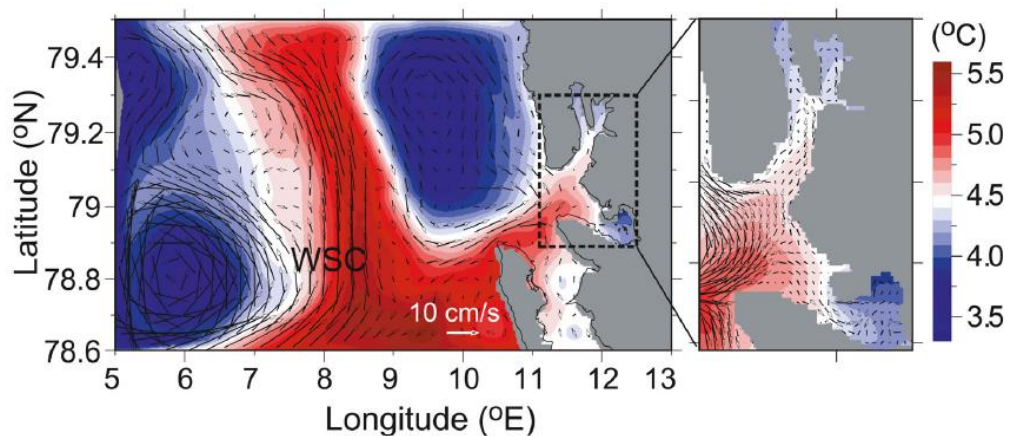


Figure 3.6: The ROMS simulation of the ocean temperature in °C (shade) overlaid by currents in cm s^{-1} (arrows) averaged over 0–150 m in August from 2013 to 2018 for (left) the entire model domain and (right) the Kongsfjorden-Krossfjorden region enlarged.

Even though interannual variations existed in the magnitude of the simulated ocean temperature and currents during 2013–2018, the general spatial pattern of the heat flow and circulation in the study region remained the same throughout the simulations. Hence, a climatology of the ocean temperature and current simulations was created for August 2013–2018 (Fig. 3.6) to show the general spatial circulation and heat flow pattern in the study region. The warm Atlantic waters from the WSC enter Kongsfjorden along its southern shore, while its direct entry to Krossfjorden happens through the fjord's eastern side. This entry of Atlantic water supplies heat to the inner part of the fjords. The heat flow that reaches the head of Kongsfjorden can further result in glacial calving and melting, resulting in cold waters from the inner Kongsfjorden flowing out of the fjord through its northern boundary. A part of this outflow enters Krossfjorden through its eastern side, along with the direct Atlantic water entry into the fjord. The flow through the eastern Krossfjorden takes a U-turn at the head and is seen reaching the glacier at the head of the western part of the fjord, Lillehookbreen, which can further result in glacial calving and melting. There is an outward flow along the western Krossfjorden, which can cause the freshwater outflow from Krossfjorden to the open ocean area. The observed spatial and vertical temperature and salinity distributions along Kongsfjorden and Krossfjorden prove this circulation pattern and heat flow pathway in the region.

3.2.3. Water masses

The Temperature (T)-Salinity (S) diagrams were used to decipher the water masses. The water mass classification of Cottier et al. (2005), adapted to conditions north of Svalbard, which is influenced more by melting sea ice, which corresponds closer to the present-day situation, has been followed for the study. Considering the geographical proximity of Kongsfjorden and Krossfjorden, this study adapted the same water mass characteristics ranges (Cottier et al., 2005) to delineate the water masses. The water masses and their generation mechanisms are detailed in section 1.2.3 in Chapter 1.

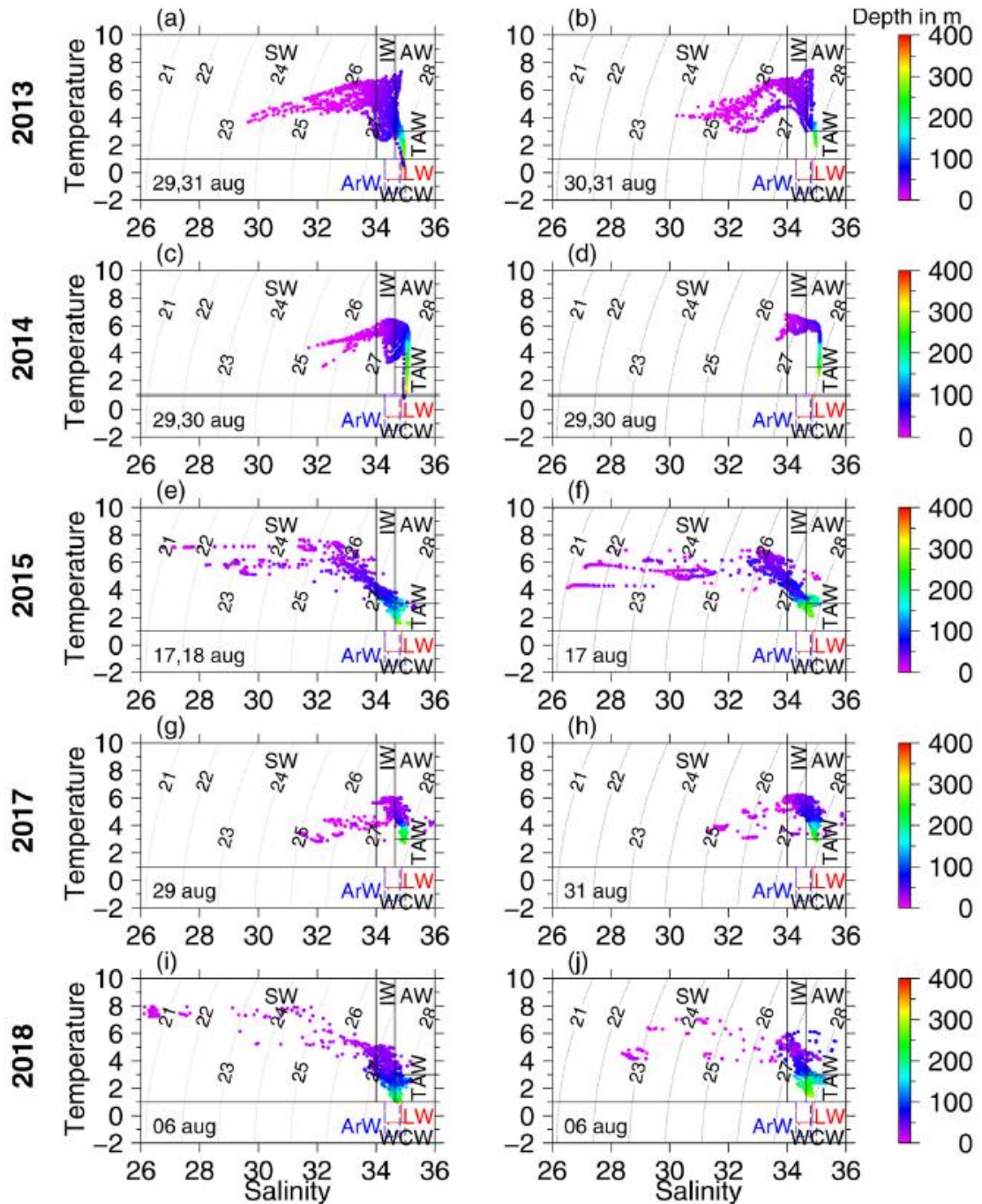


Figure 3.7: T-S diagrams in (left) Krossfjorden and (right) Kongsfjorden in August of (a, b) 2013, (c, d) 2014, (e, f) 2015, (g, h) 2017, and (i, j) 2018. The colour bar denotes the depth range of existence of each water mass.

The SW, IW, TAW, and AW were found in Krossfjorden (Fig. 3.7 (a), (c), (e), (g), (i)) and Kongsfjorden (Fig. 3.7 (b), (d), (f), (h), (j)). The SW layer thickness was the same (2013, 2014, and 2017) or higher (2015 and 2018) in Kongsfjorden than that in Krossfjorden (Table 3.2). This corresponds with the shallow MLD due to more

stratified waters in Kongsfjorden than in Krossfjorden. But, unlike Krossfjorden, LW was not found in Kongsfjorden in 2013 (Fig. 3.7 (b)). However, as the CTD profiles in Kongsfjorden were unavailable from the outer and innermost locations in August 2013, it is difficult to confirm whether LW was there in the data gap locations in 2013. The AW was dominant again in 2014 (Fig. 3.7 (d)) and the TAW in 2015 (Fig. 3.7 (f)). The AW became prominent again in 2017 (Fig. 3.7 (h)). In 2018, the SW was warmer and more saline (Fig. 3.7 (j)) than that in Krossfjorden (Fig. 3.7 (i)). TAW became dominant again in 2018. The interannual variability in the dominance of AW and TAW had the same pattern inside Kongsfjorden as it was in Krossfjorden (Table 3.2), with the dominance of AW in 2013, 2014, and 2017 and TAW in 2015 and 2018.

The dominance of AW in 2014 and TAW in 2015 was reported earlier (Promińska et al. 2017) inside both Kongsfjorden and Hornsund (Fig 1.1). Thus, it can be inferred that all three fjords in west Svalbard had the same interannual variations concerning the Atlantic-origin waters found in the fjords. Moreover, the AW in the summer of 2014 was found at 25–250 m in Kongsfjorden in their study, which corresponds with the 200 m thickness of AW that is in the present observations. 2014 summer has been characterised as an exceptional year over the analysed period from 2001 to 2015 as large areas of the fjords were filled with pure Atlantic water (Promińska et al., 2017).

In the present study, the entire water column in the fjords in 2014 and 2017 showed similar hydrography with the dominance of AW. The increased inflow of Atlantic water into the fjords as indicated by the dominance of AW, which has been studied extensively in the west Svalbard fjords. This can be due to the northerly winds resulting in an increased cross-shelf exchange of the Atlantic water (Cottier et al., 2005), pushing away the coastal waters and replacing them with the Atlantic water at the deeper levels (Cottier et al., 2007). This can also be caused by the wind curl caused by the southerly wind events along the WSC that add relative vorticity to the current, forcing it up the slope and topographic steering, leading to the geostrophic advection of the Atlantic water into Kongsfjorden (Nilsen et al., 2016).

Another notable observation is that when the SW layer was deeper, the TAW was prominent and when the SW layer was thinner, the AW was prominent (Table 3.2).

This indicates a mechanism that compensates between the opposing surface water and deep Atlantic water flows in the fjords during the observational period. The mechanism has been studied in detail in Kongsfjorden. The surface layer of water was moved by the local down-fjord winds blowing along the fjord axis. The freshwater concentration along the northern shore and its flow out of the fjord resulted from the wind blowing out of Kongsfjorden (Ingvaldsen et al., 2001; Sundfjord et al., 2017). The inflow of more saline, warmer Atlantic water over the bottom, along the southern shore, into the fjord compensated for such outflow (Moffat, 2014; David and Krishnan, 2017). The increase in the glacier melting and the freshwater discharge due to this circulation, further enhance the fjord water exchange. Higher SW layer thickness inside the fjord indicates a lesser inflow of Atlantic water into the fjord and, hence, the formation of TAW. When the SW layer thickness is thinner inside the fjord, it indicates more Atlantic water inflow into the fjord and, thus, the formation of AW. Even during the years of low AW volume, such as 2015 and 2018, Kongsfjorden had more AW than Krossfjorden (Table 3.2).

The AW was observed below ~50 m in the fjords (Fig. 3.7). The AW layer thickness was 15–30 m more in Kongsfjorden than in Krossfjorden (Table 3.2), indicating more Atlantic water intrusions into Kongsfjorden. The comparison of Kongsfjorden with Hornsund (Fig. 1.1), a fjord at the southern tip of the Svalbard archipelago and thereby a first entry point of the warm Atlantic waters from the WSC, also showed Kongsfjorden as warmer and more saline (Promińska et al., 2017). Kongsfjorden had approximately two times higher inflow of Atlantic water than Hornsund. The weaker frontal formation at the mouth, the wider and deeper trough connected to the WSC, and the lesser freshwater content than Hornsund were pointed out as the reasons for more AW inside Kongsfjorden than in Hornsund. However, the difference in the AW volume between Kongsfjorden and Krossfjorden (Table 3.2) is not as high as that between Kongsfjorden and Hornsund. Hence, the increased AW observed in Kongsfjorden compared to that in Krossfjorden could be due to the fjord being the first point of entry of the Atlantic waters among the twin fjord system with deeper outer fjord bathymetry and higher fjord volume than Krossfjorden (Svendsen et al., 2002; Promińska et al., 2017).

3.2.4. Freshwater volume

The range of freshwater volume (FWV) calculated (calculation details in Chapter 2) in Kongsfjorden is 0.11–1.45 km³ (the range of freshwater content, FWC, is 0.4–5.4 m), and that in Krossfjorden is in the range of 0.08–0.71 km³ (FWC is 0.2–4.5 m) in August from 2013 to 2018 (Table 3.2). Hence, the FWV in Kongsfjorden is 1.4–2.0 times more than that in Krossfjorden, except in 2018, when it was the same. The larger FWV in Kongsfjorden than in Krossfjorden indicates more glacial discharge into Kongsfjorden. This creates a stronger vertical gradient in the temperature and salinity in the upper Kongsfjorden, which further results in shallower MLD in Kongsfjorden compared to that in Krossfjorden, as seen in the earlier section. Svendsen et al. (2002) reported an annual freshwater discharge of 1.4 km³ into Kongsfjorden with a strong 30% annual variation. This matches with the maximum FWV estimation in Kongsfjorden during the period. However, the present Kongsfjorden FWV estimations appear ~3.5 times higher than the total summer (July-August-September) FWV reported by Promińska et al. (2017) from 2001 to 2015 (0.03–0.42 km³). The FWV in Krossfjorden is almost in the same range as reported in Hornsund (0.06–0.81 km³) by Promińska et al. (2017). The interannual variability in the FWV corresponds with the variations in the SW layer thickness. The FWV was the largest in August 2015, when the SW layer was the deepest in the fjords. On the other hand, the FWV was the smallest in 2014, when the SW layer was very thin.

The more heat advection into Kongsfjorden due to the higher Atlantic water intrusions, the deeper SW layer, and the larger FWV shows how reflective the fjord is for the ongoing changes from the open ocean, atmospheric, and terrestrial forcings, both temporally and spatially. It also shows the importance of monitoring and understanding Kongsfjorden to assess the changes. Hence, a more detailed analysis of Kongsfjorden variabilities for a decade from 2011 to 2020.

3.3. Interannual variabilities in summer (2011 – 2020)

3.3.1. Summer Kongsfjorden hydrography

The CTD profiles from July to September every year have been averaged from 2011 to 2020 to study the summer Kongsfjorden hydrography variations.

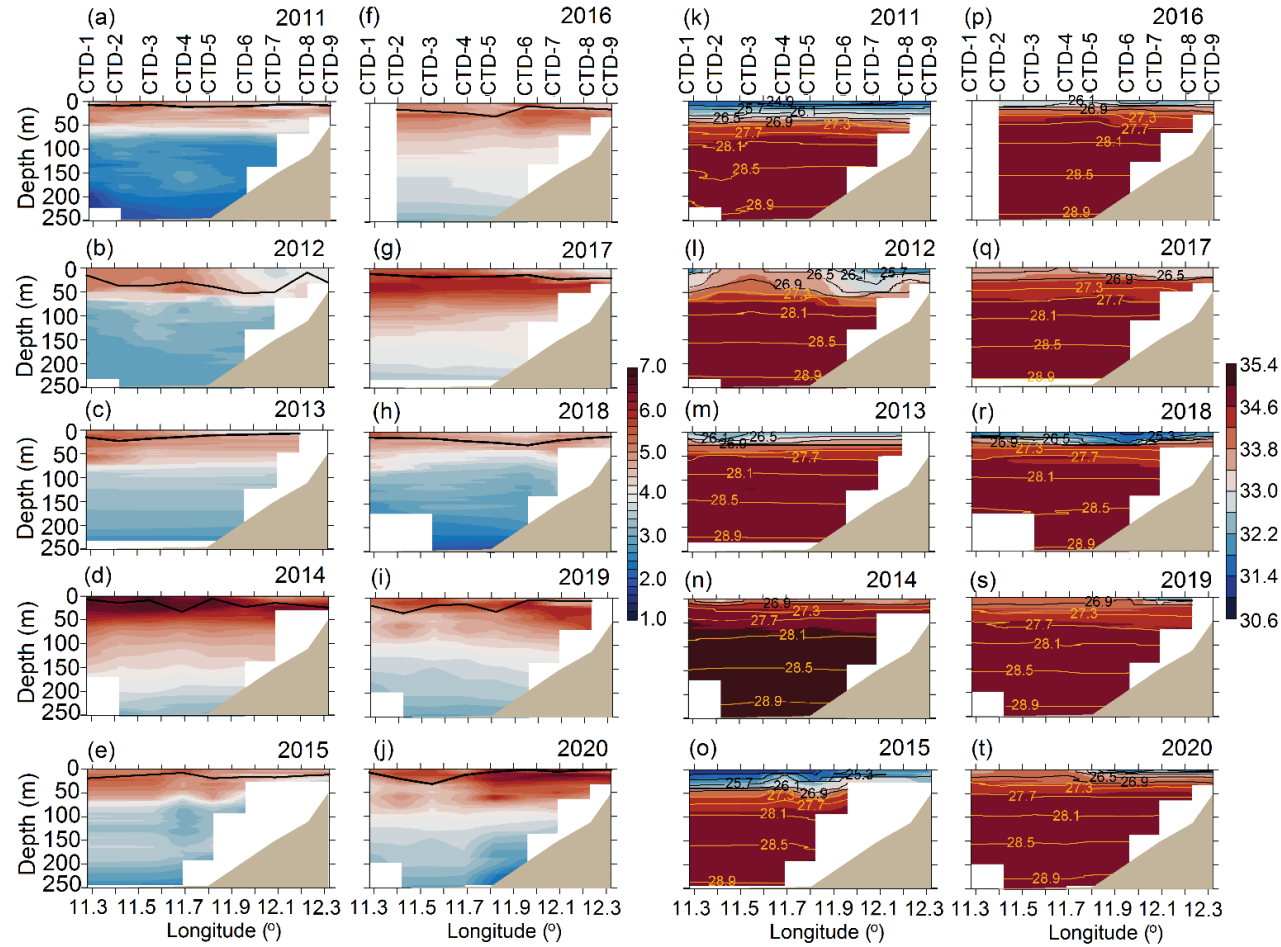


Figure 3.8: Vertical summer mean ((a) to (j)) temperature in °C from 2011 to 2020, respectively (shade), overlaid by the mixed layer depth (black contour) and ((k) to (t)) salinity in psu from 2011 to 2020, respectively (shade), overlaid by sigma-t contours, in Kongsfjorden. The light grey colour shaded portion in the diagrams is the bathymetry while the longitude increases from the open ocean side to the fjord head.

The summer vertical distribution of temperature, salinity, and density in Kongsfjorden (Fig. 3.8) showed the characteristic lateral variation from the glacier to the open ocean side, particularly within the MLD over the years. The MLD was generally shallower at the inner fjord compared to the outer fjord locations. Also, the water temperatures within the MLD in the inner fjord were 0.5 to 1.5 °C cooler than that in the MLD of the outer region. However, the water temperature in the inner Kongsfjorden was higher than 3 °C during the ten summers of the present study period.

The upper water temperature exceeded 6 °C in 2014 (Fig. 3.8 (d)), 2017 (Fig. 3.8 (g)), 2019 (Fig. 3.8 (i)), and 2020 (Fig. 3.8 (j)). On the other hand, the lower water column below 60 m was warmer in 2014 (Fig. 3.8 (d)), 2016 (Fig. 3.8 (f)), 2017 (Fig. 3.8 (g)), 2019 (Fig. 3.8 (i)), and 2020 (Fig. 3.8 (j)). Surface freshening was more prominent in 2011 (Fig. 3.8 (k)), 2015 (Fig. 3.8 (o)), and 2018 (Fig. 3.8 (r)), with the less saline layer found throughout the transect. Even though a similar freshening was observed in 2012 (Fig. 3.8 (l)), it was limited to the inner fjord. The highest salinity (> 35 psu) was found below 100 m in 2014 (Fig. 3.8 (n)). The summer mean variations from 2013 to 2018 are similar to the peak summer variations in August during the same years in the earlier section, indicating the dominance of peak summer months in the overall summer season variabilities.

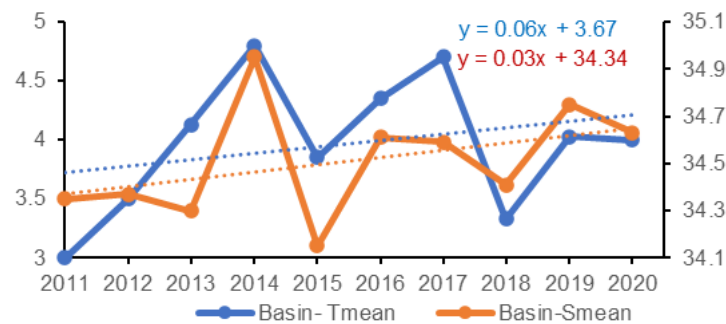


Figure 3.9: Kongsfjorden basin average summer temperature (°C) and salinity in psu from 2011 to 2020.

The summer mean Kongsfjorden water temperature (Fig. 3.9) varied between 3.0°C in 2011 and 4.8°C in 2014, while the salinity ranged between 34.15 psu in 2015 and 34.95 psu in 2014. Thus, 2014 was the year with the highest temperature and salinity during the study period. In an earlier interannual study (Promińska et al., 2017) in

summer using cruise-based CTD data from 2001-2015, the mean water temperature in Kongsfjorden ranged between 1.06°C in 2010 and 4.54°C in 2014, while the salinity ranged between 34.00 psu in 2004 and 35.00 psu in 2014. The high salinity at 100 m in 2014 and low salinity in 2015 corresponded well with our observations. The study further compared the temperature and salinity ranges, which were, on average, 0.92°C and 0.46 psu higher, respectively, than that in the southwest Svalbard fjord of Hornsund. As observed in our summer data, 2014 was the warmest and most saline, even though the mean temperature and salinity values were slightly different from the earlier study. This is due to the difference in depths and number of profiles used for the calculations. The summer water column temperature increased at a rate of 0.06 °C/yr and the salinity increased by 0.03 psu/yr showing an overall rise in the temperature and salinity over the last decade.

The temperature-induced fjord summer water column structure varied spatially, showing how temperature inversions are prone to the inner fjord (Fig. 3.8). While the outer fjord was a two-layer system with warmer-less saline surface water and cooler-high saline deep water, the inner fjord acts as a three-layer system with cold-less saline surface water, warm-saline water in the intermediate layer and cold-high saline water at deeper layers. The temporal evolution of the three-layer system and temperature inversion were further investigated, using the monthly average CTD profiles from July to October in the inner Kongsfjorden closer to the Kronebreen glacier, which has one of the highest flux rates in Svalbard (Fig. 3.10 (a)). All the available October CTD profiles from the inner fjord in 2012, 2013, 2015, 2016, 2018, and 2020 are included. The inner fjord was warmer from 2013 to 2017 and again in 2019 and 2020. 2011, 2012, and 2018 were comparatively colder. The warm core and, thereby, the temperature inversion depth each year was below 20 m, except in 2020 when it was between 10-20 m. Holmes et al. (2019) also observed the warmest water temperatures near Kronebreen at 20-25 m depths in 2016 and 2017.

The temperature inversions were strong in August and September (Fig. 3.10 (a)), coinciding with the time of maximum freshwater discharge (Svendsen et al., 2002) into the fjord. The strongest inversion, as seen in the interannual August profiles discussed in section 3.2, was in August 2017, where the depth showed a continuous

deepening from July to September. The depth range of inversion, which was 30-40 m in July, deepened to 40-50 m in August, 50-60 m in September and further down in October.

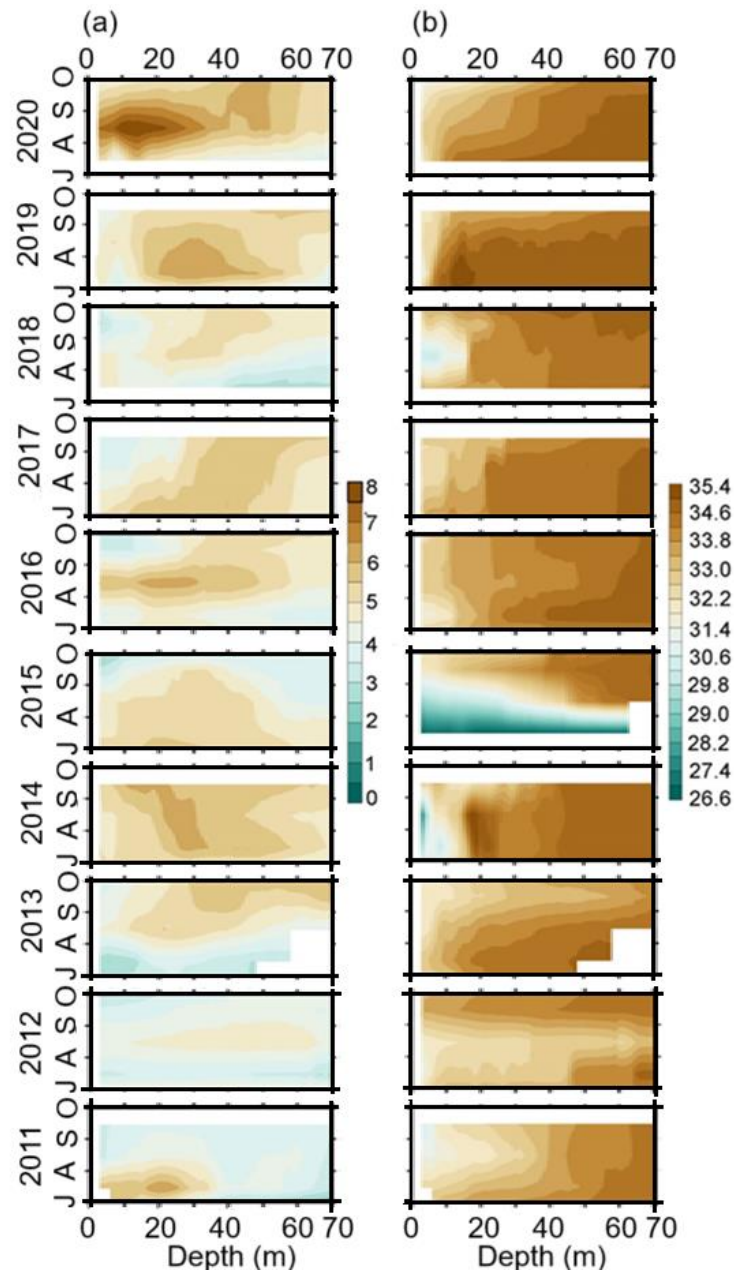


Figure 3.10: Vertical distribution of monthly mean (a) temperature ($^{\circ}\text{C}$) and (b) salinity (psu) from July to October from 2011 to 2020 in the inner Kongsfjorden. The years in increasing order are arranged from bottom to top.

The temperature inversions were strongest in August and September, and the layer deepened further in the fall season, as was found in October 2015, 2016, 2018, and 2020 (Fig. 3.10 (a)). Besides the larger freshwater discharge into the inner

Kongsfjorden, the climatological decrease in air temperature, increase in winds (Cisek et al., 2017), loss in net surface heat flux (Maturilli et al., 2015), and increase in precipitation (Osuch and Wawrzyniak, 2016) at Ny-Ålesund from August to October add to the upper-layer cooling and conversion of the two-layer fjord to the three-layer system.

De Rovere et al. (2022) addressed the interannual variation using time series temperature and salinity data from 2010 to 2020, covering the same period as the present study, but in the inner Kongsfjorden. The study reported an overall warming of the water column in inner Kongsfjorden by $0.13\text{ }^{\circ}\text{C/yr}$ at 35 m, which was double that at 85 m depth ($0.06\text{ }^{\circ}\text{C/yr}$), and a salinification by 0.3 from 2010 to 2020. The depth-averaged temperatures increased over the summers by $0.26\text{ }^{\circ}\text{C/yr}$ and were relatively stable in winters near Kronebreen glacier. The increase in water temperatures close to Kronebreen increases melting and frontal ablation (Luckman et al., 2015) and affects this tidewater glacier's stability and mass balance (Holmes et al., 2019). The studies related the increased frontal ablation rates to the increased volume of warm Atlantic waters and the air temperature, sea surface temperature, and sea ice concentration, making it important to consider atmospheric and oceanographic variables while explaining glaciological behavioural changes.

The inner fjord salinity (Fig. 3.10 (b)) showed the lowest of 26.6 in the upper 60 m in July 2015, gradually increasing till September. The inner fjord water column below 20 m showed high salinity consistent from July to September 2014 and 2017. De Rovere et al. (2022) showed that in January 2017 a linear regression change point, after which the temperature and salinity decreased. They found very little Atlantic water in the inner Kongsfjorden at the start of the decade, which increased substantially, leading to huge intrusions in 2016 and 2017. They concluded that the northward flowing Atlantic water current along the western Svalbard archipelago from lower latitudes drove local conditions in the inner Kongsfjorden profoundly over the decade. To understand the role of the Atlantic water in our study, the summer water masses are deciphered from 2011 to 2020.

3.3.2. Water masses

The SW, IW, AW, and TAW are the four water masses identified during the observational period (Fig. 3.11). Over the ten summers, three scenarios are found concerning the water masses in Kongsfjorden. The colder and lower saline summer of 2011 (Fig. 3.11 (a)) showed a strong presence of SW, IW, and TAW with no AW. The summers of 2012 (Fig. 3.11 (b)), 2013 (Fig. 3.11 (c)), 2015 (Fig. 3.11 (e)), 2018 (Fig. 3.11 (h)), and 2020 (Fig. 3.11 (j)) showed a reduction in SW compared to 2011. Besides TAW, AW was also found during these years. In the warm-saline years of 2014 (Fig. 3.11 (d)), 2016 (Fig. 3.11 (f)), 2017 (Fig. 3.11 (g)), and 2019 (Fig. 3.11 (i)), when the SW was warmer and saline compared to the other years, the fjord showed a strong AW dominance with less or almost no TAW. The thickness of the SW and AW layers reinforced these three scenarios observed over the years (Table 3.3). The SW depth has been taken as the vertical limit of 34 isohaline from the surface (Svendsen et al., 2002).

The summers from 2011 to 2020 in Kongsfjorden showed the same pattern with a thinner AW layer when the SW layer was thick, and vice versa, as observed in August 2013-2018 (Table 3.2). The SW layer thickness (Table 3.3) was the lowest in 2014, followed by 2019, and highest in 2011 and 2015. The AW layer thickness was the highest in 2014 and lowest in 2011. This pointed out an inverse relation between these two water masses formed by two distinct processes- one at the surface and the other at the subsurface-deep layers of the Kongsfjorden water column. The interannual variations in the water masses near Kronebreen glacier (De Rovere et al., 2022) slightly differed from those deciphered for the entire Kongsfjorden. The earlier study conducted at locations much closer to Kronebreen than our innermost CTD locations showed no AW in 2013 and more AW in August 2016 than 2014.

AW showed elongated presence in the summer and autumn of 2016 whereas the highest AW was found in 2014 compared to 2013 and 2016. However, different water masses have been identified from the inner and outer parts of Kongsfjorden over many years (Promińska et al., 2017). This difference reinforces the necessity to study water masses as close to glaciers as possible if the objective is to understand the glacier's behaviour against oceanographic properties.

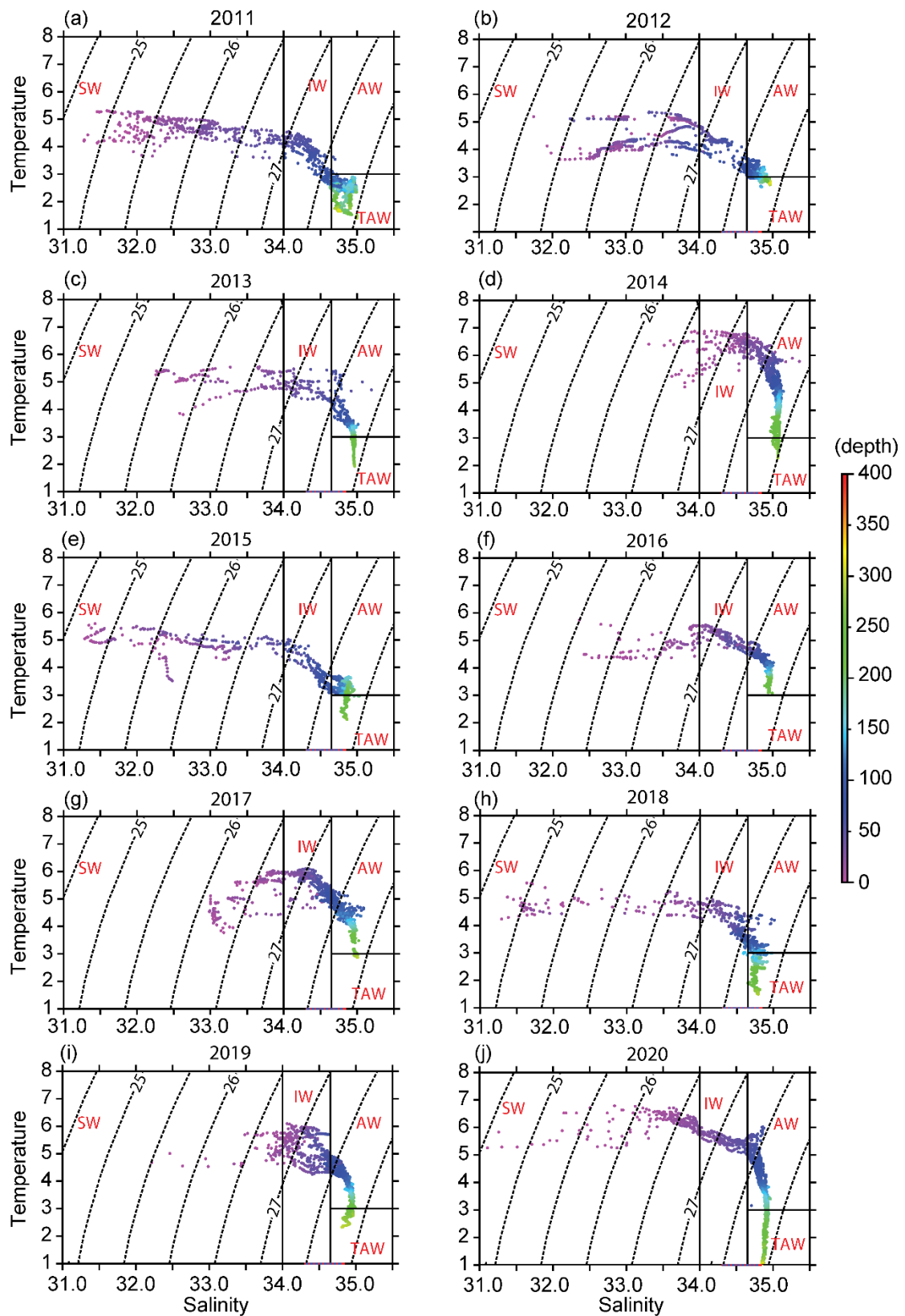


Figure 3.11: T-S diagrams showing water masses in Kongsfjorden from the summer CTD surveys conducted from 2011 to 2020 by the panels from (a) to (j), respectively. The colour bar denotes the depth range (m) of existence of each water mass.

Table 3.3: Summer mean thickness of the SW, AW, and combined AW+TAW layer with depth range of AW presence, and the maximum and mean of the temperature and salinity of the AW+TAW in Kongsfjorden.

Year	Mean SW layer thickness (m)	Mean AW layer thickness (m)	Depth range of AW (m)	Mean AW+TAW layer thickness (m)	T _{max} of AW+TAW (°C)	S _{max} of AW+TAW (psu)	T _{mean} of AW+TAW layer (°C)	S _{mean} of AW+TAW layer (psu)
2011	45	0	-	182	3.04	35.00	2.40	34.82
2012	40	77	82-159	164	3.65	34.97	2.97	34.87
2013	28	119	60-179	131	5.39	35.196	3.23	34.90
2014	7	202	25-227	219	6.78	35.36	4.24	35.03
2015	45	111	80-191	141	3.69	35.03	3.20	34.81
2016	21	167	60-227	167	4.93	35.01	4.04	34.89
2017	23	140	75-215	147	5.23	35.09	3.84	34.92
2018	26	66	75-141	179	4.38	35.01	2.86	34.77
2019	12	168	60-228	193	4.90	34.99	3.92	34.85
2020	26	129	60-189	203	6.03	34.96	3.59	34.83

The scenarios defined by Tverberg et al. (2019) and the range of intermediate depth of the AW layer during the observational period (Table 3.3) suggest that ‘winter deep’ winter scenarios prevailed in 2011, 2015, and 2018 when the SW layer thickness was higher. In ‘winter deep’ winter scenarios, the fjord bottom convection occurs with absent or limited advection of AW, leaving dense bottom waters. This causes summer AW intrusions at intermediate depths. According to the classification, most of the years in the decade belong to the winter open scenario, which was the warmest and saline when AW thickness was the highest. In the 2014 ‘winter open’ winter scenario, AW was seen from shallow to deep depths in Kongsfjorden. In the ‘winter open’ winters with no sea ice formation and AW intrusions, the AW will get cooled and transferred toward the bottom through convection, inducing further summer AW advection at shallow depths. The ‘winter open’ winters are due to atmospheric and oceanic warming and the reduction in sea ice cover. This scenario became more frequent in west Svalbard fjords after the anomalously large AW intrusion onto the West Spitsbergen shelf in February 2006. Further, the variations and trends in the SW and AW were analysed as these water masses reflect the changes in the two major

ongoing processes in the Arctic Ocean- the former with respect to the glacier melting discharge and the latter concerning the Atlantic water inflows to the Arctic Ocean.

3.3.3. Variations and trends in the SW and AW

The maximum temperature, $T_{\max}$, (Fig. 3.12 (a)) of the SW showed an increase at the rate of $0.08\text{ }^{\circ}\text{C/yr}$ while the minimum salinity, $S_{\min}$, (Fig. 3.12 (b)) showed a decrease at the rate of 0.04 psu/yr indicating the surface warming and freshening, respectively, in Kongsfjorden during summers of 2011 to 2020. Meanwhile, the mean summer temperature, T_{mean} , (Fig. 3.12 (a)) and salinity, S_{mean} , (Fig. 3.12 (b)) of the SW increased at the rates of $0.05\text{ }^{\circ}\text{C/yr}$ and 0.04 psu/yr , respectively. On the other hand, $T_{\max}$ of the AW (Fig. 3.12 (c)) increased by $0.16\text{ }^{\circ}\text{C/yr}$, double the rate of increase of SW. The T_{mean} of the AW also showed an increase of $0.07\text{ }^{\circ}\text{C/yr}$ at a higher rate than that of the SW. This higher rate of warming of the AW compared to the SW found from 2011 to 2020 in Kongsfjorden is in contrast to De Rovere et al. (2022) who reported the near-surface warming (35 m) at $0.13\text{ }^{\circ}\text{C/yr}$, almost double the rate compared to the near bottom (85 m), i.e., $0.06\text{ }^{\circ}\text{C/yr}$ between 2010 and 2020 in the inner Kongsfjorden. This is due to the temperature inversions that generally occur at shallower depths in the inner fjord, which increase the temperature compared to the deeper levels, as seen in the above section, while our calculations are for the entire basin, where the signature of the subsurface warming is more prominent. This again reinforces the fact that the profiles of the inner fjord need to be analysed separately for focused studies about the inner fjord processes. The $S_{\max}$ of the AW (Fig. 3.12 (d)) increased by 0.006 psu/yr while its S_{mean} increased by 0.008 psu/yr , one-fifth the rate of increase of the SW (0.04 psu/yr) indicating minimal changes in salinity happening in the subsurface to deep levels in the fjord compared to the surface salinity changes. In brief, the rate of summertime AW warming was higher than that of the SW, while the rate of salinification of the AW was lower than that of the SW.

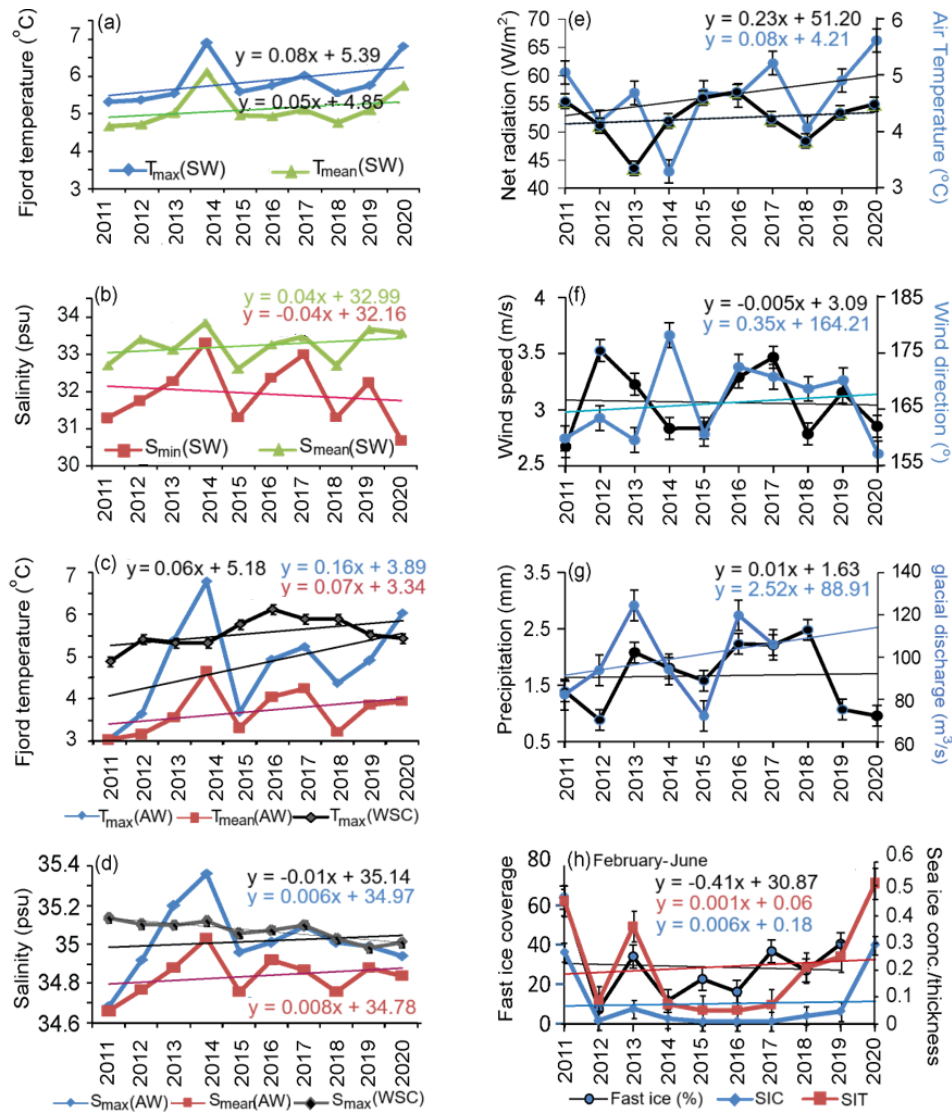


Figure 3.12: Summer mean of (a) maximum and mean water temperature, and (b) minimum and mean salinity of SW in Kongsfjorden from the CTD survey from 2011 to 2020. Summer mean of maximum and mean (c) water temperature in Kongsfjorden overlaid by the maximum annual temperature in the WSC and (d) salinity of the AW in Kongsfjorden overlaid by the maximum annual salinity in the WSC [<http://mosj.no/en/climate/ocean/temperature-salinity-fram-strait.html>] from 2011 to 2020. Summer mean (e) net surface radiation flux and air temperature, (f) wind speed and direction, (g) total precipitation and glacial discharge, and (h) fast ice coverage, sea ice concentration, and thickness in Kongsfjorden from 2011 to 2020. Except for the glacial discharge in Kongsfjorden (Pramanik et al., 2018), fast ice coverage from Gerland et al. (2020), and sea ice concentration and thickness from the CMEMS simulations, all the meteorological data in the figure is from the Ny-Ålesund weather station.

Besides, the SW and AW variations showed the same result: the coldest and least saline in 2011 and the warmest and most saline in 2014. Tverberg et al. (2019) found similar scale warming in 2006–2007 and the 1980s and 1990 as in 2014. The summer AW presence has not been as strong in earlier years as in 2012–2014. They identified this as winter open years with AW and TAW in most of the water column. In addition, there was also shallow summer AW advection. The present data and observations are an addition to the data, from 1994 CTD observations by Tverberg et al. (2019).

The interannual variations and trends observed in the SW and AW were investigated in detail analysing atmospheric (Fig. 3.12 (e, f, g)), cryospheric (Fig. 3.12 (g, h)), and oceanic (Fig. 3.12 (c, d)) factors at the same period. The net surface radiation and air temperature (Fig. 3.12 (e)) showed an increasing trend during summer 2011–2020, which corresponds with the increase in SW temperature. In contrast, the summer wind speed (Fig. 3.12 (f)) decreased over the ten years. The mean summer wind direction (Fig. 3.12 (f)) was southeasterlies (downfjord) from 2011 to 2013, 2015, and 2020 while it was south-southeasterlies in 2014 and then again from 2016 to 2019. The strong south-south easterlies in 2016, 2017, and 2019 of higher AW volume in Kongsfjorden support the conservation of volume mechanism in the AW intrusion (Moffat, 2014; David and Krishnan, 2017). The downfjord winds that lead to the wind-driven surface freshwater outflow from Kongsfjorden will enhance the compensating inward flow of the Atlantic waters to the fjord. However, 2014 which had the highest AW showed weaker winds than the other years of high intrusion. The total precipitation (Fig. 3.12 (g)) showed an increasing trend in summer 2011–2020 at the rate of 0.01 mm/yr, while the glacial discharge into Kongsfjorden till 2017 (data available only till 2017) increased by 2.52 m³/s per year. This increasing trend corresponded with the increase in freshening of the SW, which was found in terms of the $S_{\min}$. The precipitation was highest in 2018, followed by 2016, 2017, and 2013, while it was lowest in 2012, 2019, and 2020. The lowest temperature (Fig. 3.12 (a)) and salinity (Fig. 3.12 (b)) of SW in 2018 correspond with the highest precipitation. 2013 also fits this pattern with a high glacial discharge (Fig. 3.12 (g)). Even with high precipitation and glacial discharge, 2016 and 2017 showed a warmer-saline condition. Hence, the sea-ice conditions in the cold season before these summers were checked. The fast ice coverage data from Kongsfjorden (Fig. 3.12 (h)) from February to June

2011-2019 was compared with sea ice concentration and thickness obtained from the CMEMS reanalyses ($r^2 = 0.64$) first to understand the sea ice conditions in 2020 also. Based on this, both data were used to describe the variabilities in the summer of 2011-2020. The fast ice coverage showed a decline (0.41 %) per year till 2019, which correlated with the sea ice concentration and thickness changes during the same period. The low rate is due to a gradual increase in the sea ice coverage, concentration, and thickness observed from 2015. The fast ice coverage, sea ice concentration, and thickness (Fig.3. 12 (h)) were highest in 2011 and 2020, followed by 2013, all the years with low AW in summer inside Kongsfjorden. Thus, when the precipitation, glacial discharge, and fast ice coverage were highest in 2013, we found considerably higher SW layer thickness and thinner AW layer (Table 3.3). In 2014 summer, when the AW layer thickness was the highest, the sea ice coverage before the summer was the lowest. The summer precipitation and glacial discharge were also low. These three factors combined created the conducive situation for the maximum Atlantic water intrusion to Kongsfjorden and hence, the highest temperature and salinity observed in both the SW and AW.

