

Faculté des sciences

Impact of the sea ice accelerated melting on the velocity structure of the East Greenland Current

Mémoire présenté en vue de l'obtention du grade académique de Master [120] en sciences physiques, finalité approfondie.

Author : Lambotte Augustin

Supervisors : Fichet Thierry,

Ponsoni Leandro (Flanders Marine Institute - VLIZ)

Readers : Massonnet François, Goosse Hugues

School of Physics, ELIC

Academic year 2023-2024

Abstract

The East Greenland Current (EGC) flows along the east coast of Greenland, connecting the Arctic and Atlantic Oceans. This current is characterized by zonal density gradients, indicative of a baroclinic component. The EGC transports sea ice from the Arctic Ocean, which melts along its path, releasing low salinity water at the ocean's surface. With ongoing global warming, which is particularly pronounced in the Arctic region, the extent of sea ice has experienced a substantial reduction. In this master's thesis, we examine the impact of accelerated sea ice melting on the velocity structure of the EGC. We investigate how changes in sea ice melt alter the release of low salinity water at the ocean's surface and consequently affect the baroclinic component of the EGC. Utilizing the Ocean Reanalysis System v5 (ORAS5) from the European Centre for Medium-Range Weather Forecasts, we first validate this dataset against various observational data over the period from January 2011 to December 2018. During this period, we observe a weakening of the northern part of the EGC and a strengthening in the south. By computing zonal density gradients and employing the dynamical method, we demonstrate that these observed changes are partially attributed to variations in the baroclinic component, particularly evident in the northern region. We find that alterations in density profiles are due to an increase in salinity along the east coast of Greenland. Furthermore, we present evidence suggesting a potential causal relationship between the decline in sea ice melt over the region and the subsequent rise in salinity. Finally, we establish a correlation between decreasing sea ice transport through Fram Strait and diminishing sea ice melt in the EGC region, indicating a foreseeable trajectory for its future evolution.

Acknowledgments

I would like to personally thank some people for their support, without which this work would not have been possible.

First and foremost, I express my deepest gratitude to my advisors, Professor Fichet and Doctor Ponsoni, for their invaluable guidance throughout these ten months of work. Thank you for your consistent and insightful recommendations and feedback, as well as for continually pushing me further. Your support was crucial to the success of this project. Furthermore, the positive atmosphere of our meetings and discussions made this work truly enjoyable. I am grateful for the opportunity to continue working with you, Professor Fichet, and I look forward to future collaborations with you, Doctor Ponsoni.

Thank you to my climate physics classmates, Amaury, Elise, and Thomas, for all the enjoyable moments we shared during these two years.

I also extend my thanks to my parents for their support throughout this past year and during my entire university journey. You provided me with the best study conditions and have supported me throughout all these years. Additionally, thank you to my big brother, who was definitely one of the reasons I began these studies.

Finally, thank you Éléonore for always being by my side and making all these hours of work so pleasant.

Contents

List of figures	x
List of tables	xi
List of aronyms	xiii
Introduction	1
1 The East Greenland Current	5
1.1 A brief history	5
1.2 Structure, drivers and variability	6
1.3 Sea ice in the EGC region	9
1.4 Recent changes	10
2 Data used in this study	13
2.1 ORAS5	13
2.1.1 Derived variables	14
2.2 Observational data	17
2.2.1 CS2	17
2.2.2 OSI SAF	18
2.2.3 Copernicus	19
3 Validation of ORAS5	21
3.1 Sea ice	21
3.1.1 Thickness, concentration, volume, and extent	21
3.1.2 Transport	23
3.2 Ocean	28
3.2.1 Currents	28
3.2.2 Temperature and salinity distribution	29
3.3 Summary and discussion	33
4 Results	37
4.1 Changes in the EGC velocity structure	38
4.1.1 Summary	39
4.2 Changes in the EGC drivers	41
4.2.1 Zonal density gradients	41
4.2.2 Quantitative analysis with the dynamical method	41
4.2.3 Wind stress	44
4.2.4 Summary	46
4.3 Role of the sea ice in the ocean density field perturbation	47
4.3.1 Salinity as driver of the density field changes	47
4.3.2 Sea ice trends	49

4.3.3	Correlation between SIM and SSS	54
4.3.4	SIM and depth integrated salinity changes: quantitative as- essment	55
4.3.5	Confounding variables?	59
4.3.6	Summary	62
4.4	Future evolution of SIM in the EGC region	64
	Conclusion	67
	References	71