The Atlantic water intrusions to Kongsfjorden are generally associated with the horizontal density gradient and intense winds on the shelf (Svendsen et al., 2002; Tverberg et al., 2019). However, De Rovere et al. (2022) reported that the long-term local temperature variability in the inner Kongsfjorden near the Kronebreen glacier during 2010–2020 is consistent with the northward temperature evolution of WSC thereby driving local conditions in the inner fjord. Hence, the high Atlantic water intrusion in 2014 was further related to the variations in the WSC (Fig.3. 12 (c, d)) in the shelf region outside Kongsfjorden that showed a close correspondence with the high temperature and salinity in 2014. Thus, the highest AW in 2014 was due to the advection of warmer waters from the WSC as compensation for the southeasterly-driven surface freshwater outflow from Kongsfjorden under a conducive physical environment already created by the lowest fast ice coverage prior to the summer. Even though the WSC temperature (Fig.3. 12 (c)) was higher from 2015 to 2018 than in 2014, the fjord temperature showed correspondence only in 2016 and 2017 and not in 2015 and 2018. This mismatch in 2015 and 2018 with WSC was due to the higher local sea ice conditions that existed prior to these summers compared to 2014 (Fig.3.

12 (h)), besides high precipitation and glacial discharge during the summers. High AW in 2016 and 2017, where fast ice coverage was more than in 2014, the compensating flows were primarily controlled by the stronger southeasterlies, than in 2014, pushing the surface cold waters out of the fjord and pulling in the warm waters through the deeper levels. The situation was the same in 2019 as well. Also noticeable is that the interannual variation in the SW temperature and salinity were similar to the AW, indicating the increasing influence of the AW in the upper layers of the water column over the years through vertical mixing and also due to the entry of the Atlantic water through the fjord surface. The variabilities and the causative factors were further analysed in detail using the mooring time series data in section 3.4.

3.3.4. The long-term mean of 2011-2020 v/s previous summer records in Kongsfjorden

The long-term summer mean temperature and salinity for 2011-2020 were calculated and plotted to understand how the last decadal condition of Kongsfjorden was when compared to the earlier periods using the longest continuous summer record of fjord water temperature available in Kongsfjorden (Tverberg et al., 2019) from 1994 to 2014. Due to the overlap of four years from 2011 to 2014 in the estimations of the present study and Tverberg et al. (2019), we also present the mean summer temperature and salinity for 2015 to 2020 as well.

The long-term summer mean water temperature along the Kongsfjorden transect with depths from 50-300 m in 2011-2020 (Fig. 3.13 (a)) and 2015-2020 (Fig. 3.13 (b)) was 0.5-0.8 °C higher than 1994-2014 (Fig. 3.13 (c)). The surface layer $T_{\max}$ in 2011-2020 was higher than 5.5 °C compared to 1994-2014, when it was 5 °C. The $T_{\min}$ at the bottom was 2 °C in 2011-2020 and less than 1.5 °C in 1994-2014. The lateral gradient was more prominent in the surface temperature (Fig. 3.13 (a, b)) than in salinity (Fig. 3.13 (d, e)). In the inner fjord, the long-term temperature of 2011-2020 (Fig. 3.13 (a)) showed a surface temperature of greater than 4.5 °C that has increased till the outer fjord to higher than 5.5 °C, showing a 1°C temperature difference between the inner and outer fjord. A subsurface high-temperature core higher than 5 °C reached the inner fjord in 2011-2020 (Fig. 3.13 (a)). 2015-2020 mean (Fig. 3.13 (b)) showed a 0.5 °C warmer subsurface temperature core in the inner fjord compared to 2011-2020 (Fig.

3.13 (a)). This subsurface temperature that reaches the inner Kongsfjorden was 1.5 °C (2011-2020) and 1 °C (2015-2020) higher, respectively, compared to the 3.5-4.0 °C in 1994-2014 (Fig. 3.13 (c)).

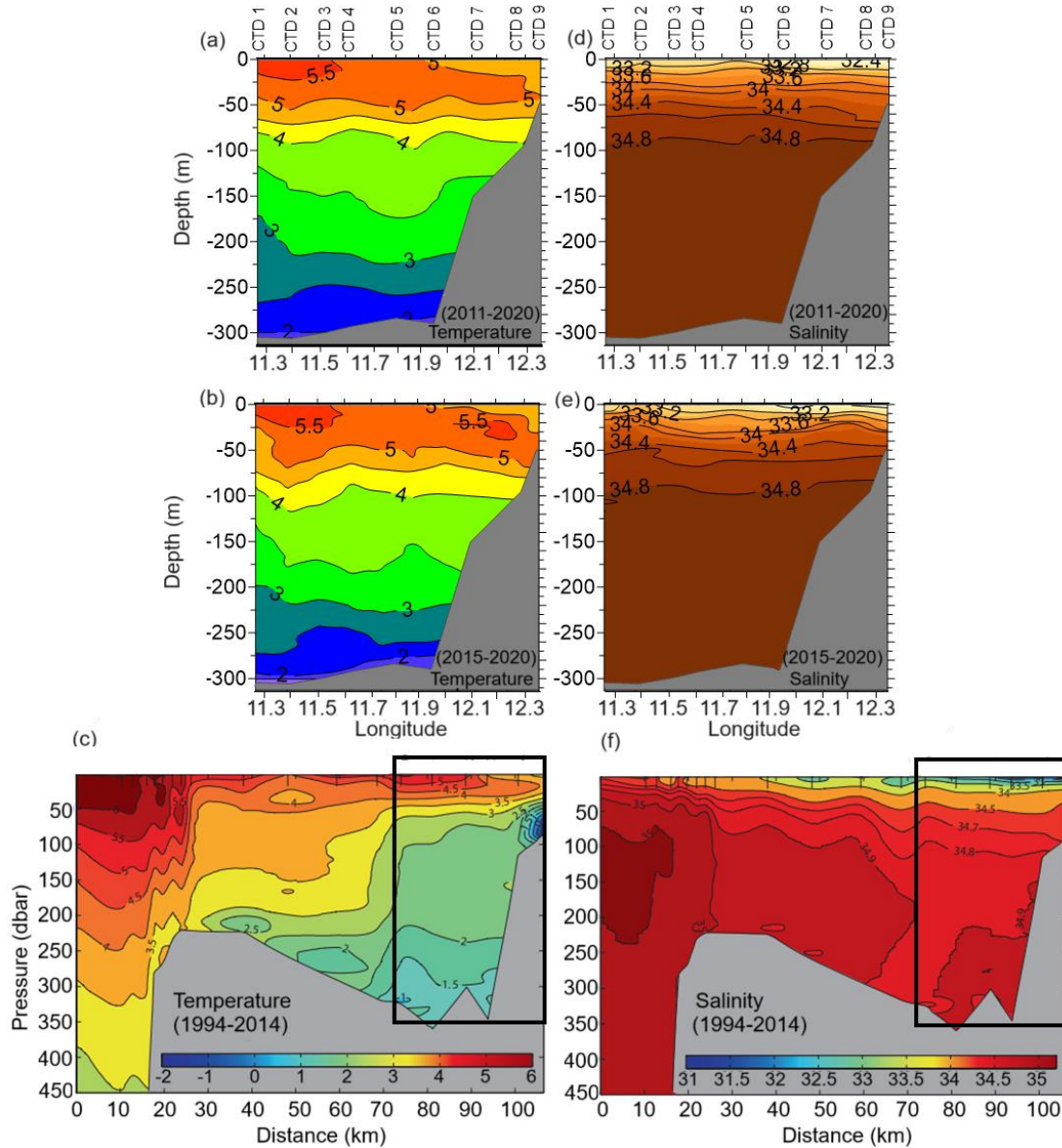


Figure 3.13: Long-term summer mean water column temperature in °C from (a) 2011 to 2020 and (b) 2015 to 2020 from the NCPOR data, and (c) from 1994 to 2014 adapted from Tverberg et al., (2019) along the Kongsfjorden transect. Long-term summer mean water column salinity in psu from (d) 2011 to 2020 and (e) 2015 to 2020 from the NCPOR data, and (f) from 1994 to 2014 adapted from Tverberg et al., (2019) along the Kongsfjorden transect. The black rectangles in (c) and (f) represent the transect portion that NCPOR surveyed.

This rise in the surface and subsurface temperature in the inner fjord during the last decade can have larger implications in terms of enhanced subglacial melting compared to the previous decades (Luckman et al., 2015), which adds to the total glacial melting and forms part of a positive feedback that will enhance the ongoing warming. The lowest salinity in the upper layer of the inner fjord (32.3-32.4 psu) was 0.1-0.2 psu lower than Tverberg et al. (2019). The highest salinity, greater than 34.9 psu, was found mostly below 200 m in the middle zone of the fjord in 1994-2014, while the highest salinity in 2011-2020 was 0.1 psu lower than in 1994-2014.

3.4. Interannual variabilities in summer, fall, winter, and spring from the mooring data (2014-2020)

The temporal evolution and yearly variations of the vertical distribution of the daily time series temperature, salinity, and water masses from 24 July 2014 to 31 December 2020 were studied using the mooring data. The vertical distribution of the water column temperature and salinity from 24 July 2014 to 31 December 2020 is given in Fig. 3. 14. Even though there is a deep, unrestricted connection of Kongsfjordrenna in Kongsfjorden that connects it to the WSC, it has a relatively shallow connection between the outer and inner parts of the fjord itself at 60 m (Luckman et al., 2015). Hence, whatever changes and processes above 60 m are transferred directly between the inner and outer fjord. The highest values of the density gradient were always found shallower than 60 m, especially during the summer when the density stratification of the upper water column was at its highest. The maximum summer MLD from 2011 to 2020 was also close to 60 m. Therefore, in our further analysis, 60 m was considered to delineate the upper and lower layers of the Kongsfjorden water column at the mooring location. Hence, the depth ranges above and below 60 m hereafter will be referred to as the upper and lower layers of the Kongsfjorden water column.

3.4.1. Temporal evolution of Kongsfjorden hydrography

During summer 2014 (Fig. 3. 14), a $T_{\max}$ of 6.6 °C in the upper 22-40 m and salinity higher than 35.12 psu from 24 July 2014 were found, which corresponded with the temperature-salinity values and depths of the presence of AW from the summer CTD survey (Table 3.2). The temperature and salinity decreased, especially up to ~ 60 m, due to the summertime freshening of the fjord surface waters. Higher than 3 °C

temperature was seen till 04 December 2014, after which the temperature decreased, delaying the winter onset till January 2015. However, less than zero (-1.0°C) winter water temperature (Fig. 3. 14 (a)) and salinity of 34.6 psu were found in February 2015 (Fig. 3. 14 (b)), followed by a rise in temperature. Further, temperature increased to 1°C by 13 May 2015 at deeper levels extending upwards to ~ 70 m. This temperature range of TAW continued its presence till 20 June, after which it increased to $> 3^{\circ}\text{C}$ and maintained the high temperatures and salinity till the first week of February 2016, mostly below 60 m.

A $T_{\max}$ of 6.5°C was found in the upper depth during mid-August in the summer of 2015 (Fig. 3. 14 (c)). The 2015 summer saw a large freshening (Fig. 3. 14 (d)), as found in the CTD survey data, with a $S_{\min}$ of 33.00 psu at 23 m depth from September to mid-October 2015. A strong halocline/pycnocline was seen up to 60 m from August to September, which then deepened to ~ 80 m in early October and then shallowed and disappeared by mid-October 2015. The average vertical salinity gradient during the period of strong halocline in the summer of 2015 was 0.021 psu/m. The winter temperature of 2016 (Fig. 3. 14 (c)) was much higher, with a temperature higher than 2°C and a salinity of ~ 34.8 psu in February 2016 (Fig. 3. 14 (d)). The lowest temperature of $\sim 1^{\circ}\text{C}$ was seen by mid-March and then again in April 2016. The temperature increased from February 2016 to mid-January 2017 (Fig. 3. 14 (c, e)). Summer freshening in 2016 (Fig. 3. 14 (f)) appeared higher than in 2014 but lower than in 2015. The summer 2016 halocline/pycnocline deciphered in the mooring data was as weak as in 2014. A warm temperature core of 5.5 - 6.0°C was below 60 m from mid-November to mid-December in 2016 (Fig. 3. 14 (e)). The water temperature showed a considerable increase again by mid-January 2017. The elongated and continuous presence of the warm water raised the winter temperatures in 2017, as in 2016. During winter 2017, the temperature was 1.5°C , and salinity was ~ 34.9 psu in February 2017 (Fig. 3. 14 (f)). A $T_{\min}$ of 0.5°C was found at the deeper depths from the end of March to early April. Higher than 3°C fjord water temperature was observed in the entire water column from mid-May 2017 (Fig. 3. 14 (e)) till 05 January 2018 (Fig. 3. 14 (g)). The summer water temperature in 2017 (Fig. 3. 14 (g)) was higher than in 2015 and 2016. However, the halocline and pycnocline in summer 2017 (Fig. 3. 14 (h)) were weak compared to 2016 and 2014.

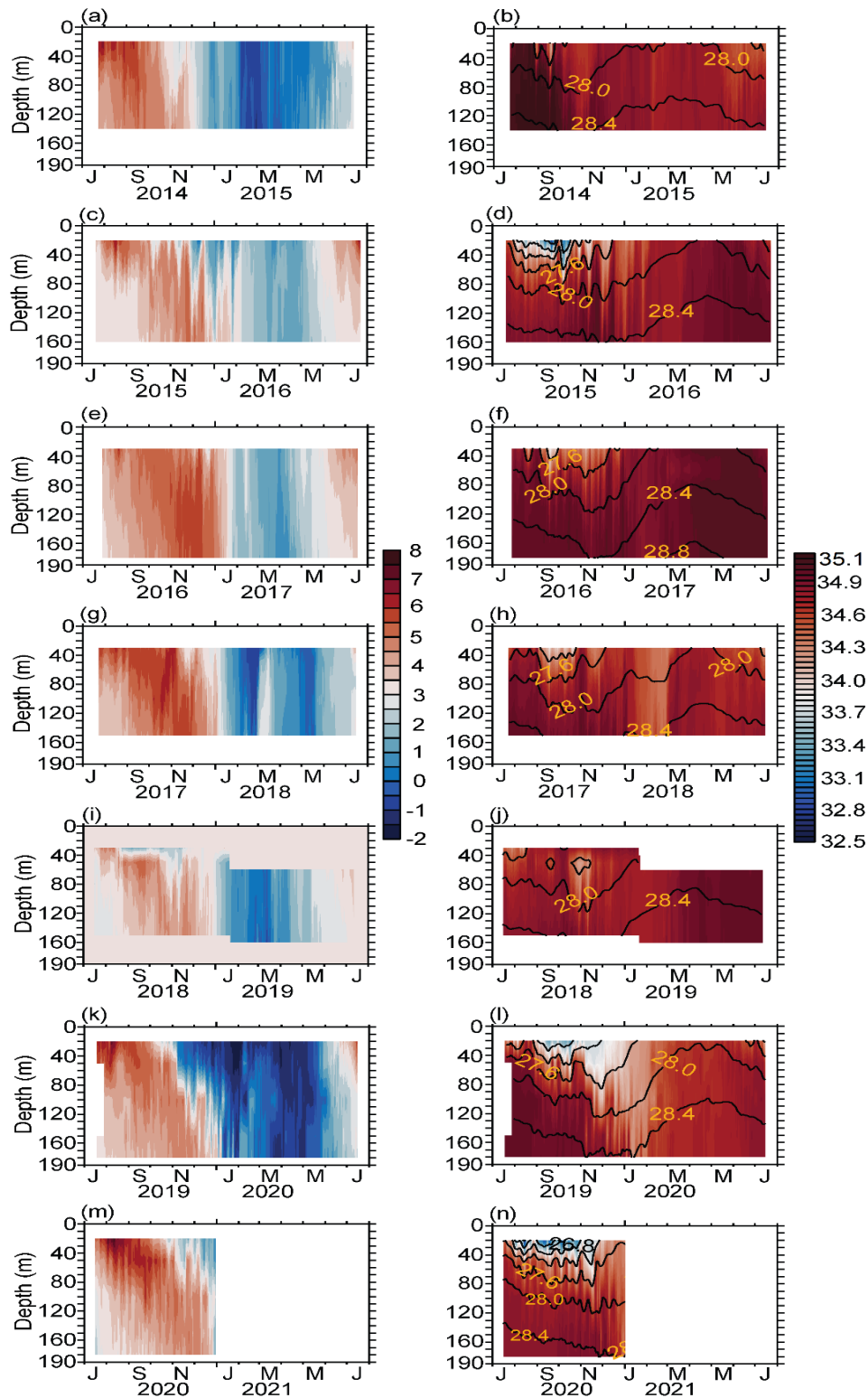


Figure 3.14: Time series of vertically interpolated daily mean (left panels) water temperature in °C and (right panels) salinity in psu overlaid by daily mean sigma-t in Kongsfjorden from the IndARC subsurface mooring for the deployment-retrieval periods of (a, b) 2014-2015, (c, d) 2015-2016, (e, f) 2016-2017, (g, h) 2017-2018, (i, j) 2018-2019, (k, l) 2019-2020, and (m, n) 2020-2021.

The water column temperature in the fall 2017 ((Fig. 3. 14 (g)) was higher than the previous years, with a maximum temperature range of 6.0-6.5 °C. The temperature decreased drastically afterwards, and a less than zero (-0.5 to -1.0 °C) winter water temperature was found by February 2018. The water column salinity (Fig. 3. 14 (h)) also showed a sharp decline during the same period, culminating in the lowest salinity of 34.40 psu in February compared to the previous years. The winter of 2018 was similar to the winter of 2015. However, colder waters were found again in April and continued until the first week of May 2018 after which the temperature increased. Higher than 3 °C temperature was seen till December 2018 (Fig. 3. 14 (i)). In 2018, a colder water layer was seen from the end of August to mid-October till 45 m depth (Fig. 3. 14 (i)), which deepened further. A higher presence of colder waters was found at these depths compared to the other years. This summertime surface cold water layer has not been picked in other years. The temperature in February 2019 (Fig. 3. 14 (i)) was in the range of 0.0-0.5 °C, whereas the salinity was ~34.80 psu (Fig. 3. 14 (j)). However, this lower temperature was seen in February and was more prominent by mid-March. The summertime upper layer freshening in 2019 (Fig. 3. 14 (k)) was strong, but not as stronger as was in 2015. The halocline and pycnocline (Fig. 3. 14 (l)) deepened from 40 m in July to 70 m in September till mid-October. The average vertical salinity gradient during the period was 0.017 psu/m, lower by 0.004/m than in the 2015 summer. The halocline/pycnocline shoaled after mid-October.

A stronger and longer presence of cold waters was seen in the fjord from January 2020 till May (Fig. 3. 14 (k)), followed by warmer waters from July to December. The winter water column of 2020 was the longest among the observational years. The temperature became less than zero degrees in December 2019 at the upper waters, which deepened until May 2020. However, warmer waters at the lowest depth in January and February 2020 make the mean winter water temperature 0.73 °C. A large range of temperature from -1.5 °C to 1.0 °C was seen from the upper to lower water column in February 2020, whereas the salinity was 34.50 psu (Fig. 3. 14 (l)), which was the lowest compared to other February during the observational years. The summer surface layer freshening was strong again in 2020 (Fig. 3. 14 (n)), and the strong halocline/pycnocline layer depth deepened from 40 m in July to 65 m in September and 70 m in October, after which it shoaled in November. The average

vertical salinity gradient was 0.0178 psu/m. Thus, the mooring data reinforced our finding of warmer years of 2014, 2016, 2017, 2019, and 2020 and colder years of 2015 and 2018. The increase in the number of warm years shows more prominence of warm Atlantic waters to Kongsfjorden. The two fjords, Kongsfjorden and Isfjorden (Tverberg et al., 2019; Skogseth et al., 2020) are reported to be shifting from typical Arctic fjords to the Atlantic type. The present study reinforces the change.

3.4.2. Trends and patterns in summer, fall, winter, and spring

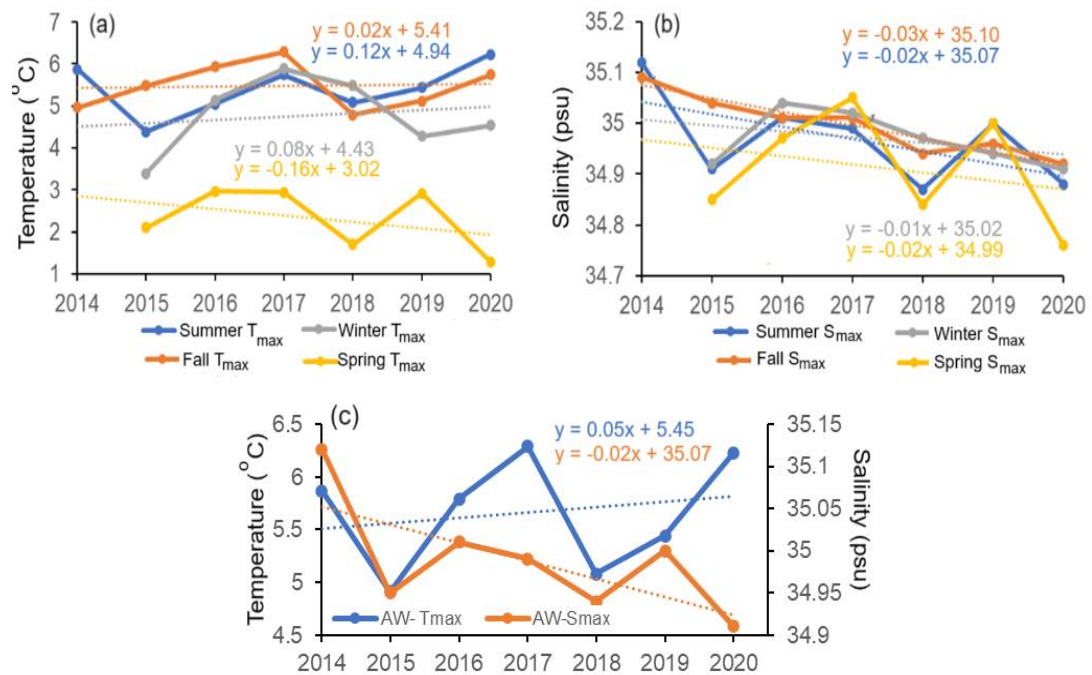


Figure 3.15: Season-wise maximum (a) temperature [°C] and (b) salinity (psu) below 60 m in Kongsfjorden from the mooring data from 2014 to 2020. (c) Maximum Atlantic water temperature and salinity during August to October when the Atlantic water intrusions were maximum in Kongsfjorden from 2014 to 2020.

Even though the trends for six to seven years are not statistically significant, the pattern of change during summer, fall, winter, and spring seasons, and the interannual nature embedded were checked. The maximum water temperature in summer, winter, and fall (Fig 3.15 (a)) increased from 2014 to 2020 at a rate highest in summer (lesser by 0.04 °C/yr compared to the trend from the CTD survey data), followed by winter and then fall. The difference in the summer rates from the mooring data and CTD survey

is due to the number of years considered for both calculations and the fact that the mooring data are continuous from a single location in Kongsfjorden. In contrast, the survey data is from the CTD profiles over the entire Kongsfjorden at certain times. The maximum spring temperatures showed a decrease over the period. The $S_{\max}$ (Fig 3.15 (b)) decreased at the highest rate in fall, followed by summer and spring, which had the same rate. Winter showed the lowest rate of decline in $S_{\max}$.

Besides the seasonal averages, the variation in the $T_{\max}$ and $S_{\max}$ of the AW layer below 60 m during the months of maximum intrusions observed (August to October) from 2014 to 2020 were studied. This particular time period was adopted for the analysis to directly compare with Tverberg et al. (2019). The distinction between the upper and lower layers at 60 m reinforces our understanding that the upper layer is affected primarily by atmospheric and surface processes, while the lower layer is affected primarily by dynamical fjord processes, including the Atlantic-origin waters that can bring in more heat and salt to the system. $T_{\max}$ showed an increase at the rate of 0.05 °C/yr (Fig 3.15 (c)), while $S_{\max}$ (Fig 3.15 (c)) showed a decrease. Otherwise, the interannual AW temperature and salinity variations match the July to September AW variations from the summer CTD survey (Fig. 3. 12 (c, d)). The increasing trend in the temperature corresponds to the increase in winter water temperatures reported by the review studies conducted by Tverberg et al. (2019) and Cottier et al. (2021) in Kongsfjorden.

Cottier et al. (2021) have compared the rate of changes in temperature and salinity at the outer, middle, and inner parts of Kongsfjorden and also with two fjords at different locations in the Svalbard archipelago- Isfjorden (Fig. 1.1) to further south of Kongsfjorden in the west Svalbard and Rijpfjorden (Fig. 1.1) on the northernmost Svalbard in proximity to the multiyear Arctic sea ice region. The mooring data in the outer part of Kongsfjorden from 2002 to 2018 showed the same rate of increase of 0.13 °C/yr in the 50-bottom mean and maximum temperatures in both the warm and cold seasons. However, in the study where they defined September to November as the warm season and March to May as the cold season, the central Kongsfjorden from 2014 to 2019 showed a temperature increase of 0.05 °C/yr in the warm season and 0.15 °C/yr in the cold season. This temperature rise rate is the same as that for August-

October 2014-2020 (Fig. 3. 15 (c)). However, the present winter temperature rise rate was $0.08\text{ }^{\circ}\text{C/yr}$ from the central Kongsfjorden, which was almost half the rate reported in the outer Kongsfjorden (Cottier et al., 2021), which might be due to the different locations in a place where the lateral gradient is large. The present study showed a summer rate of increase in the water temperature of $0.12\text{ }^{\circ}\text{C/yr}$. The inner fjord temperature from 2013 to 2020 (Cottier et al. 2021) showed a rise of $0.03\text{ }^{\circ}\text{C/yr}$ in the warm season and a decrease ($-0.09\text{ }^{\circ}\text{C/yr}$) in the cold season. When Isfjorden in west Svalbard showed an increase in the rate of temperature (the rate of the warm season is more than double the cold season) in both seasons, the colder and sea ice dominated Rijpfjorden at north Svalbard from 2006 to 2020 showed a decrease in temperature of $0.01\text{ }^{\circ}\text{C/yr}$ during the warm season while the temperature during the cold season was quite stable without any change. These changes in Rijpfjorden are due to its geographical location closer to the pole. The study showed that the temperatures in the west Svalbard fjords increased even in the coldest period of the year, which is the main reason for less sea ice reported recently in these fjords. The study concluded that it aligns with our present results, which show a higher volume of Atlantic water inside Kongsfjorden since 2014. Skogseth et al. (2020) also showed that the AW flow to the west Svalbard fjords had increased in recent summer, winter, and fall.

3.4.3. Water masses

We investigated the causative factors responsible for the observed temporal variations. The daily time series temperature (Fig. 3.16 (a)) at the upper (31 m) and lower (140 m) common depths of mooring data availability has shown a time difference of around three months between the T_{max} each year, except in 2015 which showed a four-month difference between August (31 m) and December (140 m). The T_{max} at 31 m each year was observed in August, except in 2014 and 2016, when it was in July and September, respectively. The peak temperatures at 31 m and 140 m occurred in August and November, respectively, in the four latter years of mooring observation.

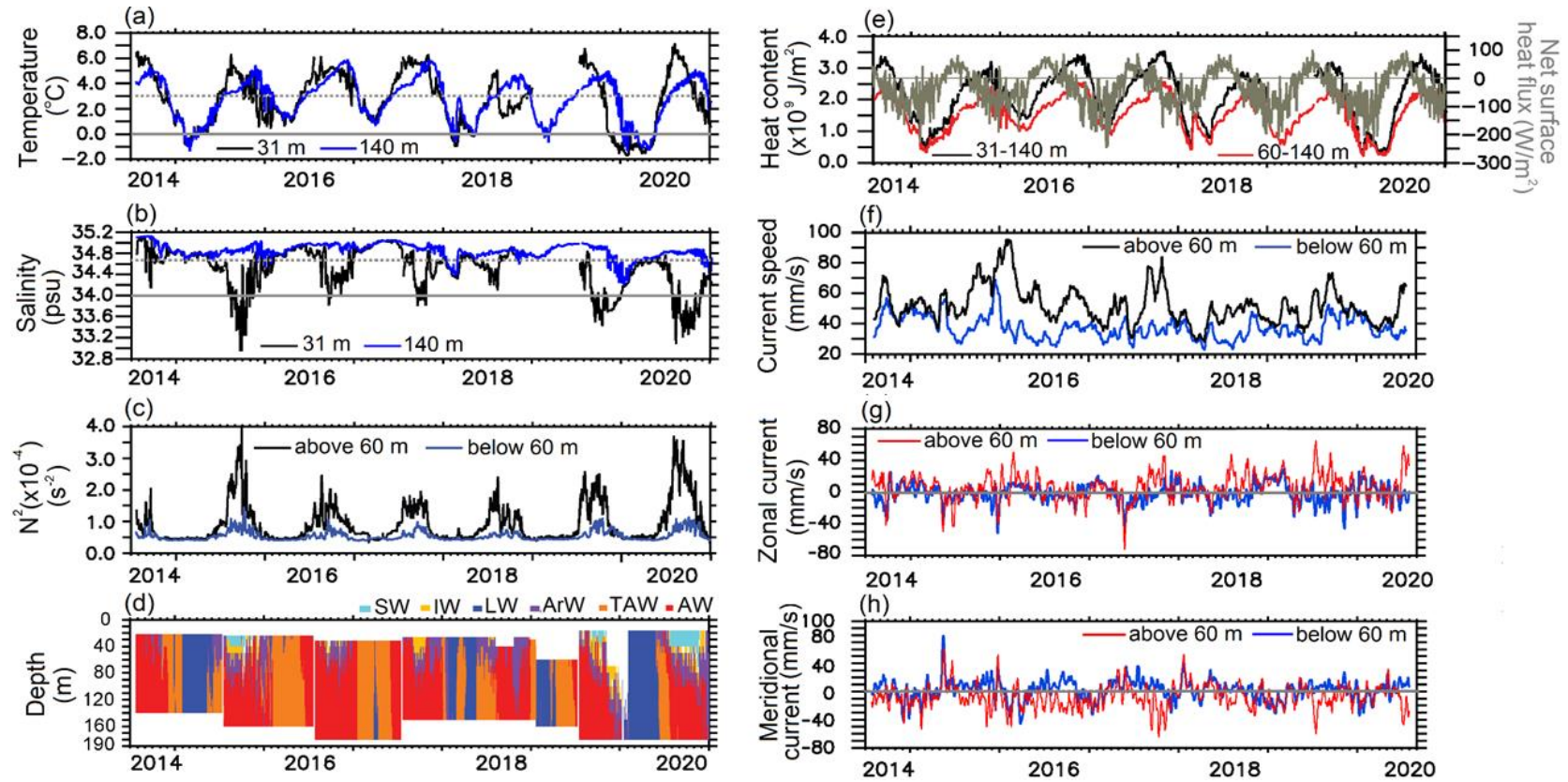


Figure 3.16: The daily time series water (a) temperature and (b) salinity at 31 m and 140 m depths, (c) depth average of the Brunt-Vaisala frequency calculated for water column above and below 60 m, (d) water masses distribution, and (e) the fjord heat content calculated for 31-140 m and 60-140 m overlaid with the net surface heat flux. The daily time series of (f) current speed, (g) zonal current velocity, and (h) meridional current velocity for depths above and below 60 m. The horizontal continuous and dotted grey lines in (a) and (b) denote 0 and 3 $^{\circ}\text{C}$, and 34 and 34.65, respectively. The horizontal grey line demarcates the zero net surface heat flux in (e) while it represents the zero current velocity in (g) and (h). Except for the ERA-interim surface heat flux data in (e), all the other data are from the mooring.

The period between the $T_{\max}$ each year between these upper and lower depths coincided with the sharp decline and subsequent rise in salinity at 31 m, except in 2018 (Fig. 3.16 (b)), which exhibited a continuous increase in salinity. During the same time, $T_{\max}$ at 31 m showed a rapid decline followed by a rise. Salinity at 31 m showed a seasonality with lower values in summer-autumn and higher values in winter-spring. The 140 m salinity showed a similar seasonality but embedded in a low-frequency variation. Salinity at both depths was the same during the colder months in winter and early spring. The upper salinity decline was the highest in 2015, followed by 2020 and 2019. On the other hand, the upper salinity decline was lowest in 2014 and 2018 and almost similar in 2016 and 2017.

The Brunt-Vaisala frequency, N^2 , the measure of stratification, calculated for the water column (Fig. 3.16 (c)), showed the highest stratification in summer, above 60 m. The interannual variation in stratification closely follows the salinity variations. The upper 60 m layer stratification was 2-5 times higher than below 60 m over the observational period. Maximum stratification was found in 2015, and 2020 in both layers, and the lowest stratification was found during the summers of 2014 and 2017. De Rovere et al. (2022) stated four phases in the stratification of the inner Kongsfjorden water column- the first phase being the start of stratification in spring and summer, where the upper water column warms at a higher pace from the heat gain from the atmosphere, second phase during the time of AW intrusion between peak summer to mid-autumn, third phase begins afterward when temperatures decrease at upper depths with warmer bottom waters and fourth phase during winter with coldest waters and deep winter convection as a result of heat loss to the atmosphere. However, our mooring data showed a depth differentiation in the water column stratification. Above 60 m, the mildest winter to mid-spring stratification, a substantial increase in stratification started from mid-spring, peaks by August/September, and then declined through the fall (Fig. 3.16 (c)). Below 60 m, the stratification increases and peaks coincided with the pure AW intrusion (Fig. 3.16 (d)) and the decline coincided with AW modification to TAW or LW.

The water column was homogenous and hence weakly stratified in winter- early spring, which was evident as the lowest N^2 and no difference between both layers. The

highest stratification in 2015 in the upper 60 m coincided with the vertical arrangement of SW with IW and ArW one below the other within the depth (Fig. 3.16 (d)). 2019 and 2020 also had the same vertical upper layer water mass structure. The upper layer water masses found were the IW over the ArW in 2016 and 2017. The longer and more dominant presence of the TAW and AW was seen in 2016 and 2017. Longer residences of the TAW and AW in 2016 and 2017 were also reported in the inner Kongsfjorden (De Rovere et al., 2022). In 2018, TAW was also seen in the upper 60 m beside the IW and ArW. IW was seen in the summer and autumn months of all years except 2014. After its presence in the 2015 winter, LW presence in the fjord during winter and spring increased from 2018 to 2020, which is a reason for the decreasing winter and spring mean temperatures, as seen in the earlier section. On the other hand, the continuous and long residence of Atlantic-origin waters observed at the deeper levels in Kongsfjorden is vital in understanding recent shifts in the fjord, such as changes in the community structure and species of the fjord biota (Fredriksen et al., 2014; Sinha et al., 2017; Hop et al., 2019; Assmy et al., 2022), in addition to the changes in physical processes such as water mass modification, stratification, subglacial melting, heat-salt entrainment and mixing. In the inner Kongsfjorden, AW was not seen at the beginning of the time series, starting in 2010, which shows that pure AW reached the inner fjord in the later years (De Rovere et al., 2022), which is picked up in our mooring observations (Fig. 3.16 (d)).

3.4.4. Role of Atlantic water advection in the interannual variability

With this observed increase in the AW and TAW and the other recent reports of increasing Atlantification inside Kongsfjorden against the increasing heat transport towards the Arctic Ocean by the WSC from the lower latitudes of the North Atlantic Ocean, the fjord heat content (HC) was estimated. The daily time series HC was calculated for the depth range of 31 to 140 m to include the maximum subsurface fjord depth and also from 60 to 140 m, focusing on the depth of the Atlantic water layer (Fig. 3.16 (e)). Here, it is assumed from the observations that surface heating and salinity effects are limited to the upper layer and do not modify significantly in the lower layer.

Both the fjord HC estimations (Fig. 3.16 (e)) showed an overall and continuous increase from March/April (except in February 2015 and May 2018), peaked by the end of summer or fall (September/October/November), and then declined. The HC variations against the net surface radiation available to the fjord were compared over the same period. The net surface radiation flux is the sum of the shortwave, longwave, sensible, and latent heat fluxes available from the ERA-interim reanalyses data averaged over Kongsfjorden. The net surface radiation flux gained by the fjord was the highest in July, followed by a decrease. The fjord started losing heat to the atmosphere by the end of September/October. The highest heat loss from the fjord and the lowest fjord HC coincided with the peak winter month (February/March) each year. However, a difference of one to three months was found between the peaks in the fjord HC and the net surface radiation flux from 2015 to 2020 (Table 3.4). It was observed that the fjord HC continued rising even after the net surface radiation started declining each year (Fig. 3.16 (e)).

Table 3.4: The month of maximum fjord HC and net surface radiation in Kongsfjorden, with the difference of the HC between these months, and mean net surface radiation over the period between the months of maximum fjord HC and net surface radiation from 2015 to 2020. The fjord HC is calculated from the mooring data, and the net surface radiation data is from the ECMWF ERA-interim reanalyses.

Year	Time of Maximum HC	Time of Maximum net surface radiation	Difference in the HC (31-140 m) between the times (Jm^{-2})	Difference in the HC (60-140 m) between the times (Jm^{-2})	Mean net surface radiation over the period (Jm^{-2})
2015	November	July	7.45E+08	6.24E+08	-105496
2016	November	July	7.19E+08	6.26E+08	-34480
2017	October	July	9.79E+08	7.46E+08	-102613
2018	September	July	6.09E+08	4.97E+08	-121867
2019	September	July	5.13E+08	4.01E+08	-138709
2020	October	July	7.17E+08	6.07E+08	-110080

Moreover, the fjord continued gaining heat even when the net surface radiation became negative, losing heat flux from the fjord, indicating an internal oceanic source of heat supply. This can be seen in the difference in the HC between the periods of the

highest HC and net surface heat radiation and the mean available net surface heat radiation over these periods every year (Table 3.4). The mean available net surface heat radiation shows an overall heat loss, confirming an internal oceanic source of heat supply in the subsurface and deeper layers that increased the heat content of Kongsfjorden.

The temporal distribution of the water masses (Fig. 3.16 (d)) showed that these periods of increasing fjord HC and net negative surface radiation coincided with the periods of Atlantic water presence in Kongsfjorden, showing that the warm Atlantic water as the heat source to Kongsfjorden. Since our model simulations could not delineate such high-resolution features in the circulation presently inside Kongsfjorden, the simulations of currents by Sundfjord et al. (2017) were referred to understand circulation patterns inside Kongsfjorden at different depths near our mooring location. The surface 0-5 m current simulations for July to September (Fig. 5, Sundfjord et al., 2017) in the mooring location are towards the northwest, out of the fjord. At 100 m, the current flow showed an overturning just north of our mooring location, resulting in a northward flow from the southern shore of Kongsfjorden with the Atlantic water finally reaching the mooring location. The upward-looking mooring ADCP (depths of each deployment period given in chapter 2) could not decipher the surface currents. Hence, the current data considered for the above 60 m average is from 30 to 60 m. The increase in current speed (Fig. 3.16 (f)) with the current above 60 m flowing towards the southeast (Fig. 3.16 (g, h)) showed an inflow towards the head of Kongsfjorden during the warm periods of increase in the fjord HC. The southeast subsurface current (Fig. 3.16 (g, h)), opposite in direction to the northwest surface current (Sundfjord et al., 2017), might be a counter-current bringing in warm waters from the Kongsfjorden mouth to the mooring location. Below 60 m, we found a northward flowing meridional current (Fig. 3.16 (h)). This northward flow towards the mooring location can transport the Atlantic water to the mooring location. Thus, our observations from July to September matched the simulations (Sundfjord et al., 2017). However, the currents below 60 m became prominently southeast (upfjord) from October till the end of the cold season, implying flow from the outer side to the interiors of Kongsfjorden.

Another study that attempted to quantify and relate Atlantic water as the major source of HC in Kongsfjorden is Cottier et al. (2005). In the study, they calculated the HC in Kongsfjorden in June and July at an interval of 11 days. The difference in the HC was $7.8 \times 10^8 \text{ Jm}^{-2}$, demanding a net surface heat flux of 820 Wm^{-2} , considering the only heat source in Kongsfjorden as the available net surface heat flux. However, the mean available surface heat flux over the period was 52 Wm^{-2} ($4.9 \times 10^7 \text{ Jm}^{-2}$), an order of magnitude less than the heating required, confirming that the increase in HC is due to warm water advection into the fjord. The mean net surface heat flux was positive and higher as they calculated for a shorter period of 11 days from June to July when the net surface radiation is in its increasingly positive phase. Hence, for comparison with this earlier finding, the fjord HC between 01 June and 31 July was estimated, which confirmed that the mean net surface heat flux available over the period (2015- 41.63 Wm^{-2} , 2016- 51.02 Wm^{-2} , 2017- 41.56 Wm^{-2} , 2018- 45.76 Wm^{-2} , 2019- 45.13 Wm^{-2} , and 2020- 53.11 Wm^{-2}) was one order of magnitude less than the net surface heat flux demanded by the water column to acquire heat from June to July (2015- 207.01 Wm^{-2} , 2016- 148.00 Wm^{-2} , 2017- 106.44 Wm^{-2} , 2018- 159.38 Wm^{-2} , 2019- 214.03 Wm^{-2} , and 2020- 274.17 Wm^{-2}). The observed fjord heat flux combines the net surface heat flux with the advected heat flux. The analysis reinstated the heat advection by AW as the significant contributor to the Kongsfjorden heat flux changes and the interannual variability that the fjord water column undergoes. During winter, the fjord flux declines due to the decreased advected heat flux and net surface heat loss compared to the other seasons. The flux analysis also proved to be a promising way of detecting periods of AW advection in Kongsfjorden.

3.4.5. Role of sea ice in the interannual variability

The mean fast ice coverage from February to June 2015 was $\sim 23 \%$ (Fig. 3.12 (h)), with a coverage of 60% in February 2015 (Gerland et al., 2020). Hence, the coldest winter water column in 2015. In 2016, the mean fast ice coverage from February to June was lower ($\sim 16 \%$) than in 2015, while the February coverage was nearly zero (Gerland et al., 2020). This describes the conducive environment of a lesser density gradient resulting in larger AW intrusions to Kongsfjorden in winter and the successive seasons, which was seen in the observation as warm winter and summer in 2016 (Fig. 3.15). Even though in 2017, the mean ice coverage was higher ($\sim 37 \%$)

than in 2015 (Fig. 3.12 (h)), the fjord ice season was the shortest, and the fjord ice extent was lower (Johansson et al., 2020). These culminated in the warmest and most saline winter to spring in 2017 (Fig. 3.15). The mean fast ice coverage in 2018 (27 %) (Fig. 3.12 (h)) was again lesser and comparable to 2015 conditions. However, the deep TAW intrusion episodes in winter 2018 (Fig. 3.16 (d)) are why the mean water column temperature higher in winter 2018 was found (Fig. 3.15). However, the overall effect of the winter sea ice conditions was carried over to spring, summer, and fall, resulting in lower temperatures. The shorter ice season in 2019, similar to 2017, led to the warmer fjord water column in the following spring, summer, and fall. In 2020, the combination of the early onset of a prominent upper layer cooling in autumn (Fig. 3.14 (f)) and the highest sea ice concentration/thickness (Fig. 3.12 (h)) resulted in colder and fresher winter and spring (Fig. 3.15). Visual observations reported a persistent sea ice cover of the fjord from late January to May, and cold temperatures are found at MDI mooring in the inner Kongsfjorden, although sea ice cover data from Kongsfjorden is not yet available for winter 2020, (De Rovere et al., 2022). However, the warmer summer and fall of 2020 (Fig. 3.15) were the one exception from the conditions assessed for the earlier years. Thus, the interannual variations in the Kongsfjorden water column temperature and salinity depended on the interannual condition of sea ice, such as its coverage, concentration, thickness, and length of sea ice season. The years with the shortest fjord ice season and lower coverage and concentration resulted in the most warm and saline years, and vice versa.

3.4.6. Role of atmospheric forcings in the interannual variability

Even though the subsurface mooring missed the surface data and thereby the variations, with the range of the summertime MLD, surface freshening, halocline-pycnocline depths, and SW reaching up to 50-60 m and deep homogenous water column during the winter-spring seasons, the time series variation of the atmospheric variables was also checked to understand its role in the observed variations.

The air temperature (Fig. 3.17 (a)) peaks and lows corresponded with the net surface radiation flux highs and lows (Fig. 3.16 (e)). As per the general pattern in the Arctic region, the wind speed (Fig. 3.17 (b)) increased from fall, peaked by winter, and then decreased by summer, thus having the highest peak in winter and lowest in summer.

The southwesterly winds (across the fjord) were found generally prominent from May to June each year (Fig. 3.17 (c)). The southeasterlies (downfjord) were prominently observed during summer. Among the summers from 2014 to 2020, the mean summer wind was the highest in 2016 and 2017 (~ 3.5 m/s) (Fig. 3.12 (f), 3.17 (b)). The summer stratification also was the lowest in these two years, out of which in 2016, the mean summer stratification above 60 m was the lowest (stability parameter (E) is $\sim 1.1 \times 10^{-5} \text{ m}^{-1}$). SW and IW, the water masses seen in the more stratified surface water column, were not seen in the upper layer mooring data during the summer of 2016 (Fig. 3.16 (d)). When the summer mean wind speed was the lowest (2.9 m/s) in 2020 (Fig. 3.17 (b)), upper water column stratification was the highest ($2.1 \times 10^{-5} \text{ m}^{-1}$), almost double the stratification in 2016. This was evident as prominent SW, IW, and ArW were seen in the summer of 2020 (Fig. 3.16 (d)).

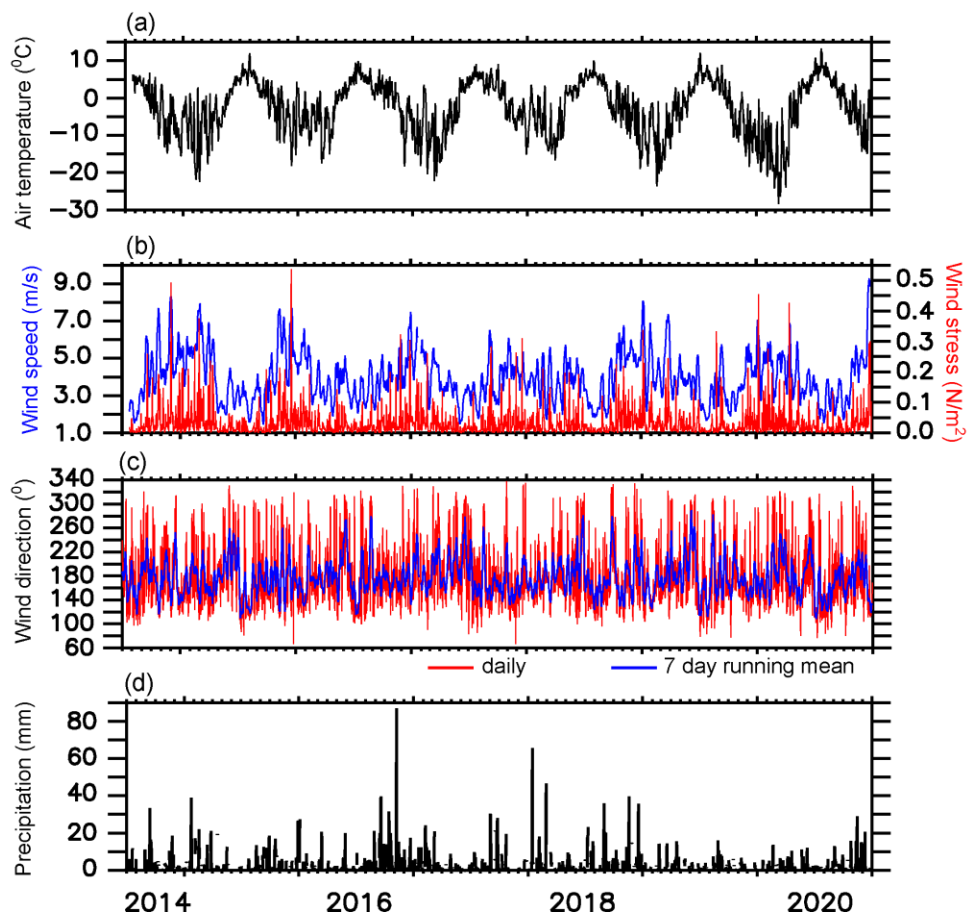


Figure 3.17: The daily (a) air temperature, (b) wind speed and stress, (c) wind direction, and (d) total precipitation from July 2014 to December 2020 from the Ny-Ålesund weather station.

The highest wind stress was found in winter when the wind speed was higher (Fig. 3.17 (b)). The direction of winds changed between southeasterlies (downfjord) and southwesterlies (across the fjord) during winter (Fig. 3.17 (c)). The highest mean winter wind speed in 2015 (5.0 m/s) (Fig. 3.17 (b)) coincided with the lowest stratification for the full depth up to 140 m ($4.8 \times 10^{-6} \text{ m}^{-1}$), one order lesser than in summer, indicating a vertically more homogenous water column compared to the other years. As the earlier section shows, the 2015 winter was the coldest and least saline among the observational years. On the other hand, when the mean winter wind speed (Fig. 3.17 (b)) was the lowest in 2020 (4.0 m/s), the stratification ($5.5 \times 10^{-6} \text{ m}^{-1}$) was the highest. Thus, wind had a higher impact in the upper 60 m in summer and to 180 m depth in our observations. The total precipitation was highest during September to March (Fig. 3.17 (d)). Out of the observational years, 2016/17 had the highest total precipitation, followed by 2018/19. The precipitation during 2018/19 might have impacted the low salinity and temperature in SW, as seen in the summer CTD survey. 2019/20 had the least precipitation but was not found to have an impact on the SW variation. With the interannual variability detailed for the seasons, the time series data was then analysed to understand the lower period intraseasonal oscillations in the fjord hydrography, which is the least studied.

3.5. Intraseasonal Variability

The intraseasonal oscillations (ISOs) that vary from 10 to 100 days (Lau and Waliser, 2005), the range of timescale close to the typical timescales for the weather (up to 15 days) and climate (from a season and beyond) are important and integral part of the weather and climate events. The intraseasonal variability in the ocean is mainly due to different processes caused by atmospheric and reasons largely intrinsic to the ocean at intraseasonal timescales. For the intraseasonal study, the time-series mooring data was analysed for the uppermost and lowermost common depths of 31 m and 140 m, respectively, throughout the deployment-retrieval periods from 2014 to 2020. A suite of spectral analyses has been used to understand the intraseasonal variability.

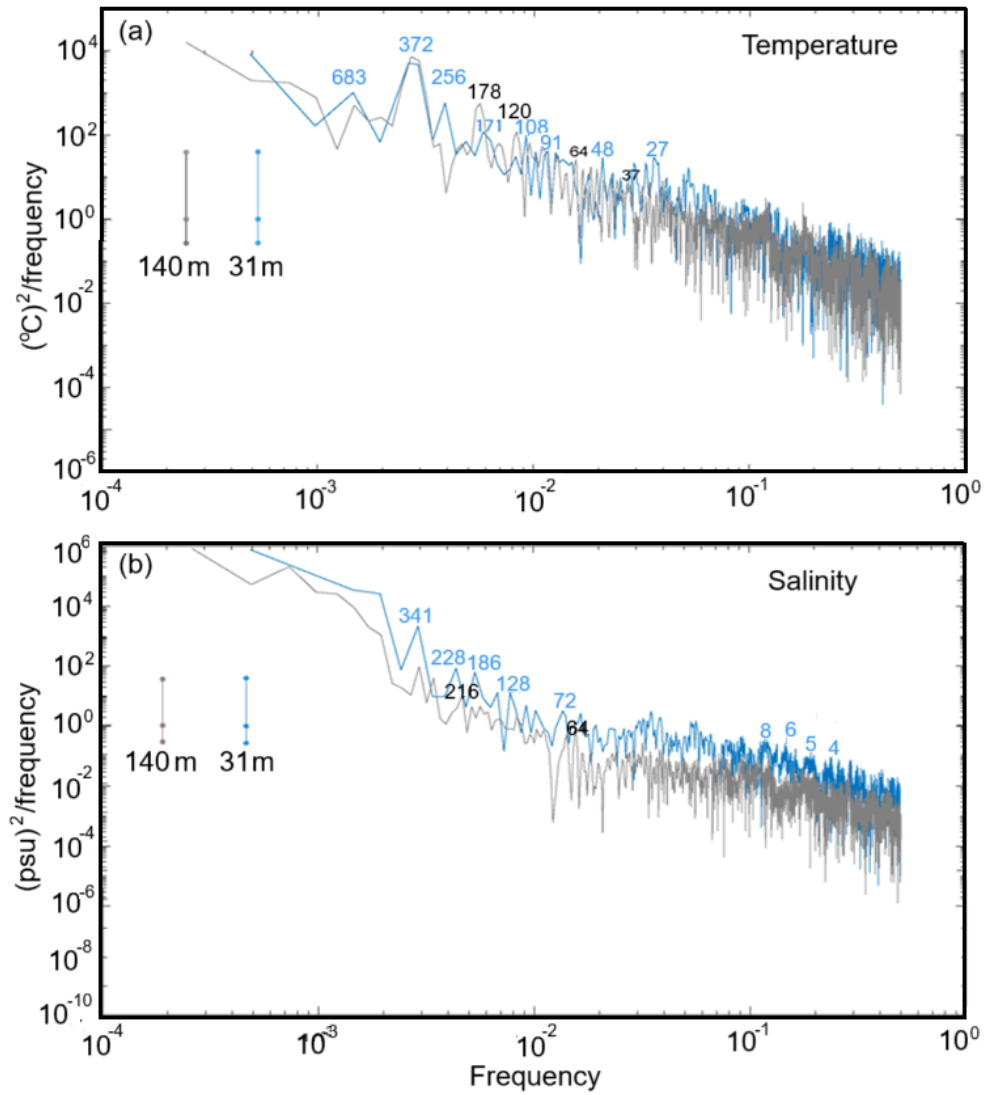


Figure 3.18: Chi-squared analysis spectra of the Kongsfjorden water (a) temperature and (b) salinity at 31m (blue) and 140 m (grey) depths with a 95% confidence level.

The chi-squared spectrum with a 95 % confidence level of the upper layer water temperature at 31 m (Fig. 3. 18 (a)) showed prominent peaks at 372, 256, 108, 91, 48, and 27 days, while the lower water temperature at 140 m showed peaks at 372, 178, 120, 64, and 37 days. Salinity at these depths (Fig 3. 18 (b)) showed significant annual and semiannual periodicities. Besides, the upper layer salinity showed prominent peaks at 128 and 72 days, while the lower layer salinity showed 72- and 64-day peaks. Moreover, the upper layer salinity showed distinctive peaks at high frequencies such as 8, 6, 5, and 4 days; the lower layer salinity lacked significance at these periodicities. Thus, apart from the annual and semiannual periodicities observed in the fjord water temperature and salinity, the water column from 31 m to 140 m also varied at different

intraseasonal time scales; the variabilities were more significant in the water temperature than the salinity.

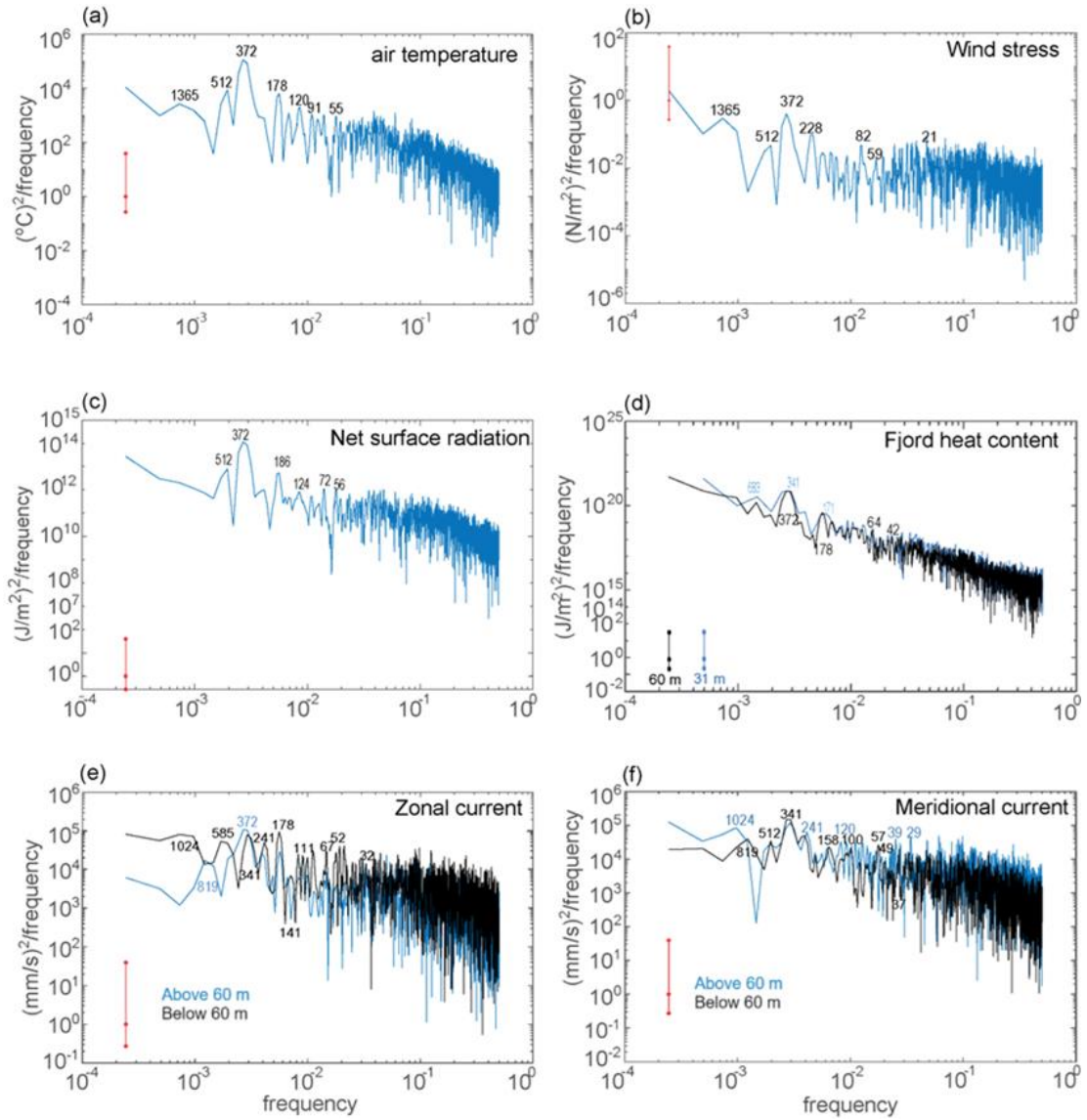


Figure 3.19: Chi-squared analysis spectra of daily weather station (a) air temperature and (b) wind stress, (c) daily net surface radiation from ERA-interim reanalyses over Kongsfjorden, (d) daily fjord heat content calculated for depths below 31 m (blue) and below 60 m (black), and (e) zonal and (f) meridional currents above (blue) and below 60 m (black) depth with 95% confidence level.

De Rovere et al. (2022) reported variabilities of several days or weeks in the inner Kongsfjorden upper water column from 2010 to 2020 which became less noticeable by the time it reached the bottom of the water column (85-92 m). However, the ISOs

were prominent at 140 m as in 31 m from our mooring data at the central Kongsfjorden. While the identified peaks at annual and semiannual periodicities are primarily due to the year-to-year and overall shifts from warm to cold seasonal variations in the hydrography, the reasons for the intraseasonal variations are not yet understood in detail in any of the earlier studies. Hence, the atmospheric and oceanic forcings were analysed to understand these intraseasonal variabilities better.

The daily weather station Ny-Ålesund air temperature (Fig. 3. 19 (a)) and wind (Fig. 3. 19 (b)) closer to the mooring location showed significant peaks at 512 and 372 days out of which 372-day periodicity was observed in the water temperature too while it was seen in salinity at a slightly lower period of 341 days. The 120-day periodicity observed in the air temperature was seen in the water temperature, whereas in 31 m, salinity showed a slightly higher period of 128 days whereas it was not significant at 140 m. While the air temperature (Fig. 3. 19 (a)) showed significant peaks at 91 and 55 days, the wind stress (Fig. 3. 19 (b)) showed 82, 59, and 21-day periodicities corresponding with the intraseasonal periodicities observed in the fjord water temperature and salinity. Further, the net surface radiation (Fig. 3. 19 (c)), a totality of the shortwave, longwave, sensible, and latent heat radiations, over Kongsfjorden showed significant periodicities such as 512, 372, 186, 124, 72, and 56 days. Out of these, 512-day periodicity was significant in the net surface radiation only, while the intraseasonal periodicities were from the long wave, sensible, and latent heat radiations, as the peaks were not as significant in the shortwave radiation. The higher and intraseasonal periodicity peaks in the net surface radiation were present in the water temperature and salinity.

The oceanic contribution to the variabilities was then investigated. The spectral analysis of the fjord heat content (Fig. 3. 19 (d)) estimated below 31m and 60 m showed significant peaks at near annual (341 and 372 days) and semiannual (178 days) periodicities. However, peaks at 64 and 42 days were prominent in the fjord heat content below 60 m. Considering this depth delineation, we analysed the fjord current above 60 m and below (Fig. 3. 19 (e, f)). While the intraseasonal peaks at 67 and 52 days were found in the zonal current (Fig. 3. 19 (e)), the intraseasonal peaks at 57 and 49 days were significant in the meridional current (Fig. 3. 19 (f)).

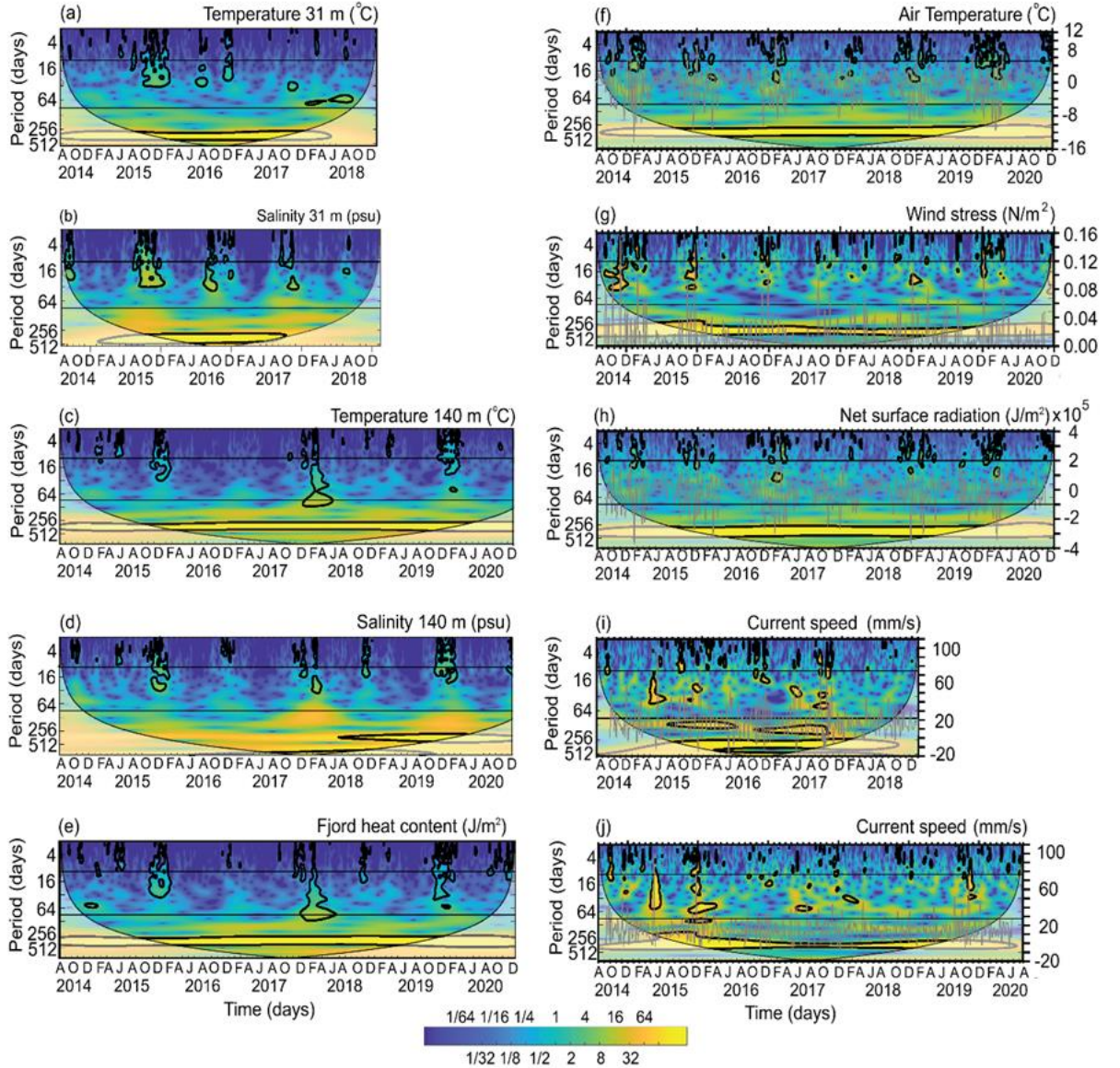


Figure 3.20: Continuous wavelet spectra of daily- 31 m (a) temperature and (b) salinity, 140 m (c) temperature and (d) salinity, and (e) 31-140 m fjord heat content from the mooring data. Continuous wavelet spectra of daily (f) air temperature, (g) wind stress, (h) net surface radiation, and current speed (i) above 60 m and (j) below 60 m depth from the mooring ADCP. The light grey line plots overlaid from (f) to (j) denote the corresponding 10-100 days band passed variables of each subplot. The two horizontal black lines on all the subplots in the figure demarcate a 10-100-day period while the contours in them denote the 95% confidence level.

Overall, the zonal and meridional currents corresponded with the higher than 100 days and intraseasonal periodicities found in the water temperature and salinity. The

corresponding intraseasonal periodicities observed in the fjord water temperature and salinity with the periodicities in the air temperature, wind stress, radiation, and current showed the role of atmospheric and oceanic factors in triggering these periodic variabilities. However, the chi-squared spectrum cannot tell the temporal evolution of these intraseasonal variations or their causative factors, which is crucial to understand. To understand this, a continuous wavelet analysis of the hydrography variables at the uppermost and lowermost depths and the atmospheric-oceanic causative factors was done. It was also tried to see to what extent the atmospheric factors could influence the shallow waters of the fjord system.

The continuous time series data from 24 July 2014 to January 2019 has been taken at 31 m for the continuous wavelet spectral analysis as the data had larger gaps in 2019 and 2020. The spectrum of the 31 m water temperature (Fig. 3. 20 (a)) showed significant energy around 10 to 32 days from September 2015 to January 2016, August to September, and then from November to December 2016. Even though high energy between 32 days and 100 days was seen from 2014 to 2015, it was not significant from 2016 till October 2017. These periodicities were found to be significant again from November 2017 to August 2018, during which the 10-32 day band was not prominent. These periods of significantly higher and lower ISO bands at 31 m coincide with frequent modifications in the water masses- mainly the SW, IW, and ArW (Fig. 3. 16 (d)). On the other hand, salinity at 31 m (Fig. 3. 20 (b)) showed significantly high energy in the 10-32 day band from August to December, from the peak summer to the beginning of winter, every year from 2014 to 2018.

The 140 m water temperature (Fig. 3. 20 (c)) showed a significant 10-32 day band, predominantly during winter from November 2015 to February 2016, December 2017 to March 2018, and December 2019 to March 2020, which are periods of more frequent alternative presence of Atlantic origin water masses and ArW/LW water masses compared to the other years. High energy was also seen in the higher periodicities above 32 days during the fall and winter seasons every year. The lowest energy at this high period band was seen at 140 m water temperature in 2018. The fjord water during these periods is weakly stratified (Fig. 3. 16 (c)) thereby leaving the entire water column also vulnerable to the effects of atmospheric forcings. Hence, the

changes in the water masses alone cannot describe the ISOs. A similar delineation and pattern in the lower and higher bands of the intraseasonal periodicities as in the water temperature were deciphered in the 140 m salinity (Fig. 3. 20 (d)). However, such a clear seasonal preference for the intraseasonal oscillations was absent at 31 m temperature (Fig. 3. 20 (a)) but was present in salinity (Fig. 3. 20 (b)). Moreover, the band between 10-32 days was more significant than the higher periods in the ISOs in the salinity at both depths. The ISOs had the least energy during the spring to peak summer in the salinity while such season-wise delineation was not found in the temperature at 31 m (Fig. 3. 20 (a)). On the other hand, the ISOs had the least energy during the spring to summer in both the temperature (Fig. 3. 20 (c)) and salinity (Fig. 3. 20 (d)) at 140 m.

The ISOs deciphered from the fjord heat content (Fig. 3. 20 (e)) were similar to that found in the temperature and salinity at 140 m. Out of the atmospheric forcings, the energy at the intraseasonal periodicity was highest in the wind stress (Fig. 3. 20 (g)), followed by the air temperature (Fig. 3. 20 (f)) and net surface heat radiation (Fig. 3. 20 (h)). The significant high energy at the 10-32 day band in the wind stress (Fig. 3. 20 (g)) during the very weakly stratified late fall to winter coincided with the water temperature and salinity at 31 m and 140 m. However, whether the wind stress is that trigger needs to be confirmed to understand whether the effects of ISOs in the wind stress reach up to 140 m. The lowest energy was found during spring and summer at the lower periods of the intraseasonal period. The increase in fjord water stratification during these seasons might have interfered with the atmospheric forcings to cause resonance at 31 m and 140 m depths. The atmospheric forcings also did not explain more significant ISOs during the winters of 2015/16, 2017/18, and 2019/20 at 140 m compared to the other years. Hence, the mean ocean currents above 60 m (Fig. 3. 20 (i)) and below (Fig. 3. 20 (j)) were studied. The current speed above 60 m (Fig. 3. 20 (i)) corresponded with the lower band intraseasonal variations in the temperature at 31 m (Fig. 3. 20 (a)) in 2015 and 2016. The significant high energy in the intraseasonal current was observed during the winters of 2015/16, 2017/18, and 2019/20 below 60 m (Fig. 3. 20 (j)), indicating the role of intrinsic oscillations in the current responsible for the temperature and salinity variations. Thus, the wind stress and current are found to play major roles in triggering the ISOs in the fjord water temperature and salinity.

We further confirmed this by doing a wavelet coherence analysis of these causative factors with the fjord hydrography variables to understand to what extent and how they covaried. The arrows pointing right in the analysis show an in-phase relation and vice versa.

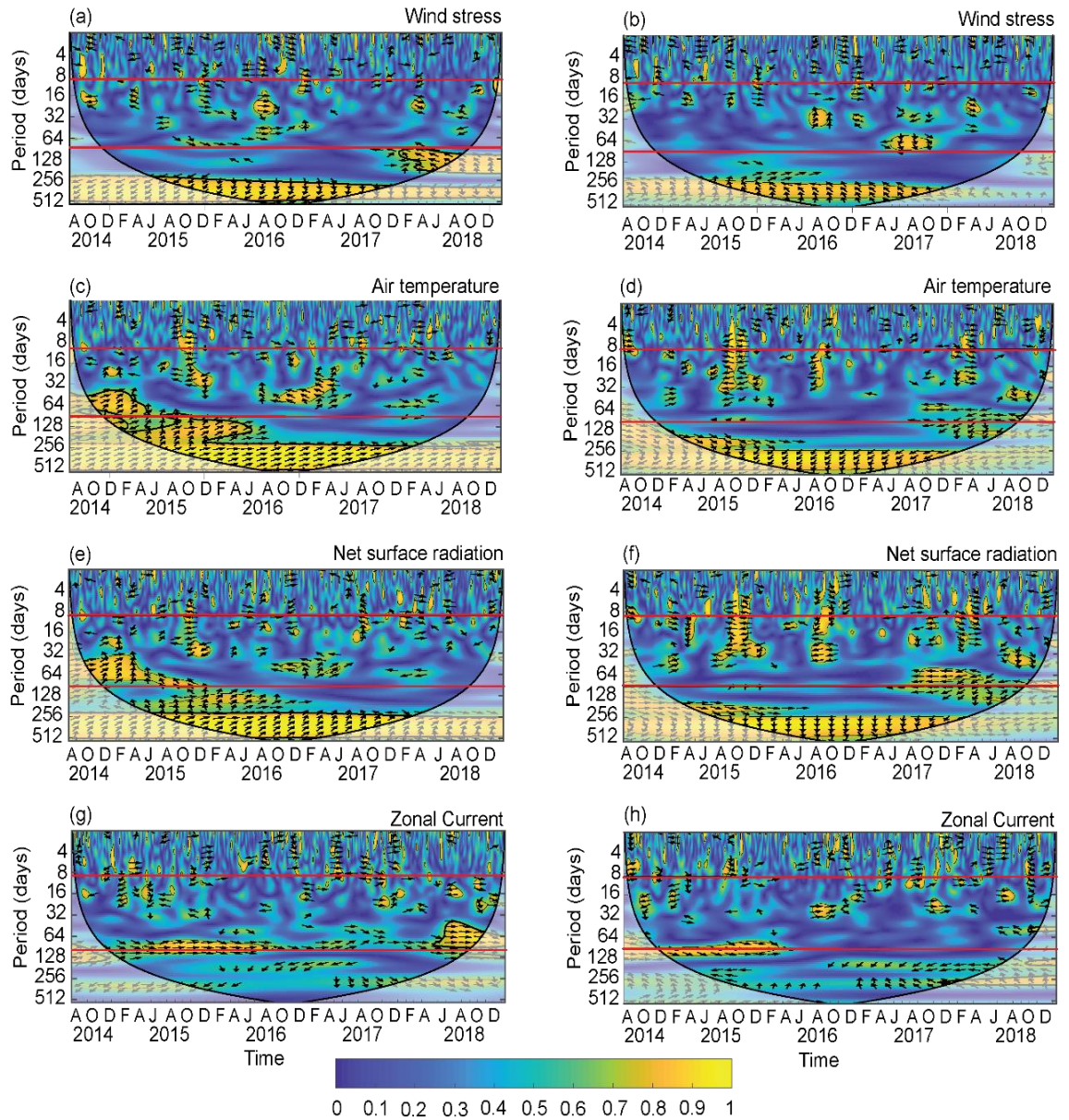


Figure 3.21: Wavelet coherence spectra of daily (left) water temperature and (right) salinity at 31 m with the daily (a, b) wind stress, (c, d) air temperature, (e, f) net surface heat radiation, and (g, h) zonal current velocity above 60 m, respectively. The two horizontal red lines in the subplots demarcate a 10-100-day period, while the contours in them denote the 95% confidence level.

The wavelet coherence of the 31 m fjord temperature (Fig. 3. 21 (a)) showed the highest and strongest in-phase covariation with the wind stress at the intraseasonal periodicities in 2015 and 2016 and for the oscillations less than 10 days. This was found throughout the observation period. In contrast, 31 m salinity (Fig. 3. 21 (b)) showed strong in-phase covariation with the wind stress only at the lower periodicities of 10-24 days, except in 2018 for periodicities of 10-64 days. Except for higher periods of ISO in 2014, the air temperature (Fig. 3. 21 (c)) and net surface heat radiation (Fig. 3. 21 (e)) showed no in-phase relation with the upper layer water temperature. Similarly, they showed a stronger covariation (Fig. 3. 21 (d, f)) with salinity only in 2018.

The water temperature at 31 m showed a strong in-phase relation with the zonal current (Fig. 3. 21 (g)) at the highest extreme period of > 96 days during the entire observation period, except in 2018, where significant periodicity was found for a wider high period ISO spectrum > 48 days when the current was the strongest southeast when compared to the other years (Fig. 3. 16 (g, h)). However, the zonal component of the current was the strongest and hence the wavelet coherence was more pronounced with the zonal current. Significant in-phase covariation was also observed with salinity and zonal currents (Fig. 3. 21 (h)) at > 96 days periodicity during 2014 and 2015; and at ~ 10 -32 days period in the summers of 2016, 2017, and 2018. Higher than 256 days in the water temperature showed significant in-phase relation with air temperature (Fig. 3. 21 (c)) and net surface radiation (Fig. 3. 21 (e)) over the entire observational period. In brief, the ISOs at 31 m fjord water temperature were predominantly controlled by the ISOs in the wind stress and the zonal current. While the lower period ISOs in the temperature and salinity were triggered by the wind stress, the higher periodicities were found to be driven by the current.

At the deeper water level of 140 m, even though the wind stress (Fig. 3. 22 (a, b)), air temperature (Fig. 3. 22 (c, d)), and net surface heat radiation (Fig. 3.22 (e, f)) showed periods of covariation during intraseasonal and less than 10 days periodicities, significant and covarying in-phase relation of the temperature and salinity existed only with the zonal current (Fig. 3.22 (g, h)) during the winters of 2015/16, 2017/18, and 2019/20. These were the winters where the external warm and local water masses

showed their presence alternatively and frequently compared to the other winters (Fig. 3. 16 (d)), indicating varying current direction within the season as can be seen in the current (Fig. 3. 16 (g, h)) and hence significant ISOs. The relation with the meridional current during the winter was also similar. This shows the dominant role of fjord currents at the deeper depths in triggering the ISOs compared to the atmospheric forcings.

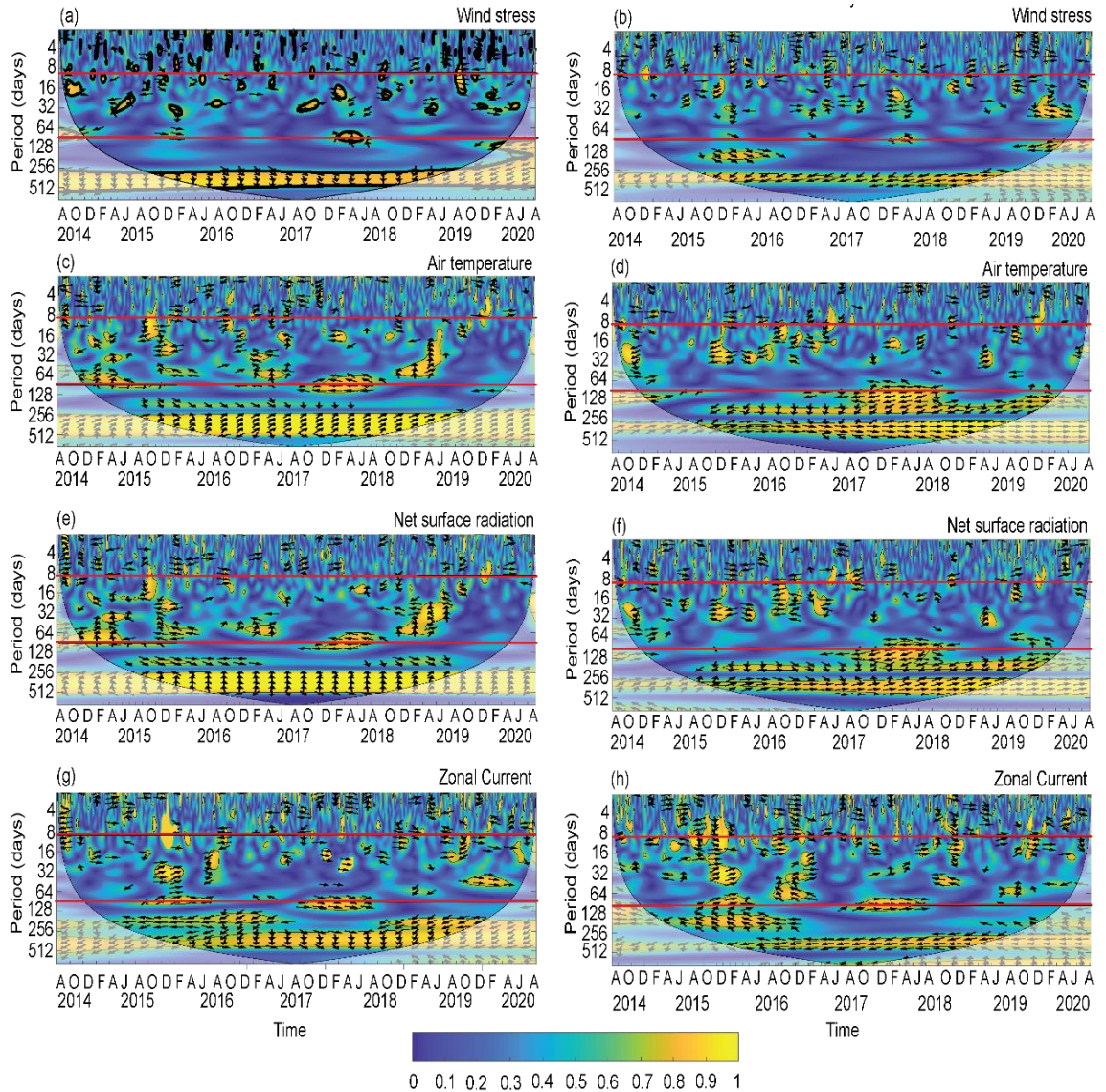


Figure 3.22: Wavelet coherence spectra of daily (left) water temperature and (right) salinity at 140 m with the daily (a, b) wind stress, (c, d) air temperature, (e, f) net surface heat radiation, and (g, h) zonal current velocity below 60 m, respectively. The two horizontal red lines demarcate the 10-100-day period, while the contours denote the 95% confidence level.

For higher than 256 days periodicities, the 140 m temperature showed strong in-phase covariation with the wind stress (Fig. 3. 22 (a)) and air temperature (Fig. 3. 22 (c)), while the salinity showed strong in-phase covariation with air temperature (Fig. 3. 22 (d)) and net surface radiation (Fig. 3. 22 (f)). On the other hand, salinity covaried almost out of phase with the wind stress (Fig. 3. 22 (b)) and zonal current (Fig. 3. 22 (h)) at 140 m. Thus, the wind stress and current speed played major roles in imbibing ISOs in the fjord water column.

The ISOs were generally significant from peak summer to winter at the upper layer while it was seen during fall and winter at the deeper fjord. The ISOs at the deeper waters were found to be driven majorly by the current and these periods are the periods of maximum AW intrusion in the last decade. Due to a homogenous water column during the seasons, wind stress also seems to have influenced the ISO at deeper layers.

3.6. Conclusion

The CTD survey data in summer 2011-2020 and time series subsurface mooring data in 2014-2020 in Kongsfjorden were used to study the interannual physical variations in the fjord water column. The fjord showed a warming ($0.06\text{ }^{\circ}\text{C/yr}$) and salinification (0.03 psu/yr) in summer. While 2011 was the coldest and least saline summer, 2014 was the warmest and most saline in the decade. The water masses in Kongsfjorden showed different scenarios in summer with respect to the surface and Atlantic water masses. When the SW layer was thick, the AW layer was thin, and vice versa. 2014 had the thickest AW and thinnest SW while 2011 had the thickest SW with no AW. The rate of AW warming, $0.07\text{ }^{\circ}\text{C/yr}$, was $0.02\text{ }^{\circ}\text{C}$ higher and the rate of salinification of the AW, 0.008 psu/yr , was one-fifth than that of the SW. The increasing net surface radiation and air temperature corresponded with the increase in SW temperature. On the other hand, the increasing trend in precipitation and glacial discharge justified the SW freshening, which was inferred from the decrease in the minimum salinity trend. The highest fast ice coverage in the cold period prior to the summer with the highest precipitation and glacial discharge in 2013 led to considerably higher SW layer thickness and thinner AW layer. The mean south-south easterlies that prevailed in 2014, 2016, 2017, and 2019, the years of high AW volume, confirmed the wind-driven mechanism of the surface freshwater outflow from Kongsfjorden enhancing the flow

of the Atlantic water into the fjord. Besides, in 2014, the sea ice coverage, concentration, and thickness were the lowest, which created a conducive situation for the maximum Atlantic water intrusion to Kongsfjorden. The role of local sea ice conditions in the Atlantic water variability inside Kongsfjorden was also evident in the close correspondence of the fjord temperature with the WSC in the shelf in 2014, 2016, and 2017 which also had the least local sea ice coverage. Also noticeable is that the interannual variation in the SW temperature and salinity were similar to the AW, indicating the increasing influence of the AW in the upper layers of the water column over the years through vertical mixing and also due to the entry of the Atlantic water through the fjord surface. Finally, the long-term mean temperatures over the water column depths for the summers of 2011-2020 were 0.5-0.8 °C higher than that in 1994-2014 (1.5-5.0 °C) reported by Tverberg et al. (2019).

The circulation in the Kongsfjorden-Krossfjorden and the shelf-slope region was studied using the ROMS model simulations for 2013-2018, the period of freshwater data availability. The simulations showed the entrance of the warm Atlantic waters from the WSC through the southern shore of Kongsfjorden. The circulation pattern shows the heat supply by these warm waters to the head, leading to glacial calving and melting. A portion of the freshwater outflow from the Kongsfjorden, through its northern side, enters Krossfjorden through its eastern side in the direction of the direct Atlantic water entry to the fjord. Mixing freshwater from Kongsfjorden decreases the temperature in the eastern Krossfjorden. Hence, the temperature is lower in Krossfjorden than in Kongsfjorden. Laterally, the Kongsfjorden water column showed two temperature scenarios- a two-layer system in the outer Kongsfjorden having warmer-less saline surface water above cold-high saline deep water and a three-layer system in the inner fjord having warm-saline water sandwiched between cold-less saline surface water and cold-saline deep water. The temperature inversion in the first two layers in the inner fjord deepened from July to October, with the strongest inversions occurring in August-September, coinciding with the maximum freshwater discharge in the fjord surface. Besides the freshwater discharge, decrease in air temperature, increase in winds, net surface heat flux loss, and increase in precipitation from August to October add to the upper-layer cooling and thereby the formation of temperature inversion and a three-layer temperature structure in the inner fjord.

The daily time series data from the mooring from 2014 to 2020 were used to understand the variabilities, especially in winter, spring, and fall in Kongsfjorden. Even though 2020 was the longest winter with longer residence of LW, the coldest winter temperature of $-1.0\text{ }^{\circ}\text{C}$ was found in February 2015, and the lowest salinity of 34.4 psu in February 2018. The temperature increased in summer, winter, and fall from 2014 to 2020, at a rate highest in summer followed by winter and then fall. The rate of winter temperature rise found at the mooring location, which is the central Kongsfjorden, was $0.08\text{ }^{\circ}\text{C/yr}$, half the rate reported in the outer part of the Kongsfjorden which confirmed sharp lateral gradient in the fjord hydrography. But, the rate of warming in winter was higher than that of the summer warming. However, the spring temperature showed a decrease over the period due to the delay and extension of winter overlapping the defined spring months that was observed frequently in most of the years during the decade. The maximum temperature of the AW layer below 60 m during months with maximum intrusions (August-October) showed a temperature increase of $0.05\text{ }^{\circ}\text{C/yr}$ from 2014 to 2020, the same rate found below 60 m in July-September averages as well. The rate corresponded to the recent reports of an increase in winter water temperatures by Tverberg et al. (2019) and Cottier et al. (2021) in Kongsfjorden. However, the water temperature showed a three-month difference, generally between the temperature peaks at 31 m in August and lowers at 140 m in November each year. The period between these peaks coincided with a sharp decline and subsequent rise in salinity, especially at 31 m. Salinity at 140 m showed seasonality as in 31 m but embedded in a low-frequency variation. Stratification maximum in summer was 2-5 times higher in the upper 60 m layer than below it. The highest stratification was found in 2015 and 2020, with the largest surface freshening with a vertical arrangement of SW, IW, and ArW one below the other, and the lowest stratification in 2014 and 2017, the warmest years of the decade. The stratification increases and peaks below 60 m coincided with the pure AW intrusion while its decline coincided with the modification of AW to TAW or LW. After its presence in the 2015 winter, LW's presence in the fjord during winter and spring increased from 2018 to 2020, which is a reason for the decreasing winter and spring mean temperatures during these three years. The longest and most dominant presence of the TAW and AW in 2016 and 2017 resulted in the warmest winters.

The fjord HC became maximum one to three months after the peak in the net surface radiation flux during 2015-2020. The fjord HC continued rising even after the net surface radiation started declining and even when the fjord was losing heat to the atmosphere. These periods of increasing fjord HC and declining surface radiation coincided with the periods of Atlantic water presence in Kongsfjorden. The increase in southeast current in 30-60 m at the mooring location, a counter-current to the surface northwest current, points towards an inflow to the fjord during increasing HC. During the same time, below 60 m, we found a northward flowing meridional current towards the mooring location, which transports the Atlantic water from the southern shore to the mooring location. From October till the end of the cold season with maximum AW intrusion, we found prominent southeast currents below 60 m, implying the transport of heat by the warm Atlantic water to Kongsfjorden. Thus, the observed fjord heat flux combines the net surface heat flux with the advected heat flux. The flux analysis also proved to be a promising way of detecting periods of AW advection in Kongsfjorden.

The interannual variations in the Kongsfjorden water column temperature and salinity depended on the interannual condition of sea ice, such as its coverage, concentration, thickness, and length of sea ice season. The years with the shortest fjord ice season and lower coverage and concentration resulted in the most warm and saline years, and vice versa. The high fast ice coverage in 2015 led to the coldest winter water column. The shortest fjord ice season and low ice extent resulted in the warmest and most saline winter to spring in 2017. On the other hand, even with a comparable mean fast ice coverage to 2015 conditions, the deeper level TAW intrusion episodes in winter 2018 resulted in a warmer winter than in 2015. Thus the overall effect of the winter sea ice conditions was carried over to the spring, summer, and fall hydrography environment in the fjord. The direction of winds changed between southeasterlies (downfjord) and southwesterlies (across the fjord) during winter. The highest wind speed with the lowest stratification, one order lesser than in summer, added to the coldest and vertically more homogenous water column than the other years. On the other hand, when the mean winter wind speed was the lowest in 2020, the stratification was the highest. Thus, winds had a higher impact in the upper 60 m in summer and almost the full water column in winter.

The least investigated ISOs in Kongsfjorden hydrography were studied using the mooring time series observations. Besides the strong annual and semiannual periodicities, the spectral analysis of the water temperature (91, 48, and 27 days at 31 m; 64 and 37 days at 140 m) and salinity (72 days at 31 m; 72 and 64 days at 140 m) showed distinctively prominent but nearby periodicities of ISOs dominating at 31 m and 140 m. These variabilities were more significant in the water temperature than the salinity. The upper layer salinity showed distinctive peaks at high frequencies such as 8, 6, 5, and 4 days too. The continuous wavelet spectrum showed that the ISOs were prominent generally from late summer to winter at 31 m, where frequent modifications in the water masses occurred. At 140 m, the ISOs were stronger in the fall-winter when the fjord water was weakly stratified, leaving the entire water column vulnerable to the effects of atmospheric forcings to reach deeper depths. The ISOs had the least energy during the spring to summer at both depths. The increase in fjord water stratification during these seasons might have interfered with the atmospheric forcings to cause resonance at 31 m and 140 m depths, especially at the lower periodicities. The wind stress and current speed played major roles in imbibing ISOs in the fjord water column. While the lower period ISOs in the temperature and salinity at 31 m were triggered by the wind stress the current drove the higher periodicities. The ISOs at the deeper waters were found to be driven majorly by the current and these periods of significant ISOs at deeper levels are the periods of maximum AW intrusion in the last decade. Due to the existence of a homogenous water column during these seasons, wind stress also seems to have influenced the ISOs at the deeper layer.

The study showed the continuous and long residence of Atlantic-origin waters at the deeper Kongsfjorden which is important with respect to the shifts the fjord is undergoing, such as changes in physical processes, the community structure and species of the fjord biota, etc. The increase in the number of warm years over the decade shows the prominence of warm Atlantic waters to Kongsfjorden and its rapid shift to Atlantic type. The updated observational information on Kongsfjorden hydrography can add to the understanding of the physical and biogeochemical processes in the west Svalbard fjords. In addition, the inferences will help address modeling challenges in the region and the coherent depiction of the dynamic Arctic fjord processes.

Chapter 4

Biogeochemical variabilities in Kongsfjorden and the role of physical processes

4.1. Background

The recent increase in the Atlantic water inflow to Kongsfjorden (Węśławski et al., 2017), sea-ice decline, retreating glaciers, and increased melting (Promińska et al., 2018) not only affect its physical characteristics but also its biogeochemistry. The glacial melt contributes to particulate matter affecting the underwater light conditions (Piquet et al., 2014). The freshwater discharged into the fjord decreases salinity and is responsible for the surface water mass. It also mixes with warm-saline water from the ocean and plays a role in forming other water masses (Svendsen et al., 2002). Meanwhile, as seen in the earlier chapters, the warm Atlantic water entering the fjord also modifies the water masses and makes them warmer (David and Krishnan, 2017). Heterogeneity has increased in the water column in the last two decades due to the mixing of different water masses, visible in physical and biogeochemical properties at various depths (Hop and Wiencke, 2019). The altered physical and biogeochemical characteristics, including light, nutrients, temperature, and mixing, affect phytoplankton bloom development in the fjord.

The seasonal phytoplankton bloom, also known as the spring bloom, occurs yearly in the entire Arctic Ocean. Solar insolation and sea-ice cover in the Arctic play a major role in determining the start and length of this productive season (Oziel et al., 2019). Phytoplankton is subjected to a protracted duration of darkness due to the limitation of insolation during the polar night. The spring brings light, resulting in thermal stratification and kickstarting the bloom (Winder and Cloern, 2010). Once stratification becomes stronger in summer, the spring bloom exhausts surface nutrients and dies (Piquet et al., 2010). Subsequently, an active phytoplankton community that balances light and nutrient conditions (Brown et al., 2015; Barbieux et al., 2019) is found below the MLD (Coupel et al., 2011; Ardyna et al., 2013). This subsurface

chlorophyll maximum (SCM) is caused by sinking surface biomass and low light acclimated algal cells with increased chl *a*. The SCM is expected to bestow primary production and biogeochemical cycles (Ardyna et al., 2013). Thus, the interconnected physical, chemical, and biological factors such as light, temperature, mixing, nutrient availability, respiration, and grazing pressure are responsible for phytoplankton bloom (Cabré et al., 2016; Rohr et al., 2017).