List of Figures

1.1	Petermann’s chart of the Arctic. Figure from Tammiksaar et al. (1999)	6
1.2	Nordic Seas bathymetry and main currents. Figure from Isachsen et al. (2014)	7
2.1	Time line of different forcings and assimilation data used in ORAS5, figure from (Zuo et al., 2019).	14
3.1	SIT averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and CS2. Right panels: the difference for the two months between ORAS5 and CS2.	23
3.2	SIC averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and CS2. Right panels: the difference for the two months between ORAS5 and CS2.	24
3.3	Mean SIT, mean SIC, SIV and SIE for both ORAS5 (red) and CS2/OSI SAF (blue). The dashed line represents the average over 8 years and the vertical bars the standard deviations. The area over which the SIT and SIC were averaged and SIE and SIV calculated is displayed in red on the inset map.	24
3.4	SID averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and OSI SAF. Right panels: the difference for the two months between ORAS5 and OSI SAF.	26
3.5	Comparison of Fram Strait sea ice transport between ORAS5 - Min et al. (2019) for M1 and M2 - and Ricker et al. (2018) for R. Upper: The monthly time series. Below: The seasonal cycle calculated by averaging each month between 2011 and 2017. Error bars indicate standard deviation over the 6 years. The gate used for the calculation is displayed in red in the inset map.	27
3.6	ORAS5 ocean surface velocity averaged between 2011 and the end of 2018. Left: March. Right: September.	29
3.7	Four vertical transects of the mean southward velocity component magnitude. The transect location is indicated in the inset map.	30
3.8	Four vertical transects of the summer (July, August and September) mean southward velocity component magnitude. The transect location is indicated in the inset map.	30
3.9	Four vertical transects of the winter (January, February and March) mean southward velocity component magnitude. The transect location is indicated in the inset map.	31

3.10	Bathymetry of the Nordic Seas. The Blosseville basin is delimited in red.	31
3.11	T-S diagram from ORAS5 at 78.5°N and 3°W. All: averaged over each month between 2011 and the end of 2018. Winter: January, February, and March. Summer: July, August, and September. The colour indicates the depth.	34
3.12	SSS climatological mean (between 2011 and the end of 2018) map for the months of March and September from ORAS5 and Copernicus. Right panels: the difference for the two months between ORAS5 and Copernicus.	34
3.13	SSD anomaly climatological mean (between 2011 and the end of 2018) map for the months of March and September from ORAS5 and Copernicus. Right panels: the difference for the two months between ORAS5 and Copernicus.	35
4.1	Upper left: Trend of KE. Upper right: Trend of EKE. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. Bottom: Scatter plot of the significant KE trends and EKE trends (both significant and non-significant).	40
4.2	Trends of the southward velocity component between 2011 and the end of 2018 along four transects spanning the EGC path. Positive (negative) trends indicate a strengthening (weakening) of the southward flow. Statistically significant trends are marked with a dot. . . .	40
4.3	Left: Map of the KE trend between 2011 and the end of 2018 with the 5 transects (T1, T2, T3, T4, T5). Right: Trend of the zonal ocean density gradients between 2011 and the end of 2018 for each transect for each depth level above 200 m. Statistically significant trends are marked with a dot.	42
4.4	Change of the southward component of the current between 2011 and the end of 2018. Right: As computed by the dynamical method. Left: As given in ORAS5. Statistically significant changes are marked with a dot.	44
4.5	Left: Trends of surface wind stress magnitude between 2011 and the end of 2018. Center: Trends of angle between 2011 and the end of 2018, given in anti-clockwise degree per year. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. Right: Initial direction and magnitude of the wind stress.	45
4.6	R factor as defined by eq. 4.10 averaged between 2011 and the end of 2018 for the five transects as defined in Fig. 4.3.	49
4.7	Trends of salinity, potential temperature and potential density anomaly between 2011 and the end of 2018 at the surface and at 50 m depth. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC.	50
4.8	SIT, SSS and SSD anomaly averaged over winter (January, February and March) and Summer (July, August, September).	50

4.9	Upper: Trends of SIT and SIC between 2011 and the end of 2018. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. Bottom: Sea ice transport through the four gates displayed on the maps above. A one year smoothing average was applied to the time series and the linear trends are displayed, with the corresponding slopes indicated on the top of the figures.	52
4.10	Evolution of SIV and SIE calculated over the area delineated in red between 2011 and the end of 2018. A one year smoothing average was applied to the time series. The linear trends are displayed, with the corresponding slopes indicated on the top of the figures.	53
4.11	Left: Correlation between SIM and SSS. Right: Trends of SIM between 2011 and the end of 2018. The statistically significant correlations and trends are marked with a black dot. The mean path of the EGC is delineated with a solid black line.	56
4.12	The map: SSS trends between 2011 and the end of 2018, with the statistically significant trends marked with a black dot and definition of the boxes A1 to A6 . The time series: SIM (in blue) and mean SSS (in red) for each of the 6 boxes. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.	57
4.13	Correlation between SIM and depth averaged salinity. Left: Time series of SIM (blue) over boxes B1 and B2 and depth averaged salinity (red) above 200 m depth in the same boxes. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed. Right: Trends of salinity averaged above 200 m depth and boxes definition.	60
4.14	ST transport as calculated with eq. 4.14 through two transects displayed in red in the inset map. A one year smoothing average has been applied to the time series.	62
4.15	SIV (in blue) in the red area in the inset map and sea ice transport (red) through the gate G1 in the inset map. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.	65
4.16	The map: SIT trends between 2011 and the end of 2018, with the statistically significant trends marked with a black dot, definition of the boxes A1 to A6 and the gate G1 . The time series: SIM (in red) for each of the 6 boxes and sea ice transport through the gate G1 . A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.	66