According to Ji et al. (2010), the phytoplankton bloom phenology is a sensitive indicator of climate change. The availability of underwater light in the Arctic is affected by the decline, early break-up, and late freezing of sea ice (Nicolaus et al., 2012). The thinning of sea ice and early break-up results in huge under-ice blooms (Arrigo et al., 2012) and early-shifted blooms (Kahru et al., 2011). High wind occurrences in conjunction with delayed sea-ice freezing enable higher latitude summer or fall bloom (Ardyna et al., 2014), suggesting a poleward shift of bloom features (Yool et al., 2015). Recent and intense blooms have been observed in the Arctic with warm water species composition (Hegseth and Sundfjord, 2008). Kongsfjorden ecosystem, like the rest of the Arctic Ocean, is changing its bloom phenology. A shift in bloom timing in the period between March–June has been observed under varying sea-ice conditions and Atlantic water inflow (Hegseth and Tverberg, 2013; Hodal et al., 2012; Piquet et al., 2014; van De Poll et al., 2016; Assmy et al., 2023). Freshwater and particulate input from meltwater changes stratification and underwater light availability, particularly in the inner fjord influencing spatial variability of bloom (van De Poll et al., 2016). Blooms are also observed in late summer and autumn (Caroppo et al., 2017; Hegseth et al., 2019). The phytoplankton community dynamics were affected by both Atlantic water intrusion and glacial melt in Kongsfjorden (Hegseth and Sundfjord, 2008; Piquet et al., 2014; Assmy et al., 2023). The existing understanding of Arctic ecosystems and, thereby, Kongsfjorden is a bottom-up controlled system with seasonality in primary production. Accordingly, the Arctic polar night is considered a time with minimum biological activities due to limited nutrients. However, Berge et al. (2015) documented the system showing high biological activity and interactions across trophic levels during polar night. They found that the diversity and presence of juvenile stages in some habitats were enhanced in winter than in the generally productive and sunlit seasons.

Most of the biogeochemistry studies in Kongsfjorden are in the productive spring and summer, which provided key information on phytoplankton bloom phenology. Even though some studies cover nearly the entire annual cycle, including winter and fall, they don't address the bloom phenology (Iversen and Seuthe, 2011; Seuthe et al., 2011; Darnis et al., 2017). Even with these detailed snapshots from discrete datasets, a complete and continuous story is missed. This affects the understanding of what happens during the period of data gaps, such as what happens during the polar night and the role of winter convective mixing towards the interannual variability of phytoplankton bloom phenology (Mayot et al., 2017), etc. Phytoplankton responds to climate scale variations and small-scale variabilities like heatwaves, winds, and upwelling events (Winder and Cloern, 2010). Understanding such unexplored seasons is important to understand the shifting Kongsfjorde biogeochemical system. Learning more about the annual and interannual variabilities will help delineate how the system works now and where it can go, especially when Arctic regions are experiencing accelerated environmental change. The lack of year-round and continuous biogeochemical observations kept a gap in describing its annual cycle and variability in Kongsfjorden.

The present chapter explores the biogeochemical variables in all seasons from 2015 to 2018, with a special focus on the annual cycle of biological (chlorophyll) and chemical (nitrate) variability using year-round in-situ sensor-based sub-surface data from the mooring. Section 4.2 showcases the annual and interannual variabilities in the biogeochemistry of Kongsfjorden. Section 4.3 addresses the interannual variations in the spring and secondary phytoplankton blooms, their development, and their coupling to the oceanic and atmospheric forcings. Section 4.4 describes the role of Atlantic water intrusions to the fjord nutrient variability while section 4.5 concludes the results of the study.

4.2. Biogeochemical variabilities in Kongsfjorden

The daily average of the time series subsurface biogeochemical variables monitored from the mooring at two upper layer depths above 50 m from 2015 to 2018 was used for the study. The details of the sensors with the variables and depth at which they were measured for each mooring deployment-retrieval period are given in Table 4.1.

Table 4.1: Details of moored biogeochemical and physical sensors and observed variables with their year-wise nominal depths of deployment from 2015 to 2018.

Sl. No.	Sensors	Variables	Year	Depths
1	ECO-FLNTUS	1. Chlorophyll <i>a</i> Fluorescence 2. Turbidity	2015-2016	25 m; 35 m
			2016-2017	32 m; 45 m
			2017-2018	27 m; 36 m
2	SUNA V2 UV	Nitrate	2015-2016	37 m
			2016-2017	47 m
			2017-2018	40 m
3	ECO-PAR	Photosynthetically Active Radiation	2015-2016	25 m, 35 m
			2016-2017	45 m
			2017-2018	27 m; 36 m
4	SBE-37 CTD-IMP-ODO MicroCAT	1. Temperature, 2. Conductivity (salinity), 3. Dissolved Oxygen (DO)	2015-2016	24 m, 34 m
			2016-2017	31 m, 41 m (no DO at 41 m)
			2017-2018	25 m; 38 m
5	Teledyne 300 KHz upward-looking ADCP	current	2015-2016	30-50 m
			2016-2017	
			2017-2018	

4.2.1. Annual cycle

4.2.1.1. 2015-2016

The daily average chl *a* concentration (Fig. 4.1 (a)) varied between 0.05-7.19 $\mu\text{g/L}$ at 25 m and between 0.04–4.99 $\mu\text{g/L}$ at 35 m. The lowest chl *a* was observed during December 2015-March 2016, while the highest chl *a* on 18 May. Dissolved Oxygen (DO) (Fig. 4.1 (b)) at 24 m varied between 6.62-8.85 ml/L, while at 34 m varied between 6.47-8.29 ml/L. Even though the DO values at 34 m were lesser than at 24 m, their changes over the period closely corresponded with minimum DO values found in November and maximum in April-May. The nitrate concentration at 37 m (Fig. 4.1 (c)) varied between 1.01-13.30 μM . While the lowest nitrate values were found in

August, the highest was in April. The photosynthetically active radiation (PAR) was measured at the same depths (Fig. 4.1 (d)) as the chl *a* was highest by mid-April with smaller peaks by the end of April and early-May. PAR at 25 m varied between 0.04-8.85 $\mu\text{mol photons/m}^2 \text{ s}$ and at 35 m varied between 0.04-3.59 $\mu\text{mol photons/m}^2 \text{ s}$. Turbidity (Fig. 4.1 (e)) at 25 m varied between 0.15-5.17 NTU and at 35 m varied between 0.20-9.19 NTU, showed higher turbidity at 35 m. Turbidity at both depths as chl *a* showed higher values in June-September 2015, with the highest observed in July 2016.

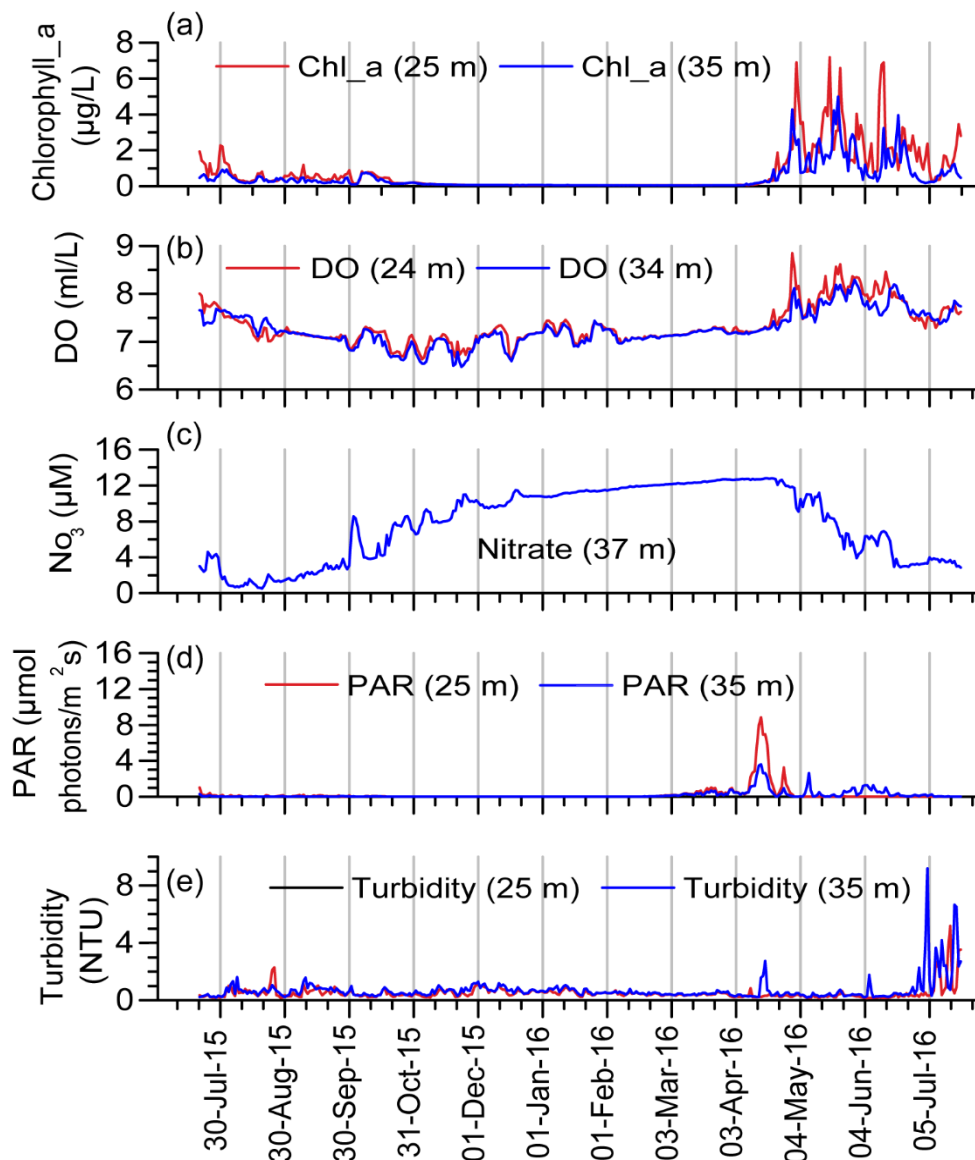


Figure 4.1: Daily (a) chlorophyll *a* (chl *a*) at 25 m and 35 m, (b) DO at 24 m and 34 m, (c) nitrate at 37 m, (d) PAR at 25 m and 35 m, and (e) turbidity at 25 m and 35 m during 2015-2016 from the mooring.

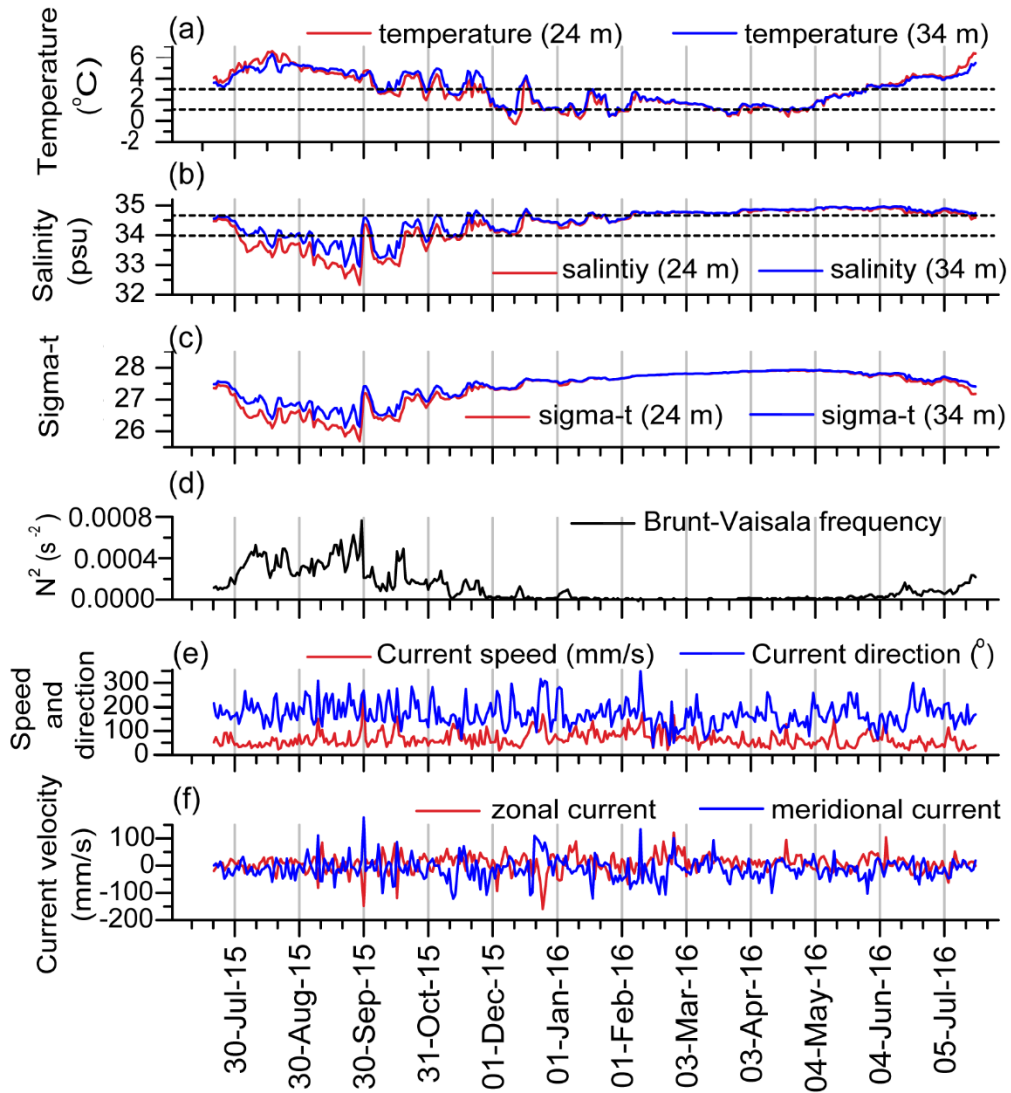


Figure 4.2: Daily water (a) temperature, (b) salinity, and (c) sigma-t at 24 m and 34 m during 2015-2016 from the mooring. Daily (d) Brunt-Vaisala frequency between 24 and 34 m, (e) current speed and direction, and (f) zonal and meridional current velocity during 2015-2016 from the mooring. The two horizontal black lines in (a) demarcate 1 and 3 °C and that in (b) demarcate 34.00 and 34.65 psu.

Water temperature (Fig. 4.2 (a)) at 24 m varied between -0.30–6.59 °C and at 34 m varied between 0.43-6.33 °C. The lowest temperature at 24 m and 34 m was found on 12 December 2015 and 26 January 2016, respectively. Even with the highest temperature found at 24 m depth, the negative temperatures occurred at the same level, making the lower depth warmer. However, in July-August 2015 and July 2016, the water temperature at 34 m was lower than the warm upper layer. Temperature was high in July and August with the highest values at both the depths on 17 August 2015.

Low temperatures were found during December-January. Salinity (Fig. 4.2 (b)) varied between 32.33-34.95 psu at 24 m and 32.95–34.97 psu at 34 m. At both depths, salinity was highest in May and lowest in September. Sigma-t (Fig. 4.2 (c)) varied between 25.69-27.92 at 24 m and 26.12-27.92 at 34 m. Changes in density resonated salinity variations over time at both depths, with the highest density in May and the lowest in September. The Brunt-Vaisala frequency, N^2 (Fig. 4.2 (d)), calculated from the density varied between $0-7.6 \times 10^{-4} \text{ s}^{-2}$. N^2 showed an overall increase from 24 July to 29 September 2015. The highest N^2 was found on 29 September, which then decreased and became close to zero by the first week of January 2016. The subsurface currents (Fig. 4.2 (e)) were weaker in August 2015 followed by an overall increase in speed (Fig. 4.2 (e)) from September 2015 to mid-February 2016 after which it slowed down again till April flowing towards the southeast (Fig. 4.2 (f)). Early and end-May saw higher current speed followed by a slow down again.

The daily net surface radiation, air temperature, wind speed, direction, and stress it imparted to the fjord, precipitation, and glacial discharge (Fig. 4.3) during the same time as the fjord biogeochemical and physical observations were then analyzed to understand the role of atmospheric factors in the annual biogeochemical cycle. The net surface radiation and air temperature (Fig. 4.3 (a)) closely corresponded concerning seasonal and intraseasonal variations. The net surface radiation varied between $-6.94-2.83 \times 10^5 \text{ J/m}^2$ while the air temperature varied between $-18.43-11.93^\circ\text{C}$. The highest radiation and air temperature was observed in July 2015 and 2016. While the lowest net surface radiation was found by late December 2015 and mid-March 2016, the lowest air temperature was found by mid-December 2015 and then again by mid-March 2016. They decreased from July to December 2015 and then showed an overall increase till July 2016. The variability was more from December 2015 to March 2016. Winds (Fig. 4.3 (b)) strengthened from September 2015 to January 2016, followed by a decrease. Winds varied between 0.45-18.95 m/s during the observational period. Storms that were higher than 10 m/s were seen from October 2015 to January 2016. Wind stress (Fig. 4.3 (c)) varied between 0.0004-0.54 N/m^2 . The highest wind speed coincided with the strongest wind stress found by mid-December. Winds were prominently southeasterlies (downfjord), while the second dominant was southwesterlies. Wind stress was higher from October 2015 to February

2016 with an average of 0.07 N/m^2 for the period. Glacial discharge (Fig. 4.3 (d)) varied between $0.1\text{-}376.30 \text{ m}^3/\text{s}$ and was higher in July-August 2015 and June-July 2016. The highest glacial discharge was seen by mid-August 2015. The discharge decreased from July to September 2015 while it increased from June to July 2016. Total precipitation varied between $0.10\text{-}26.70 \text{ mm}$ (Fig. 4.3 (d)) and was higher in September 2015 and early January 2016.

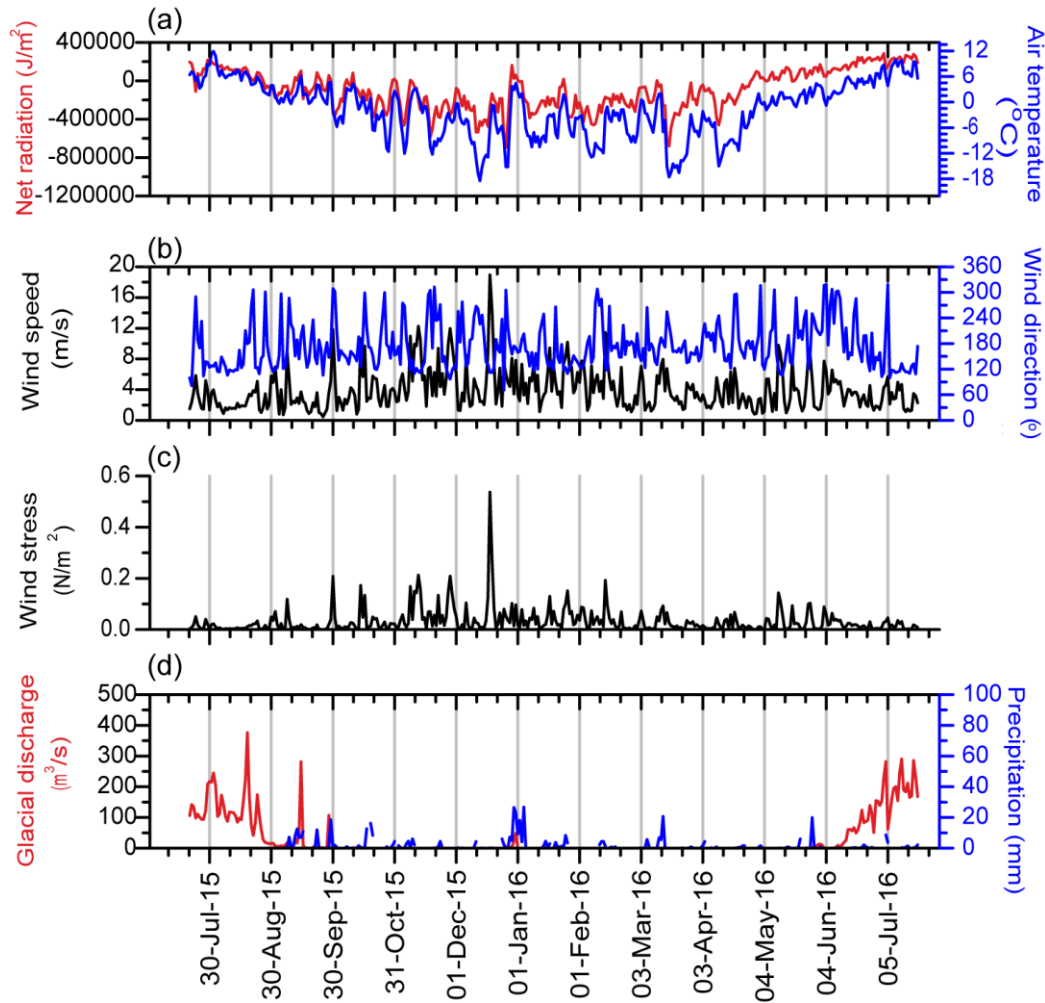


Figure 4.3: Daily meteorology data from the Ny-Ålesund weather station- (a) net surface radiation and air temperature, (b) wind speed and direction, (c) wind stress, and (d) glacial discharge and total precipitation during the mooring data period of 2015-2016. Glacial discharge data source: Pramanik et al. (2018).

4.2.1.2. 2016-2017

The chl *a* concentration (Fig. 4.4 (a)) varied between 0.04–7.68 $\mu\text{g/L}$ at 32 m and between 0.04–6.69 $\mu\text{g/L}$ at 45 m. The lowest chl *a* was observed during December 2016–March 2017 while the highest chl *a* in early May, the pattern of high and low during the same period as observed during 2015–2016. DO (Fig. 4.4 (b)) at 31 m varied between 6.30–8.07 ml/L and corresponded to the variations of highs and lows observed in chl *a*. Nitrate concentration at 47 m (Fig. 4.4 (c)) varied between 4.40–15.46 μM . The lowest nitrate values were found in late September 2016 followed by August 2016. The highest nitrate was in mid-April 2017 as in 2016.

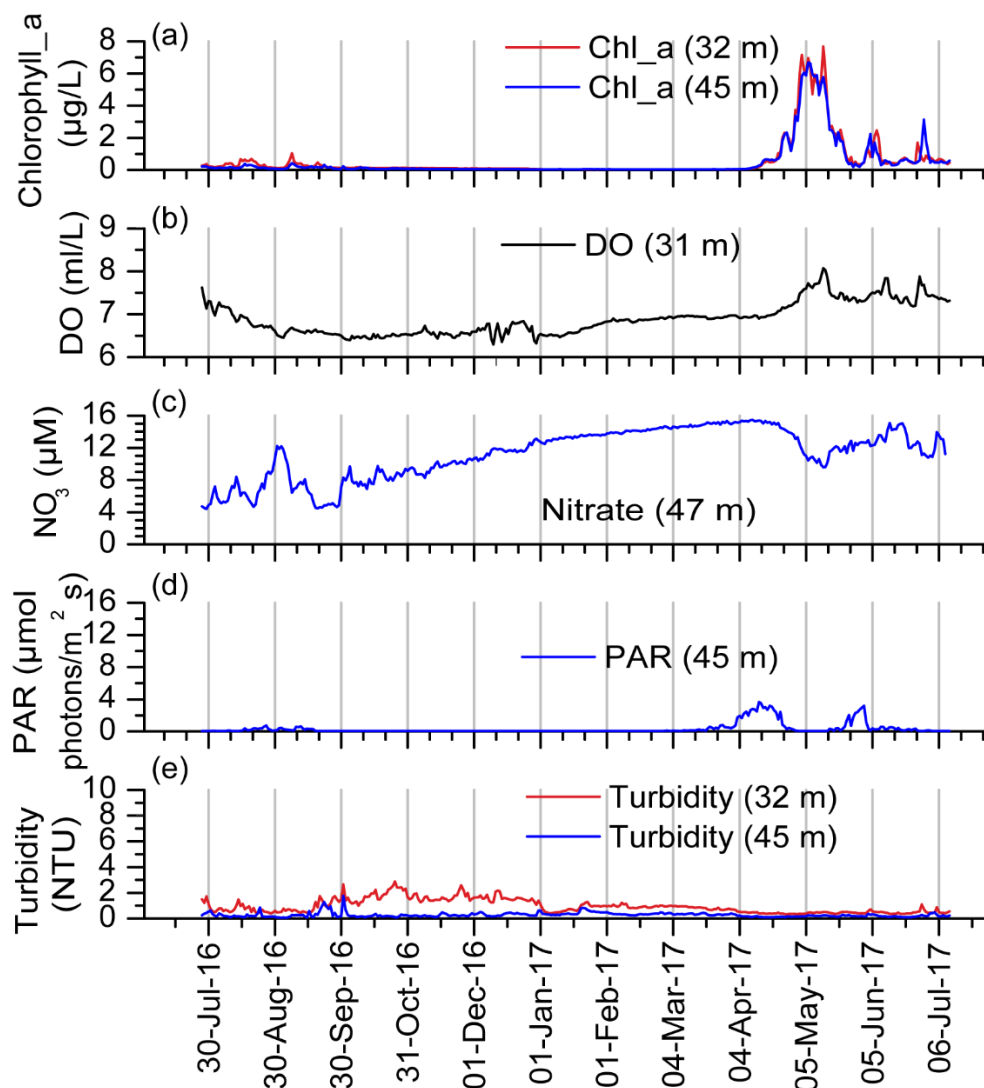


Figure 4.4: Daily (a) chlorophyll *a* (chl *a*) at 32 m and 45 m, (b) DO at 31 m, (c) nitrate at 47 m, (d) PAR at 45 m, and (e) turbidity at 32 m and 45 m during 2016-2017.

Since the PAR sensor at 32 m failed to work, the PAR measurements during 2016-2017 were available only for the deeper 45 m (Fig. 4.4 (d)). PAR at 45 m varied between 0.06-3.65 $\mu\text{mol photon/m}^2\text{s}$ and was highest by mid-April 2017 when the highest nitrate was found followed by a similar highest PAR value again by end-May. Turbidity (Fig. 4.4 (e)) levels at 32 m (0.33-2.88 NTU) was higher than that at 45 m (0.08-1.75 NTU). Turbidity at both depths showed higher values from July-December 2016 with the highest of 2 NTU observed on 23 October 2016 at 32 m.

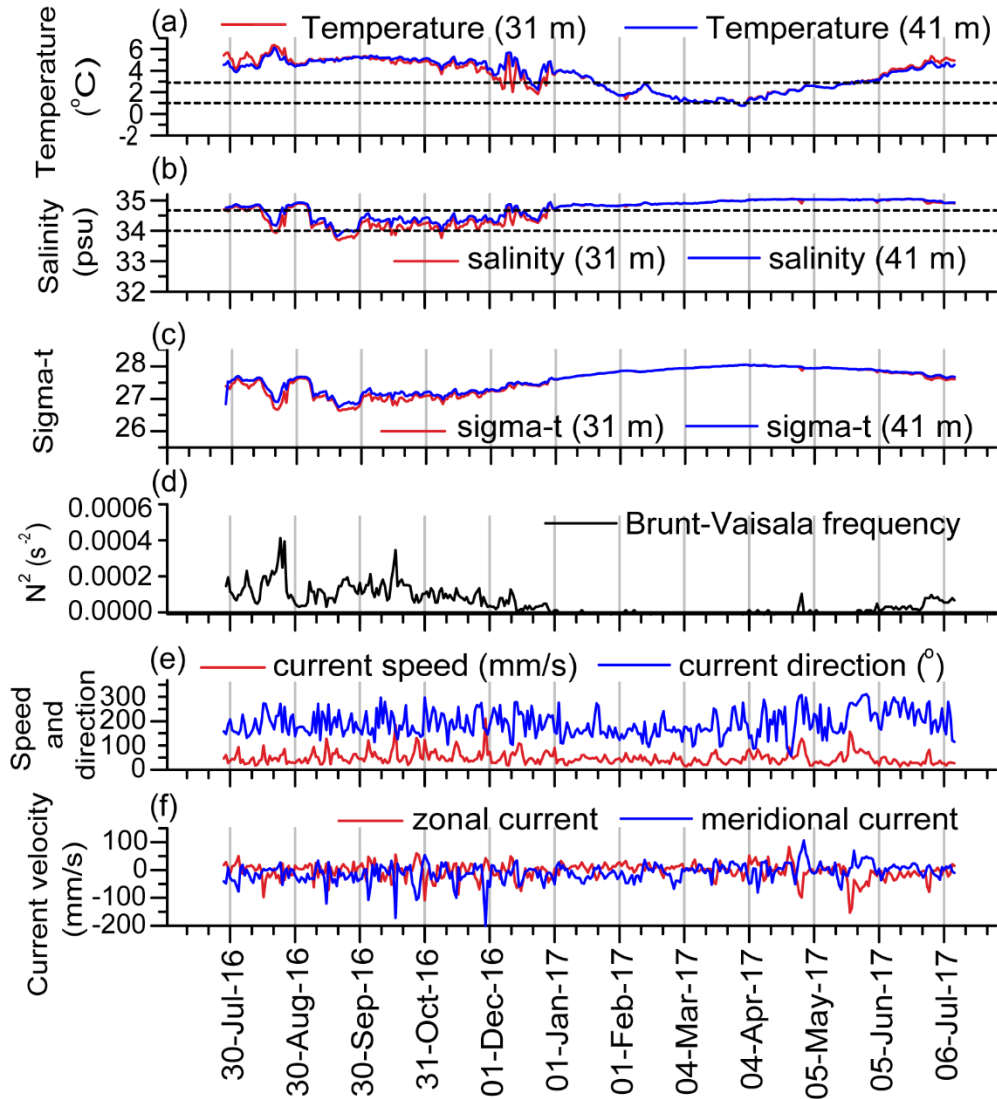


Figure 4.5: Daily water (a) temperature, (b) salinity, and (c) sigma-t at 31 m and 41 m. Daily (d) Brunt-Vaisala frequency between 31 and 41 m, (e) current speed and direction, and (f) zonal and meridional current velocity during 2016-2017. The two horizontal black lines in (a) demarcate 1 and 3 $^{\circ}\text{C}$ and that in (b) demarcate 34.00 and 34.65 psu.

Water temperature during 2016-2017 (Fig. 4.5 (a)) varied between 0.77-6.38 °C at 31 m and 0.76-6.1 °C at 41 m, indicating warmer waters at the upper layer where chl *a* was also higher than the lower depth. The temperature was highest at the end of August 2016 while it was highest at mid-August 2015. Low temperatures were found during December 2016 and January 2017. The lowest temperature was found at both depths by end-March 2017, while the lowest temperature was found earlier in December 2015-January 2016. Salinity (Fig. 4.5 (b)) varied between 33.69-35.05 psu at 31 m and 33.82–35.05 psu at 41 m. At both depths, salinity was highest in mid-June while it was highest one month before in 2016 and lowest on 20 September 2016 the same month as in 2015. Changes in density (Fig. 4.5 (c)) resonated with overall salinity variations at both depths. Sigma-t varied between 26.63-28.05 at 31 m and 26.74-28.05 at 41 m. However, the highest density was found on 04 April 2017 at both the depths and the lowest on 20 September 2016 as in salinity. The upper water column of the fjord acquired the highest density one month before in 2017, compared to 2016. However, the lowest density was in September 2016 and 2017. The N^2 (Fig. 4.5 (d)) varied between $0-4.1 \times 10^{-4} \text{ s}^{-2}$ with the highest N^2 found on 22 August 2016. N^2 showed an overall decrease from July to December 2016, while the N^2 values were higher from July to December 2015, which became zero by January 2016. The subsurface currents were stronger from August 2016 to January 2017 (Fig. 4.5 (e)). The currents fluctuated between southeasterly and southwesterly directions from July to December 2016 (Fig. 4.5 (e, f)), after which, from January to February 2017 currents slowed down and flowed southeast into the fjord, which was found in the period of subsurface Atlantic water intrusions (Fig. 3.16 (d), chapter 3). Currents became faster afterward till May to decrease again.

The atmospheric factors and glacial discharge during 2016-2017 were analyzed (Fig. 4.6). The net surface radiation (Fig. 4.6 (a)) varied between $-10.18-3.13 \times 10^5 \text{ J/m}^2$ and air temperature varied between $-22.3-8.0 \text{ }^\circ\text{C}$. The net radiation and air temperature decreased from July 2016 to March 2017, then increased till July. The highest radiation and air temperature was observed in mid-September 2016. It was found in the earlier years of 2015 and 2016 in July (11 °C). The lowest net surface radiation was found on 10 March 2017, coinciding with the lowest air temperature.

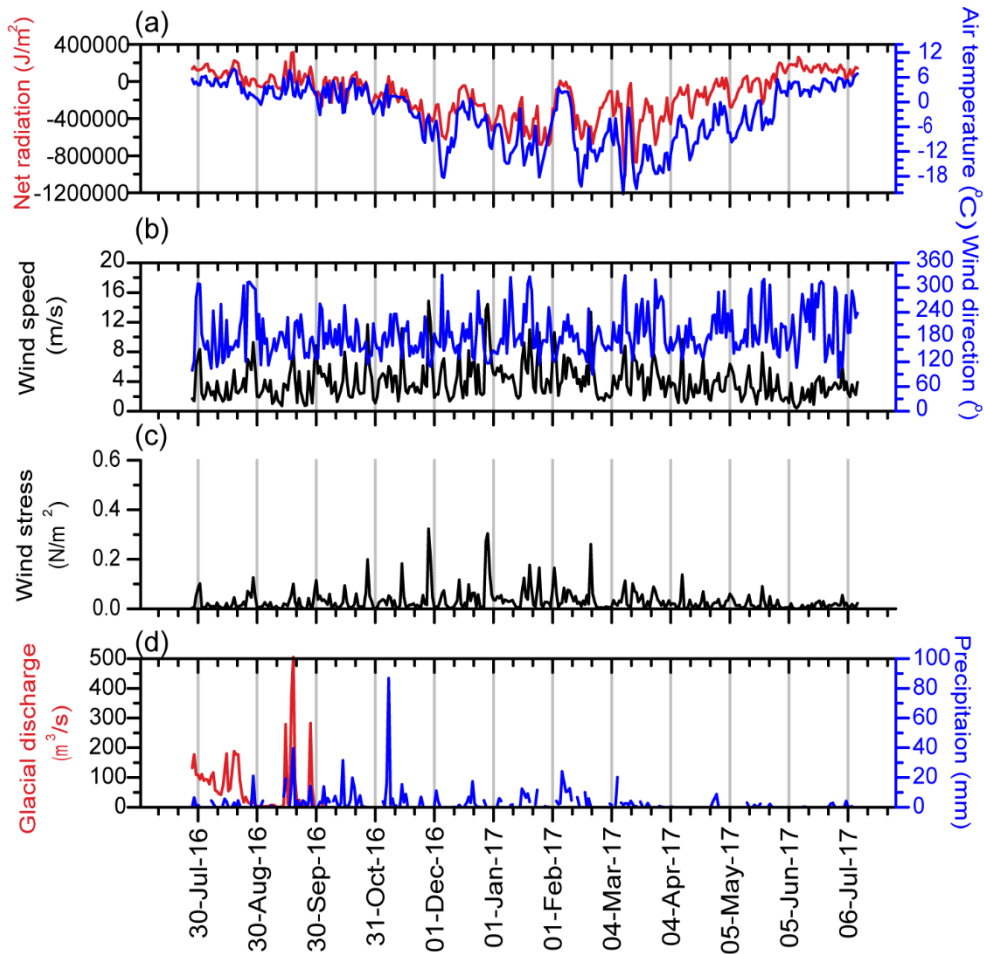


Figure 4.6: Daily meteorology data from the Ny-Ålesund weather station- (a) net surface radiation and air temperature, (b) wind speed and direction, (c) wind stress, and (d) glacial discharge and total precipitation during the mooring data period of 2016-2017.

Winds varied between 0.44-18.45 m/s (Fig. 4.6 (b)). Stronger winds were observed from October 2016 to February 2017 compared to the rest of the months. Winds higher than 10 m/s were seen by the end of October, November, and December in 2016, as well as in mid-January and early February in 2017. These periods of storm events were similar to that in 2015-2016. Wind stress (Fig. 4.6 (c)) varied between 0.0003-0.32 N/m². The highest wind speed coincided with the strongest wind stress and was found on 30th November. Winds were frequently changing between southeasterlies and southwesterlies (Fig. 4.6 (b)). Wind stress (Fig. 4.6 (c)) was higher during the cold months. Glacial discharge (Fig. 4.6 (d)) varied between 0.1-505.6 m³/s and was found to be higher in July-August 2016 than in 2015 with the peak value seen on 19

September 2016. There was no glacial discharge data for 2017. Total precipitation (Fig. 4.6 (d)) varied between 0.1-86.8 mm and was higher from September to December 2016, with the highest value on 07 November 2016.

4.2.1.3. 2017-2018

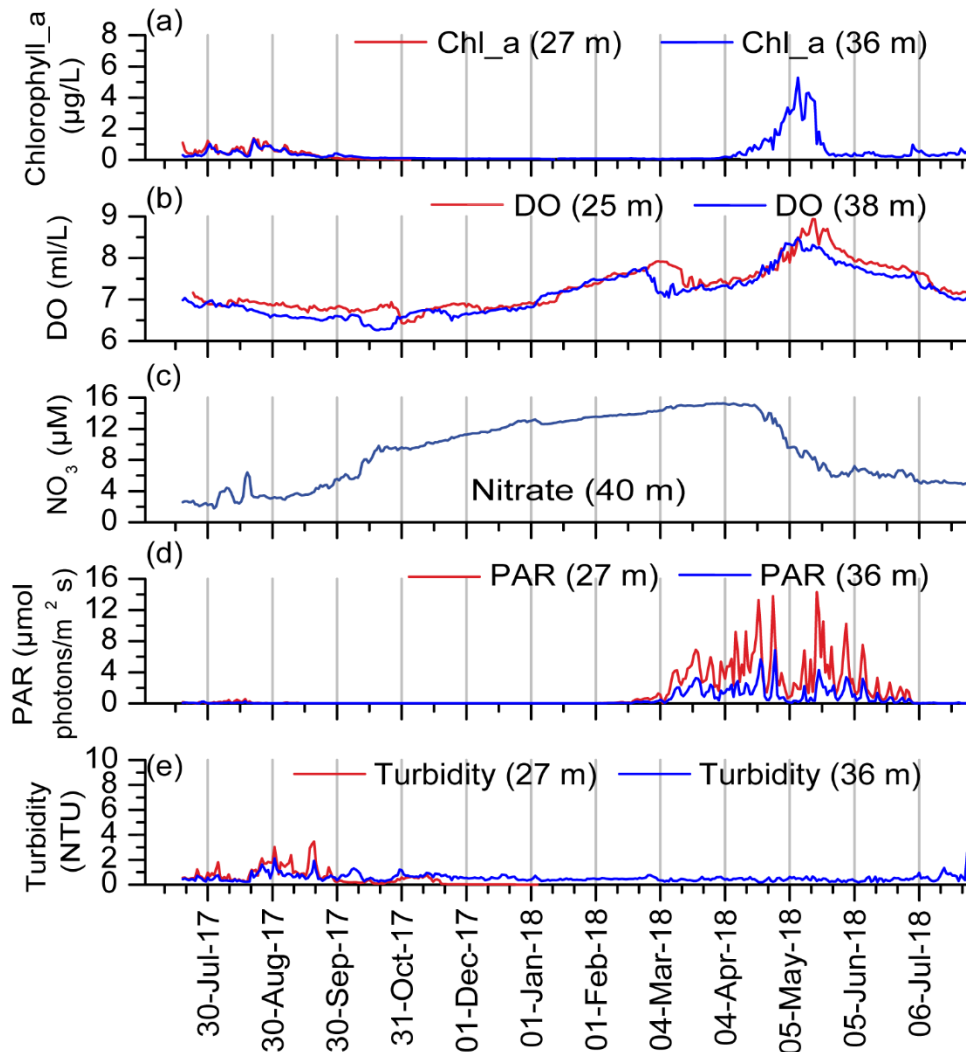


Figure 4.7: Daily (a) chlorophyll *a* (chl *a*) at 27 m and 36 m, (b) DO at 25 m and 38 m, (c) nitrate at 40 m, (d) PAR at 27 m and 36 m, and (e) turbidity at 27 m and 36 m during 2017-2018 from the mooring.

Since the battery of the chl *a* sensor at 27 m was drained by October 2017, the data was only available till that period. The chl *a* concentration at 27 m from July to October 2017 (Fig. 4.7 (a)) varied between 0.01-1.39 µg/L. The chl *a* concentration at 36 m for the full observational period varied between 0.02–5.28 µg/L. The lowest chl

a was observed from mid-November 2017 to March 2018 while the highest chl *a* was in early May 2018 as it was in 2017.

DO at 25 m was generally higher than at 38 m (Fig. 4.7 (b)). DO at 25 m varied between 6.42-8.94 ml/L while at 38 m between 6.26-8.49 ml/L. Almost a two-week lag was found in the DO at 25 m when compared to the DO variations observed at 38 m. Moreover, the DO at 25 m showed more correspondence with the variations in chl *a* at 38 m. Nitrate at 40 m (Fig. 4.7 (c)) varied between 1.80-15.28 μM . The lowest nitrate values were found in early-August 2017. The highest nitrate was in early-April 2018, a week earlier than in 2016 and 2017. PAR (Fig. 4.7 (d)) at 27 m varied between 0.04-14.31 $\mu\text{mol photon/m}^2\text{s}$ and at 36 m varied between 0.04-6.84 $\mu\text{mol photon/m}^2\text{s}$. PAR was highest on 26 April, 01 May, and then again on 20 May 2018. There was a low PAR phase between 01 and 20 May. However, the overall increase in PAR from mid-February to mid-May 2018, the productivity at 36 m picked up only from April 2018 when nitrate decreased. Turbidity (Fig. 4.7 (e)) levels were the lowest during the same period the high PAR values were seen. Turbidity at 27 m from July to mid-October 2017 varied between 0.001-3.45 NTU and was higher than that at 36 m. Turbidity at 36 m for the entire observational period varied between 0.18-2.55 NTU. Turbidity showed higher values from July-December 2017 and then in July 2018.

Water temperature during 2017-2018 (Fig. 4.8 (a)) varied between -1.68-6.42 $^{\circ}\text{C}$ at 25 m and -1.55-6.46 $^{\circ}\text{C}$ at 38 m, indicating warmer waters at the lower layer which was just the opposite of 2016-2017. This temperature inversion was found from mid-September to October, mid-end November 2017, and from end-February to mid-March in 2018. The temperature was highest by mid-August at 25 m and mid-October at 38 m. Low temperatures were found during December and January. The lowest temperature at 25 m was in early-March while it was by mid-February at 38 m. Salinity (Fig. 4.8 (b)) varied between 33.02-34.79 psu at 25 m and 33.69-34.84 psu at 38 m. Salinity was highest from mid-March to April at 25 m while it was highest in mid-July and late September at 38 m. Changes in density (Fig. 4.8 (c)) were similar to salinity variations at both depths. Sigma-t varied between 26.13-27.89 at 25 m and 26.81-28.05 at 38 m. While the highest density was found from early to mid-April 2018 at

both the depths and the lowest on 30 September 2017. However, the lowest density was in September in all three years from 2015 to 2017.

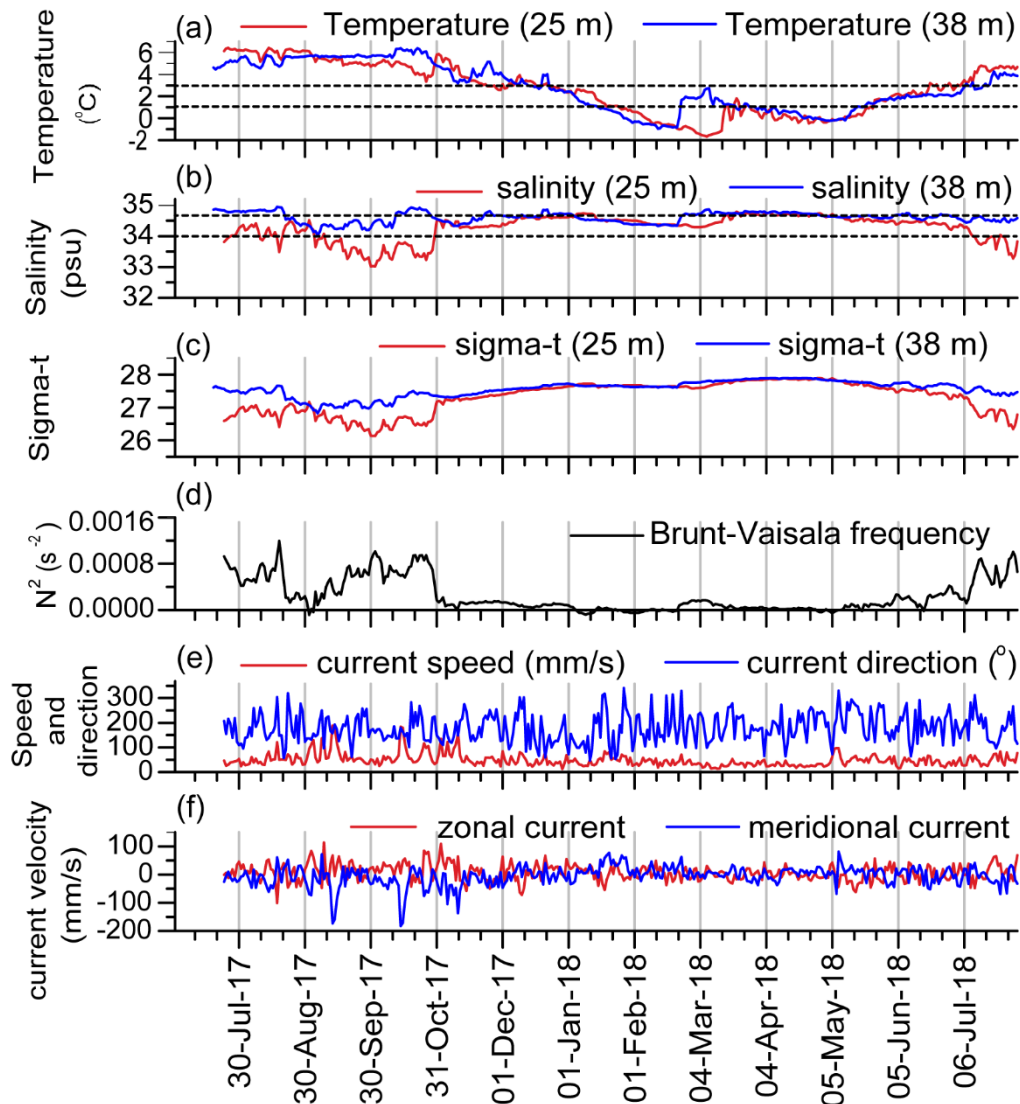


Figure 4.8: Daily water (a) temperature, (b) salinity, and (c) sigma-t at 25 m and 38 m. Daily (d) Brunt-Vaisala frequency between 25 and 38 m, (e) current speed and direction, and (f) zonal and meridional current velocity during 2017-2018 from the mooring. The two horizontal black lines in (a) demarcate 1 and 3 °C and that in (b) demarcate 34.00 and 34.65 psu.

The N^2 (Fig. 4.8 (d)) varied between $1.4\text{--}8.6 \times 10^{-4} \text{ s}^{-2}$ with the highest N^2 was found on 19 August 2016. N^2 showed an increase from July to 19 August 2016 after which it declined to zero by 02 September followed by an increase again and reduced to zero by 10 November 2017. A weak stratification was observed thereafter till early January

2018, which then showed a slight increase in mid-January. Then it increased from near zero values from mid-February but again declined from 04 March. An overall steep increase in stratification was then found from 07 May to July 2018. Stratification was relatively weak when productivity picked up in April 2018 (Fig. 4.7 (a)). However, the peak chl *a* was observed when the salinity increased. One reason for the lag observed in the enhancement of productivity at lower depths with an earlier onset of high PAR values can be the missing upper chl *a* observations where the interplay between the physics and biogeochemistry might have been more evident and present. The subsurface current speed was higher from mid-August to mid-September 2017 and then again from mid-October to mid-November (Fig. 4.8 (e)). The currents were prominently southeasterly with southwesterlies as the second dominating current direction. These two directions changed often throughout the period (Fig. 4.5 (e, f)).

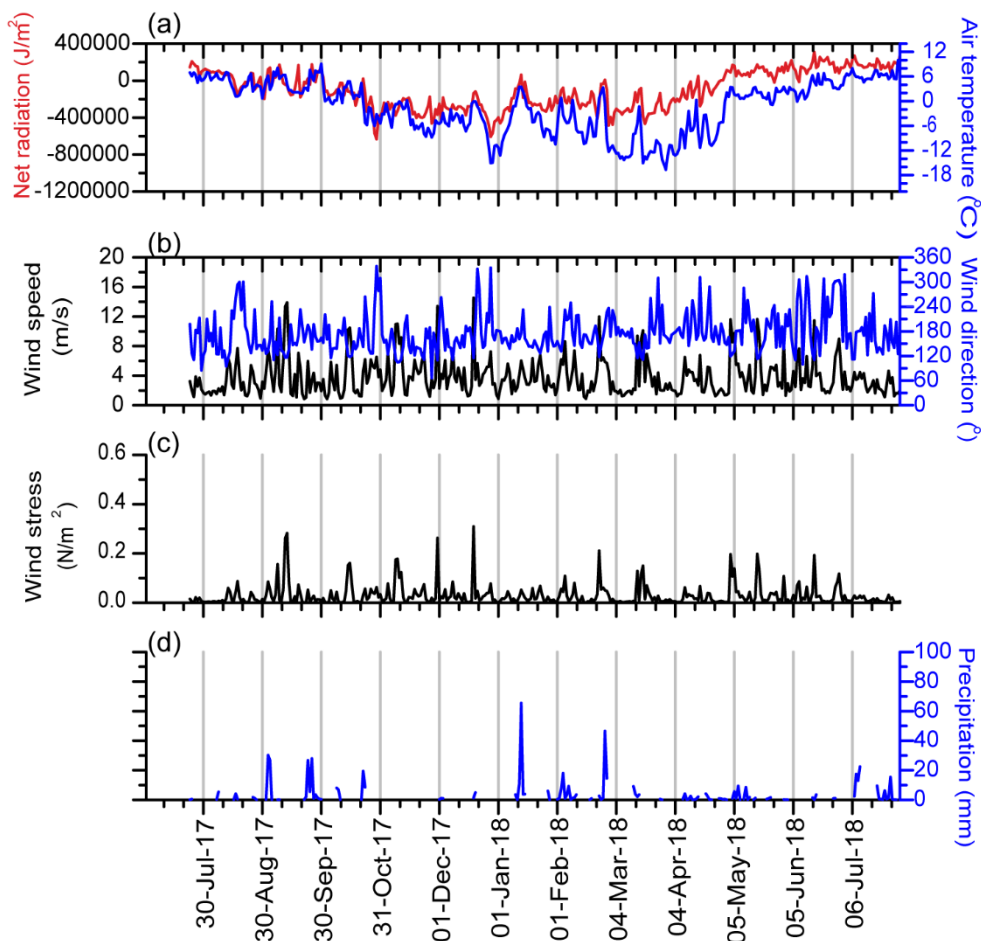


Figure 4.9: Daily meteorology data from the Ny-Ålesund weather station- (a) net surface radiation and air temperature, (b) wind speed and direction, (c) wind stress, and (d) total precipitation during the mooring data period of 2017-2018.

While the net surface radiation varied between $-6.3 - 3.1 \times 10^5 \text{ J/m}^2$, air temperature varied between $-16.7 - 10 \text{ }^\circ\text{C}$ (Fig. 4.9 (a)). The net surface radiation and air temperature decreased from July to end-December 2017 which then increased till July 2018. However, the air temperature showed a similar lowest magnitude in March 2018 as in December 2017. Even though there was a slight decrease in the net surface radiation in March, it was not as low as in December 2017. Wind speed (Fig. 4.9 (b)) varied between 0.86-14.84 m/s. Winds higher than 10 m/s were seen by mid-September, mid-October, and mid- and end-November in 2017. During 2017-2018, these stronger than 10 m/s wind events were found again in mid-February, mid-March, end-April, mid-May, and in mid-June 2018. The number of storm events observed during 2017-2018 was higher compared to the previous two years. The wind stress varied between 0.001-0.31 N/m^2 (Fig. 4.9 (c)). There was no glacial discharge data for 2017-2018. The total precipitation that varied between 0.1-65.5 mm (Fig. 4.9 (d)) was higher in September 2017 and then in January-February and July 2018.

4.2.2. Interannual variations

The observation period has been split into four seasons- summer (June - September), fall/autumn (October – November), winter (December – March), and spring (April – May) to understand the interannual variations of biogeochemistry in detail. The results of the correlation analysis have been ranked as good ($r^2 = 0.2-0.5$), moderate ($r^2 = 0.5-0.8$), and strong ($r^2 = 0.8-1.0$).

4.2.2.1. Summer

Overall, the chl *a* at 25 m and 35 m (Fig. 4.1 (a)) in summer 2015 was $< 1.0 \text{ } \mu\text{g/L}$, except on 24 July and 01 August when the concentration was higher than $2.0 \text{ } \mu\text{g/L}$ at 25 m. DO (Fig. 4.1 (b)) corresponded to the variations in chl *a*. Nitrate (Fig. 4.1 (c)) increased during the time between the two chl *a* peaks. The PAR values were also higher during the peaks (Fig. 4.1 (d)). The peaks coincided with IW from its characteristic temperature (Fig. 4.2 (a)), salinity (Fig. 4.2 (b)), and density (Fig. 4.2 (c)) ranges and also with the increase in stratification (Fig. 4.2 (d)). Moreover, low (southeasterlies) winds (Fig. 4.3 (b)) exerted low wind stress (Fig. 4.3 (c)) during the chl *a* peaks while higher southwesterly winds (5 m/s) with higher wind stress were observed between the chl *a* peak. High winds during the same time cause mixing and

nitrate supply from the lower to upper water column. From the direction of the high winds, which is up the fjord, there is a chance of more direct entry of open waters to Kongsfjorden. The strong southeasterly currents (Fig. 4.2 (e, f)) to the fjord during the time confirmed the warm water entry to the fjord. Salinity and density also were seen to increase during the peaks. The glacial discharge (Fig. 4.3 (d)) during the chl *a* peaks was also higher compared to the time in between, showing yet another source of nutrient supply through meltwater. The pattern and correspondence of chl *a* variations with the above-mentioned factors were similar throughout the summer. Another but smaller enhancement in summer chl *a* was found from 17-19 August 2015 but was associated with higher DO, stratification, and a decrease in nitrate and winds. The water mass during this peak turned out to be SW which is the result of glacial melt, as evident in the highest glacial discharge over the entire observational period (375 m³/s). The statistical analysis of summer 2015 (Table 4.2) shows a good correlation between 25 m chl *a* with PAR and DO at the same depth level, and also with salinity, density, and stratification at both depths. The correlation was higher at 25 m than at 35 m.

Early June 2016 showed a low chl *a* (Fig. 4.1 (a)) with its response in DO (Fig. 4.1 (b)) decreased by 0.5 ml/ L associated with high nitrate (by 4 µM) (Fig. 4.1 (c)) was observed during high northwesterly winds (Fig. 4.3 (b, c)) and low stratification (Fig. 4.2 (d)). As the winds subsided by the second week of June, chl *a* increased. This period of higher productivity continued till 28 June at 35 m and 05 July at 25 m. During this productive period, the highest chl *a* concentration of 6.90 µg/L and 3.95 µg/L was found at 25 m and 35 m, respectively. After this, chl *a* decreased again to ~ 0.27 µg/L by the end of the observational period in 2015-2016, with its response seen as decreased DO. Even though the depths of biogeochemical variables monitored (Table 4.1) during 2016-2017 were different from 2015-2016, the 32 m which is the upper depth during 2016-2017 was quite close to 35 m during 2015-2016. Hence, the continuity and interannual comparison were done by focusing on these nearby depths. The lower depth during 2016-2017 was 45 m. The chl *a* (Fig. 4. 4 (a)) continued decreasing with slight enhancements found around mid-August and early-September in 2016 with a corresponding increase in DO at 31 m (Fig. 4. 4 (b)). The enhancement was higher at 32 m than at 45 m.

Table 4.2: Correlation (r^2) of chl *a* with other biogeochemical variables (nominal depths of deployment in 2015-2016 are 25 m and 35 m, in 2016-2017 are 32 m and 45 m, and in 2017-2018 are 27 m and 36 m), ocean variables (nominal depths of deployment in 2015-2016 are 24 m and 34 m, in 2016-2017 are 31m and 41 m, and in 2017-2018 are 25 m and 38 m), and atmospheric variables in summer. The nitrate sensor was at 37 m in 2015, 47 m in 2016, and 40 m in 2017. Suffixes (u) and (l) after the variables represent the upper and lower depths mentioned above. The moderate and good correlations ($p < 0.0001$) are in bold red and black fonts, respectively.

Correlated variables [Unit]	2015		2016		2017	
	Chl_a (25 m)	Chl_a (35 m)	Chl_a (32 m)	Chl_a (45 m)	Chl_a (27 m)	Chl_a (36 m)
Chl_a (l) [$\mu\text{g/L}$]	0.60	---	0.35	---	0.26	---
Turbidity (u) [NTU]	0.05	0.04	0.02	0.03	0.00	0.00
Turbidity (l) [NTU]	0.12	0.06	0.04	0.00	0.00	0.00
PAR (u) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.23	0.02	0.17	0.13	0.00	0.00
PAR (l) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.19	0.01	0.03	0.01	0.00	0.00
NO_3 [μM]	0.02	0.02	0.09	0.13	0.03	0.06
DO (u) [ml/L]	0.33	0.22	0.45	0.42	0.10	0.16
DO (l) [ml/L]	0.16	0.23	0.02	0.25	No data	No data
Zonal current [mm/s]	0.04	0.00	0.04	0.00	0.02	0.07
Meridional current [mm/s]	0.04	0.00	0.02	0.05	0.00	0.01
Current speed [mm/s]	0.02	0.00	0.00	0.00	0.01	0.00
Current direction [$^\circ$]	0.01	0.00	0.06	0.00	0.00	0.01
Brunt Vaisala frequency [s^{-2}]	0.22	0.16	0.04	0.03	0.05	0.07
Sigma-t (u)	0.30	0.19	0.17	0.12	0.14	0.16
Sigma-t (l)	0.24	0.14	0.17	0.12	0.19	0.18
Salinity (u) [psu]	0.24	0.19	0.13	0.09	0.17	0.17
Salinity (l) [psu]	0.21	0.15	0.14	0.09	0.19	0.18
Temperature (u) [$^\circ\text{C}$]	0.09	0.00	0.19	0.13	0.01	0.03
Temperature (l) [$^\circ\text{C}$]	0.17	0.02	0.24	0.16	0.06	0.08
Runoff [m^3/s]	0.17	0.33	0.00	0.00	No data	No data
Precipitation [mm]	0.03	0.12	0.06	0.05	0.04	0.03
Wind speed [m/s]	0.03	0.03	0.00	0.01	0.03	0.01
Wind direction [$^\circ$]	0.002	0.00	0.00	0.00	0.00	0.01
Wind stress [Nm^{-2}]	0.01	0.00	0.01	0.02	0.02	0.02
Net surface Radiation [Jm^{-2}]	0.09	0.20	0.12	0.15	0.02	0.02
Air Temperature [$^\circ\text{C}$]	0.18	0.36	0.01	0.02	0.01	0.03

A continuous increase of nitrate from 5 μM to 12 μM was found in 10 days from 20 August that remained the same till 03 September and then decreased to 7 μM by 08 September. The period of increase in nitrate (Fig. 4.4 (c)) was found when chl *a* and DO decreased. This rapid increase in nitrate coincided with the transformation of IW to AW from 20 August till 03 September followed by a reverse modification from AW to IW during 04-08 September when the chl *a* declined.

Stratification (Fig. 4.5 (d)) increased from 13 August (Brunt-Vaisala frequency, $N^2 = 5 \times 10^{-5} \text{ s}^{-2}$) to 20 August ($N^2 = 4 \times 10^{-4} \text{ s}^{-2}$) coincided the rise in nitrate. After this N^2 declined reaching $3 \times 10^{-5} \text{ s}^{-2}$ on 30 August followed by an increase afterwards. The winds (Fig. 4.6 (b)) were frequently changing between southwest and northwest with their speed increasing from 2.2 m/s on 20 August to 9.8 m/s on 29 August during when the chl *a* decline and increase in nitrate were seen. Afterward, the wind became more stable southeasterlies and showed a decline. The glacial discharge and precipitation were higher from mid- to end-September in 2016 (Fig. 4.6 (d)). Nitrate did not show any increase during this period of high glacial discharge and this might be because the runoff effects did not impact the depth of nitrate observation. The 32 m chl *a* showed a moderate correlation (Table 4.2) with DO at the same depth. It showed a good correlation with PAR at the same depth and with temperature, salinity, and density at 32 m and 45 m. The 45 m chl *a* also showed good correlation with PAR at 32 m, nitrate at 40 m, DO and water temperature at both the depths, and also with the net surface radiation. The chl *a* showed better correlations with net surface radiation at deeper layers in the summers of 2015 and 2016.

In summer 2017, during early- and end- June chl *a* (Fig. 4.4 (a)) showed enhanced values with an average of 2.75 $\mu\text{g/L}$. While the peak was first seen at 45 m on 05 June, the peak at 32 m was found after one day on 07 June. On the contrary, the second peak was first found at 32 m on 25 June (2.00 $\mu\text{g/L}$) after which the peak at 45 m was found after 3 days on 28 June (3.50 $\mu\text{g/L}$). The rise in DO at 31 m (Fig. 4.4 (b)) was not concurrent with the first chl *a* enhancement, whereas it was during the second chl *a* peak in June. Also, DO showed a rise by 0.5 ml/L after the first peak in chl *a* declined. No DO observations were made closer to 45 m during 2016-2017 mooring period. Nitrate (Fig. 4.4 (c)) showed a decrease during both the peaks with an increase in

between the peaks. AW was present at both depths during June and July 2017 as seen in the characteristic combination of temperature (Fig. 4.5 (a)), salinity (Fig. 4.5 (b)), and density (Fig. 4.5 (c)). Stratification was higher during both the peaks (Fig. 4.5 (d)), the second stratification ($N^2 = 1 \times 10^{-4} \text{ s}^{-2}$) being double stronger than the first.

During the 2017-2018 deployment-retrieval period, the observational depths were 27 m and 36 m which were closer to the depths in 2015-2016 and 36 m is closer to the 35 m and 32 m observations made in 2015-2016 and 2016-2017, respectively. Two chl *a* peaks (Fig. 4.7 (a)) of 1 $\mu\text{g/L}$ were seen on 18 and 30 July at 27 m with a corresponding increase in DO at 25 m (Fig. 4.7 (b)). However, only the second peak was present at the lower 36 m. SW was present during the first peak and IW during the second peak at 25 m (Fig. 4.9 (a, b, c)). At 38 m, AW was found from end-July to mid-August. Stratification (Fig. 4.6 (d)) was higher ($N^2 = 9 \times 10^{-4} \text{ s}^{-2}$) during the first peak on 18 July while it declined to $4.2 \times 10^{-4} \text{ s}^{-2}$ by 30 July. However, a second chl *a* peak was observed on 30 July. Nitrate (Fig. 4.7 (c)) did not show a strong correspondence during these two peaks. On the other hand, nitrate showed a strong correspondence with a decrease in response to the increase in chl *a* during mid- and end-August, with a rise in nitrate between the peaks. Nitrate peaked at 5 μM once the mid-August chl *a* peak declined and then again to 7.5 μM after the end-August chl *a* peak declined. PAR values (Fig. 4.7 (d)) were higher during mid-August at 27 m than 36 m. During mid-August, 25 m fjord had IW while 38 m, it was AW (Fig. 4.6 (a, b, c)). During end-August, both the depths showed IW. The chl *a* peaks in July and August coincided with low winds (Fig. 4.9 (b, c)). September saw a continuous decline in chl *a* (Fig. 4.7 (a)) and DO (Fig. 4.7 (b)) with a rise in nitrate (Fig. 4.7 (c)). PAR values remained close to the detection limit in September (Fig. 4.7 (d)). An increase in SW freshening was found at 25 m in September while IW was present at 38 m (Fig. 4.9 (a, b, c)). Stratification (Fig. 4.8 (d)) showed an overall increase in September corresponding to the upper-level freshening. The increase in wind speed (Fig. 4.9 (b)) and stress (Fig. 4.9 (c)) in September coincides with the decrease in chl *a* and corresponding variations observed in other biogeochemical variables. Total precipitation of 30 mm (Fig. 4.9 (d)) was also seen during early- and end- of September which might have added to the observed freshening. The summer 2017 chl *a* showed a good correlation (Table 4.2) with DO, salinity, and density at both observation depths.

Table 4.3: Year-wise range, average, and standard deviation of biogeochemical variables in summer from the daily mooring data. The number of data points in 2015 is 73 and in 2016 and 2017 is 122.

Summer	Variables [Unit]	Range	Average	Standard deviation
2015	Chl 25 m [$\mu\text{g/L}$]	0.22 – 2.28	0.64	0.43
	Chl 35 m [$\mu\text{g/L}$]	0.18 – 0.93	0.36	0.18
	Turb 25 m [NTU]	0.19 – 2.29	0.55	0.38
	Turb 35 m [NTU]	0.20 – 1.61	0.65	0.31
	NO_3 37 m [μM]	1.01 – 5.11	2.53	1.03
	PAR 25 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 1.03	0.10	0.13
	PAR 35 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.29	0.05	0.03
	DO 24 m [ml/L]	6.87 – 8.01	7.29	0.26
	DO 34 m [ml/L]	6.83 – 7.68	7.30	0.21
2016	Chl 32 m [$\mu\text{g/L}$]	0.11 – 6.90	1.03	1.24
	Chl 45 m [$\mu\text{g/L}$]	0.07 – 3.95	0.51	0.66
	Turb 32 m [NTU]	0.16 – 5.17	0.79	0.78
	Turb 45 m [NTU]	0.08 – 9.19	0.79	1.37
	NO_3 47 m [μM]	2.85–12.21	5.62	2.17
	PAR 32 m [$\mu\text{mol photons/m}^2 \text{ s}$]	No data		
	PAR 45 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 1.30	0.25	0.28
	DO 31 m [ml/L]	6.46 – 8.46	7.21	0.57
	DO 41 m [ml/L]	No data		
2017	Chl 27 m [$\mu\text{g/L}$]	0.12 – 2.46	0.65	0.39
	Chl 36 m [$\mu\text{g/L}$]	0.17 – 3.14	0.58	0.43
	Turb 27 m [NTU]	0.22 – 3.45	0.85	0.65
	Turb 36 m [NTU]	0.10 – 2.11	0.50	0.38
	NO_3 40 m [μM]	1.80–15.05	6.68	4.62
	PAR 27 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.72	0.10	0.11
	PAR 36 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 3.20	0.15	0.33
	DO 25 m [ml/L]	6.69 – 7.88	7.09	0.30
	DO 38 m [ml/L]	6.49 – 7.04	6.72	0.16

In summer 2018, chl *a* (Fig. 4.7 (a)) showed a slight increase in early-July which was also seen in DO and as a decrease in nitrate. A monthly average of 6 μM nitrate (Fig. 4.7 (c)) was present in June 2018 (0.4 $\mu\text{g/L}$ chl *a*) which was lower compared to 13 μM in 2017 (1.0 $\mu\text{g/L}$ chl *a*), and higher than the 5 μM in 2016 (1.8 $\mu\text{g/L}$ chl *a*). June

chl *a* in 2018 is lower than these two years, even though the PAR values were also higher in June 2018 (Fig. 4.7 (d)) compared to the other observational years at both depths. IW ((Fig. 4.8 (a, b, c)) was present at both depths in June whereas it was IW in early-July when the chl *a* increased. Later at the upper 25 m depth, IW was modified to SW while IW continued at 38 m. Stratification (Fig. 4.8 (d)) was higher during the peak in early-July followed by a decrease. From 07 July, stratification has increased considerably due to freshening, as evident in the lower salinity (Fig. 4.8 (b)) when the winds (Fig. 4.9 (b)) decreased. Precipitation (Fig. 4.8 (d)) of 10-20 mm was also seen during the time which added to the freshening.

Thus, the range and average of summer chl *a* (Table 4.3) is higher in the warm years of 2016 and 2017 compared to the cold year of 2015 (Chapter 3). Also, higher turbidity and nitrate values were seen in the warm years (Table 4.3). The subsurface nitrate in 2016 and 2017 was more than 2-3 times, respectively, compared to 2015. The deeper depths of nitrate observations in 2016 and 2017 can be one reason for higher nitrate values, as the nitrate concentration increases with depth. However, the nitrate between the depths 37 m in 2015 and 40 m in 2017 which are closer also had a three time fold difference. High turbidity in summer which is seen during the declining phase in chl *a* indicates input from the dying productivity and also indicates more sediment-laden meltwater in the fjord that affects the subsurface depths. The meltwater is a source of nutrients for the fjord (Svendsen et al., 2002). Also, the years 2016 and 2017 proved warmer and saline due to higher AW intrusion than in 2015 (chapter 3). The AW is another known source of nutrient supply to Kongsfjorden (Hegseth et al., 2019). Thus, high nitrate levels in these warm years can be a combination of these two sources. The higher chl *a* in 2016 and 2017 is one reason for the higher turbidity observed afterward in the same years. DO (Table 4.3) showed a decrease in warm years 2016 and 2017 when compared to cold 2015, supporting the classic DO-water temperature relation of less oxygen in warmer waters (Hanelt et al., 2004).

4.2.2.2. Fall/Autumn

Early October 2015 showed a decrease in chl *a*, (Fig. 4.1 (a)) also evident as a decrease in DO (Fig. 4.1 (b)) coincided with a nitrate peak of 8.6 μM (Fig. 4.1 (c)).

Table 4.4: Correlation (r^2) of chl *a* with other biogeochemical variables (nominal depths of deployment in 2015-2016 are 25 m and 35 m, in 2016-2017 are 32 m and 45 m, and in 2017-2018 are 27 m and 36 m), ocean variables (nominal depths of deployment in 2015-2016 are 24 m and 34 m, in 2016-2017 are 31m and 41 m, and in 2017-2018 are 25 m and 38 m), and atmospheric variables in fall. The nitrate sensor was at 37 m in 2015, 47m in 2016, and 40 m in 2017. Suffixes (u) and (l) after the variables represent the upper and lower depths. The strongest, moderate, and good correlations ($p < 0.0001$) are in bold blue, red, and black fonts, respectively.

Correlated variables [Unit]	2015		2016		2017	
	Chl_a (25 m)	Chl_a (35 m)	Chl_a (32 m)	Chl_a (45 m)	Chl_a (27 m)	Chl_a (36 m)
Chl_a (l) [$\mu\text{g/L}$]	0.90	---	0.39	---	0.63	---
Turbidity (u) [NTU]	0.05	0.08	0.01	0.03	0.11	0.00
Turbidity (l) [NTU]	0.11	0.15	0.00	0.28	0.01	0.06
PAR (u) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.43	0.39	0.00	---	0.00	0.00
PAR (l) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.34	0.29	0.00	---	0.00	0.00
NO_3 [μM]	0.79	0.69	0.54	0.30	0.60	0.84
DO (u) [ml/L]	0.42	0.45	0.09	0.04	0.14	0.00
DO (l) [ml/L]	0.40	0.46	No data	No data	0.03	0.01
Zonal current [mm/s]	0.02	0.02	0.00	0.00	0.00	0.03
Meridional current [mm/s]	0.01	0.01	0.01	0.00	0.02	0.00
Current speed [mm/s]	0.04	0.05	0.01	0.02	0.00	0.01
Current direction [$^\circ$]	0.04	0.04	0.01	0.00	0.01	0.04
Brunt Vaisala frequency [s^{-2}]	0.06	0.01	0.03	0.01	0.28	0.52
Sigma-t (u)	0.72	0.68	0.12	0.04	0.51	0.61
Sigma-t (l)	0.71	0.74	0.12	0.06	0.50	0.46
Salinity (u) [psu]	0.67	0.65	0.01	0.00	0.54	0.51
Salinity (l) [psu]	0.62	0.67	0.01	0.01	0.50	0.23
Temperature (u) [$^\circ\text{C}$]	0.01	0.04	0.32	0.16	0.04	0.29
Temperature (l) [$^\circ\text{C}$]	0.06	0.10	0.23	0.07	0.01	0.23
Runoff [m^3/s]	0.31	0.38	No data	No data	No data	No data
Precipitation [mm]	0.00	0.03	0.01	0.01	0.00	0.00
Wind speed [m/s]	0.10	0.10	0.03	0.01	0.01	0.06
Wind direction [$^\circ$]	0.00	0.02	0.01	0.01	0.01	0.01
Wind stress [Nm^{-2}]	0.08	0.07	0.04	0.01	0.00	0.04
Net surface Radiation [Jm^{-2}]	0.28	0.32	0.24	0.12	0.35	0.34
Air Temperature [$^\circ\text{C}$]	0.28	0.29	0.24	0.08	0.29	0.28

Chl *a* concentration then rose to $\sim 0.8 \mu\text{g/L}$ by 03 October which then continued decreasing throughout the season reaching $0.07 \mu\text{g/L}$. This decrease was reflected in the DO which decreased from $\sim 8 \text{ ml/L}$ to 7 ml/L . During the same time, nitrate rapidly increased reaching values high up to $11.5 \mu\text{M}$, but was embedded with high intraseasonal variabilities. A very strong negative correlation was found between chl *a* and nitrate at 25 m and 35 m depths (Table 4.4), with the correlation slightly higher at 25 m. Even with these high nitrates, the fall chl *a* was low which is due to the limitation of light (Fig. 4.1 (d)). The characteristic rapid decrease in temperature (4.8°C to 1.4°C at 24 m; 5.0°C to 1.7°C at 34 m) (Fig. 4. 2 (a)) and increase in salinity (32.97 to 34.52 psu at 24 m; 33.24 to 34.83 psu at 34 m) (Fig. 4. 2 (b)), and density (26.32 to 27.46 kg m^{-3} at 24 m; 26.49 to 27.57 kg m^{-3} at 34 m) (Fig. 4. 2 (c)) were observed. Hence, chl *a* showed a strong negative correlation with salinity and density (Table 4.4). Meanwhile, nitrate showed a strong positive correlation with salinity ($r^2 = 0.66\text{-}0.76$; $p < 0.0001$) and density ($r^2 = 0.77\text{-}0.83$; $p < 0.0001$), showing the relation between dense water masses and increase in nitrate. The correlation was higher at the upper depth of 25 m (Table 4.4). No significant correlation was found with temperature.

The chl *a* values was close to the detection limit throughout fall 2016 (Fig. 4.4 (a)) whereas DO (Fig. 4.4 (b)) showed 6.5 ml/L with very minimal variations. Nitrate showed a rapid rise with intraseasonal oscillations in it (Fig. 4.4 (c)) as was in fall 2015. The almost zero PAR (Fig. 4.4 (d)) during the period describes the lack of detectable chl *a*. A strong negative correlation was found between chl *a* and nitrate which was higher at 32 m (Table 4.4). The temperature decreased from 5.8°C to 4.0°C at 31 and 41 m, showing an almost homogenous water column (Fig. 4. 5 (a)). The rate of decrease in temperature (0.03°C/day) was just half of that in 2015. Salinity did not show any considerable increase. Instead, it frequently varied between 34.00-34.65 psu (Fig. 4. 5 (b)), and density (Fig. 4. 5 (c)) also was found with strong intraseasonal oscillations in October but by the end of November, the density became steady. Overall, the sigma-t increased from 27.00 to 27.35. These relations were found to be the weak correlation of chl *a* with salinity and density (Table 4.4). However, temperature showed a good correlation which was higher at the upper 32 m. On the other hand, nitrate showed a good positive correlation with density ($r^2 = 0.39$; $p <$

0.0001) at both depths and a negative correlation with the 31 m temperature ($r^2 = 0.27$; $p < 0.0001$) showing a thermally driven water mass modification as in fall 2015. The correlation was less significant with the temperature at 41 m.

Table 4.5: Year-wise range, average, and standard deviation of biogeochemical variables in the fall season from the daily mooring data. Number of data points is 61.

Fall	Variables [Unit]	Range	Average	SD
2015	Chl 25 m [$\mu\text{g/L}$]	0.07 – 0.82	0.25	0.22
	Chl 35 m [$\mu\text{g/L}$]	0.07 – 0.77	0.21	0.19
	Turb 25 m [NTU]	0.20 – 1.07	0.42	0.19
	Turb 35 m [NTU]	0.28 – 1.19	0.56	0.23
	NO_3 37 m [μM]	4.31–11.50	8.02	2.05
	PAR 25 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.08	0.04	0.01
	PAR 35 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.05	0.04	0.002
	DO 24 m [ml/L]	6.64 – 7.31	6.97	0.19
	DO 34 m [ml/L]	6.47 – 7.26	6.85	0.22
2016	Chl 32 m [$\mu\text{g/L}$]	0.09 – 0.20	0.12	0.02
	Chl 45 m [$\mu\text{g/L}$]	0.06 – 0.23	0.08	0.03
	Turb 32 m [NTU]	1.00 – 2.88	1.75	0.42
	Turb 45 m [NTU]	0.10 – 1.75	0.25	0.22
	NO_3 47 m [μM]	6.90–10.65	8.97	0.80
	PAR 32 m [$\mu\text{mol photons/m}^2 \text{ s}$]	No data		
	PAR 45 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.06 – 0.06	0.06	0.00
	DO 31 m [ml/L]	6.40 – 6.73	6.51	0.06
	DO 41 m [ml/L]	No data		
2017	Chl 27 m [$\mu\text{g/L}$]	0.01 – 0.10	0.04	0.02
	Chl 36 m [$\mu\text{g/L}$]	0.08 – 0.35	0.13	0.06
	Turb 27 m [NTU]	0.03 – 0.81	0.28	0.21
	Turb 36 m [NTU]	0.26 – 1.31	0.67	0.25
	NO_3 40 m [μM]	5.38–11.24	9.02	1.74
	PAR 27 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.04	0.04	0.00
	PAR 36 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.04	0.04	0.00
	DO 25 m [ml/L]	6.42 – 6.93	6.75	0.13
	DO 38 m [ml/L]	6.26 – 6.76	6.56	0.16

Fall in 2017 showed a similar scenario of almost no detectable chl *a* (Fig. 4.7 (a)) as in 2016. DO showed variations with time and with depth (Fig. 4.7 (b)) wherein a

decrease from 6.65 to 6.25 ml/L was found during mid-October in DO at 38 m coincided with a rapid increase in 40 m nitrate from 6 to 10 μM (Fig. 4.7 (c)). However, the relation was absent at 25 m. DO increase with nitrate in November. The intraseasonality was lower in 2017 than in 2015 and 2016. A very strong negative correlation was found between chl *a* and nitrate (Table 4.4), and the correlation was higher at 36 m, which might be due to the close-by depths of observations. PAR was low as in the previous years (Fig. 4.7 (d)). A rise in temperature by 0.6 °C (Fig. 4.8 (a)) was seen at the same time the decrease in DO was observed at 38 m. The temperature difference (Fig. 4.8 (a)), salinity (Fig. 4.8 (b)), and density (Fig. 4.8 (c)) decreased by the end of October, and in November the water column density was almost the same at both the depths. A strong negative correlation of chl *a* was found with salinity and density as was found in 2015 (Table 4.4). Nitrate showed a moderate positive correlation with salinity ($r^2 = 0.24\text{--}0.51$; $p < 0.0001$) and density ($r^2 = 0.54\text{--}0.66$; $p < 0.0001$) the correlation being stronger at the upper depth of 25 m. Nitrate showed a moderate negative correlation with temperature also ($r^2 = 0.34\text{--}0.44$; $p < 0.0001$) in fall 2017. The chl *a* concentration (Table 4.5) was highest in fall 2015 while almost the same concentration was found around 30 m depth in 2016 and 2017. DO (Table 4.5) in the cold year 2015 was higher compared to the warm years 2016 and 2017 while nitrate concentration increased from 2015 to 2017. When PAR was below the detectable level in all three years, turbidity was higher in 2016 and 2017 (Table 4.5).

4.2.2.3. Winter

Very low production (0.04–0.07 $\mu\text{g/L}$) was found throughout winter 2015/16 (Fig. 4.1 (a)) whereas DO (Fig. 4.1 (b)) and nitrate (Fig. 4.1 (c)) has shown an overall increase. DO levels, even though low ($\sim 6.6\text{--}7.4$ mL/L), have shown an overall increase and had strong intraseasonal variability from December 2015 to January 2016 (Fig. 4.1 (b)). In the absence of productivity, this increase in DO levels in winter can be due to the intrusion of the cold Arctic waters from the Spitsbergen Polar current on the shelf region entering Kongsfjorden when the density matches. This understanding is supported by the strong southeast currents in winter 2015 (Fig. 4.2 (e, f)) bringing waters to the fjord. Nitrate (Fig. 4.1 (c)) increased from 9.97 to 12.62 μM and showed a strong negative correlation with chl *a* ($r^2 = 0.77$ to 0.78 ; $p < 0.001$).

Table 4.6: Correlation (r^2) of chl *a* with other biogeochemical variables (nominal depths of deployment in 2015-2016 are 25 m and 35 m, in 2016-2017 are 32 m and 45 m, and in 2017-2018 are 27 m and 36 m), ocean variables (nominal depths of deployment in 2015-2016 are 24 m and 34 m, in 2016-2017 are 31m and 41 m, and in 2017-2018 are 25 m and 38 m), and atmospheric variables in winter. The nitrate sensor was at 37 m in 2015, 47 m in 2016, and 40 m in 2017. Suffixes (u) and (l) of variables represent the upper and lower depths. The strongest, moderate, and good correlations (p -value < 0.0001) are in bold blue, red, and black fonts, respectively.

Correlated variables [Unit]	2015/16		2016/17		2017/18
	Chl_a (25 m)	Chl_a (35 m)	Chl_a (32 m)	Chl_a (45 m)	Chl_a (36 m)
Chl_a (l) [$\mu\text{g/L}$]	0.84	---	0.39	---	---
Turbidity (u) [NTU]	0.47	0.54	0.75	0.35	0.32
Turbidity (l) [NTU]	0.55	0.53	0.02	0.04	0.02
PAR (u) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.12	0.10	0.04	0.00	0.00
PAR (l) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.12	0.11	0.04	0.00	0.01
NO_3 [μM]	0.78	0.77	0.75	0.29	0.00
DO (u) [ml/L]	0.00	0.01	0.12	0.01	0.00
DO (l) [ml/L]	0.05	0.01	No data	0.04	0.07
Zonal current [mm/s]	0.01	0.04	0.11	0.00	0.04
Meridional current [mm/s]	0.01	0.03	0.01	0.00	0.01
Current speed [mm/s]	0.05	0.03	0.07	0.00	0.03
Current direction [$^\circ$]	0.03	0.07	0.14	0.08	0.09
Brunt Vaisala frequency [s^{-2}]	0.33	0.28	0.59	0.17	0.02
Sigma-t (u)	0.77	0.80	0.78	0.27	0.00
Sigma-t (l)	0.75	0.79	0.75	0.26	0.00
Salinity (u) [psu]	0.67	0.75	0.84	0.36	0.01
Salinity (l) [psu]	0.58	0.71	0.69	0.38	0.01
Temperature (u) [$^\circ\text{C}$]	0.05	0.09	0.16	0.01	0.00
Temperature (l) [$^\circ\text{C}$]	0.00	0.01	0.38	0.05	0.01
Runoff [m^3/s]	0.01	No data	No data	No data	No data
Precipitation [mm]	0.00	0.01	0.02	0.02	0.04
Wind speed [m/s]	0.00	0.00	0.00	0.04	0.01
Wind direction [$^\circ$]	0.02	0.03	0.00	0.00	0.01
Wind stress [Nm^{-2}]	0.00	0.01	0.00	0.04	0.00
Net surface Radiation [Jm^{-2}]	0.01	0.00	0.04	0.00	0.00
Air Temperature [$^\circ\text{C}$]	0.01	0.00	0.07	0.00	0.00

PAR (Fig. 4.1 (d)) was off the detection limit due to the general absence of light. However, in the winter of 2015/16, the PAR started increasing from 19 February 2016. The low winter water temperature varied between -0.3 and 4.3 °C, with the upper depth of 24 m depth being colder (Fig. 4.2 (a)). There was an overall decrease in water temperature and an increase in salinity and density from December to March. Salinity increased from 34.03 to 34.88 psu while density increased from 27.31 to 27.81 kg m⁻³. However, the water column was homogenous with almost no depth-wise variations in salinity and density. Stratification became zero from the second week of January and remained the same till the end of the season, indicating mixing between the depths (Pond and Pickard, 1983). Chl *a* showed a good positive correlation with turbidity and stratification while it showed a strong correlation with salinity and density, the correlations being stronger in the 24 m (Table 4.6).

During winter 2016/17, the chl *a* concentration varied between 0.04–0.10 µg/L (Fig. 4.4 (a)). DO (Fig. 4.4 (b)) has shown a 0.5 ml/L increase over the season with higher intraseasonal variations in December. Nitrate (Fig. 4.4 (c)) has shown a steady rise from ~ 11.00 to 15.00 µM. Nitrate showed a strong negative correlation with chl *a* at 32 m and a good correlation at 45 m too (Table 4.6). PAR (Fig. 4.4 (d)) was close to the detection limit till end-February 2017 after which it started rising. Turbidity during winter 2016/17 (Fig. 4.4 (e)) was higher than in 2015/16 with an average value of 1.03 NTU at 32 m which is almost double that in the previous winter. The winter water temperature decreased from 4.0 in December to the lowest temperature of 1.0 °C in March (Fig. 4.5 (a)). Salinity (Fig. 4.5 (b)) increased from 34.25 to 34.95 psu while density (Fig. 4.5 (c)) increased from 27.35 to 27.95 kg m⁻³. The water column became homogenous from January to March 2017, and hence, there was zero stratification (Fig. 4.5 (d)) at the same time. Chl *a* showed a strong positive correlation with turbidity at 32 m, a good correlation with turbidity at 45 m, and a good to moderate correlation with stratification (Table 4.6). A strong correlation was found between salinity and density at 31 m, while a good correlation was found between the same variables at 41 m and temperature at 31 m.

Table 4.7: Year-wise range, average, and standard deviation of biogeochemical variables in winter from the daily mooring data. Number of data points is 121.

Winter	Variables [Unit]	Range	Average	SD
2015/16	Chl 25 m [$\mu\text{g/L}$]	0.05 – 0.07	0.06	0.005
	Chl 35 m [$\mu\text{g/L}$]	0.04 – 0.07	0.05	0.008
	Turb 25 m [NTU]	0.26 – 1.09	0.52	0.17
	Turb 35 m [NTU]	0.34 – 1.30	0.60	0.20
	NO_3 37 m [μM]	9.47 – 12.70	11.42	0.86
	PAR 25 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 1.02	0.14	0.11
	PAR 35 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 0.79	0.22	0.15
	DO 24 m [ml/L]	6.62 – 7.46	7.17	7.13
	DO 34 m [ml/L]	6.60 – 7.44	7.14	7.14
2016/17	Chl 32 m [$\mu\text{g/L}$]	0.04 – 0.10	0.06	0.02
	Chl 45 m [$\mu\text{g/L}$]	0.04 – 0.06	0.05	0.003
	Turb 32 m [NTU]	0.43 – 2.18	1.03	0.37
	Turb 45 m [NTU]	0.17 – 0.83	0.37	0.12
	NO_3 47 m [μM]	10.38– 15.32	13.45	1.27
	PAR 32 m [$\mu\text{mol photons/m}^2 \text{ s}$]	No data		
	PAR 45 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.06 – 0.81	0.12	0.14
	DO 31 m [ml/L]	6.30 – 6.97	6.76	0.17
	DO 41 m [ml/L]	No data		
2017/18	Chl 27 m [$\mu\text{g/L}$]	No data		
	Chl 36 m [$\mu\text{g/L}$]	0.02 – 0.10	0.08	0.01
	Turb 27 m [NTU]	No data		
	Turb 36 m [NTU]	0.26 – 0.80	0.46	0.11
	NO_3 40 m [μM]	11.24– 15.28	13.43	1.11
	PAR 27 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 6.89	0.90	1.66
	PAR 36 m [$\mu\text{mol photons/m}^2 \text{ s}$]	0.04 – 3.24	0.39	0.74
	DO 25 m [ml/L]	6.69 – 7.92	7.28	0.37
	DO 38 m [ml/L]	6.64 – 7.76	7.17	0.32

In winter 2017/18, the chl *a* concentration varied between 0.02–0.10 $\mu\text{g/L}$ at 36 m (Fig. 4.7 (a)). DO (Fig. 4.7 (b)) at 25 m and 38 m showed a rise from December (6.7 ml/L) to mid-February (7.7 ml/L) after which the 38 m DO decreased to 7.3 ml/L by early March while 25 m DO decreased to 7.4 ml/L by end of March. Intraseasonal oscillations were found throughout the season. Nitrate increased from 12.00 to 15.5 μM (Fig. 4.7 (c)). In winter 2017, nitrate showed no significant correlation with any

other oceanic or atmospheric variables (Table 4.6). PAR (Fig. 4.7 (d)) was close to the detection limit till early-February 2018 after which it increased rapidly compared to both the earlier winters. Turbidity during winter 2017/18 around 30 m was lower when compared to the previous winters (Fig. 4.7 (e)). The winter water temperature decreased from 3.0 in December to -0.85 °C in mid-February 2018 (Fig. 4.5 (a)) followed by a temperature inversion with 38 m temperature rising back to 3 °C by early March and decreasing further to 1 °C by the end of March. This temperature inversion is the reason for the decrease in 36 m DO observed during the same time. The temperature at 25 m reached a minimum of -1.85 °C by early-March followed by a rise till the end of March reaching 1°C. Salinity (Fig. 4.5 (b)) and density (Fig. 4.5 (c)) increased, but the rate of increase was lesser when compared to the other two years. Stratification (Fig. 4.5 (d)) showed a slight increase in December and during the time of temperature inversion. Once the inversion ended, the water column went back to a homogenous condition. No significant correlation was found between chl *a* and any other biogeochemical, oceanic, and atmospheric variables (Table 4.6).

With the generally lowest seasonal values of chl *a* values in winter, the chl *a* around 30 m between the years showed a maximum value of 0.07 µg/L in 2015/16 and 0.10 µg/L in 2016/17 and 2017/18 (Table 4.7). DO values were lowest in winter 2016/17 while it was comparable in 2015/16 and 2017/18 (Table 4.7). Average nitrate was also lowest for 2015/16 with comparable values in the other two winters.

4.2.2.4. Spring

Chl *a* (Fig. 4.1 (a)) started rising on 02 April 2016 at 35 m and 03 April 2016 at 25 m with the rate of increase being faster at 35 m than at 25 m. From 22 April, chl *a* values became higher than 0.5 µg/L and increased rapidly at a faster rate at 25 m than 35 m which marked the start of the spring bloom phenomenon. Chl *a* of 6.89 µg/L at 25 m and 4.28 µg/L at 35 m were observed till end-April before it started declining around the second week of May. Concurrent with this chl *a* drop, high winds of 5–10 m/s were found around 08–14 May (Fig. 4.3 (b)). Once the wind speed decreased around mid-May chl *a* increased, indicating a revival of the bloom with distinctively high peaks in bloom on the 18th (7.19 µg/L) and 23rd May (6.59 µg/L) at 25 m followed by 20 (4.22 µg/L) and 22 May (4.98 µg/L).

Table 4.8: Correlation (r^2) of chl *a* with other biogeochemical variables (nominal depths of deployment in 2015-2016 are 25 m and 35 m, in 2016-2017 are 32 m and 45 m, and in 2017-2018 are 27 m and 36 m), ocean variables (nominal depths of deployment in 2015-2016 are 24 m and 34 m, in 2016-2017 are 31m and 41 m, and in 2017-2018 are 25 m and 38 m), and atmospheric variables in spring. The nitrate sensor was at 37 m in 2015, 47m in 2016, and 40 m in 2017. Suffixes (u) and (l) represent the upper and lower nominal depths of the variables. The strongest, moderate, and good correlations (p-value < 0.0001) are in bold blue, red, and black fonts, respectively.

Correlated variables [Unit]	2016		2017		2018
	Chl_a (25 m)	Chl_a (35 m)	Chl_a (32 m)	Chl_a (45 m)	Chl_a (36 m)
Chl_a (l) [$\mu\text{g/L}$]	0.33	---	0.96	---	---
Turbidity (u) [NTU]	0.05	0.00	0.03	0.03	0.11
Turbidity (l) [NTU]	0.01	0.01	0.07	0.07	0.11
PAR (u) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.19	0.17	0.56	0.55	0.14
PAR (l) [$\mu\text{mol photons/m}^2 \text{ s}$]	0.25	0.12	0.56	0.54	0.14
NO_3 [μM]	0.41	0.41	0.63	0.61	0.11
DO (u) [ml/L]	0.49	0.80	0.73	0.69	0.31
DO (l) [ml/L]	0.44	0.73	No data	No data	0.60
Zonal current [mm/s]	0.11	0.04	0.01	0.02	0.12
Meridional current [mm/s]	0.01	0.02	0.00	0.00	0.12
Current speed [mm/s]	0.05	0.00	0.17	0.19	0.16
Current direction [$^\circ$]	0.09	0.06	0.02	0.02	0.13
Brunt Vaisala frequency [s^{-2}]	0.03	0.30	0.00	0.00	0.02
Sigma-t (u)	0.06	0.24	0.14	0.12	0.00
Sigma-t (l)	0.06	0.19	0.16	0.14	0.00
Salinity (u) [psu]	0.26	0.09	0.02	0.02	0.04
Salinity (l) [psu]	0.28	0.16	0.06	0.06	0.07
Temperature (u) [$^\circ\text{C}$]	0.28	0.31	0.16	0.14	0.17
Temperature (l) [$^\circ\text{C}$]	0.22	0.27	0.17	0.16	0.48
Runoff [m^3/s]	0.01	0.02	No data	No data	No data
Precipitation [mm]	0.00	0.06	0.00	0.00	0.03
Wind speed [m/s]	0.00	0.01	0.01	0.02	0.05
Wind direction [$^\circ$]	0.00	0.02	0.03	0.02	0.01
Wind stress [Nm^{-2}]	0.00	0.00	0.00	0.01	0.06
Net surface Radiation [Jm^{-2}]	0.38	0.48	0.03	0.01	0.27
Air Temperature [$^\circ\text{C}$]	0.41	0.44	0.03	0.01	0.27

A second break in chl *a* was found from 25 May to 10 June again coinciding with high winds of 5–8.5 m/s (Fig. 4.3 (b)). The chl *a* is strongly and positively correlated with DO. Maximum values of DO (8.3–8.8 ml/L) were at the same time as chl *a* peaks in May. The chl *a* showed a good correlation with subsurface PAR, more significantly at the upper layer (Table 4.8). The PAR which was up to 8.85 and 3.59 $\mu\text{mol photons/m}^2\text{s}$, respectively, at 25 and 35 m in April then declined as a response to the chl *a* increase by late-April and May. Nitrate (Fig. 4.1 (c)) increased and reached its annual peak of 13.3 μM from 17 to 20 April 2016. Afterward, it decreased when chl *a* started to rise. With the highest rate of decrease in early-May, nitrate reached 4.4 μM by end-May. This relation of nitrate with chl *a* can be seen as a moderate negative correlation (Table 4.8).