4.17	Summary of the main observations and relationships highlighted in this study. A green (red) box indicates a decrease (increase) in the given variable. A dark (light) green arrow indicates a strong (weak) relationship. The red arrow indicates a relationship at odds with the observed change. Figures and table references are displayed alongside the corresponding results.	70
------	--	----

List of Tables

2.1	ORAS5 native variables used in this study.	15
3.1	Data used for Fram Strait sea ice transport calculation in Min et al. (2019) (M1 and M2) and Ricker et al. (2018) (R).	26
3.2	Water masses definition from Rudels (2002)	33
4.1	Linear trend (between 2011 and the end of 2018) slope and p-value for SIM over boxes A1 to A6 as defined in Fig. 4.12 and boxes B1 and B2 as defined in Fig. 4.13.	56
4.2	Change SIM (ΔSIM), depth averaged salinity (ΔS), expected change in depth averaged salinity related to change in SIM ($\Delta S_{\Delta\text{SIM}}$) and relative importance of $\Delta S_{\Delta\text{SIM}}$ in ΔS (R_S). Results are for time periods $t_1^{B1} \rightarrow t_2^{B1}$, $t_2^{B1} \rightarrow t_3^{B1}$, $t_1^{B2} \rightarrow t_2^{B2}$, $t_2^{B2} \rightarrow t_3^{B2}$ and area B1 and B2 as defined in Fig. 4.13.	60

Acronyms

AAW	Arctic Atlantic Water
AIW	Atlantic Intermediate Water
AW	Atlantic Water
CS2	CryoSat-2
DW	Deep Water
ECMWF	European Centre for Medium-Range Weather Forecasts
EGC	East Greenland Current
EKE	eddy kinetic energy
GTS	Global Telecommunication System
JMFZ	Jan Mayen Fractural Zone
KE	kinetic energy
NDW	Nordic Deep Water
NWP	Numerical Weather Prediction
ORAS5	Ocean Reanalysis System 5
OSI SAF	Ocean and Sea Ice Satellite Application Facility
PIOMAS	Pan-Arctic Ice Ocean Modeling and Assimilation System
PSW	Polar Surface Water
PSW_w	warm Polar Surface Water
PW	Polar Water
RAW	Recirculating Atlantic Water
SIC	sea ice concentration
SID	sea ice drift
SIE	sea ice extent
SIM	sea ice melt
SIT	sea ice thickness

SIV	sea ice volume
SSD	sea surface density
SSS	sea surface salinity
SST	sea surface temperature
ST	salt transport
uPDW	upper Polar Deep Water
WSC	West Spitsbergen Current

Introduction

The Earth's climate has undergone continuous changes over its billions of years of existence, transitioning through warm periods characterized for instance by lush forests in Greenland, as well as glacial periods marked by extensive ice caps in Scandinavia. Paleo-climatologists are striving to reconstruct this history and understand the causes of these variations. They have identified a multitude of processes at work on different timescales, ranging from the Earth's geomorphological evolution to variations in astronomical parameters (Zachos et al., 2001). Over the last few centuries, our ability to measure our climate has increased considerably and we have observed that, since the beginning of the industrial era, the Earth's global temperature has been rising at an accelerating rate (Mann et al., 1999; Marcott et al., 2013). This observed climate change is not of the same nature as previous ones, and cannot be explained by the natural mechanisms that usually govern the Earth's climate but is due to human activities and the resulting greenhouse gas emissions. This anthropological cause has been debated for decades, mainly because it calls into question our societal model, but it is no longer in doubt today, *"Human activities, principally through emissions of greenhouse gases, have unequivocally caused global warming, with global surface temperature reaching 1.1°C above 1850-1900 in 2011-2020."* (Calvin et al., 2023).

Global warming results in the disruption of all aspects of the climate and poses a serious threat to nature and humanity. The sea level has already risen by 0.2 m between 1901 and 2018, and projections for the end of the century range from 0.43 m to 0.84 m. This increase will lead to extreme sea level events, like those we observe today once a century, occurring almost every year in many places. Beyond 2100, this rise will continue for centuries, and the level will not return to normal for thousands of years (Intergovernmental Panel On Climate Change, 2022). Additionally, the observed intensification of the water cycle will continue, with increased variability, monsoonal precipitation and extreme droughts and floods. Projections also indicate a rise in the frequency of cyclones and storms, heat waves and aridity (Calvin et al., 2023). The capacity of oceans and land to absorb carbon is expected to decline, and ocean acidification to intensify (Calvin et al., 2023). It is therefore crucial to understand our climate system and how it will evolve in the future if we are to withstand the risks of this century and beyond.

Although warming is global, not all regions of the world are warming at the same rate. The strongest signal is found in the Arctic, where surface temperatures have increased 2 to 3 times faster than the global average (Chylek et al., 2009). This phenomenon, known as *Arctic Amplification*, appears to be an inherent feature of our climate system that is well reproduced both by climate models (Smith et al., 2019)

and in paleoclimate archives (Miller et al., 2010). It is explained by several processes called *feedbacks*, i.e. when a change in the climate system causes an effect that either reinforces (positive feedback) or counteracts (negative feedback) the initial change. Taylor et al. (2013) studied these feedbacks and their relative importance in polar amplification. The most significant of these is the surface albedo feedback, in which melting sea ice reduces the local albedo, thereby amplifying the initial warming (Curry et al., 1995).

The accelerated changes are unmistakably evident in the Arctic waters and sea ice dynamics. The Arctic halocline, crucial for insulating sea ice from the underlying warm Atlantic Water (AW), is exhibiting clear signs of weakening, heightening the vulnerability of the sea ice cover to further melting processes in the eastern Arctic Ocean (Polyakov et al., 2020). Through continuous satellite-based measurements, Lindsay and Schweiger (2015) observed a notable decline of 65% in the annual mean thickness of Arctic sea ice from 1975 to 2012. The rate of volume loss is estimated at $2,900 \text{ km}^3 \text{ decade}^{-1}$ (Schweiger et al., 2011) and the sea ice extent loss at $12 \% \text{ decade}^{-1}$, reaching a minimum in September 2012 down to $3.63 \times 10^6 \text{ km}^2$. This sea ice diminution caused an increase of solar heat input into the Arctic Ocean (Perovich et al., 2007). The sea ice decline is observable in all seasons, but peaks in September when sea ice is at its minimum, and it is not about to stop. The first summer ice free condition will likely occur before 2050, but could even occur in the 2020s or 2030s, regardless of the emission scenario (Jahn et al., 2024; Notz and Community, 2020). This represents a significant change unprecedented in the climate system for several thousand years. The last instance of the Arctic Ocean being seasonally ice-free occurred during the Marine Isotope Stage 5e (MIS5e), approximately 130,000 to 115,000 years ago (Vermassen et al., 2023).

Sea ice is one of the largest biomes on Earth. Seals and walruses depend on it to rest and give birth, while it is used as a hunting platform by the polar bears. It is also a foraging ground for many mammals, as Arctic foxes and even whales under the sea ice. Despite the low temperatures and high salinity, we find plenty of life in the brine fraction on the sea ice bottom, from micro-organisms to larger fishs (Arrigo, 2014). The algae and phytoplankton living in and under sea ice are of prime importance for the primary production in the Arctic Ocean. This entire ecosystem is threatened by the sea ice melt, reducing the accessibility of food for the mammals that evolve in this area, or disrupting phytoplankton production by increasing the length of the melt season, which impacts all the species that depend on it (Post et al., 2013).

In addition to the local impact of disappearing sea ice, the rapid changes taking place in the Arctic has a longer-range influence, both in the atmosphere and in the ocean. Sea ice acts as a powerful insulator between these two environments. Its absence leads to increased heat, moisture and momentum transfer at the sea-air interface, resulting in higher air temperatures, humidity and cloud cover over the Arctic Ocean. This could reduce the stability of the local troposphere and affect the large-scale pressure pattern over Eurasia, possibly favoring the negative phase of the North Atlantic Oscillation and Arctic Oscillation (Vihma, 2014). Sea ice cover influences the oceanic thermohaline circulation. When sea ice grows, it releases

salt due to brine rejection, and conversely, when it melts, it results in a discharge of low salinity water at the top of the ocean. These changes in salinity have an impact on the density profile of the water column, particularly near the freezing point where the thermal expansion coefficient is around 30 times smaller than the haline coefficient (Aagaard and Carmack, 1989). This has an impact on the deep convection, an important climatic process that takes place notably in the Greenland Sea and plays a role in the meridional overturning circulation (Liu and Fedorov, 2019; Li et al., 2021; Manabe and Stouffer, 1994; Rudels et al., 2012; Killworth, 1983).

The Arctic Ocean has only two connections with other oceans. With the Pacific via the Bering Strait and with the Atlantic via the Nordic Seas, where seawater can pass either through the shallow Barents Sea or the very deep Fram Strait. The latter, with its depth of up to 2650 m, is the deepest Arctic connection and the only gateway to the World Ocean for the Arctic’s deep and intermediate waters (De Steur et al., 2014; Untersteiner, 1988) and therefore the most important Arctic gateway. Indeed, the southward outflow through Fram Strait accounts for 90% of heat exchange and 75% of mass exchange between the Arctic and the rest of the World Ocean, transporting sea ice formed upstream in the Arctic and driven by transpolar drift (Wadhams, 1983; Aagaard and Greisman, 1975), being the greatest outlet of Arctic sea ice (Aagaard and Carmack, 1989; Kwok et al., 2004). The Fram Strait accounts for approximately 25% of the total annual freshwater export from the Arctic (Serreze et al., 2007). Sensitivity experiments conducted with numerical models have revealed that the oceanic circulation of the Atlantic is responsive to fluxes of low-salinity surface water, emphasizing the role of sea ice export from the Arctic basin in the dynamics of climate change (Smith and Gregory, 2009). To the south, the Denmark Strait is another important point of interaction between the Arctic and the Atlantic. The mean water volume transport in the Denmark Strait is about 5 Sverdrup (Sv, $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) (Worthington, 1970), which represents roughly half of the total dense overflow across the Greenland-Scotland ridge (Dickson et al., 2008). This water flows southwards as an organized geostrophic current from Fram Strait to Denmark Strait along the Greenland coast, linking the Arctic to the Atlantic. This current is known as the East Greenland Current (EGC).