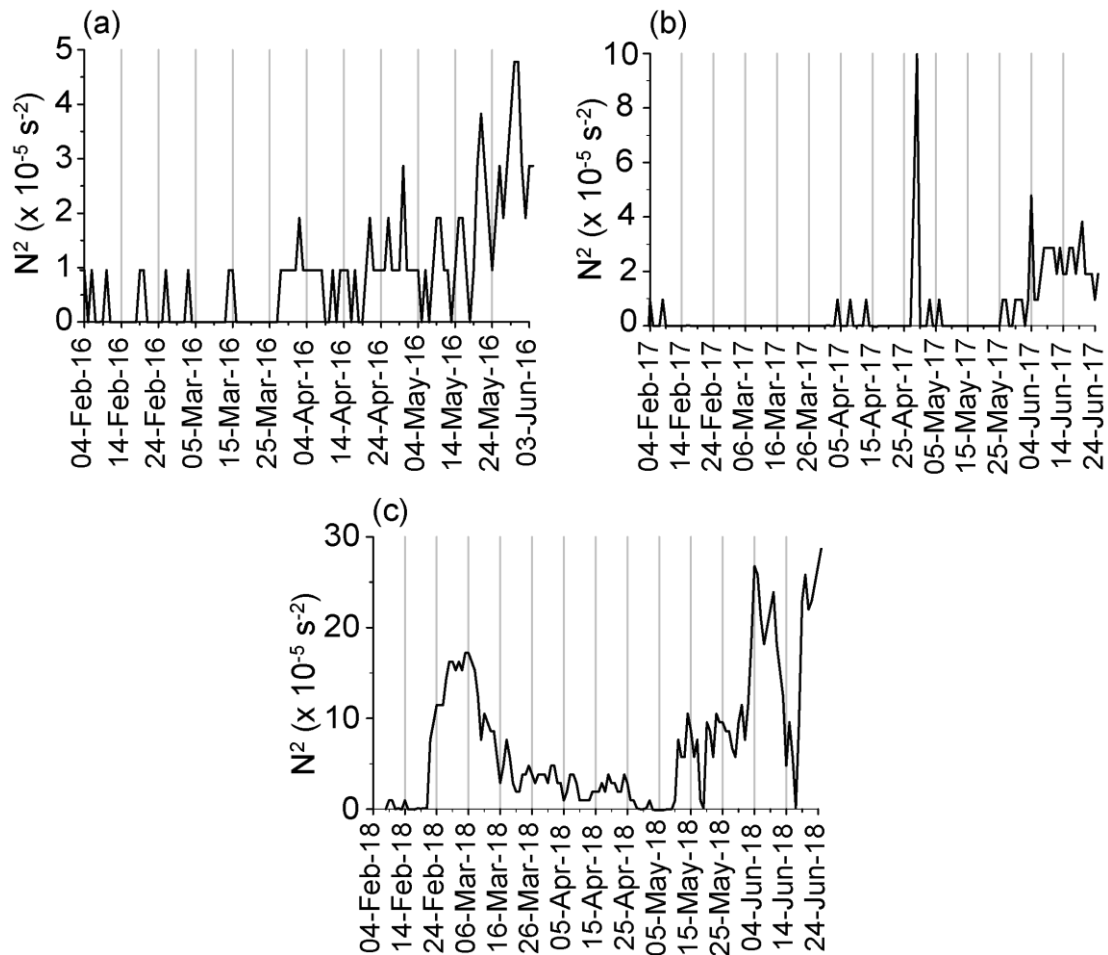


Figure 4.10: Daily Brunt-Vaisala frequency enlarged for the period from February to June in (a) 2016, (b) 2017, and 2018 from the mooring data.

Also, nitrate was negatively correlated to DO ($r = 0.59$ to 0.76 ; $p < 0.001$), with the highest correlation at the lower depth of 35 m. After winter, the water temperature (Fig. 4.2 (a)) remained low in April which then increased from 01 May and reached 3.3°C by end-May. However, salinity (Fig. 4.2 (b)) and density (Fig. 4.2 (c)) didn't change significantly in spring. Stratification (Fig. 4.10 (a)) increased gradually with a few breaks in between during high winds, which were linked to the spring bloom breaks as discussed above. Once the winds calmed down, stratification was reestablished.

In spring 2017, chl *a* (Fig. 4.4 (a)) started rising from April and peaked on 13 April. By 15 May, the bloom declined rapidly from 7.19 to $0.2\ \mu\text{g/L}$ on 22 May. This was concurrent with a $5\ \text{m/s}$ wind event (Fig. 4.6 (b)) on 15 May. Once the wind speed declined, the chl *a* began to rise again till end-May. PAR (Fig. 4.4 (b)) showed two peaks in early-mid April and late May coinciding with the lows of the spring bloom. DO (Fig. 4.4 (b)) in response to chl *a* increased steadily reaching the peak coinciding with the chl *a* peak followed by a decline and then again a rise from 23 May. Nitrate (Fig. 4.4 (c)) showed a mirror image of DO decreasing from April till 15 May followed by an increase till end-May. This was proved with a strong negative correlation of nitrate with the 36 m DO ($r^2 = 0.96$; $p < 0.001$). The correlation analysis (Table 4.8) also showed a moderate correlation of chl *a* with subsurface PAR, DO, and nitrate. Over the spring of 2017, the water temperature (Fig. 4.5 (a)) increased steadily from 1°C to 3°C whereas the salinity (Fig. 4.5 (b)) was $35.00\ \text{psu}$ with a sigma- t (Fig. 4.5 (c)) < 27.92 which are characteristics of transformed Atlantic water mass (Cottier et al., 2005). Even though stratification (Fig. 4.10 (b)) increased from 02 April, there were discontinuities including the time of the chl *a* peak on 13 April.

During spring 2018, 36 m chl *a* (Fig. 4.7 (a)) started rising from April and peaked by 09 May ($5.28\ \mu\text{g/L}$) which then decreased and reached a minimum of $0.11\ \mu\text{g/L}$ by 22 May after which it continued with the same concentration till end-May. DO reflected the temporal variations in chl *a*. The DO (Fig. 4.7 (b)) peak at 38 m on 09 May reflected the first and highest peak in chl *a* while the peak in 25 m DO by mid-May reflected the second peak in chl *a* around the same time. PAR (Fig. 4.7 (b)) showed two peaks in early-mid April and late May coinciding with the lows of the spring

bloom. Nitrate (Fig. 4.7 (c)) declined rapidly in spring till mid-May after which it continued at more or less the same value (6 μ M) till end-May.

Table 4.9: Year-wise range, average, and standard deviation of biogeochemical variables in spring from the daily mooring data. Number of data points is 61.

Spring	Variables [Unit]	Range	Average	SD
2016	Chl 25 m [μ g/L]	0.05 – 7.19	1.82	1.81
	Chl 35 m [μ g/L]	0.05 – 4.99	1.29	1.21
	Turb 25 m [NTU]	0.15 – 0.84	0.30	0.10
	Turb 35 m [NTU]	0.20 – 2.73	0.46	0.39
	NO ₃ 37 m [μ M]	3.90 – 13.30	10.46	2.60
	PAR 25 m [μ mol photons/m ² s]	0.04 – 8.85	1.13	2.14
	PAR 35 m [μ mol photons/m ² s]	0.04 – 3.59	0.66	0.86
	DO 24 m [ml/L]	7.19 – 8.85	7.74	0.46
	DO 34 m [ml/L]	7.16 – 8.29	7.58	0.35
2017	Chl 32 m [μ g/L]	0.05 – 7.68	2.15	2.32
	Chl 45 m [μ g/L]	0.05 – 6.69	1.99	2.15
	Turb 32 m [NTU]	0.33 – 0.71	0.45	0.07
	Turb 45 m [NTU]	0.10 – 0.34	0.18	0.05
	NO ₃ 47 m [μ M]	9.59 – 15.46	13.15	1.91
	PAR 32 m [μ mol photons/m ² s]	No data		
	PAR 45 m [μ mol photons/m ² s]	0.06 – 3.65	1.27	1.19
	DO 31 m [ml/L]	6.90 – 8.07	7.28	0.31
	DO 41 m [ml/L]	No data		
2018	Chl 27 m [μ g/L]	No data		
	Chl 36 m [μ g/L]	0.11 – 5.28	1.46	1.40
	Turb 27 m [NTU]	No data		
	Turb 36 m [NTU]	0.20 – 0.67	0.42	0.12
	NO ₃ 40 m [μ M]	5.86 – 15.28	11.25	3.50
	PAR 27 m [μ mol photons/m ² s]	0.67 – 14.31	4.80	3.30
	PAR 36 m [μ mol photons/m ² s]	0.09 – 6.84	1.56	1.37
	DO 25 m [ml/L]	7.37 – 8.94	8.00	0.47
	DO 38 m [ml/L]	7.27 – 8.49	7.85	0.38

Nitrate showed a strong negative correlation with DO ($r^2 = 0.61$ - 0.77 ; $p < 0.0001$), the correlation was higher at the upper 25 m. However, the correlation between nitrate and

chl *a* was not statistically significant. Water temperature (Fig. 4. 8 (a)) decreased slightly in April after which it increased steadily in May from -1.0 to 2 °C. Salinity (Fig. 4. 8 (b)) was close to 34.65 psu in April and then decreased to 34.4 psu by end-May. Sigma-t (Fig. 4. 8 (c)) was 27.8 in April, which was then reduced to 27.5 by the end of May. Stratification (Fig. 4. 10 (c)) which was already high and decreased gradually in April. Then a sharp increase in stratification was seen on 09 May, which coincided with the peak in chl *a*, which increased further till end-May.

Few continuous time series measurements of biogeochemical variables are available in Kongsfjorden (van De Poll et al., 2016; D'Angelo et al., 2018; Singh et al., 2019). Among these, Indian mooring data from 2015-2016 has been used to address seasonality (Singh et al. 2019). The present study including two more years of data of subsurface chl *a*, DO, nitrate, PAR, turbidity, temperature, salinity, and current observations, not only brings out the seasonality of biogeochemical variables but also highlights interannual variation in the fjord biogeochemistry and phytoplankton bloom from multiple years under different environmental conditions.

Seasonality in Kongsfjorden physics and biogeochemistry has been increased by glacier melting and Atlantic water intrusion (Hop and Wiencke, 2019). The present study reinforces the existence of four discrete seasons each year. Winter is the dark and cold season with high nutrients but almost no production. The following season which is spring on the other hand is the most productive season due to the onset and rapid rise in sunlight. Summer with maximum sunlight, melting and surface depletion of nutrients, and fall with diminishing sunlight and melting are the other two seasons. In winter, the upper water column nutrient (nitrate) replenishes due to the absence of detectable photosynthesis under no sunlight conditions. Mayot et al. (2017) reported that the nutrient redistribution in winter caused by convective mixing impacts the interannual variability of the phytoplankton bloom phenology. In spring, the onset and increase of solar radiation and subsequent thermal stratification of these nutrient-rich fjord waters lead to spring bloom (Winder and Cloern, 2010). Spring bloom is important at high latitudes as the seasonal cycle of productivity starts during this highly productive season (Sakshaug, 2004; Hodal et al., 2012). The bloom contributes hugely to the annual primary production and also plays an important part in supporting

secondary production. Hence the bloom phenology and interannual variability were investigated over the observational years.

4.3. Interannual variability of spring and summer blooms

4.3.1. Spring bloom

Singh et al. (2019) studied the phytoplankton bloom development in Kongsfjorden using one-year data during 2015-2016. In 2016, the productive season was found for a period of two and half months from the last week of April to the first week of July (Table 4.10). In 2017, the season was one and half months long from mid-April to end-May while in 2018 it was one month and a week-long from mid-April to 22 May. The 2016 spring bloom was the most elongated bloom that extended to early summer compared to the other two years. However, there were two major troughs in the 2016 spring bloom compared to the continuous spring bloom that existed in 2017 and 2018 which was discussed by Singh et al. (2019). The breaks in these three bloom periods from end-April to early-May, mid-May to early-June, and the second week of June to early July coincided with high winds disturbing stratification leading to a decrease in phytoplankton growth. Besides weakening the stratification, these strong upfjord winds can shift phytoplankton biomass towards the inner fjord (Meire et al., 2016). High winds control spring bloom timing in polar regions as resultant mixing can delay the bloom onset (Fitch and Moore, 2007; Collins et al., 2009; Vikebø et al., 2019). A nitrate replenishment due to non-utilization was observed during both bloom breaks. As the winds calmed down, the stratification was established again leading to the bloom.

At high latitudes, the bloom dynamics are controlled mainly by the interaction between nutrients, light, and mixing (Yool et al., 2015). As an indicator of phytoplankton biomass, chl *a* with other biogeochemical variables pointed out different stages of bloom development. The correlation of DO with chl *a* was strong in spring and moderate in summer. Nitrate indicated a general nutrient depletion in spring and summer with a winter replenishment. The subsurface PAR was depth-dependently responsive to chl *a* and turbidity. According to density and Brunt-Vaisala frequency from temperature and salinity, well-mixed conditions existed in winter and an

increased stratification from spring to summer. Thus, the physical and biogeochemical conditions characterize different seasons in the fjord.

Table 4.10: Year-wise start, end, and peak dates and peak chl *a* values of spring and secondary summer blooms demarcated with higher than 0.5 µg/L chl *a*.

Year	Bloom period	Peak time	Peak chl <i>a</i> value (µg/L)	Secondary bloom start	Secondary bloom end	Peak time	Peak chl <i>a</i> value (µg/L)
2015				1. 20 July 2. 17 August 3. 30 August 4. 06 October	1. 06 August 2. 19 August 3. 30 September 4. 14 October	30 July	2.28
2016	22 April- 06 July	18 May	7.19	1. 9 July 2. 14 August 3. 06 September	1. --- 2. 21 August 3. 08 September	20 July	2.80
2017	15 April- 29 May	13 May	7.68	1. 01 June 2. 17 June 3. 24 July 4. 11 August	1. 10 June 2. 19 July 3. 03 August 4. 08 September	07 June	2.46
2018	15 April- 22 May	09 May	5.28	1. 10 June 2. 03 July 3. 26 July 4. 12 August 5. 22 August 6. 12 September	1. 14 June 2. 08 July 3. 29 July 4. 14 August 5. 04 September 6. 20 September	27 August	1.45

As discussed in the winter season above, the near-zero N^2 values in winter indicate a well-mixed subsurface water column in all three winters. Convective mixing can occur as deep as the bottom, resulting in maximum MLD and nutrients spread throughout the water column (Sverdrup, 1953). Nitrate was at its maximum in early spring in the three years. In 2016, PAR values increased from the end of February followed by a rise in water temperature from April with a steep increase from the last week of April. According to Somavilla et al. (2017), the differential heating-up of surface waters in spring leads to a thermal gradient that inhibits mixing, and results in MLD shoaling. The shoaling continues in summer till a maximum temperature is reached (Brown et

al., 2015). The summer CTD observations showed < 25 m MLD, with a maximum of 50 m MLD in 2014 summer (Table 3.2, chapter 3). A 30-40 m is reported to be the prominent MLD in Kongsfjorden by Hop et al. (2002).

Singh et al. (2019) saw the euphotic zone considered as 1% of surface PAR (Hop et al., 2002) was deeper including the 35 m depth from 19 February to 10 March 2016 during the early period of light return. The euphotic zone for 2017 and 2018 also included the deeper levels of the present observations at 47 m and 40 m, respectively. During early-spring, surface water remains relatively transparent before the beginning of glacial runoff and phytoplankton growth (Oziel et al., 2019). However, not much photosynthetic activity was observed until 02 April in 2016 spite of up to $0.53 \mu\text{mol photons/m}^2 \text{ s}$ PAR at 25 m and 35 m and up to $13.2 \mu\text{M}$ nitrate at 37 m.

Further investigation showed that the stratification of the water column in 2016 has picked up from late-March to early-April. Stratification in the inner Kongsfjorden was found to be due to solar insolation, air-sea heat exchange, and ice-melt and runoff (Meire et al., 2017). Stratification, at the same time, supports phytoplankton suspension in the euphotic zone and limits the entrainment of new nutrients across the density gradient (Kulk et al., 2018), imposing counteracting effects on phytoplankton. On the onset of an almost continuous increase in stratification, the phytoplankton biomass increased from 03 April 2016 emulating a bloom scenario by 22 April once chl *a* was higher than $0.5 \mu\text{g/L}$. Unfavorable mixing conditions in the fjord delay the phytoplankton bloom. Phytoplankton fails to stay in the euphotic layer when the water column is unstable or neutral with continued mixing, which then negatively affects the growth (Chiswell et al., 2015). With warmer waters at 35 m, the initial rate of chl *a* increase was higher at 35 m. When the 35 m temperature was ~ 0.4 °C warmer than that at 25 m by 21 April, the chl *a* at 35 m was $\sim 0.6 \mu\text{g/L}$ higher than that at 25 m. This reinstated temperature-induced difference in the phytoplankton growth (Coello-Camba et al., 2014; Coello-Camba and Agustí, 2017) in the absence of light and nutrient limitation before the onset of bloom, and may need further investigation to conclude such depth-wise variation. During the first half of the spring bloom, from 22 April to end-May including the peak bloom, the water mass was TAW. However, an

overall decrease in the spring bloom happens during the period when TAW is modified to AW.

In 2017, PAR at both depths started increasing from 24 February during which the nitrate levels were high. However, chl *a* again showed a delay in its rise only from April onwards culminating in spring bloom by 15 April. As seen in 2016, the zero water column stratification during the period between the increase in PAR and chl *a* explained the delay in the rise in chl *a*. Even though stratification increased from 02 April 2017, discontinuity of stratification was found in between, especially during the peak on 13 May 2017, thus failing to explain the spring bloom peak. However, the current direction at the mooring site during the spring bloom period in 2017 was towards the northwest, indicating a possible advection of chlorophyll from the southern shore which is where the core of AW inflow happens in Kongsfjorden whereas the currents. The currents were towards the southeast, towards the inner fjord, during the 2016 spring. The water temperature during the period between the rise in PAR and chl *a* was almost stable ($\sim 1^{\circ}\text{C}$) with higher than 34.65 showing the presence of TAW. When the chl *a* started to rise in April water temperature was also increased from 1°C and by the end of the bloom, the temperature crossed 3°C resulting in the modification to AW, and water column stratification also started a steep rise. Thus, TAW, based on the characteristics of temperature, salinity, and depth values supported by the current direction, was present at the depths during the 2017 spring bloom.

Again, in 2018, PAR increased from 05 February coinciding with the seasonal high nitrate in the water column. As in the previous years, a lag was found in the chl *a* to rise which it did from 31 March. The overall stratification was higher during the bloom period, which strengthened further after the bloom. The water mass found during the bloom period was ArW, Arctic, and LW local water masses with salinity close to 34.65 psu ($T = -1.5$ to 1.0°C ; $S = 34.30$ to 34.80 psu; Cottier et al., 2005). This is supported by the dominant northwest current direction found during the period. A decrease in DO at a lower depth was seen, which corresponded with a rise in temperature and salinity at the lower depth at the same time, from the end of February to mid-March. Stratification increased during the same period due to this TAW intrusion.

Thus, one to two months delay was found between the rise in PAR and chl *a* during the observation which was mainly driven by the water column stratification. However, the northeast currents during the neutrally stratified water conditions in spring 2017 hint towards a possible advection of chl *a* from the southern Kongsfjorden. In 2016 and 2017, when the bloom evolved and peaked under the TAW scenario, the magnitude of the bloom was higher (7.19-7.68 $\mu\text{g/L}$) compared to 2018 when the bloom formation happened under the colder-less saline ArW (5.28 $\mu\text{g/L}$) scenario. Earlier studies since 2006 reported March to May as the spring bloom time in Kongsfjorden, nearby Arctic fjords, and ocean, with its peak in April (Hodal et al., 2012; Hegseth and Tverberg, 2013; Piquet et al., 2014). However, there was no detectable subsurface phytoplankton in March during 2016, 2017, and 2018. The bloom occurred late in these three years [late- (2016) and mid-April (2017, 2018)] compared to the earlier studies. This can also be because the subsurface bloom became detectable only by this time in the middle fjord where the mooring is. It is to be noted at this point that the bloom timing and location in a fjord are a product of stratification and glacial melting, driven by increased air temperature and warm Atlantic water intrusion (van De Poll et al., 2016). The timing and pattern of the 2016, 2017, and 2018 blooms are similar to the 2007 bloom at 20 m reported in Kongsfjorden as predicted for an ice-free fjord with normal winter mixing (Hegseth and Tverberg, 2013). The mean fast ice coverage from February to June in 2016, 2017, and 2018 were $\sim 20\%$, $\sim 40\%$, and $\sim 30\%$, respectively (Fig. 3.12 (h)), chapter 3). The three subsurface blooms were comparable to the surface blooms reported earlier (Hegseth et al., 2019) with peak chl *a* of 7.2 $\mu\text{g/L}$ and 4.8 $\mu\text{g/L}$ at 25 m and 35 m, respectively.

4.3.2. Secondary bloom

The blooms that happen after a short transition after the spring bloom are called secondary blooms (Juul-pedersen et al., 2015). Even though the starting time of the secondary bloom in 2015 cannot be conveyed from the observations during 2015-2016, four secondary bloom periods with $> 0.5\mu\text{g/L}$ chl *a* were delineated from July to October (Table 4.10). This indicated the secondary productivity period extended till mid-October after which chl *a* reduced to detection limits by early November. In 2016, chl *a* reduced in late June to early July indicating a short transition period from the spring bloom which was then followed by an increase in chl *a* by mid-July which is a

secondary bloom (Table 4.10). Since this measurement was paused due to the warranted yearly retrieval and the follow-up mooring deployment, the ending of this secondary bloom in 2016 was missed and thereby is unknown. Hence the blank in the secondary bloom ending time in Table 4.10. However, two more secondary blooms were detected in August and September. The chl *a* dropped to detection limits by mid-September. 2017 showed four secondary blooms from June to early September and the chl *a* reached the detection limit by early October. 2018 showed six, the highest number of secondary blooms from June to September compared to the other three summers.

The secondary blooms in 2015 occurred when the fjord was dominated by SW and IW water masses, while in 2016 it was dominated by AW except for mid-August when it was IW. In 2017, the blooms from June to July were when the fjord had AW. But in late July bloom had SW and AW while the August blooms had IW and AW. September bloom had SW and IW when the bloom started diminishing. The June blooms in 2018 occurred when the fjord had IW while the blooms from July to August happened when SW and IW were present. September blooms in 2018 occurred when the fjord had IW and TAW. The secondary bloom periods varied between 2-31 days. The production was detectable till early fall, as can be seen in 2015 and 2017. The peak value of chl *a* in the secondary bloom was the lowest in summer 2018 (Table 4.10) which was 0.6-1.9 times lower than the peaks found in the other three earlier secondary bloom seasons.

A decrease in nitrate is the principal factor in ending phytoplankton bloom (Kulk et al., 2018). Hegseth et al. (2019) reported a lack of secondary blooms in Kongsfjorden. However, they showed secondary bloom intensity comparable to spring bloom in 2006, 2007, 2010, and 2012. Higher temperatures in summer led to a decrease in DO. This temperature rise enhanced stratification thereby preventing nutrient renewal on the surface. As grazing and mortality intensified, the correspondence between chl *a* and DO was reduced in summer. Also, glacial runoff mainly induces the highest turbidity, which reduces the euphotic layer depth and limits primary production (Calleja et al., 2017). In summer, the euphotic zone depth can be <3 m near glaciers in Kongsfjorden (Keck et al., 1999). Zooplankton grazing, nutrient depletion, and light

limitation due to turbidity are found to be major factors responsible for the summer phytoplankton decrease (Winder and Cloern, 2010).

4.4. Role of Atlantic water in nitrate variations

Atlantic water inflow, local-dense water formation, glacier meltwater, and sub-glacial discharge are the major nutrient sources in Kongsfjorden (Angelo et al., 2018; Hopwood et al., 2018; van de Poll et al., 2018). In spring and summer, the surface and sub-surface depths of the stratified water column become nutrient-limited. Later, winter convective mixing introduces the nutrients again from the bottom to the upper depths. This association of nutrient-rich waters with dense-saline winter water masses was evident by the positive correlations of nitrate with density ($r^2 = 0.94 - 0.98$, $p < 0.0001$) and salinity ($r^2 = 0.74 - 0.83$). Moreover, the significant role of convective mixing towards nitrate variability was reflected in its negative correlation with the stratification. The water column started destabilizing in the fall.

During fall 2015 and winter 2015/16, nitrate at 37 m showed five prominent peaks around 02 and 28 October, 06 and 24 November, and 19 December, which coincided with Atlantic water inflows identified from its characteristic temperature ($>3^\circ\text{C}$), salinity (> 34.65 psu), and density ($< 27.92 \text{ kg m}^{-3}$) (Cottier et al., 2005), especially at the closer depth of nitrate observations which is at 35 m. Earlier studies (Hegseth and Tverberg, 2013; Piquet et al., 2014) documented interruptions in winter water cooling in Kongsfjorden by warm Atlantic water inflows. The Atlantic water is known to have more nitrate than the Fjord water (Hegseth et al., 2019). Hence, the intrusions that coincided with the nitrate peaks around the same periods are due to the nitrate supply by the Atlantic water. On the other hand, summer even after having maximum Atlantic intrusions in Kongsfjorden (Cottier et al., 2005), shows no correlation with subsurface nitrate due to complexity from the utilization of nutrients by phytoplankton and grazing. No more episodic Atlantic inflows were observed at the subsurface for the rest of the 2015/16 winter. Again, in the fall and winter of 2016/17, three nitrate enhancement events were observed at 47 m in early- and mid-October and early- and end- December 2016. When the first two just touched the salinity characteristic at 41 m, it is expected to go higher than 34.65 at 47 m, the peaks in December showed pure Atlantic water intrusion. From January 2017 onwards, both depths showed a complete

shift to AW, coinciding with the ongoing overall nitrate increase. Again, in the 2017 fall and 2017/18 winter, a nitrate peak at 40 m around 22 October coincided with the AW characteristics at 38 m. Later, the second and a smaller nitrate enhancement was seen coinciding again with the TAW which is again from the Atlantic water mass intrusion. Out of the three falls and winters during the observational period, the peaks in nitrate were more prominent in 2015 and 2015/16 and the peaks are found to be prominent when it is pure AW when compared to TAW which again reinforces that the more the AW, the more nitrate it brings to Kongsfjorden. Thus, the Atlantic water also supplied nutrients in winter, besides convective mixing.

4.5. Conclusion

Based on the subsurface chl *a*, DO, nitrate, PAR, turbidity, temperature, salinity, and current observations for four years from 2015, this study not only brings out the annual, seasonal, intraseasonal, and interannual variabilities of Kongsfjorden biogeochemistry but also throws light on the evolution of spring and secondary blooms in the fjord subsurface for multiple and distinctive years. Besides, it emphasizes the coupling between the fjord biogeochemistry and regional oceanic and atmospheric factors.

Even with high nutrient availability, winter and fall showed the least productivity because of the light limitation. Nutrients (nitrate) increased rapidly at the highest annual rate during fall. It went on increasing in winter under weak stratification and deep convective mixing, but at a rate 2-3 times lesser than that in fall. During these least productive seasons, episodes of Atlantic water were also seen increasing the nitrate concentration further from the ambient condition. With the cooling of waters, DO increased gradually from fall through winter, confirming the classic relation of more oxygen dissolving in colder waters. There were significant intraseasonal oscillations in DO and nitrate which was higher in the fall through half of the winter season which coincided with the warm water intrusions. This shows the role of Atlantic water in imbibing intraseasonal oscillation in the fjord biogeochemistry. The densest winter waters were seen after this period in the three winters observed.

Moving to spring, the fjord gradually becomes highly productive with an increase in PAR and stratification coinciding with a decline in nitrate due to utilization. The highest annual rate of decline in the nutrients was seen in the spring season. Contrary to the earlier studies from 2006 that reported March to May as the period of spring bloom in Kongsfjorden, nearby Arctic fjords, and the ocean with its peak in April, the present study showed a late occurrence from late April in 2016 and mid-April in 2017 and 2018. Also, the peaks in the bloom were found by mid-May in 2016 and 2017, and in early May in 2018. The subsurface production occurred when the water column stratified after a delay of one to two months after the rise in the subsurface PAR. The 2016 spring bloom showed two major breaks while the 2017 and 2018 blooms showed a continuous pattern with smaller ups and lows. However, the breaks observed during the bloom were induced by high local winds of greater than 5 m/s, which induces mixing and thereby weak stratification. The mixing allowed for the comeback of the bloom after the winds reduced, reestablishing the stratification. Both 2016 and 2018 blooms confirmed the mechanism. However, the 2017 spring was different in the absence of stratification during the bloom. The bloom in this spring was found to be mostly advective, as indicated by the strong northeast currents that transport chlorophyll from the southern shore of Kongsfjorden to the mooring location in central Kongsfjorden. The southern Kongsfjorden is more productive and is the core region of nutrient-rich Atlantic water. This possibility is further confirmed by the TAW present during spring 2017, which confirmed the presence of Atlantic water during the bloom. The bloom magnitudes were higher in the warm years of 2016 and 2017, when the bloom evolved and peaked under the TAW scenario compared to 2018 when the formation occurred under the colder-less saline ArW-LW scenario. Moreover, the timing and pattern of the blooms in the present study years are similar to the 2007 bloom as predicted for an ice-free Kongsfjorden with normal winter mixing. Also, these three subsurface blooms which were comparable to the surface blooms reported earlier indicated the increasing subsurface chlorophyll and hence the need to incorporate them in modelling studies.

On the contrary to winter, summer is the period of maximum sunlight but is left with nutrient-depleted surface waters due to its maximum utilization, which happened in the spring before. Also, the strongest stratification of the water column during summer

inhibits further mixing of nutrients from the deeper waters to the upper layers. Hence, with the leftover nutrients and those supplied to the fjord by the meltwater which would be at its maximum on the fjord surface during the season and the Atlantic water, summer showed secondary blooms whose magnitude is generally lower than the spring blooms. During the present observational years, the maximum summer chl *a* was less than half of the maximum chl *a* observed during the spring bloom at the subsurface. However, these secondary blooms are not continuous phenomena like most spring blooms. Instead, they exhibit as episodic events of enhancement in chl *a*. This might be due to the source of nutrients described above being episodic events rather than a continuous supply. The secondary bloom period in the summer of 2015 and 2017 extended further to the successive fall season. As seen in the spring bloom, the summer productivity was also higher in 2016 and 2017, dominated by the Atlantic water masses, and lower in years dominated by surface and intermediate water masses. This further shows the increasing role of the Atlantic water in the fjord nutrient supply and productivity and its interannual variability.

With the increase in Atlantic water intrusions, continuous monitoring of the fjord biogeochemistry and physics is needed to understand the rapid changes and also it is necessary to incorporate such observations into the modelling efforts.

Chapter 5

Role of climate modes in the observed variabilities in Kongsfjorden

5.1. Background

Chapters 3 and 4 have shown physical and biogeochemical variabilities at different timescales in Kongsfjorden. The wavelet analysis of the variables (section 3.5, chapter 3) exhibited statistically significant and high energy levels at periods higher than seasonal and annual timescales. This indicates the response of the fjord to forcings of similar high periodicity, which led to the present investigation on the influence of two major climate modes in the region, which are the North Atlantic Oscillation (NAO) and AO.

The NAO is a wintertime weather pattern formed over the North Atlantic region, with a north-south pressure difference between 'the Azores High' around 20°N to 40°N and 'Icelandic Low' around 50°N to 70°N (Barnsten and Livezey, 1987; Hurrell et al., 2003). This pressure difference regulates the strength of the westerly winds across the Atlantic. A high-pressure difference leads to a positive NAO phase and vice versa. AO, also known as the Northern Annular Mode (Thompson and Wallace, 1998) is a fluctuating shift in the atmospheric pressure between the Arctic and mid-latitudes. This climate pattern has counter clockwise winds circulating ~ 55°N latitude. The AO phase is determined from the north-south location of the midlatitude jet stream that steers storms (Dahlman, 2009). During the positive phase of AO, high pressure was observed over the northern Pacific and Atlantic Oceans and low pressure over the Arctic. The negative phase of AO will have the opposite pressure system. Even though NAO and AO are often found correlated, a clear distinction exists between them and their influences (Thompson and Wallace, 1998). As the NAO pattern arises when the pressure anomaly is specifically centred around the Atlantic mid-latitudes, it has a greater influence on the wind-driven surface ocean flow in the North Atlantic. On the other hand, AO is the periodic movement of the atmospheric mass from the Arctic to

the mid-latitude areas. It is demonstrated that the NAO and the AO are connected to changes in the AW influx and outflow of Polar Water masses. The atmosphere's days-to-decades-long NAO and AO (Thompson and Wallace 2000; Thompson et al. 2000; Woollings et al. 2015) have seasonal to multiannual effects on the ocean (Eden and Jung 2001; Rodwell and Folland 2002; Badin et al. 2003; Mecking et al. 2015; Woollings et al. 2015; Delworth et al. 2017).

With this understanding, the present chapter is an attempt to understand how the fjord hydrography and the AW inflows to the fjord responded to the variations in the NAO and AO and to what depth these influences reach. While section 5.2 in the chapter investigates the correspondence between the variations in the fjord hydrography and NAO-AO, section 5.3 shows details on the most prominent influence from the NAO to the fjord variabilities. Section 5.4 summarizes the results.

5.2. Correspondence of Kongsfjorden hydrography with the NAO and AO

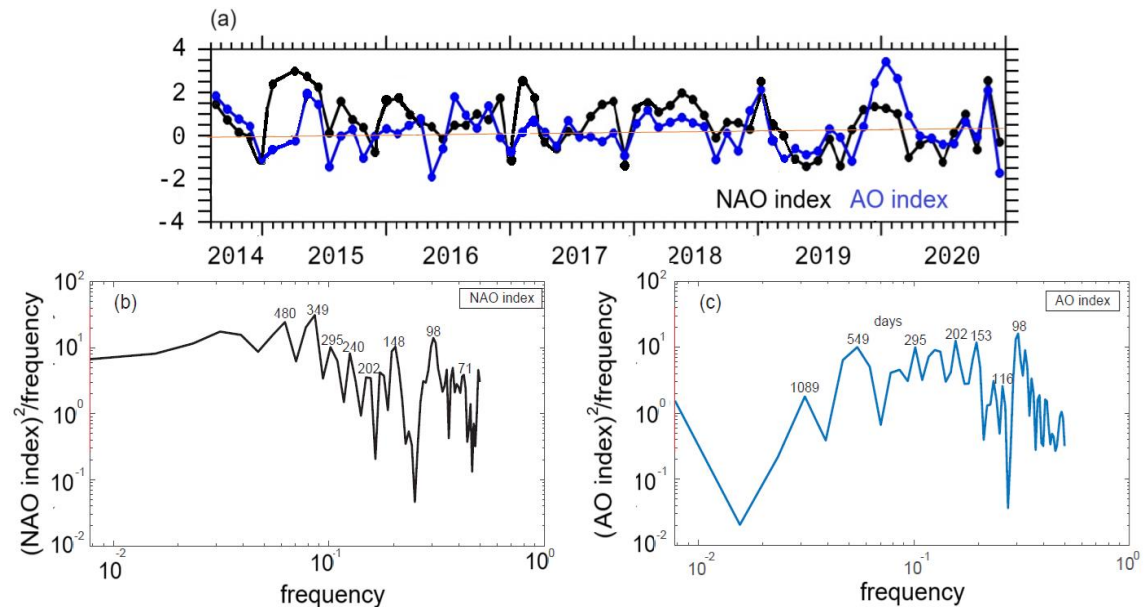


Figure 5.1: (a) The monthly NAO and AO indexes and the chi-squared analysis spectra with a 95% confidence level of the (b) NAO and (c) AO indexes from July 2014 to December 2020. The orange horizontal line in (a) denotes zero index.

Fig. 5.1 (a) shows the NAO and AO indexes during the mooring data from July 2014 to December 2020 which has been considered for the present analysis. The chi-squared spectral analysis of the NAO (Fig. 5.1 (b)) and AO (Fig. 5.1 (c)) indexes for the same period as the mooring data from July 2014 to December 2020 showed common peaks at 98, 202, and 295 days periodicities. Closer periodicities of 148 and 153 days were also seen in the NAO and AO indexes. Besides these common peaks, the NAO showed distinctive prominent peaks at periodicities such as 71, 240, 349, and 480 days, whereas the AO showed higher periodicities of 549 and 1089 days. In the spectral analysis conducted earlier in chapter 3, the Kongsfjorden water temperature at the upper 31 m and lower 140 m (Fig. 3.18 (a)) showed common peaks of 91 and 372 days, the first one being closer to the common periodicity of 98 days in the NAO (Fig. 5.1 (b)) and AO (Fig. 5.1 (c)). Besides, the upper depth temperature also showed 108 and 256 days. The higher periodicities in the above two sets of water temperature peaks are close to the 240 and 349-day periodicities observed in the NAO (Fig. 5.1 (b)). Prominent peaks at 228 and 341 days were seen in the 31 m salinity, while the 140 m salinity (Fig. 3.18 (b), chapter 3) exhibited 216, 240, and 341 days periodicities. These periodicities in the salinity are also closer to the periodicities observed in the NAO (Fig. 5.1 (b)). Thus, the water temperature and salinity at the upper and lower depths of observation in Kongsfjorden showed more common and prominent peaks in the NAO than in the AO.

The winter NAO was mostly positive throughout the observation period except in winter 2014/15 when AO was negative (Fig. 5.1 (a)). The winter NAO index was more positive than the winter AO except in 2019/20. During summer, both the climate modes changed between opposite and same phases to each other in alternate years. The general decrease in water temperature (Fig. 5.2 (a)) in the winter season, itself provides a mismatch with the positive NAO that is expected to bring more warm-saline Atlantic waters into the fjord. However, the salinity (Fig. 5.2 (b)) showed an increase in all observation winters except in 2017/18. The NAO and AO were further correlated to the temporal variation of pure AW inside Kongsfjorden (Fig. 5.2 (c)). The AW intrusions were seen strongly coinciding with the increase in current speed at both the upper and lower subsurface water column (Fig. 5.2 (d)). However, the

period of maximum AW in Kongsfjorden saw mixed phase variations in NAO and AO.

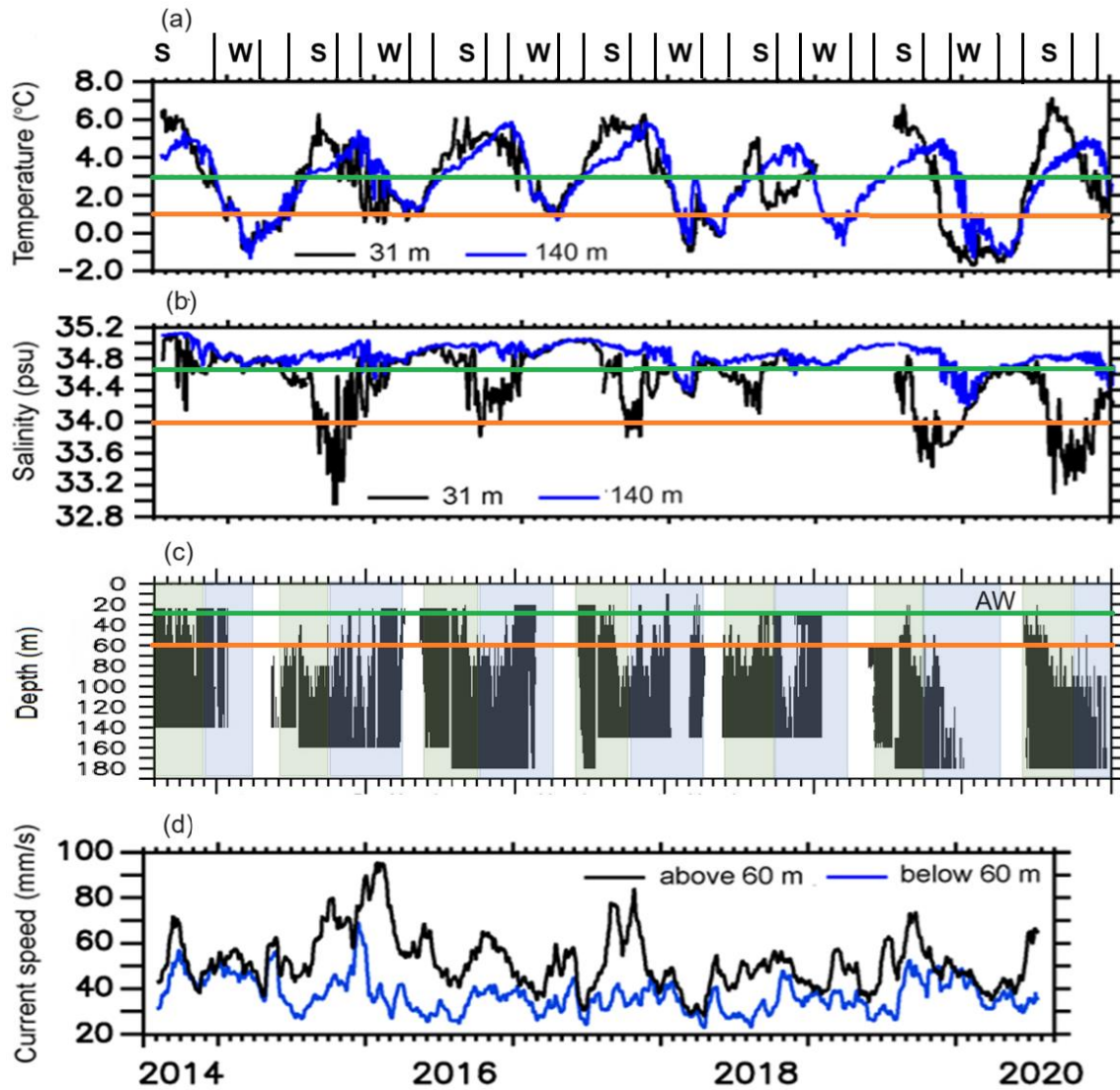


Figure 5.2: The daily time series of water (a) temperature and (b) salinity at 31 m and 140 m depths, (c) the vertical distribution of the Atlantic water mass from its characteristic temperature $> 3\text{ }^{\circ}\text{C}$, salinity $> 34.65\text{ psu}$, and $\sigma\text{-t} < 27.92$, and (d) daily current speed in the upper and lower water column defined by 60 m in Kongsfjorden from 2014 to 2020. The notations S and W marked on the upper time axis of (a) denote summer and winter seasons, as defined in Chapter 2, respectively which is applicable for all the subplots in the figure. The orange and green horizontal lines demarcate 1 and $3\text{ }^{\circ}\text{C}$ in (a), 34.00 and 34.65 psu in (b), and 27 and 60 m in (c), respectively.

The warmer months in Kongsfjorden with maximum AW intrusion, as found in chapter 3 are from August to November. This AW intrusion season in 2014 showed a positive phase of the NAO and AO. During this season, the current speed was high, and thereby the presence of AW. On the other hand, in 2015, when NAO was positive with a negative AO, even when the current speed was the highest compared to the other years, not much AW was observed inside the fjord. In brief, the positive NAO did not show a corresponding increase in AW inside the fjord. In 2016, as in 2014, both NAO and AO were positive and accompanied by a high current speed that resulted in high AW volume.

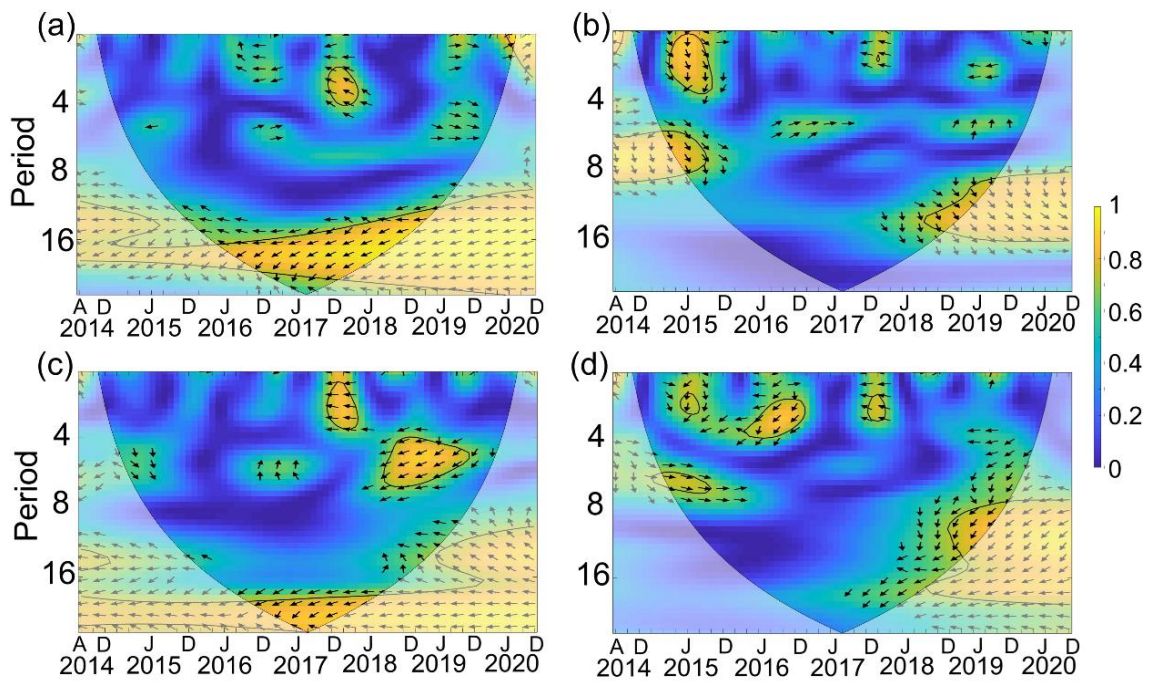


Figure 5.3: Wavelet coherence spectra between monthly AO index and (a) temperature at 31 m, (b) salinity at 31 m, (c) temperature at 140 m, and (d) salinity at 140 m.

The Atlantic water intrusion period in 2017 was similar to 2015 with a positive NAO and negative AO. However, the fjord had higher currents and more AW inside Kongsfjorden in 2017, in contrast to the low AW volume in 2015. In 2018, the NAO was weakly positive with a negative AO. However, considerable AW volume was found inside the fjord. During the warm months of 2019, when both the NAO and AO were negative, AW volume was low. However, this relation was again found disrupted in 2020 with high AW volume inside Kongsfjorden when both NAO and AO were

prominently negative. Thus, when some years showed a season-wise one-to-one correspondence of the phases of NAO-AO with the fjord water temperature, salinity, and AW, there were other years such as 2015, 2018, and 2020 where the correspondence was lacking in the same season. The correspondence with the phase changes was higher in NAO than in AO.