The EGC is of significant importance in the climate system due to its interplay within the Fram and Denmark Strait overflow dynamics, as mentioned above. This current serves as a conduit for both dense water and buoyancy, which contribute to the meridional overturning circulation. Furthermore, a portion of the mixing between Atlantic and Arctic waters, pivotal for vertical stratification and deep convection (Rudels et al., 2012), occurs within the core of the EGC, where segments of the West Spitsbergen Current (WSC) recirculate. The impacts of global warming are prominently manifest within the EGC region, evidenced by the reduction in sea ice presence (Spren et al., 2020; Sumata et al., 2022) and the warming trend of the Nordic Seas (De Steur et al., 2023; Mork et al., 2019; Tsubouchi et al., 2021).

In this context of accelerating change in the Arctic and its future impact on global climate, it is important to understand the future evolution of this region. The melting of sea ice in the EGC area has been observed for decades, and a more

precise understanding of its impact on the current would enable us to foresee its evolution, and therefore the evolution of the climate in a broader framework. For this reason, this master’s thesis will address the following research question,

“What is the impact of the sea ice accelerated melting on the velocity structure of the East Greenland Current?”

We begin in Chapter I by elucidating the EGC, delineating its principal characteristics, and discussing the existing knowledge regarding its recent evolution within the broader context of global warming. Following this, Chapter II provides an overview of the Ocean Reanalysis System 5 (ORAS5), the dataset underpinning our investigation. This chapter also discusses the various observational datasets employed in the validation process of ORAS5. In Chapter III, we undertake a validation of the reanalysis dataset, juxtaposing it against various observations of sea ice and oceanic variables. To do so, we will notably use a newly released dataset from [Landy et al. \(2022\)](#), which provides, for the first time, a year-round record of sea ice thickness spanning from 2011 to 2020. Since ORAS5 dataset spans from 1979 to the end of 2018 ([Copernicus Climate Change Service, Climate Data Store, 2021](#)), we have chosen our research period as the intersection of these two, spanning from 2011 to the end of 2018.

Following the validation of the dataset, we divide our research question in four parts, that we answer successively in Chapter IV:

1. *“What are the observed changes in the EGC velocity structure over our study period?”*
2. *“Are the observed changes in the EGC velocity structure related to shifts in its baroclinic component due to perturbations in ocean density profiles?”*
3. *“Are these density changes related to the sea ice accelerated melting?”*
4. *“What future evolution of the sea ice melting in the EGC region can we expect?”*

We examine in this chapter the evolving dynamics of the current, the sea ice distribution in the region, and the associated density distribution with the relative roles of salinity and temperature. Through our analysis, we discern that certain alterations in the current’s strength are correlated with variations in zonal ocean density gradients, which can be ascribed to salinity fluctuations. Subsequently, we demonstrate that these salinity fluctuations can partially be elucidated by changes in sea ice melt within the EGC region. Finally, we show that this change in the melting of sea ice in the area is correlated with the sea ice transport through the Fram Strait. This provides an opportunity to foresee the evolution of sea ice melt in the EGC region based on the projections of sea ice transport through the Fram Strait.

Chapter 1

The East Greenland Current

1.1 A brief history

The first recorded expedition in the EGC region goes back to the Viking time. [Coachman \(1968a\)](#) quoted from *Konungs Skuggsjà*, a book written by an unknown Viking author around 1250, the description of large sea ice extent near the Greenland, flowing with high velocity reminiscent of the sea ice drift speed along the EGC. *“As soon as one has passed over the deepest part of the ocean, he will encounter such masses of ice in the sea that I know no equal of it anywhere else in all the Earth . . . They . . . extend so far out from the land that it may mean a journey of four days or more to travel across them. There is more ice to the northeast and north of the land than to the south, southwest, and west [“land” probably refers to the settled southern part of Greenland] . . . These ice floes have peculiar habits. Sometimes they lie as quiet as can be, though cut apart by creeks or large fjords; at other times they travel with a speed so swift and violent that a ship with a fair wind behind is not more speedy; and when once in motion, they travel as often against the wind as with it.”* ([Larson, 1917](#)).

However, it wasn’t until 1820 that the first description of the EGC structure emerged when Scoresby published his book *“An account of the Arctic regions, with a history and description of the northern whale-fishery”* where we can find the following quote *“Though the set of the current in the Greenland Sea be generally towards the south-west, yet near the shore, where there are eddies and tides, it is very irregular.”* ([Scoresby, 1820](#), p.287). It seems that he was the first writer to note that the current was faster and more regular offshore than near the coast ([Coachman, 1968b](#)), a characteristics inherent to the EGC as it flows over the continental slope. Forty years later, in 1865, Petermann published charts of the Arctic (Fig. 1.1). These charts highlighted two notable features of the EGC. Firstly, it wasn’t merely depicted as an isolated current within the Greenland Sea but rather as a link between the Arctic Ocean and the Atlantic, an important step in grasping its role as a mediator between these oceans ([Coachman, 1968b](#)). Secondly, the current was observed to bifurcate north of Iceland, with the primary portion going through the Denmark Strait, while the other turned southeastward, merging with the East Icelandic Current.



Figure 1.1: Petermann’s chart of the Arctic. Figure from [Tammiksaar et al. \(1999\)](#).

The first temperature measurements in the EGC were conducted by Koldewey in 1879 ([Deutsche Nordpolar-Expedition \(2nd : 1869-1870\)](#), 1874). Although, it wasn’t until the *Norwegian Ocean expedition* and the subsequent publication of Mohn’s results in 1887 ([Mohn, 1887](#)) that the first systematic scientific overview of the EGC emerged ([Coachman, 1968b](#)). That includes large number of water temperature and density measurements along with sea surface topography computation. With these data, they distinguished two different water masses in the EGC. The first one was coming from the Arctic Ocean and was located on the western part of the current, while the second one, farther east, was warmer water coming from the WSC which joined the EGC in its southward flow. Eventually, they agreed with Petermann’s chart about the separation of the current north of Iceland. This accumulation of knowledge about the EGC and the Nordic Seas continued in the following decades thanks to several explorers as Ryder in 1891 or Koefed and Halland-Hansen in 1905 aboard the *Belgica* ([Ryder, 1895](#); [Helland-Hansen, 1907](#)).

1.2 Structure, drivers and variability

The Nordic Seas constitute a pivotal region situated north of Iceland and south of Svalbard, flanking the coasts of Norway and Greenland. As shown in Fig. 1.2, they are demarcated to the south by the imposing Greenland-Scotland ridges and to the north by the convergence of the Fram Strait-Spitsbergen-Norway. Comprising three distinct entities, namely the Greenland Sea adjacent to the Fram Strait, the Icelandic Sea positioned south of the Jan Mayen Fractural Zone (JMFZ) proximate to the Denmark Strait, and the Norwegian Sea situated to the east of the Mohn’s Ridge, these seas serve as arenas where Atlantic and Arctic waters continuously interact.

In these maritime realms, an incessant mingling of Atlantic and Arctic waters ensues. The AW arrives through three distinct branches, with the Iceland branch being the smallest, boasting a mean flux of 0.8 Sv, followed by the Faroe and Shetland branches, each experiencing a flux of 3.8 Sv (Hansen et al., 2008). Consequently, the total mean inflow of AW into the Northern seas is approximately 8.5 Sv, characterized by a mean temperature of 8.5° C and a salinity of 35.25 practical salinity units (psu, 1 psu = 1g of salt per kg of water) (Hansen et al., 2008). As these waters traverse the Norwegian Sea, they undergo a cooling process before bifurcating, either venturing further into the Arctic Ocean or partaking in recirculation dynamics (Eldevik et al., 2009; Hansen and Østerhus, 2000). Conversely, on the western periphery, a contrasting flux of cold and less saline water emanates from the Arctic region. This outflow carries with it a proportion of the aforementioned recirculated water, contributing to the intricate hydrodynamic milieu of the Nordic Seas. This continual mixing of Atlantic and Arctic waters influences vertical stratification dynamics, which are critical for deep convection processes. Notably, the intrusion of saline AW into northern latitudes prompts cooling, densification, and subsequent sinking, creating favorable conditions for deep convection phenomena.

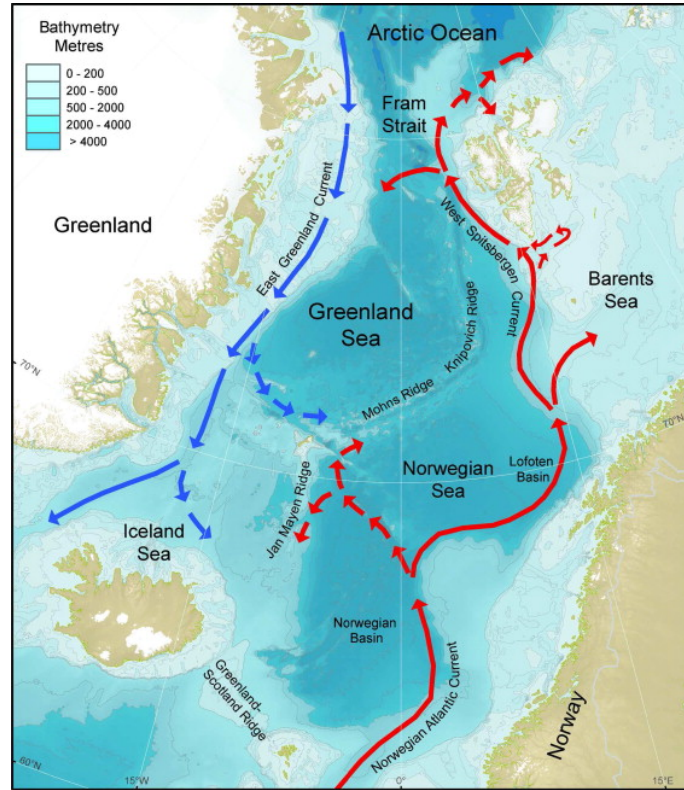


Figure 1.2: Nordic Seas bathymetry and main currents. Figure from Isachsen et al. (2014).