The correlation between the AO and water temperature at 31 m ($r^2 = 0.06$) and 140 m ($r^2 = 0.04$), and salinity at 31 m ($r^2 = 0.01$) and 140 m ($r^2 = 0.006$) were also very low during the observational years. The seasonal correlations were also of the same range. The wavelet coherence analysis of AO with the water temperature and salinity at 31 m (Fig. 5.3 (a, b)) and 140 m (Fig. 5.3 (c, d)) showed an in-phase relation in winter 2016/17 having a period of 5-6 months (≈ 150 -180 days) in the water temperature and salinity at 31 m. The summer of 2019 also showed an in-phase relationship of 31 m water temperature with the AO. The salinity at 140 m also showed some coherence at 180-210 days. Other than these, there was no significant coherence between the AO and the subsurface fjord water temperature and salinity during the observational period. Hence further investigations were focussed on the NAO.

5.3. North Atlantic Oscillation and Kongsfjorden variabilities

During its positive phase, the NAO shifts strong wind/storm tracks northward, increasing the strength of wind-driven Atlantic water flow to the Arctic Ocean via the Nordic waters (Dickson et al., 2000; Carton et al., 2011). Earlier studies (Quadfasel et al., 1993; Grotefendt et al., 1998; Serreze et al., 2000) indicated that the warming and increased extent of Atlantic water in the Arctic Ocean's 200-900 m depth were compatible with more Atlantic water inflow caused by the positive NAO. On the other hand, the negative phase of the NAO reduces warm-saline water intrusions into the Arctic. The NAO influences the eastern Arctic region through interannual changes in storm courses in the Nordic waters (Lien et al., 2013). Vinje (2000) has established a strong negative correlation of NAO with the Nordic sea ice extent for 135 years from 1860 which further allows more intrusions to the Arctic. The dominant atmospheric mode in the North Atlantic in the entire year, the NAO is more prominent in winter and contributes more than one-third of the total variance in sea level pressure (Ottersen et al., 2001). The NAO has large effects on ecology. The temperature rise and changes

in the winter direction and strength of ocean surface currents under the influence of the positive NAO in the last decades led to changes in food availability and the volume of the Norwegian Sea deep water where the overwintering copepod *Calanus finmarchicus* population resides resulting in reduction of them in the northeast Atlantic. Survival of adult little auks from 2006 to 2013 has shown a decrease, negatively correlated with the NAO, with a 2-year time lag (Hovinen et al., 2014). The study showed that the NAO changes the food quality or availability of Arctic *Calanus* copepods, which are the main prey of little auks, by the NAO-induced decrease in sea ice, ice algal production, and an earlier but shorter spring bloom.

The North Atlantic current that carries the Atlantic water northwards bifurcates into the Barents Sea and Fram Strait. The WSC flows east of the Fram Strait and on the shelf of the Svalbard archipelago towards the Arctic Ocean, carrying the warm-saline Atlantic waters. As detailed in chapter 1, the WSC is the main pathway of heat and salt transport to the Arctic region. Studies have connected the Arctic Ocean warming to changes in the NAO based on temperature variations in the WSC and the NAO index (Dickson et al., 2000; Blindheim et al., 2000; Saloranta and Haugan, 2001; Schlichtholz and Goszczko, 2005).

Very few studies tried to relate the winter NAO and the summertime AW temperatures in the WSC (Blindhelm et al., 2000; Dickson et al., 2000; Saloranta and Haugan, 2001). The first two studies showed a high correlation between the 50-500 m layer WSC temperature and the winter NAO index with no time lag. However, Blindhelm et al. (2000) reported a 2-3 year lag from the south in the Norwegian Sea. On the other hand, only a moderate correlation (0.6) with one-two year lag was found with the same data for a different period in the 50-200 m layer (Swift et al., 1997). Saloranta and Haugan (2001) found the presence and absence of coherence between the winter NAO and summer time WSC temperature and salinity variations at different periods. Schlichtholz and Goszczko (2005) documented a time-lag in the AW temperature in the WSC from the winter NAO. The study showed that the AW temperature lagged by 1 year behind the NAO while its salinity led the NAO by 1 year, indicating an obvious lead of AW salinity to its temperature. Also, with the lead-lag between the temperature-salinity in the AW and NAO, the correlations were higher compared to

the earlier studies that checked the correlation during the same time or season. The study concluded that the AW salinity can be a prognostic tool with its one-year temperature lead. The same way, the NAO phase can tell the AW temperature variations in the WSC.

With the formation of a conducive environment, as detailed in chapter 1 of this thesis, the Atlantic water has been carried through the WSC and then enters the west Svalbard fjords, including Kongsfjorden. Hurrell (1995) and Hop et al. (2002) reported that the NAO accounted for annual fluctuations in the strength of AW flow into the fjord. NAO implicates the WSC strength (Hurrell, 1995) and thereby influences the Kongsfjorden marine ecosystem through the temperature and salinity rise in the fjord from the Atlantic water advection. David and Krishnan (2017) tried to relate the summertime fjord AW variations with the phase of NAO in the same summer. The study used the monthly NAO index and established a positive relation of increase in AW inside Kongsfjorden when NAO was in a positive phase in the summer of 2013. On the other hand, the negative phase of NAO in the summers of 2011 and 2012, with no such strong northward winds, led to the low volume of AW observed in the fjord. However, with just three summers, the study could not establish whether the relation would be in the same way for long and continuous periods of observations and when checked between seasons.

Table 5.1: Correlation (r) of lead-lag analysis between the winter (December-March) NAO index and mean temperature as well as salinity (August-November) below 60 m in Kongsfjorden from 2014 to 2020. Zero lead/lag is when it is the winter NAO and successive ASON fjord variable. The +1 yr (-1 yr) denotes the ASON fjord variables after (before) a year with the winter NAO. Correlations of $p < 0.005$ are in bold letters.

ASON	NAO (DJFM) lead/lag						
		-3 yr	-2 yr	-1 yr	0	+1 yr	+2 yr
	T (31 m)	0.34	-0.54	-0.33	0.07	0.71	0.54
	T(below 60 m)	0.37	-0.41	-0.20	-0.28	0.76	0.42
	S (31 m)	-0.28	0.42	0.34	-0.29	0.75	0.56
	S(below 60 m)	-0.28	0.37	0.66	-0.29	0.79	0.42

In the present study, the NAO showed better correlations than the AO with the fjord temperature ($r^2 = 0.18 - 0.29$) and salinity ($r^2 = 0.11 - 0.23$). However, the still lower and less significant correlation with the NAO is due to the lack of correspondence observed between the NAO phases and the fjord temperature and salinity variations in the same season, as discussed in the section above. This kept forward another thought of an interseasonal (between the seasons) correspondence rather than the correspondence between the NAO and the fjord hydrography in the same season of the same year.

For the first time in Kongsfjorden, a detailed analysis is being done to find the long-term relation between NAO and the Kongsfjorden hydrography. The study conducted a lead-lag analysis (Table 5.1) as in Schlichtholz and Goszczko (2005). The temperature below 60 m to relate the NAO with the AW layer variations has been taken for the analysis. The analysis showed that the mean (ASON) water temperature and salinity at the subsurface lagged by a year from the winter NAO exhibited the highest positive correlation in the range ($r^2 = 0.50 - 0.62$; $p < 0.005$). This one-year lag in the fjord temperature is similar to that observed in the Schlichtholz and Goszczko (2005).

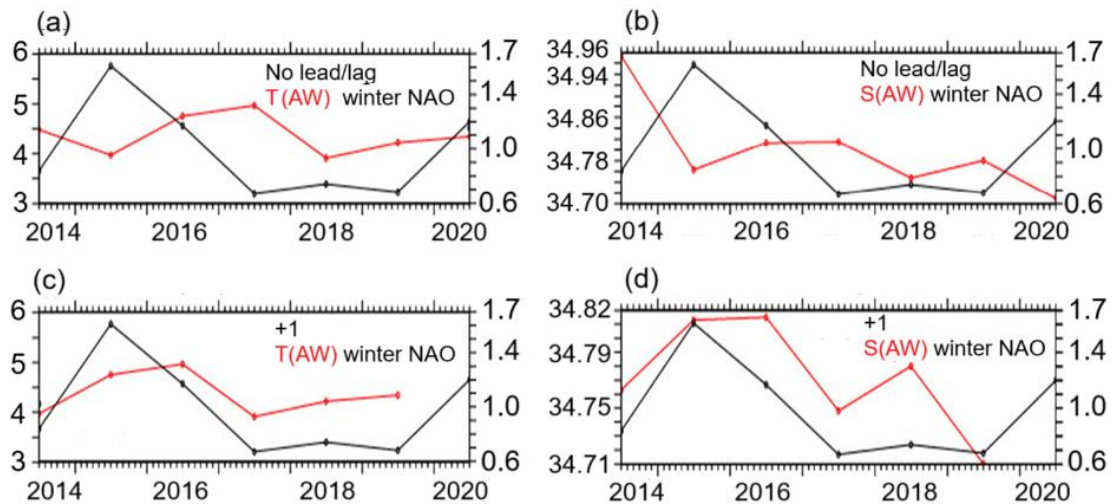


Figure 5.4: Mean Atlantic water (a) temperature ($^{\circ}\text{C}$) and (b) salinity (psu) for the August-November period with maximum warm water intrusion overlapped by winter NAO index without any lead/lag. Mean Atlantic water (c) temperature and (b) salinity for August-November overlapped by the winter NAO index led by one year.

Contrary to this earlier study, the present study did not see any advancement of one year in salinity. Instead, the present study saw salinity also having the highest correlation with the same lag of one year from the winter NAO as in temperature. The correlation was higher in the AW layer indicating the influence of NAO in the AW intrusions. A good correlation was also seen between the winter NAO index and the ASON temperature and salinity at both the depths behind the winter NAO by two years. The relation has been made more evident in the figure 5.4. Compared to no strong correspondence between the variations in the wintertime NAO and successive ASON temperature (Fig. 5.4 (a, b)), the one-year lagged Atlantic water temperature and salinity showed a higher correlation (Fig. 5.4 (c, d)) during the study period. The study highlights that the strength and phase of the NAO in winter affect the strength and volume of Atlantic water that gets inside Kongsfjorden during ASON in successive years.

The one to two-year lag between the NAO and Atlantic water properties in the WSC is found to be a result of several physical processes affecting heat and water mass transfer from the North Atlantic Ocean to the Arctic Ocean. During the positive NAO phase, the Subpolar Gyre strengthens, enhancing Atlantic water transport to the Nordic Seas. However, the northward propagation of Atlantic water takes time, causing a delay before temperature signals reach the WSC (Häkkinen et al., 2011; Årthun et al., 2012; Barton et al., 2018). Additionally, the NAO influences thermocline depth through wind stress and buoyancy forcing. Slowly propagating baroclinic Rossby waves can take one to two years to adjust thermocline depth and heat content in the Nordic Seas (Häkkinen, 1999; Curry and McCartney, 2001). NAO-driven sea ice variability also modifies ocean-atmosphere heat exchange and freshwater input, contributing to further delays in Atlantic water temperature response (Screen and Simmonds, 2010; Carmack et al., 2015). Internal ocean processes, such as eddies and mixing, further enhance this lag after NAO-driven changes (Fer et al., 2003; Våge et al., 2011). Together, these interconnected processes create the observed lag in the Atlantic water properties in the WSC which can also be responsible for similar delays observed in Kongsfjorden since the Atlantic water reaches the fjord from the WSC. Even though, the lead/lag relation between the variations in the subsurface Atlantic water properties in Kongsfjorden and the winter NAO from 2014 to 2020 is

determined in the present study, the relation may vary as reported by Saloranta and Haugan (2001) who found different lead/lag between the subsurface Atlantic water properties and the winter NAO during different periods and geographical locations along the way of the WSC on west Svalbard. With a very limited number of earlier studies in Kongsfjorden, a very detailed analysis in space and time using observations and modeling is needed to establish the physical processes involved in the relation between the Atlantic water properties and the winter NAO index. Hence, this is one of the future prospects of this thesis.

5.4. Conclusion

The chapter investigated the correspondence and influence of two major climate modes- NAO and AO in Kongsfjorden subsurface hydrography variations. The study established the prominence of the NAO with much lesser implications from the AO. NAO showed similar peaks that were prominent in the temperature and salinity variations. The temporal variations in the water temperature, salinity, and presence of AW also showed the prominent role of NAO. However, the correlation between the successive winter NAO and ASON water temperature and salinity at the upper 31 m and lower 140 m were not statistically significant.

Moreover, a corresponding change in AW volume inside Kongsfjorden to the interannual changes in the NAO index was not always seen during the observational period. Further, a lead-lag analysis done on the NAO with the water temperature and salinity showed that corresponding changes from the wintertime NAO were seen in the August-November subsurface water temperature and salinity, after a year. Subsequently, the winter NAO and a year-lagged ASON water temperature and salinity showed the highest positive correlation of 0.50 – 0.62 ($p < 0.005$). The correlation was higher in the AW layer below 60 m, showing a stronger influence of the winter NAO in the AW intrusions after a year. A secondary high correlation was seen between the NAO index and the water temperature and salinity at both depths after two years. In short, this documentation of the lag observed in the Kongsfjorden hydrography after the winter NAO provides a detailed picture of how the subpolar Atlantic influences the polar Kongsfjorden. This will help the modelling efforts while addressing the processes related to and between these two regions. The lag can be

utilized as a predictive tool for the interannual hydrography and AW variations in Kongsfjorden. However, more years of continuous time series data would be needed to establish the influences from the higher period climate mode and its relation with the fjord variabilities.

Chapter 6

Summary and Conclusion

The interannual physical variations in Kongsfjorden, a west Svalbard fjord in the eastern Arctic Ocean, were deciphered using the spatial CTD survey data in summer 2011-2020 and time series subsurface mooring data during 2014-2020. A summertime warming of $0.06\text{ }^{\circ}\text{C/yr}$ and salinification of 0.03 psu/yr were found in the fjord. 2014 was the warmest and most saline in the decade, and 2011 summer was the coldest and least saline. The surface and Atlantic water masses in Kongsfjorden showed different scenarios in summer. The AW layer was thin whenever the SW layer was thick, and vice versa. The thickest AW and the thinnest SW were observed in 2014, but 2011 had the thickest SW with no AW. The rate of AW warming (0.07°C/yr) was $0.02\text{ }^{\circ}\text{C}$ higher, and the rate of salinification of the AW (0.008 psu/yr) was one-fifth than that of the SW. The increase in SW temperature was synonymous with the increasing net surface radiation and air temperature. The increasing trend in precipitation and glacial discharge justified the SW freshening inferred from the decrease in the minimum salinity trend. High AW volume was observed in 2014, 2016, 2017, and 2019. The mean south-south easterlies that prevailed during these years showed the wind-driven mechanism of the surface freshwater outflow from Kongsfjorden, enhancing the Atlantic water flows into the fjord. The sea ice coverage, concentration, and thickness were the lowest in 2014, which created a conducive situation for the maximum Atlantic water intrusion to Kongsfjorden. The role of local sea ice conditions in the Atlantic water variability inside Kongsfjorden was also evident as it followed the fjord temperature variations with the WSC in 2014, 2016, and 2017. These years had the least local sea ice coverage. It is worth noting that the pattern of the interannual variations in the SW temperature and salinity was similar to the AW. This indicated the increasing influence of the AW in the upper layers of the water column over the years through vertical mixing and, in part, due to the entry of the Atlantic water through the fjord surface. Finally, the long-term mean temperatures over the water column depths for the summers of 2011-2020 were $0.5\text{-}0.8\text{ }^{\circ}\text{C}$ higher than that in 1994-2014 ($1.5\text{-}5.0\text{ }^{\circ}\text{C}$) reported by Tverberg et al. (2019).

The entrance of the warm Atlantic waters from the WSC through the southern shore of Kongsfjorden was revealed in the ROMS simulations. The circulation pattern showed the impact of the heat supply by these warm waters on the head, which led to glacial calving and melting. A portion of the freshwater outflow from Kongsfjorden through its northern side enters eastern Krossfjorden. While entering, this fresh colder water mixes with the warmer Atlantic water that directly enters from the open ocean through eastern Krossfjorden decreasing the water temperature in the region. Laterally, the Kongsfjorden water column showed two temperature scenarios- a two-layer system in the outer Kongsfjorden having warmer-less saline surface water above cold-high saline deep water and a three-layer system in the inner fjord having warm-saline water sandwiched between cold-less saline surface water and cold-high saline deep water. The first two layers in the inner fjord showed deepened temperature inversion from July to October. The strongest inversions occurred in August and September, coinciding with the maximum freshwater discharge into the fjord surface. A decrease in air temperature, increase in winds, net surface heat flux loss, and increase in precipitation from August to October adds to the upper-layer cooling in addition to the freshwater discharge, resulting in the formation of temperature inversion and a three-layer temperature structure in the inner fjord.

The variabilities, especially in winter, spring, and fall in Kongsfjorden, were studied using the daily time series mooring data from 2014 to 2020. Even though 2020 was the longest winter with more extended residence of LW, the coldest winter temperature of -1.0°C was found in February 2015, and the lowest salinity of 34.4 psu in February 2018. The temperature increased in summer, winter, and fall from 2014 to 2020. The rate of winter water temperature rise of 0.08°C/yr found at the mooring location in the central Kongsfjorden was half the rate reported in the outer Kongsfjorden, which confirmed a sharp lateral gradient in the fjord hydrography. The increase in winter temperatures is in agreement with the recent reports by Tverberg et al. (2019) and Cottier et al. (2021) in Kongsfjorden. The spring temperature decreased over the period due to the delay and extension of winter, which overlapped the defined spring months. Such overlaps were observed frequently over the decade. During months with maximum AW intrusions (August-October), the maximum temperature in the AW

layer below 60 m showed a rise of 0.05 °C/yr from 2014 to 2020. This rate was the same as the summer average below 60 m.

Between the temperature peaks at 31 m in August and lows at 140 m in November each year, the water temperature showed a three-month difference. Salinity at 140 m showed seasonality as in 31 m but embedded in a low-frequency variation. Stratification maximum in summer was 2-5 times higher in the upper 60 m layer than below it. The highest stratification was found in 2015 and 2020, with the largest surface freshening and with a vertical arrangement of SW, IW, and ArW, one below the other. The lowest stratification was in 2014 and 2017, the warmest years of the decade. The stratification increases and peaks below 60 m coincided with the pure AW intrusion, while its decline coincided with the modification of AW to TAW or LW. After its presence in the 2015 winter, LW's presence in the fjord during winter and spring increased from 2018 to 2020. This is a reason for the decreasing mean temperatures in winter and spring during these years. The warmest winters in 2016 and 2017 were due to the TAW and AW's longest and more dominant presence.

The interannual condition of sea ice, such as its coverage, concentration, thickness, and length of sea ice season, influences the interannual variations in the Kongsfjorden water column temperature and salinity. The years with the shortest fjord ice season and lower coverage and concentration resulted in the most warm and saline years, and vice versa. The high fast ice coverage in 2015 led to the coldest winter water column, and the shortest fjord ice season, and the low ice extent resulted in the warmest and most saline winter to spring in 2017. The overall effect of the winter sea ice conditions was carried over to the spring, summer, and fall hydrography environments in the fjord. The highest wind speed with the lowest stratification, one order lesser than in summer, added to the coldest and vertically more homogenous water column than the other years. When the mean winter wind speed was the lowest in 2020, the stratification was the highest. Thus, winds had a higher impact in the upper 60 m in summer and almost the entire water column in winter.

After one to three months of peak in the net surface radiation flux that happened mostly in July, the fjord HC became maximum. The HC continued rising even after the net

surface radiation declined and became negative indicating heat loss to the atmosphere. The periods of increasing fjord HC and declining surface radiation coincided with the periods of Atlantic water presence in Kongsfjorden. Prominent southeast currents below 60 m were observed from October till the end of the cold season with maximum AW intrusion, implying heat transport by the warm Atlantic water to Kongsfjorden. The flux analysis proved to be a promising way of detecting periods of AW advection in Kongsfjorden.

The least investigated ISOs in Kongsfjorden hydrography were studied using the mooring time series observations. The water temperature and salinity spectral analysis showed ISOs (10-100 days) at 31 m and 140 m which were more significant in the water temperature. The ISOs were prominent generally from late summer to winter at 31 m, where frequent modifications in the water masses occurred. The ISOs had the least energy at both depths from spring to summer. Wind stress and current speed played major roles in imbibing ISOs into the fjord water column. However, currents played a major role in triggering the ISOs in the deeper waters. Also, the period of significant ISOs at the deeper levels were the periods of maximum AW intrusion implying its role in the generation of ISOs in Kongsfjorden. The study showed continuous and long residence of Atlantic-origin waters at the deeper Kongsfjorden over the last decade. The increase in the number of warm years resulted in a shift of the fjord to the Atlantic type. The updated observational information on Kongsfjorden hydrography can add to the understanding of the physical and biogeochemical processes in the west Svalbard fjords. Further, the results can help coherently depict the dynamic Arctic fjord processes and address the modelling challenges in the region.

The study is one among the very few to bring out the annual, seasonal, intraseasonal, and interannual variabilities of Kongsfjorden biogeochemistry, its coupling to the physical processes, and the evolution of subsurface spring and secondary blooms for multiple and distinctive years using the continuous time series mooring observations. Significant ISOs in DO and nitrate were higher in the fall through half of the winter season and coincided with the warm water intrusions. This shows the role of Atlantic water in imbibing ISOs in the fjord biogeochemistry besides the hydrography. Winter and fall showed the least productivity because of light limitation, even though the

availability of nutrients was high. Nutrients (nitrate) increased rapidly at the highest annual rate during fall. It increased in winter under weak stratification and deep convective mixing but at a rate of 2-3 times less than that in fall. During these least productive seasons, episodes of Atlantic water were also seen increasing the nitrate concentration further from the ambient condition. After this period, the densest winter waters were seen.

The highest annual rate of decline in the nutrients was seen in the spring season. The present study showed a late occurrence of spring bloom from late April in 2016 to mid-April in 2017 and 2018. The peaks in the bloom were found by mid-May in 2016 and 2017 and in early May in 2018. The 2016 spring bloom showed two major breaks, while the 2017 and 2018 blooms showed a continuous pattern with minor ups and lows than in 2016. The breaks in the bloom were induced by high local winds greater than 5 m/s, which induces mixing and, thereby, weak stratification. The mixing allowed for the nutrient rejuvenation in the upper layer and thereby the comeback of the bloom after the winds reduced, reestablishing the stratification. Both 2016 and 2018 blooms confirmed the mechanism. However, the 2017 spring was different with the least role of stratification during the bloom. The 2017 spring bloom was found to be majorly advective, as indicated by the strong northeast currents that transported chlorophyll from the more productive and nutrient-rich southern shore of Kongsfjorden to the mooring location. The bloom magnitudes were higher in the warm years of 2016 and 2017 when the bloom evolved and peaked under the TAW scenario, compared to 2018, when the formation occurred under the colder-less saline ArW-LW scenario. Moreover, the timing and pattern of the blooms in the present study years are similar to the 2007 bloom as predicted for an ice-free Kongsfjorden with normal winter mixing.

Summer is the period of maximum sunlight and nutrient-depleted surface waters after the maximum nutrient utilization in the spring. The robust stratification during the summer inhibits further mixing of nutrients from the deeper waters to the upper layers. Summer also had secondary blooms whose magnitude is generally lower than the spring blooms. However, these secondary blooms are not continuous phenomena like most spring blooms. Instead, they are episodic events of enhancement in chl *a* as the

source of nutrients described above being episodic events rather than a continuous supply. The secondary bloom period in the summer of 2015 and 2017 extended further to the successive fall season. As seen in the spring bloom, the summer productivity was also higher in 2016 and 2017, dominated by the Atlantic-origin water masses, and lower in years dominated by SW and IW. This further shows the increasing role of the Atlantic water in the fjord nutrient supply and productivity and its interannual variability. With the increase in Atlantic water intrusions, continuous monitoring of the fjord biogeochemistry and physics is needed to understand these changes. The study also reiterates the need to incorporate such observations and dynamics of subsurface spring and secondary bloom in modelling studies.

The study further explored the correspondence and influence of two major climate modes- NAO and AO in Kongsfjorden subsurface hydrography. NAO showed similar peaks that were prominent in the temperature and salinity variations. The prominent role of NAO was observed, as indicated by the temporal variations in the water temperature, salinity, and presence of AW. However, the one-to-one correlation between the NAO and water temperature and salinity with time was not statistically significant. A corresponding change in AW volume inside Kongsfjorden to the interannual changes in the NAO index was also not always seen. A lead-lag analysis showed that corresponding changes from the wintertime NAO were seen after a year in the August-November subsurface water temperature and salinity. Subsequently, the winter NAO and the one-year lagged water temperature and salinity showed the highest positive correlation ($r^2 = 0.50 - 0.62$; $p < 0.005$) improving the correlation to higher than double the correlation obtained with no lead/lag. The correlation was higher in the AW layer below 60 m. NAO index and the two-year lagged water temperature and salinity also showed a secondary high correlation. The documentation of the lag observed in the Kongsfjorden hydrography after the winter NAO provides a detailed picture of the subpolar Atlantic influences on the polar Kongsfjorden. This study will further help in the modelling efforts for deciphering Arctic biogeochemical dynamics for a bigger picture in addressing the processes related to and between the Atlantic and Kongsfjorden.

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Publications

1. Divya, D. T., Subeesh, M. P., Anju, V. S., and Anilkumar, N. (2021). Variability in the hydrography of two proximate Arctic fjords during 2013–18. *Polar Science*, 30, 100735.
2. Divya David T, Pribin K. P, Renjith V R, and N. Anilkumar (2025). Insights into the biogeochemical variability and phytoplankton bloom dynamics from a multiyear time-series mooring in Kongsfjorden, an Arctic fjord. *Journal of Geophysical Research-Oceans* (Under Review).

Conferences

1. Divya David T, Subeesh M. P, Teesha Mathew, Renjith V. R, Anilkumar N, and M. Ravichandran. Physical and biogeochemical variability over a decade in an Arctic fjord, and its climate relevance (2024). Ocean observations, processes and modelling, World Ocean Science Congress at IIT Madras, 27-29 February 2024.
2. Divya David T, Anilkumar N, Teesha Mathew, and M Ravichandran. Role of regional climate modes in Kongsfjorden over the last decade (2023). Climate change and variability, National Conference on Polar Sciences, NCPOR Goa, 1-19 May 2023
3. Divya David T, N. Anilkumar, Teesha Mathew, Renjith V. R, and M. Ravichandran. Decennary variability and regulating processes in Kongsfjorden, an Arctic fjord (2023). OSICON-23, INCOIS Hyderabad, 23-25 August 2023.
4. Divya David T, Archana Singh, Shridhar Jawak, Subeesh M. P, and N Anilkumar (2022). Subsurface chlorophyll maximum and water column primary productivity in Kongsfjorden'. Arctic Science Summit Week 2022 in Tromsø, Norway March 26 – April 1
5. Divya David T, Pribin K. P, Teesha Mathew, Subeesh M. P, Anil Kumar N, and M. Ravichandran. Deciphering Kongsfjorden hydrography from 2011 to 2020 (2022). SIOS online conference on Earth observation and remote sensing applications in Svalbard, 12-14 October 2022

Award

Best paper award in the session ‘Ocean observations, processes and modelling’ in the World Ocean Science Congress conference at IIT Madras, 27-29 February 2024 for the paper (details in Sl. No. 5 under conferences section above)

Other publications related to the thesis

1. Bensi, M., Nilsen, F., Ferre, B., Skogseth, R., Moskalik, M., Korhonen, M., Vogedes, D., Kovacevic, V., Paladini de Mendoza, F., Ingrosso, G., Langone, L., Giordano, P., Inall, M., Mano, B., Sundfjord, A., Bailey, A., Foss, Ø., Daase, M., Castro de la Guardia, L., David T, Divya; Renner, Angelika; Dumont, Estelle; Glowacki, Oskar; Ogg, Franziska; Olsen, Helene; Dolven, Knut Ola; Jones, E. (2025). The Atlantification process in Svalbard: a broad view from the SIOS Marine Infrastructure network (ARiS). In SESS report 2024 - The State of Environmental Science in Svalbard - an annual report (pp. 138–151). Svalbard Integrated Arctic Earth Observing System.
2. Philipp Assmy, Ane Cecilie Kvernvik, Haakon Hop, Clara J.M. Hoppe, Melissa Chierici, T. Divya David, Pedro Duarte, Agneta Fransson, Laura M. García, Weronika Patuła, Sławomir Kwaśniewski, Marion Maturilli, Olga Pavlova, Agnieszka Tatarek, Jozef M. Wiktor, Anette Wold, Klara K.E. Wolf, Allison Bailey, Plankton dynamics in Kongsfjorden during two years of contrasting environmental conditions, Progress in Oceanography, 2023, 102996, ISSN 0079-6611.
3. Finlo Cottier, Ragnheid Skogseth, Divya David, Francesco De Rovere, Daniel Vogedes, Malin Daase, and Jørgen Berge (2022). Temperature and salinity time series in Svalbard fjords – ‘Integrated Marine Observatory Partnership (iMOP II)’, SESS report 2021, The State of Environmental Science in Svalbard – an annual report.
4. Archana Singh, Divya David T., Sarat C. Tripathy, Ravidas K. Naik, Interplay of regional oceanography and biogeochemistry on phytoplankton bloom development in an Arctic fjord (2020), Estuarine, Coastal and Shelf Science, Volume 243, 2020, 106916, ISSN 0272-7714.



Variability in the hydrography of two proximate Arctic fjords during 2013–18

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Fjords
Atlantic water
Glacial melting

ABSTRACT

Svalbard fjords undergo significant oceanographic and environmental changes associated with the ongoing Arctic warming and increased Atlantic water inflows. We used CTD data from 2013 to 2018 combined with numerical model simulations to decipher the hydrography and circulation in the Krossfjorden-Kongsfjorden, two west Svalbard fjords, with a focus on the Krossfjorden. The Atlantic water prominent in the fjords showed strong inter-annual variability. The Atlantic water and summer freshwater volume (FWV) were more in the Kongsfjorden, with the FWV being 1.4–2.0 times more than 0.08–0.71 km³ in the Krossfjorden. The Krossfjorden had a two-layer structure with warm less-saline surface waters above cold high-saline waters in early summer. The fjord transformed to a three-layer system with warm high-saline waters between cold less-saline surface and cold high-saline deep waters in peak summer and fall. The Krossfjorden temperature showed east-west variability, with a warmer eastern side. Numerical simulations show that the warm Atlantic water from the West Spitsbergen current enters the Krossfjorden through its eastern side, supplying heat to the inner fjord. An outflow is also observed along the western Krossfjorden. This study adds to the understanding of the Svalbard fjord processes in recent times.

1. Introduction

The Arctic Ocean is undergoing an alarming sea-ice decline in recent decades (Brennan et al., 2020). The warm water advection from the North Atlantic Ocean to the Arctic that has increased over the last decade emerged as a pivotal reason for the eastern Arctic sea-ice decline (Polyakov et al., 2017; Wang et al., 2020). Apart from the sea-ice decline, the warm waters interact with marine-terminating glaciers and lead to their retreat. The glacier mass loss and associated freshwater discharge significantly affect the global sea-level rise (IPCC, 2014).

The Arctic fjords are bounded by oceanic and terrestrial realms and respond to or exert change on these two realms (Cottier et al., 2010). The glacier-dominated fjords have a seasonal freshwater input in the inner part and warm saline input from the open ocean at its outer part (Svendsen et al., 2002; Cottier et al., 2005). Smith et al. (2015) show that the fjords regulate the carbon cycle by burying the largest amount of organic carbon per unit area globally and controlling the atmospheric carbon dioxide concentrations at times of varying ice sheet dynamics.

The Kongsfjorden-Krossfjorden is a twin fjord system situated between the latitudes of 78°40' N and 79°30' N and longitudes of 11°03' E and 13°06' E at the north-west of Spitsbergen in the Svalbard archipelago (Fig. 1). The Krossfjorden is 30 km long and 3–6 km in width, with a volume of 25 km³ (Svendsen et al., 2002). The Kongsfjorden is 20 km long, 4–10 km in width, with a volume of 29.4 km³ (Ito and Kudoh, 1997). The Krossfjorden is aligned from the north to south direction, whereas the Kongsfjorden is aligned from the north-west to south-east. Both the fjord channels intersect at Kongsfjordrenna, a deep glacial basin, as it lacks a sill in the mouth. The outer part of the Krossfjorden is up to 374 m deep (MacLachlan et al., 2010). Towards the north, the Krossfjorden bifurcates into two minor fjords with water depths of ca. 250–100 m (Sexton et al., 1992): Lilliehookfjorden and Mollerfjorden by Kong Haakons Halvøya. The maximum depth of Kongsfjorden is 400 m (Harms et al., 2007). The glacier discharge from a tidewater glacier-Lilliehookbreen at the head of the Krossfjorden strongly influences the fjord system. Meanwhile, there are five tidewater glaciers in Kongsfjorden: Kronebreen, Kongsvegen, and Kongsbreen at the head and

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Conwaybreen and Blomstrandbreen at the north coast (Fig. 1).

Located to the east of the West Spitsbergen Current (WSC), the northernmost branch of the North Atlantic current that transports heat to the Arctic, the fjords are important both climatically and oceanographically (Cottier et al., 2005). Due to the location of the fjords, the overall rise and the variability in the Atlantic water inflows to the eastern Arctic can be identified as change in the intruded Atlantic water volume inside the fjords (Svendsen et al., 2002; David and Krishnan, 2017; Tverberg et al., 2019). The warm Atlantic waters that reach the fjords result in calving, melting, and mass loss of marine-terminating glaciers and more melt-water in the fjords (Luckman et al., 2015). This melt-water is the primary freshwater source in summer and even

small in quantity in winter. Additional sources are snowmelt, river runoff, groundwater discharge, and precipitation (Cottier et al., 2005). Therefore the hydrographic properties of the Kongsfjorden-Krossfjorden system are mainly affected by these Atlantic and glacier melt-waters that result in strong physical gradients in the fjord system (Svendsen et al., 2002; Sundfjord et al., 2017).

Comprehensive reviews have focused on modern oceanographic and physical conditions (Svendsen et al., 2002; Tverberg et al., 2019) as well as marine biology and ecosystems in the Kongsfjorden-Krossfjorden system (Hop et al., 2002, 2006; Wiencke and Hop, 2016). Among the Kongsfjorden-Krossfjorden, the Kongsfjorden remains one of the most extensively studied fjords in the west Svalbard. A recent study based on a

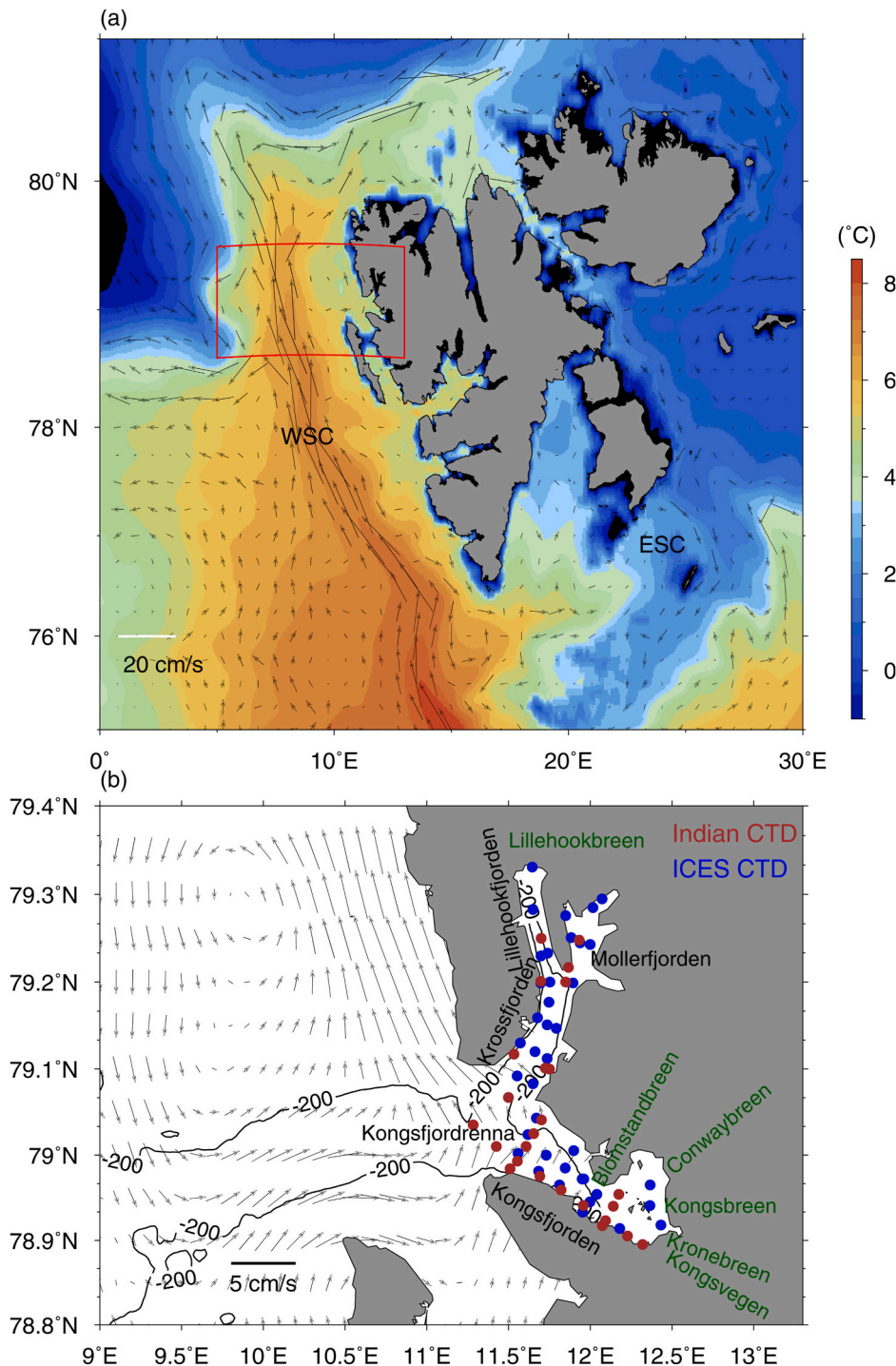


Fig. 1. (a) Map of the Svalbard archipelago with the 0–50 m depth-averaged ocean temperature (°C) overlaid by the currents (cm s⁻¹) averaged for the month of August over the period 2002–2017. The red rectangle denotes the ROMS model domain. (b) The study region, Krossfjorden-Kongsfjorden, marked with the Indian (blue dots) and ICES (red dots) CTD measurement locations. Isobath in the diagram denotes 200 m depth. The ocean temperature in (a) and ocean current in (a) and (b) are from the GLORYS12V1 product, the CMEMS global ocean eddy-resolving (1/12° horizontal resolution, 50 vertical levels) reanalysis [https://resources.marine.copernicus.eu/?option=com_csw&view=details&product_id=GLOBAL_REANALYSIS_PHY_001_030]. The bathymetry and topography in the diagrams are from the SRTM30 PLUS estimated topography, 30 s, global, v11 data [https://topex.ucsd.edu/WWW_html/srtm30_plus.html]. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

sediment core from the Krossfjorden reported an increase in August sea surface temperature of 0.6 °C over 60 years from 1953 to 2014 (Guru-vayoorappan et al., 2019). They reasoned it with increased CO₂ forcing and air-sea interaction rather than an increasing influence from the Atlantic water. No detailed hydrographical studies have been carried out in the Krossfjorden as in the Kongsfjorden. Hence, it is important to understand the recent variabilities in the Krossfjorden and compare it with the Kongsfjorden.

The present study focused on the evolution of the Krossfjorden hydrography during summer-fall seasons from 2013 to 2018. The inter-annual variability in the Krossfjorden-Kongsfjorden system, and a comparison between these two fjords were also carried out. The paper is organized as follows. Section 2 describes the data, model, and methods used for the analysis. Section 3 shows the hydrography, mixed layer variability, water masses, and freshwater volume in the Krossfjorden-Kongsfjorden system. Section 4 discusses the evolution of the hydrography, heat flow, and circulation in the Krossfjorden, followed by inter-annual variability in the fjords and comparison between them. Section 5 summarizes and concludes the results.

2. Data and methods

2.1. Conductivity-Temperature-Depth data

We used the in-situ water temperature and salinity data from hydrography surveys in the Krossfjorden-Kongsfjorden using SBE 19plus V2 SeaCAT Conductivity-Temperature-Depth (CTD) profiler during the summer-fall seasons (June–October) for six years from 2013 to 2018. The accuracy of temperature was ± 0.005 °C, whereas conductivity was ± 0.0005 S/m. The CTD processing was done following the Seabird recommended methodology, as described in David and Krishnan (2017). The upper 2 m CTD values were not considered to avoid any discrepancies that might have happened in the soaking depth handling by multiple researchers operating the profiler and the skin layer errors.

Apart from this, we also used historical CTD data from the International Council for the Exploration of the Sea (ICES) [<http://ocean.ices.dk/HydChem/>]. The ICES database is a repository of CTD data from different sources, such as expeditions conducted in the region by multiple research institutions and ships of opportunity. We used the ICES CTD data from 22 locations in the Krossfjorden and 17 locations in the Kongsfjorden in August 2013 and 2014. The spatial distribution of the CTD locations in the Krossfjorden-Kongsfjorden is shown in Fig. 1. The geographical locations and the station depths are given in Table 1. The CTD survey conducted with the total number of profiles available each month is presented in Table 2.

Even though the data was not available every month in the Krossfjorden from June to October each year, this six-year data covered at least one month among the three peak summer months (July–September). The fjord temperature and salinity profiles collected in each month were averaged to obtain the monthly means for the study. The data was available for August month in all the years except 2016. Hence the interannual variability is discussed based on the August data. The fall data were available only in October of 2015 and 2016.

2.2. Water mass classification

We studied the water masses in the fjords from the water temperature (T) and salinity (S) using T-S diagrams. Considering the geographical proximity of the Krossfjorden and the Kongsfjorden, we adapted the same water mass characteristics ranges of the Kongsfjorden (Cottier et al., 2005) to delineate the water masses in the Krossfjorden also (Table 3), expecting negligible changes in the range of the characteristic temperature-salinity of the water masses between the fjords. The surface water mass (SW), intermediate water mass (IW), and local water mass (LW) are the internal water masses that are formed inside the fjords. The SW is formed due to glacial melting during late spring and

Table 1

The geographical locations of the Indian CTD survey conducted during the summer-fall of 2013–2018 with the depth at each location in the Krossfjorden and the Kongsfjorden.

Krossfjorden			
CTD Locations	Latitude (°N)	Longitude (°E)	Station depth (m)
I1	79.0667	11.5000	296
I2	79.1167	11.5333	282
I3	79.1013	11.7245	175
I4	79.1000	11.7500	224
I5	79.2009	11.6968	255
I6	79.2000	11.8500	244
I7	79.2167	11.8667	169
I8	79.2500	11.7000	265
I9	79.2476	11.9334	184
Kongsfjorden			
1.0	79.0354	11.2836	345
2.0	79.0099	11.4245	356
3.0	78.9931	11.5547	326
3.1	78.9840	11.5090	322
3.2	79.0100	11.6080	246
3.3	79.0250	11.6540	227
3.4	79.0410	11.7020	290
4.0	78.9755	11.6915	313
5.0	78.9587	11.8224	293
6.0	78.9408	11.9576	301
7.0	78.9228	12.0937	189
7.1	78.9170	12.0740	281
7.2	78.9400	12.1410	196
7.3	78.9540	12.1750	64
8.0	78.9050	12.2278	193
9.0	78.8951	12.3201	49

summer in the upper few meters of the fjord. It exhibits a wide range of temperature due to high particulate content resulting in warming from insolation, while the salinity generally is below 34 (Cottier et al., 2005) (Table 3). The IW is formed by the mixing and entrainment between the upper SW and the lower Atlantic water mass (AW) or Transformed Atlantic water mass (TAW). The LW is formed during autumn and winter through surface cooling and convection. Hence it has a low temperature (Table 3). The remnants of the LW are found at deeper levels inside the fjord water column during summer in some years. The AW and TAW are the external water masses of the warmer-saline Atlantic Ocean origin that enter the fjords from the WSC. More details on the formation mechanisms of these water masses are detailed in Svendsen et al. (2002), Cottier et al. (2005), and Nilsen et al. (2008).

2.3. Mixed layer depth

The density gradient, $d\rho/dz$ where ρ is the ocean density and z is the depth, was computed from the temperature, salinity, and depth profiles from the CTD data in the fjords. The mixed layer depth (MLD) was then estimated using the criterion of depth at which the maximum density gradient was found (Meire et al., 2016).

2.4. Freshwater content

We estimated the freshwater content (FWC) in the fjords following Dølven (2015),

$$FWC = \int_z (S_{ref} - S/S_{ref}) dz$$

where S is the measured salinity, S_{ref} is the reference salinity, and z is the depth in meters of the water column. Here, the S_{ref} was taken as 34.65 that demarcated the external and internal fjord water masses. The total freshwater volume (FWV) in the Krossfjorden was estimated by multiplying the surface area of the fjord, 135 km² (length of 30 km multiplied by average width of 4.5 km), adapted from Svendsen et al. (2002) with the FWC. The FWV was estimated for the Kongsfjorden by multiplying

Table 2

The dates of the Indian and ICES CTD data availability with the total number of profiles in each month during the summer-fall of 2013–2018 in the Krossfjorden and the Kongsfjorden.

Year	Month	Krossfjorden			Kongsfjorden		
		Dates of CTD observation		Number of profiles	Dates of CTD observation		Number of profiles
		Indian	ICES		Indian	ICES	
2013	August		21, 30, 31	22		29, 31	17
2014	July	15		9			
	August		29, 30	21		29, 30	10
2015	August	3, 17		17	17, 18		15
	October	24		9			
2016	June	29		9			
	July	14		9			
	October	12		9			
2017	July	4, 27		12			
	August	31		9	29		16
	September	21		9			
2018	July	19, 23		9			
	August	6, 10, 17		19	6		6

Table 3

The ranges of temperature (°C) and salinity characteristics of the water masses in the Krossfjorden and the Kongsfjorden (adapted from Cottier et al., 2005).

Water mass in Krossfjorden	Temperature (T)	Salinity (S)
Internal water masses		
Surface water mass (SW)	>1	<34
Intermediate water mass (IW)	>1	34 < S < 34.65
Local water mass (LW)	<1	>34.3
External water masses		
Atlantic water mass (AW)	>3	>34.65
Transformed Atlantic water mass (TAW)	1 < T > 3 °C	>34.65

the surface area of the fjord, 267.81 km² (Prominska et al., 2017) with the FWC.