The EGC charts a southerly course along the eastern shoreline of Greenland, spanning approximately 1500 km as it traverses the region between the Fram and Denmark Straits, covering the Greenland and Icelandic Seas. It follows the continental margin, where the water depth transition from a few hundred meters to two kilometers in the Greenland Sea. As it progresses southwards, passing the JMFZ,

the ocean floor gradually ascends, with depths no longer exceeding one kilometer. Near the Denmark Strait, after the Blosseville Basin, depths decrease to around 500 m, marking a notable shallowing of the seafloor.

To the west of Fram Strait, a flow of cold, low-salinity water descends southwards from the Arctic Ocean, signifying the inception of the EGC. Concurrently, along the eastern flank, AW courses northwards within the WSC. Between latitudes 79°N and 79°50'N, a portion of the WSC undergoes recirculation, merging with the EGC dynamics (Schauer et al., 2008). To the north of 79°N, a baroclinic flow prevails, characterized by minimal seasonal fluctuations. However, upon reaching 78°50'N, the WSC recirculation induces a notable increase in barotropic flow rates, raising from 5.8 ± 1.8 Sv to 8.7 ± 2.5 Sv, accompanied by seasonal variability (De Steur et al., 2014). This warm and saline water descends beneath the lighter surface layers, giving rise to a distinct vertical stratification pattern as outlined by Coachman (1968b). The stratification includes three primary layers: Polar Water (PW), Atlantic Intermediate Water (AIW), and Deep Water (DW).

PW extends from the surface down to approximately 150 m, characterized by temperatures ranging from 0°C to the freezing point. During summer, a temperature minimum is often observed around the 50 m depth, while in winter, temperatures remain uniformly near the freezing point. Beneath the PW lies the AIW, which extends from approximately 150 m to 800 m depth. This layer exhibits positive temperature and salinity values that increase with depth, typically reaching a salinity range between 34.88 and 35 psu above the 400 m mark. Below this depth, the salinity of the AIW remains relatively constant. Finally, the DW layer is found below 800 m depth, characterized by temperatures dropping below 0°C and a salinity range between 34.87 and 34.95 psu.

This recirculation phenomenon, characterized by the intrusion of AW into the EGC, extends slightly southwards until the primary core of the EGC is established. Coachman (1968b) measurements during the winter of 1965 revealed that, among the various water masses contributing to the EGC, the AIW is the predominant component, accounting for 21.3 Sv out of a total outflow of 31.5 Sv. Notably, Woodgate et al. (1999) quantified this process at 75°N, reporting an average annual transport of 21 ± 3 Sv, with seasonal fluctuations ranging from a minimum of 11 Sv in summer to a maximum of 37 Sv in winter. The observed seasonal variability of volume transport is primarily attributed to the presence of AIW within the EGC and is not observed in the Fram or Denmark Straits (De Steur et al., 2014). Consequently, Woodgate et al. (1999) postulated that the observed variability stems from fluctuations in the WSC, driven by wind patterns within the Greenland Sea gyre. Indeed, De Steur et al. (2014) estimated the wind-driven Sverdrup transport and demonstrated that the seasonality was largely attributable to the wind stress curl.

Through an examination of vertical speed profiles derived from 1965 winter measurements, Coachman (1968a) concluded that horizontal motion persists across all depths, indicative of a predominantly barotropic mode. However, contrasting findings by Woodgate et al. (1999) suggested that the barotropic mode accounted for 70% of the total variance in the annual mean. This discrepancy is explained by the

summer baroclinicity observed by [Coachman \(1968a\)](#) through measurements of the slope of the isopycnals. This corresponds to the aforementioned description of current variability, a small surface baroclinic current with low variability flowing from Arctic and a stronger submerged AIW barotropic current displaying high seasonal variability, peaking in winter. Furthermore, [Paquette et al. \(1985\)](#) documented the presence of steeply inclined isopycnals near the polar front - defined as the boundary between PW and the AW and AIW - which could contribute to the high baroclinic speed jet near the sea ice edge found by [Foldvik et al. \(1988\)](#).

As the EGC progresses towards the JMFZ, a segment of the warm AIW veers eastwards in a cyclonic motion, following the Gyre, while the remainder continues its southward trajectory within the EGC ([Coachman, 1968a](#)). Around 70°N, the EGC bifurcates into two distinct cores: the *shelfbreak EGC* and the *separated EGC*. The former follows the relatively shallow continental slope of the Icelandic Sea, while the latter courses along the northern expanse of the Blosseville Basin ([Våge et al., 2013](#); [Håvik et al., 2017](#)). Upon reaching the Denmark Strait, a significant proportion of the EGC remains composed of AIW, as evidenced by [Eldevik et al. \(2009\)](#), who determined that it constitutes 64% of the overflow water observed in the Denmark Strait. Subsequently, upon traversing the Denmark Strait, the intermediate and deep water masses within the EGC, denser than their North Atlantic counterparts, descend towards abyssal depths of the ocean ([Håvik et al., 2017](#)).