2.5. ROMS model

The hydrography and the circulation in the Krossfjorden and the Kongsfjorden were simulated using the three-dimensional numerical model- Regional Ocean Modeling System (ROMS) version 3.6 (Shchepetkin and McWilliams, 2003, 2005, 2009). The model domain is comprised of the two fjords and their adjacent shelf-slope region (Fig. 1). The open boundary at the west was extended to the deep ocean to incorporate the WSC, the primary oceanic heat source of the two fjords. The horizontal resolution was set as ~500 m. The model grid was prepared using the 'romspre-processing toolbox' gridding package developed by ROMS-AGRIF (ftp://ftp.legos.obs-mip.fr/pub/romsagrif/DA TA_ROMS/user_manual/doc_roms_agrif_v2.1_19_07_2010.pdf). The model topography was obtained from the Shuttle Radar Topography Mission (SRTM) 30+ bathymetry data with a resolution of 30 arc seconds. This bathymetry was developed by merging satellite-derived and ship acoustic sounding bathymetry (Becker et al., 2009). SRTM data incorporated the Arctic bathymetry from the International Bathymetric Chart of the Ocean (IBCAO). Forty vertical sigma layers were used in the model with high-resolution ensured in the upper water column (~O(10 m)) and low resolution towards the deeper ocean.

Hourly wind and atmospheric flux at the surface extracted from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA5 reanalysis data [https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels?tab=overview] were used to force the model. The boundary and initial conditions were given from the high-resolution (1/12°) Hybrid Coordinate Ocean Model (HYCOM) and Navy Coupled Ocean Data Assimilation (NCODA) global 1/12° reanalysis product (https://hycom.org/data/glb08pt08/expt-19pt1). The daily sea level, temperature, salinity, and velocity components were given at three open boundaries at the north, west, and south. Glacier discharge from the five tidewater glaciers in the

Kongsfjorden was given from the subsurface snow model output (Pramanik et al., 2018) as a point source in the model. Since the snow model output is available only for the tidewater glaciers around the Kongsfjorden, the discharge was set zero in the Krossfjorden. Nonlocal K-Profile Parameterization (KPP) mixing scheme (Large and Gent, 1999) was used to parameterize the sub-grid scale vertical mixing with background eddy viscosity of 10⁻⁵ m²s⁻¹ and diffusivity of 10⁻⁶ m²s⁻¹. Horizontal mixing in the model was parameterized by Smagorinsky type viscosity and diffusion schemes (Muschinski, 1996; Griffies et al., 2000). The background eddy viscosity and diffusivity were set as 3 m²s⁻¹ and 10 m²s⁻¹, respectively. The model run was carried out for the years 2002–2018. The model output for the period 2013–2018 was used for the analysis.

3. Results

3.1. Hydrography

The CTD profiles during the summer-fall of 2013–2018 were analyzed to understand the hydrography distribution and its variability in the Krossfjorden and the Kongsfjorden. Fig. 2 shows the water temperature, salinity, and density averaged over each fjord every month from June to October. The 3 m water temperature (hereafter mentioned as near-surface water temperature) in the Krossfjorden varied between 3 and 6.8 °C during the summer of 2013–2018. The highest near-surface temperature was recorded in August 2015 and the lowest in September 2017. This result is consistent with earlier studies that the summertime (June–September) surface temperature in the fjord generally increases from June, peaks in July–August, and decreases (Cottier et al., 2005; Hegseth and Tverberg, 2013; Sundfjord et al., 2017). The mean vertical water column temperature in the Krossfjorden varied between 0 and 7 °C.

The onset of summer increases the surface water temperature in the fjord and gradually heats the lower water column from above (Sundfjord et al., 2017). This results in warmer less-saline waters over colder high-saline waters. As the melting of the marine-terminating glaciers increases in the summer (Svendsen et al., 2002), the freshwater discharged into the fjord surface forms a thin cold freshwater layer over the existing warm upper layer resulting in a temperature inversion. The temperature inversions with a warm subsurface layer between the colder surface and deep waters were observed in the Krossfjorden from August onwards in all the years except 2018. The inversion was the strongest in August and September 2017 compared to that in the other observational periods. The temperature inversion layer deepened further compared to the respective summer months with the inversions in the fall season (October in 2015 and 2016). There was no temperature inversion in 2018. The 3 m salinity (hereafter mentioned as near-surface

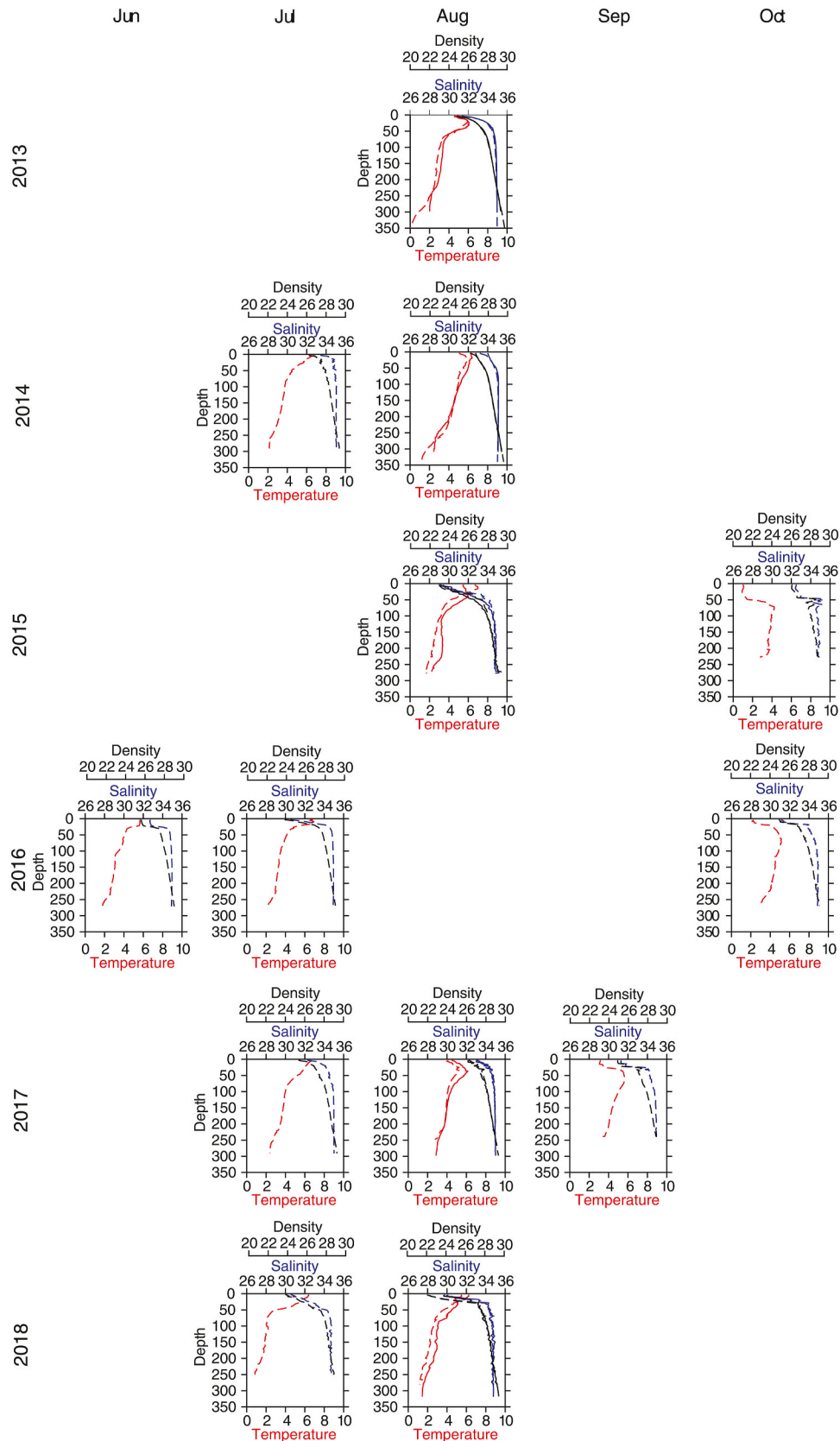


Fig. 2. The monthly fjord averaged vertical water temperature in °C (red), salinity (blue), and sigma-t in kg/m^3 (black) in the Krossfjorden-Kongsfjorden for the period 2013 to 2018. The dashed lines represent the Krossfjorden profiles, and the continuous lines in August diagrams represent the Kongsfjorden profiles. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

salinity) varied between 28.0 and 33.3 (Fig. 2). Vertically, the salinity varied between 28.0 and 35.0, with the minimum always observed at the surface. The salinity below 100 m was always higher than 34.6, with not much variability as in the upper layer. The water column density closely followed the variations in the salinity (Fig. 2).

The inter-annual variability in the Kongsfjorden and its comparison to the Krossfjorden was done for August (Fig. 2). The mean vertical water temperature in the Kongsfjorden was ~ 0.1 – 1 °C warmer than in the Krossfjorden. The exceptions are the upper 30 m in 2015 and 2018 that was ~ 1.5 °C and 1 °C warmer, respectively, in the Krossfjorden. The temperature inversions were observed in the Kongsfjorden with the deepest and strongest inversion in August 2017, as was found in the Krossfjorden. The inversion layer in the Kongsfjorden was generally deeper than that in the Krossfjorden. It was 40 m in the Kongsfjorden and 30 m in the Krossfjorden in August 2017. However, in 2017, the temperature increased at the same rate of 0.04 °C/m within the inversion layer in both the fjords. No temperature inversion was found in August 2018, as was in the Krossfjorden. The salinity difference between the fjords was significantly less, especially below 100 m (Fig. 2). However, the upper 15–20 m in August of 2014 and 2018 was more saline (1–2) in the Kongsfjorden than in the Krossfjorden. The density differences between the fjords were closely reflecting the variations observed in the salinity.

3.2. Mixed layer depth

The generally deep winter MLD in the Kongsfjorden and the Krossfjorden shallows during the summer, due to the upper water column stratification induced by the surface melt-water discharge into the fjords (Svendsen et al., 2002). In the present data, the MLD was 30 m in the Krossfjorden in June 2016 (Table 4). Over the entire observational period, a very shallow MLD in the range of 4–7 m was seen in July. Deeper MLD in the range of 8–33 m was found in August and September. The MLD further deepened in fall than the respective summers. Hence a shallow mixed layer characterizes the Krossfjorden in the early summer and a gradual deepening that resulted in a deeper mixed layer in the late summer and fall. The MLD showed large inter-annual variability in both the fjords (Table 4). The Krossfjorden MLD in August varied between 8 and 33 m from 2013 to 2018. The deepest MLD was in 2015 and the shallowest in 2013. In August, the Kongsfjorden MLD varied between 6 and 18 m, except in 2014 when it was 50 m. The MLD was the shallowest in August 2013 as it was in the Krossfjorden. In brief, the MLD in the Kongsfjorden was 2–23 m shallower than that in the Krossfjorden, except in August 2014, where the Kongsfjorden MLD was deeper.

3.3. Spatial distribution of hydrography

The western Krossfjorden is reported with large calving (Den Norske Los, 1992) and more freshwater discharge (Ingvaldsen et al., 2001), making it colder compared with the eastern Krossfjorden. Hence, we examined the spatial water temperature (Fig. 3) and salinity (Fig. 4) distributions along the western and eastern coasts of the Krossfjorden.

In August 2013, the upper 60 m eastern Krossfjorden was 0.5 – 1.0 °C warmer than the western side (Fig. 3 (a)). But the salinity (Fig. 4 (a)) did not show any difference. A similar temperature difference between the eastern and western Krossfjorden was observed in 2014 in the entire water column (Fig. 3 (b)). As it was in the previous years, the eastern fjord was 1 – 1.5 °C warmer than its western side (5 – 5.5 °C) in August 2015 also (Fig. 3 (c)). The upper 25 m water column in the Krossfjorden was colder in 2017 compared to the earlier observations (Fig. 3 (d)). However, the temperature contrast between both sides was present with robust cooling in the western Krossfjorden. The increase in the upper water temperature from the northern (head) to southern (mouth) end along the western transect is also evident in 2017. The upper 50 m salinity in August 2017 (Fig. 4 (d)) also showed a prominent freshening in the western Krossfjorden compared to the eastern part. The core of the

Table 4
Monthly variations of the Mixed Layer Depth (MLD, m), Surface water mass (SW) characteristics (thickness in m, maximum temperature (T_{max}) in °C, and minimum salinity (S_{min})), Atlantic water mass (AW) characteristics (thickness in m, maximum temperature (T_{max}) in °C, and maximum salinity (S_{max})), and freshwater volume (FWV) in km^3 from 2013 to 2018 in the Kongsfjorden and the Krossfjorden.

Year	Months	Krossfjorden						Kongsfjorden										
		MLD (m)	Thickness of SW	Thickness (m)	T _{max} of SW (°C)	S _{min} of SW	Thickness of AW	Thickness (m)	T _{max} of AW (°C)	S _{max} of AW	FWV (km ³)	MLD (m)	Thickness of SW	Thickness (m)	T _{max} of SW (°C)	S _{min} of SW	Thickness of AW	T _{max} of AW (°C)
2013	August	8	60	6.8	6.8	29.6	75	7.2	34.8	0.19	6	50	6.9	30.2	100	7.5	35.0	0.3
2014	July	4	25	7.4	7.4	32.5	200	6.6	35.3	0.02	50	25	6.3	33.6	200	6.5	35.2	0.11
2015	August	9	25	6.2	6.2	32.0	200	6.5	35.2	0.08	10	100	7.0	26.5	50	3.9	35.0	1.45
	August	33	90	7.6	7.6	26.7	20	3.8	35.0	0.71								
2016	October	49	50	3.0	3.0	31.7	230	4.6	35.3	0.33								
	June	30	50	6.6	6.6	31.5	110	4.5	35.0	0.16								
2017	July	6	30	8.0	8.0	29.4	150	6.0	35.0	0.16								
	October	18	40	4.8	4.8	28.9	175	5.5	35.1	0.26								
2018	July	7	50	7.3	7.3	30.7	170	5.6	35.1	0.13								
	August	16	30	4.8	4.8	31.4	230	6.0	35.8	0.09	11	30	6.0	31.2	230	6.0	35.9	0.21
2019	September	27	70	5.5	5.5	29.9	0.35	200	5.5	0.35								
	July	4	40	7.4	7.4	29.3	0.43	No AW	No AW	0.43								
2020	August	29	50	8.0	8.0	26.1	20	4.4	35.1	0.64	18	60	7.0	28.4	35	4.8	35.4	0.64

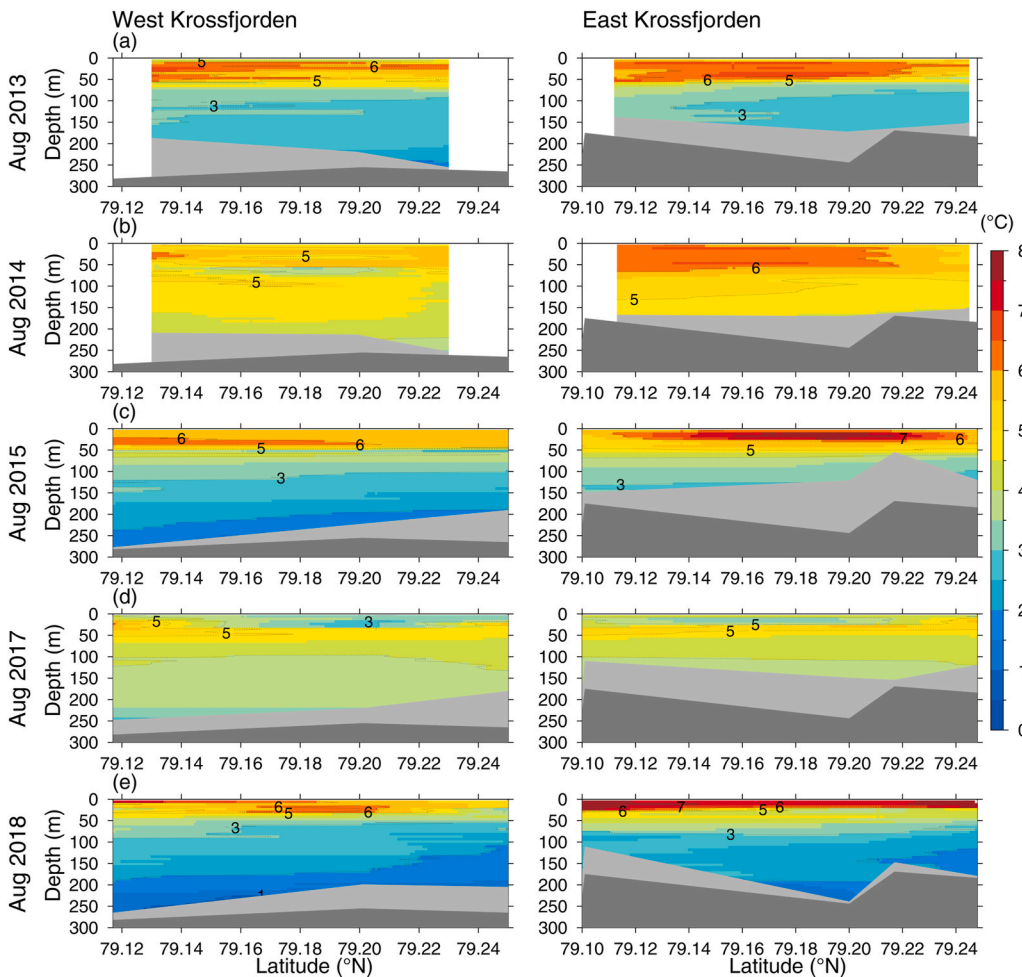


Fig. 3. The monthly spatial distribution of the water column temperature ($^{\circ}\text{C}$) along the western (left) and eastern (right) Krossfjorden in August from 2013 to 2018. Lower to higher latitudes in the x-axis corresponds to the direction in each side of the Krossfjorden from its outer to inner fjord area. The contour lines are the isotherms. The dark grey shade in the plots indicates the CTD location depth, and the light grey shade represents no data region. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

upper layer warming was up to 15 m and 50 m in the western and eastern sides, respectively, in August 2018, indicating a warmer upper water column in the east Krossfjorden (Fig. 3 (e)). Both the eastern and western Krossfjorden had colder deep waters than the previous years. Hence the east Krossfjorden was 0.5–1.5 $^{\circ}\text{C}$ warmer than the west Krossfjorden in August from 2013 to 2018.

Fig. 5 shows the spatial distribution of water column temperature and salinity along the Kongsfjorden. The CTD profiles were not available at the outer and innermost locations of the Kongsfjorden in 2013 (Fig. 5 (a), (b)) and 2014 (Fig. 5(c), (d)). In 2015, the CTD profiles were not available in the outermost locations of the Kongsfjorden (Fig. 5 (e), (f)). In August 2013, a core of maximum water column temperature was found around 30–40 m (Fig. 5 (a)) with cooler waters above and below it, as was in the Krossfjorden (Fig. 3 (a)), which is the temperature inversion discussed in the hydrography section. The isohaline of 34.5 was ~ 30 m shallower (Fig. 5 (b)) than that in the Krossfjorden (Fig. 4 (a)), indicating more volume of high saline deep waters in the Kongsfjorden. Compared to 2013, the water column, especially below 50 m, was warmer in 2014 (Fig. 5 (c)) while the upper 50 m was fresher (Fig. 5 (d)). The spatial distribution of the water temperature (Fig. 5 (c)) and the salinity in 2014 (Fig. 5 (d)) was very similar to that found in the eastern Krossfjorden (Figs. 3 (b), 4 (b)). However, the temperatures in the upper Kongsfjorden were 0.5–1 $^{\circ}\text{C}$ higher than that in the Krossfjorden (Fig. 3 (b)). But the salinity distribution in the Kongsfjorden (Fig. 5 (d)) and the Krossfjorden (Fig. 4 (b)) were similar.

The Kongsfjorden showed a temperature inversion, as discussed in the hydrography, in 2015 (Fig. 5 (e)) centered around 30 m (6.5–7 $^{\circ}\text{C}$), similar to that found in the eastern Krossfjorden (Fig. 3 (c)). The upper

layer salinity was lower in 2015 (Fig. 5 (f)) for both the fjords than in 2013 and 2014, as stated in the hydrography section. The head to mouth lateral gradient in the temperature (Fig. 5 (g)) and salinity (Fig. 5 (h)) is seen in the upper 50 m Kongsfjorden in August 2017 as was seen in the Krossfjorden (Fig. 3 (d)). The water column hydrography distribution appeared similar to 2014. As discussed earlier, August 2018 had colder waters than the previous years except 2015, especially below 50 m, in both the fjord systems (Figs. 5 (i), 3 (e)). The spatial distributions in August 2018 (Fig. 5 (i), (j)) appeared similar to 2015 (Fig. 5 (e), (f)).

3.4. Water masses

The SW, LW, IW, TAW, and AW were found in the Krossfjorden during the observational period (Fig. 6 (a), (c), (e), (g), (i)). In August 2013, the Krossfjorden had SW, IW, AW and TAW (Fig. 6 (a)). Even though not significant, a very thin layer with remnants of LW was also found during the period. In August 2014, AW became prominent besides the SW and IW (Fig. 6 (c)). The AW was found at a depth range between 50 and 250 m. In 2015, the SW was warmer and fresher than in 2013 and 2014 (Fig. 6 (e), Table 4). Moreover, the SW layer was deeper than the earlier two years. TAW was prominent in 2015 with a very thin AW layer (Table 4). The SW in August 2017 was colder than the previous years (Fig. 6 (g)). The AW dominated the fjord in August 2017 as it was in 2014. The SW was the warmest and freshest in August 2018 among the observational years (Fig. 6 (i)). The TAW was dominant again in 2018 as it was in 2015.

The SW, IW, TAW, and AW were found in the Kongsfjorden during the observational period (Fig. 6 (b), (d), (f), (h), (j)). The SW layer

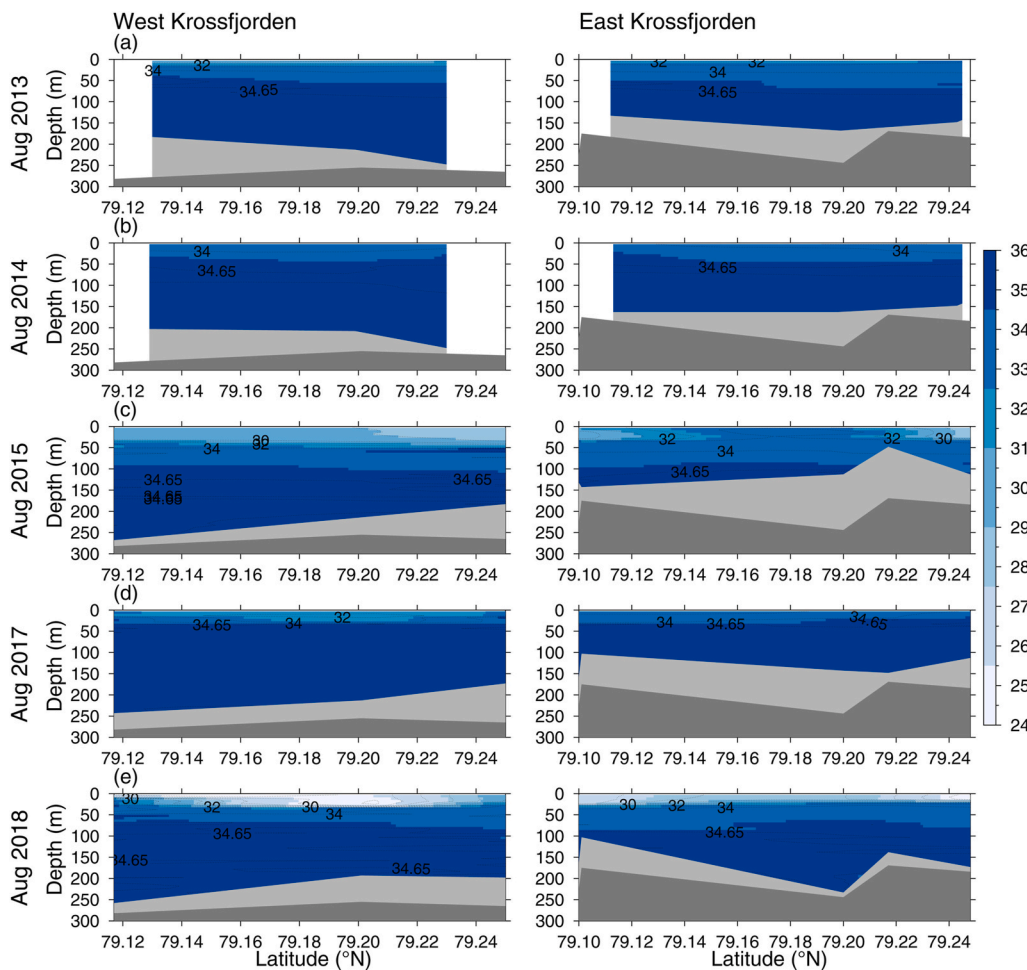


Fig. 4. The monthly spatial distribution of the water column salinity along the western (left) and eastern (right) Krossfjorden in August from 2013 to 2018. Lower to higher latitudes in the x-axis correspond to the direction in each side of the Krossfjorden from its outer to inner fjord area. The contour lines are the isohalines. The dark grey shade in the plots indicates the CTD location depth, and the light grey shade represents no data region. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

thickness was the same (2013, 2014, and 2017) or higher (2015 and 2018) in the Kongsfjorden than that in the Krossfjorden (Table 4). The SW, IW, AW, and TAW were found in 2013 (Fig. 6 (b)). But, unlike the Krossfjorden, LW was not found in the Kongsfjorden. As the CTD profiles in the Kongsfjorden were not available from the outermost and innermost locations in 2013 and 2014, we cannot conclude whether LW was there in the innermost locations of the fjord in 2013. The AW was dominant again in 2014 (Fig. 6 (d)) and the TAW in 2015 (Fig. 6 (f)). The AW became prominent again in 2017 (Fig. 6 (h)). In 2018, the SW was warmer and saline (Fig. 6 (j)) than that was in the Krossfjorden (Fig. 6 (i)). TAW became dominant again in 2018. The inter-annual variability in the dominance of AW and TAW had the same pattern inside the Kongsfjorden as it was in the Krossfjorden (Table 4). In both the fjords, the AW was seen prominent when the SW thickness was thinner, and the TAW was prominent when the SW thickness was thicker.

3.5. Freshwater volume

The FWV in the Krossfjorden is in the range of 0.02–0.71 km³ for the entire period of observations from June to October while it is in the range of 0.08–0.71 km³ (range of FWC is 0.2–4.5 m) in August, during the period from 2013 to 2018 (Table 4). The range of FWV in the Kongsfjorden is 0.11–1.45 km³ (range of FWC is 0.4–5.4 m) (Table 4). The FWV in the Kongsfjorden is 1.4–2.0 times more than that in the Krossfjorden except in 2018 when it was the same in both the fjords. The inter-annual variability in the FWV corresponds with the variations in the thickness of the SW layer. The FWV is the largest in August 2015, when the SW layer is the deepest in the fjords. On the other hand, the FWV is the smallest in 2014 when the SW layer is very thin.

4. Discussion

4.1. Evolution of the Krossfjorden hydrography

The present study shows a two-layer structure water column in the early summer (June–July) in the Krossfjorden with warm less-saline surface waters above the cold high-saline waters below it (Fig. 2). A well-mixed surface layer of ~30 m in June was followed by a seasonal pycnocline that acted as a barrier for mixing between the surface and deep waters. This mixed upper layer shallowed in July. It is reported (Svendsen et al., 2002; Cottier et al., 2005) that the freshwater discharge in the Krossfjorden-Kongsfjorden would be maximum in the peak summer months (August–September) from the snow melting, surface, and submarine melting of glaciers, calving, river runoff, groundwater discharge, and precipitation. This freshwater discharged into the fjord surface forms an additional thin cold less-saline layer over the existing warm uppermost layer in the two-layer system. Thus the two-layer water column in the Krossfjorden transforms to a three-layer system from August onwards. Warm high-saline waters were found between the cold less-saline surface and the cold high-saline deep waters balancing the density stratification. This set-up of warmer water temperatures between the colder waters up and down leads to the temperature inversion layer in the fjord. Therefore, a temperature inversion is a significant component in the three-layer fjord structure. This was also observed in the fall season. The climatology at Ny-Ålesund shows a general drop in the air temperature and increase in the winds (Cisek et al., 2017) along with the net surface heat flux loss (Maturilli et al., 2015) and increased precipitation (Osuch and Wawrzyniak, 2016) from August to October. These factors lead to the upper layer cooling that converts the two-layer

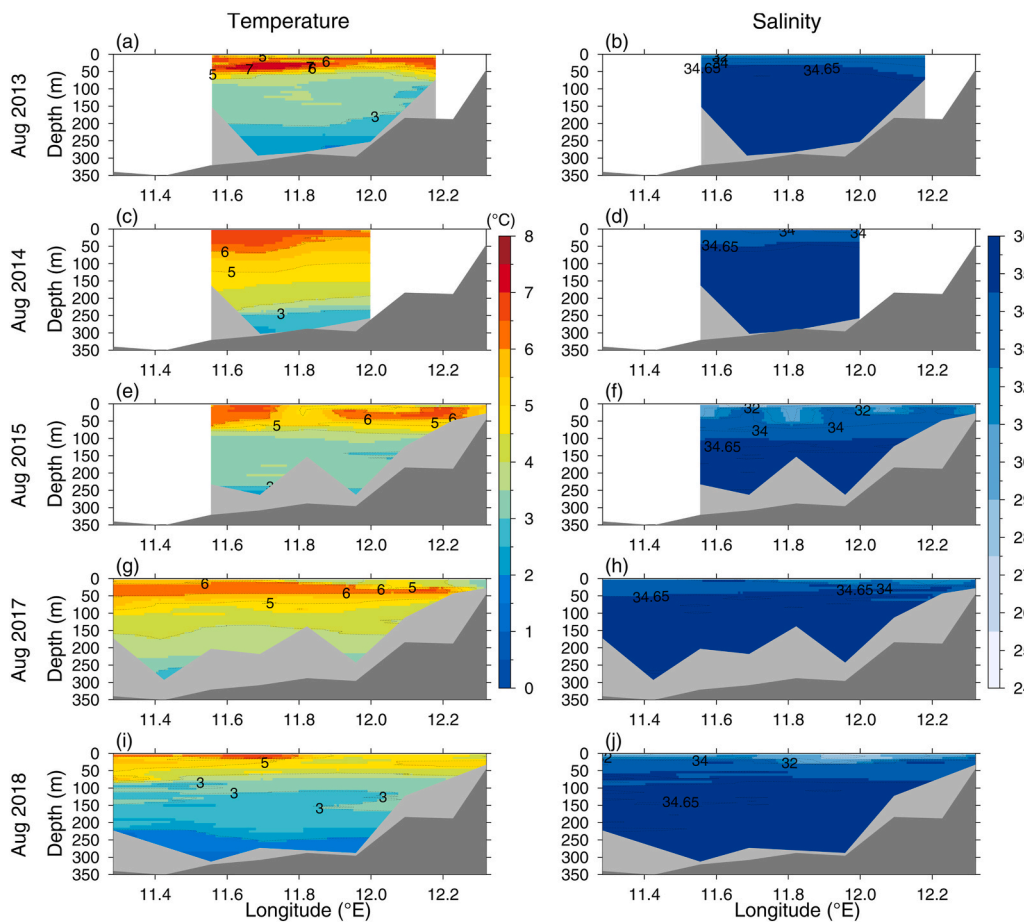


Fig. 5. The monthly spatial distribution of the water column (left) temperature and (right) salinity along the Kongsfjorden in August from 2013 to 2018. Lower to higher longitudes in the x-axis corresponds the direction from its outer to inner fjord area. The contour lines in the temperature diagrams are the isotherms, and in the salinity diagrams are the isohalines. The dark grey shade in the plots indicates the CTD location depth and the light grey shade represents no data region. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

fjord structure to the three-layer system.

4.2. Heat flow and circulation in the Krossfjorden

The spatial distribution of temperature-salinity in the Krossfjorden in August showed that the eastern fjord was warmer (Fig. 4) and saline (Fig. 5) than its western side. Since there is no current data available in the Krossfjorden, we used ROMS model simulation to understand the fjord circulation. Since the inter-annual variations are investigated in the present study during August, and also it being one of the peak summer months in the fjords with high Atlantic water intrusions (Svendsen et al., 2002; Tverberg et al., 2019), we analyzed the model output for August month during 2013–2018. The simulated temperature, circulation pattern, and the Atlantic water inflows closely follow the earlier model simulations (Ingvaldsen et al., 2001; Svendsen et al., 2002; Cottier et al., 2005; Arild et al., 2017) even without the freshwater discharge information in the Krossfjorden. Even though the inter-annual variations existed in the magnitude of the simulated ocean temperature and currents during 2013–2018, the general pattern of the heat flow and circulation in the study region remained the same throughout the simulations. Hence we made a climatology of the ocean temperature and current simulations for August 2013–2018 (Fig. 7).

The warm Atlantic waters from the WSC enter the Kongsfjorden along its southern shore, while its direct entry to the Krossfjorden happens through its eastern side (Fig. 7). This flow through the eastern Krossfjorden supplies heat to the inner Krossfjorden. The flow is seen reaching the glacier at the head of the western part of the fjord, Lillehökbreen, which can further result in glacial calving and melting. There is an outward flow along the western Krossfjorden which can cause the freshwater outflow from the fjord to the open ocean area. The

cold waters from the inner Kongsfjorden flow out of the fjord through its northern boundary. A part of this outflow enters the Krossfjorden through its eastern side along with the direct Atlantic water entry into the fjord.

4.3. AW in the fjords

The AW was generally observed below ~50 m in the fjords during the observational period (Fig. 6). The AW layer thickness was 15–30 m more in the Kongsfjorden than in the Krossfjorden during most of the observational period (Table 4), indicating more Atlantic water intrusions into the Kongsfjorden. An earlier comparison of the Kongsfjorden with the Hornsund (Prominska et al., 2017), a fjord further south of it, also showed the Kongsfjorden as the warmer and more saline. The Kongsfjorden had approximately two times higher inflow of Atlantic water than the Hornsund. The weaker frontal formation at the mouth, the wider and deeper trough connected to the WSC, and the lesser freshwater content than the Hornsund were pointed out as the reasons for more AW inside the Kongsfjorden than that in the Hornsund (Prominska et al., 2017). However, the difference in the AW volume between the Krossfjorden and the Kongsfjorden (Table 4) is not as high as that between the Kongsfjorden and the Hornsund. Hence the increased AW observed in the Kongsfjorden compared to that in the Krossfjorden could be due to the fjord being the first point of entry of the Atlantic waters among the twin fjord system with deeper outer fjord bathymetry and higher fjord volume than the Krossfjorden (Svendsen et al., 2002; Prominska et al., 2017). The more heat advection into the Kongsfjorden due to the higher Atlantic water intrusions than the other Svalbard fjords gathered more scientific attention towards this fjord. The S_{max} of AW showed a gradual increase from 2013 to 2018 in both the

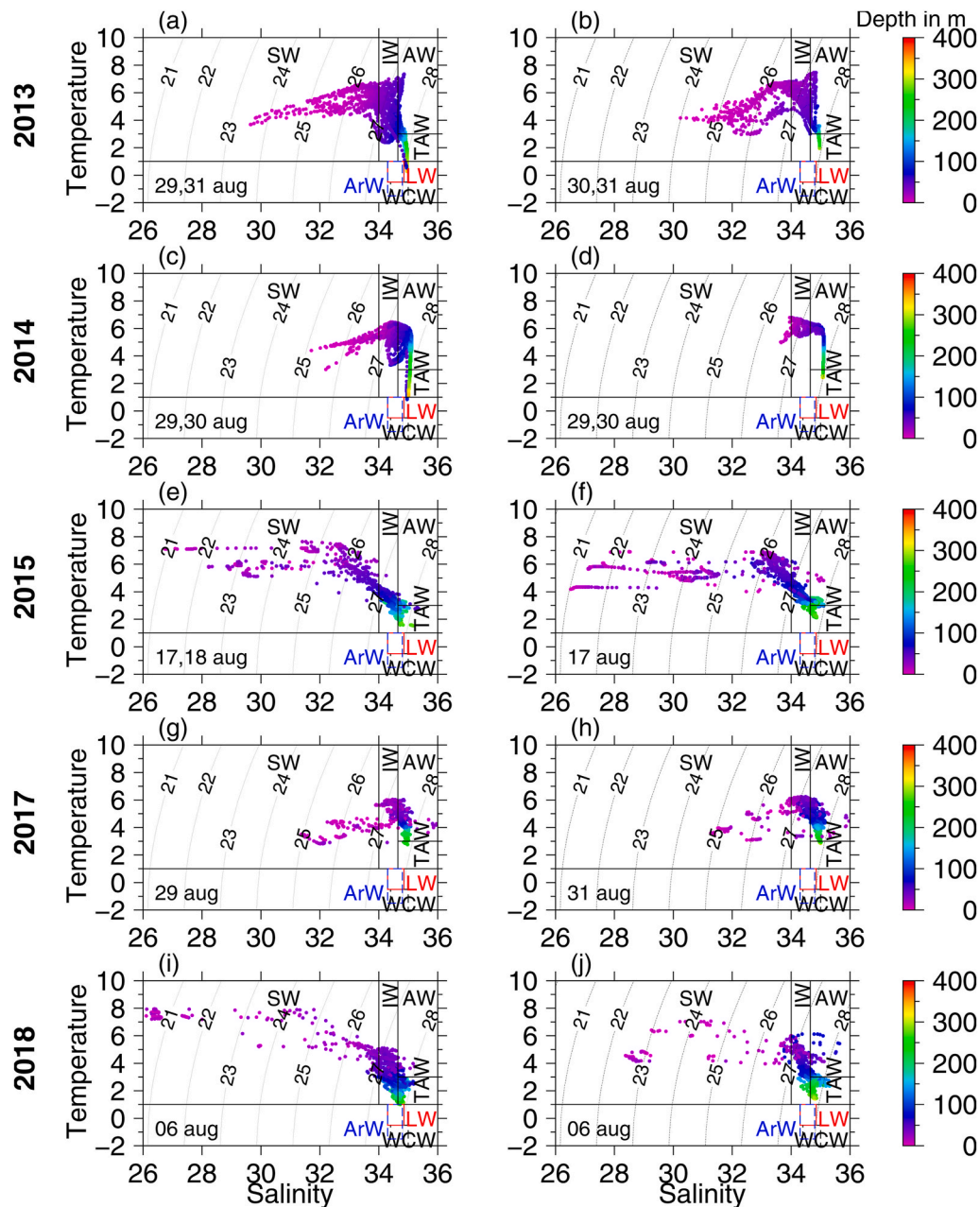


Fig. 6. The Temperature (°C) - Salinity diagrams in August from 2013 to 2018 in the (left) Krossfjorden and (right) Kongsfjorden. The colour bar denotes the depth scale at which the water masses existed. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

fjords with an inter-annual variation embedded while the $T_{\max}$ was highly variable. This increase in salinity over the period indicates an increase in the Atlantic origin waters and its extended residence period in the fjord system.

4.4. Inter-annual variability of the Atlantic water

The inter-annual variability of AW and TAW is the same inside both the fjords, with the dominance of AW in 2013, 2014, and 2017, and TAW in 2015 and 2018 (Table 4). Prominska et al. (2017) also reported the dominance of AW in 2014 and TAW in 2015 inside both the Kongsfjorden and the Hornsund, indicating that all these three fjords in the west Svalbard had the same inter-annual variations in the Atlantic origin waters inside it. Moreover, the AW in summer 2014 was found at 25–250 m in the Kongsfjorden in their study, which corresponds with the 200 m thickness of AW that we found in our observations in the

Kongsfjorden. 2014 summer has been characterized as an exceptional year over the analyzed period from 2001 to 2015 as large areas of the fjords were filled with pure Atlantic water (Prominska et al., 2017). In the present study, the entire water column in 2014 and 2017 in the fjords showed similar hydrography with the dominance of AW. The dominance of the AW indicates the increased inflow of the Atlantic water into the fjords and is studied extensively in the west Svalbard fjords. This can be due to the northerly winds resulting in an increased cross-shelf exchange of the Atlantic water (Cottier et al., 2005), pushing away the coastal waters and replacing them with the Atlantic water at the deeper levels (Cottier et al., 2007). This can also be caused by the wind curl caused by the southerly wind events along the WSC that add relative vorticity to the current, forcing it up the slope and topographic steering leading to the geostrophic advection of the Atlantic water into Kongsfjorden (Nilsen et al., 2016).

Another notable observation is that the TAW was prominent when

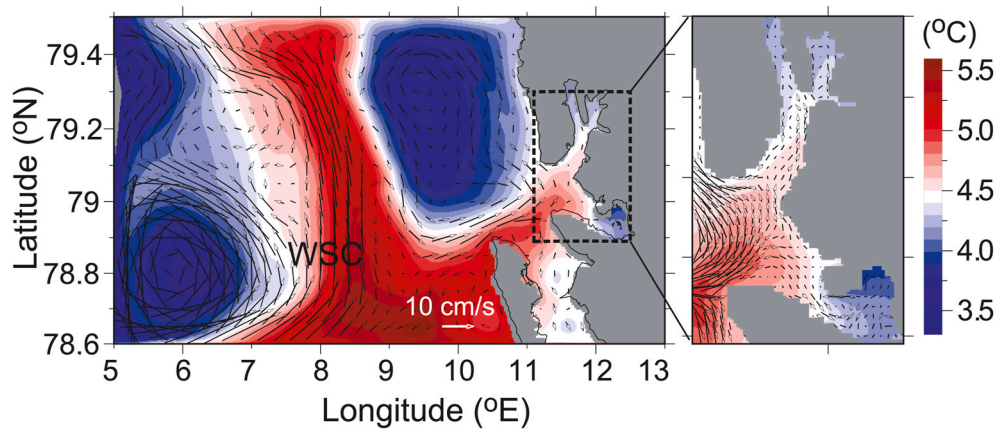


Fig. 7. The ROMS simulation of the ocean temperature in °C (shade) overlaid by the current vectors (cm s^{-1}) that are depth averaged for 0–150 m in August from 2013 to 2018 in the study region. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

the SW layer was deeper and the AW was prominent when the SW layer was thinner (Table 4). This indicates a compensation mechanism between the opposing surface water and deep Atlantic water flows in the Krossfjorden and the Kongsfjorden during the observational period. The mechanism has been studied in detail in the Kongsfjorden. Local down-fjord winds blowing along the fjord axis move the surface layer of water. In Kongsfjorden, the wind blowing out of the fjord results in the freshwater concentration along the northern shore and its flow out of the fjord (Ingvaldsen et al., 2001; Sundfjord et al., 2017). Such outflow would be compensated by the inflow of more saline, warmer Atlantic water over the bottom, along the southern shore into the fjord (Moffat, 2014; David and Krishnan, 2017). This circulation may increase the glacier melting and the freshwater discharge, further enhancing the fjord water exchange. When the SW layer thickness is more inside the fjord, it indicates a lesser inflow of the Atlantic water into the fjord and, hence, the formation of TAW. When the SW layer thickness is thinner inside the fjord, it indicates more Atlantic water inflow into the fjord and, thus, the formation of AW. Even during the years of low AW volume, such as 2015 and 2018, the Kongsfjorden had more AW than the Krossfjorden (Table 4). The lower layer water column in the Krossfjorden was cooler than that in the Kongsfjorden due to the lesser volume of AW and lesser heat delivery inside the Krossfjorden than in the Kongsfjorden.

4.5. Freshwater volume in the fjords

The FWV in the Kongsfjorden was 1.4–2.0 times more than that in the Krossfjorden, except in 2018 when it was the same in both the fjords. The larger FWV in the Kongsfjorden than that in the Krossfjorden indicates more glacial discharge into the Kongsfjorden. This creates a stronger vertical gradient in the temperature and salinity in the upper Kongsfjorden that further results in shallower MLD in the Kongsfjorden compared to that in the Krossfjorden. Svendsen et al. (2002) reported an annual freshwater discharge of 1.4 km^3 into the Kongsfjorden with a strong 30% annual variation. This matches with the maximum FWV estimation in the Kongsfjorden during the period. However, the present Kongsfjorden FWV estimations appear ~ 3.5 times higher than the total summer (July–August–September) FWV reported by Prominska et al. (2017) from 2001 to 2015 ($0.03\text{--}0.42 \text{ km}^3$). The FWV in Krossfjorden is almost in the same range as reported in Hornsund ($0.06\text{--}0.81 \text{ km}^3$) by Prominska et al. (2017).

5. Summary and conclusion

The CTD data in the summer–fall of 2013–2018 in the Kongsfjorden and Krossfjorden were used to understand the hydrography evolution in

the two fjords. The CTD profiles in August, the common month of data available in both the fjords during 2013–2018, had been utilized to examine the inter-annual variability in the fjords and to compare between the fjords.

In the early summer, the Krossfjorden water column had a two-layer structure with warm less-saline surface waters above the cold high-saline waters. In the peak summer and fall, the fjord shifted to a three-layer structure having warm high-saline waters sandwiched between the cold less-saline surface and cold high-saline deep waters. The uppermost two layers in this three-layer system show the temperature inversion. In August, the inter-annual MLD varied between 8 and 33 m in the Krossfjorden and 6–18 m in the Kongsfjorden, indicating a shallower MLD in the Kongsfjorden.

The spatial distribution of hydrography showed a warmer and saline eastern Krossfjorden when compared to the western Krossfjorden. This east-west variability in the Krossfjorden hydrography was examined using the ROMS model simulations. The simulations showed the entrance of the warm Atlantic waters from the WSC through the eastern fjord. The circulation pattern shows the heat supply by these warm waters to the head of the western part of the Krossfjorden, and this may lead to the glacial calving and melting of the glacier-Lillehökbreen. A portion of the outflow from the Kongsfjorden, through its northern side, also enters the Krossfjorden through its eastern side in the direction of the direct Atlantic water entry to the fjord.

The SW layer was generally deeper in the Kongsfjorden compared to that in the Krossfjorden. The estimated FWV in August corresponded with the SW layer thickness in the Kongsfjorden and was 1.4–2.0 times more than $0.08\text{--}0.71 \text{ km}^3$ FWV in the Krossfjorden. The AW layer thickness was more in the Kongsfjorden than in the Krossfjorden, indicating more Atlantic water intrusions into the Kongsfjorden. The large inter-annual variability observed in the volume of the Atlantic waters that entered into the fjords, the dominance of TAW and AW, were the same in both the fjords. The AW dominated in 2013, 2014, and 2017 and the TAW dominated the fjords in 2015 and 2018. When the AW was prominent, the SW layer was thinner in the fjords, and when the TAW was prominent, the SW layer was deeper in the fjords. This is due to a compensation mechanism of the surface and deep flows in the fjord (Ingvaldsen et al., 2001; Sundfjord et al., 2017; Moffat, 2014; David and Krishnan, 2017). The down-fjord wind blowing out of the fjord carries away the surface freshwater layer out of the fjord. This surface flow would be compensated by the inflow of more saline, warmer Atlantic water over the bottom. So the smaller the SW layer thickness, the larger the inflow of the Atlantic waters into the fjord resulting in the formation of AW. In contrast, the opposite conditions result in the formation of TAW.

The study shows the recent hydrographical changes in the

Krossfjorden-Kongsfjorden, the fjord systems in close proximity with almost the same geographical-physical setting. Further, it shows how these fjords responded similarly but with different magnitude to the Atlantic water inflows and freshwater discharge from the glacial melting over five years. The results can benefit in understanding the Krossfjorden circulation and other processes associated with the heat supply to the inner Krossfjorden, such as the glacier melting. The updated observational information can further add to the understanding of the physical and biogeochemical processes in the west Svalbard fjords. In addition to this, the inferences will help address modeling challenges in the region and the coherent depiction of the dynamic Arctic fjord processes.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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