1.3 Sea ice in the EGC region

In addition to water transport, the sea ice transport through the Fram Strait constitutes another pivotal mechanism connecting the Arctic and the North Atlantic regions. Utilizing the Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS), [Spreen et al. \(2020\)](#) estimated that the volume of sea ice transported through the Fram Strait represents approximately 14% of the total Arctic sea ice volume. This process significantly influences the exchange of freshwater and energy between the Arctic and North Atlantic Oceans. [Walsh and Chapman \(1990\)](#) observed that North Atlantic salinity and temperature anomalies, such as the extreme and long-lived “Great Salinity Anomaly” of the late 1960s and early 1970s, were accompanied by substantial variations in sea ice. The presence of sea ice within the EGC region curtails the heat exchange between the ocean and the atmosphere. Conversely, its disappearance amplifies these fluxes, consequently contributing to surface-based Arctic Amplification ([Serreze et al., 2009](#)). Due to the swift flow rate characterizing the EGC and the notably low temperatures of the water it transports, this area constitutes one of the southernmost extents of Arctic sea ice coverage. During the winter season, it is common for the sea ice boundary to extend as far south as 65°N in this region.

Sea ice cover in the EGC region is characterized by seasonal variations, with the maximum extent generally occurring in March and the minimum in September. During winter, sea ice blankets the current’s path, extending continuously up to the Denmark Strait and often protruding eastwards by approximately 300 km. Conversely, summer sees a substantial reduction in sea ice coverage, with much of

it melting away, rendering the EGC predominantly ice-free. This seasonal melting of sea ice leads to the release of significant amounts of low salinity water into the surrounding ocean waters, resulting in observable decreases in salinity of up to 5 psu at the ocean surface (Coachman, 1968b).

In the eastern sector of the Fram Strait, the influx of warm AW transported northwards by the WSC induces substantial melting of sea ice. Consequently, the western shoreline of Spitsbergen, situated approximately at 79°N latitude, experiences predominantly ice-free conditions, even in the presence of sea ice further south near the vicinity of Iceland.

The Greenland Sea experiences little sea ice formation, primarily receiving sea ice transported from the Arctic region, which subsequently melts within the Greenland Sea. Therefore, knowing the flux of sea ice through the Fram Strait is crucial for comprehending the dynamics of its downstream evolution. Measurements conducted by Min et al. (2019) assessed the volume of sea ice imported through the Fram Strait from September 2010 to December 2016. Their findings revealed a notable seasonal fluctuation in sea ice import, ranging from $244 \pm 43 \text{ km}^3$ in autumn 2016 to $973 \pm 59 \text{ km}^3$ in spring 2012. Moreover, interannual variability was observed, with annual sea ice import levels ranging from $1974 \pm 291 \text{ km}^3$ in 2013 to $2491 \pm 280 \text{ km}^3$ in 2012.

1.4 Recent changes

In recent decades, the EGC region has undergone significant and rapid transformations attributed to the effects of global warming. Pronounced warming in the Nordic Seas has been observed (Mork et al., 2019; Tsubouchi et al., 2021). This warming trend is predominantly ascribed to the heightened heat transport originating from the Atlantic, which has substantially influenced the regional climatic conditions (Tsubouchi et al., 2021). Concurrently, there has been a notable reduction in the transport of sea ice through the Fram Strait, attributed to sea ice thinning (Spreen et al., 2020). These changes have a non negligible impact on the dynamics of the EGC and its associated environmental processes. Consequently, this section will present these changes in order to subsequently analyze the evolutionary dynamics of the EGC.

The study conducted by Selyuzhenok et al. (2020) analyzed sea ice in the Greenland Sea utilizing PIOMAS reanalysis from 1979 to 2016. Their investigation revealed a notable decrease in sea ice volume (SIV) of 113 km^3 per decade. Intriguingly, during the same period, they observed an increase in sea ice transport through the Fram Strait, amounting to 115 km^3 per decade. However, it becomes apparent that the positive trend in sea ice transport halted sometime between 2000 and 2010. This observation contrasts with the results of Zhang et al. (2017), who found no clear trend in SIV import for the period spanning 1979 to 2012. Additionally, Spreen et al. (2020) investigated a narrower timeframe from 1994 to 2014, discovering a decrease in sea ice import of $54 \pm 4 \text{ km}^3$ per month per decade, primarily due to sea ice thinning. Synthesizing these findings suggests that, while sea ice transport through

the Fram Strait increased towards the end of the last century, it is now exhibiting signs of decline. The minimum sea ice import through the Fram Strait was reached in 2018, when it was reduced to less than 40% compared to the period 2000-2017, due to the simultaneous presence of thin and slow sea ice (Sumata et al., 2022).

In recent decades, the Nordic Seas have undergone a pronounced warming, as evidenced by observations documented by Mork et al. (2019). Their analysis revealed a warming trend extending to depths of 1000 m, around $0.05^{\circ}\text{C year}^{-1}$ between 2011 and the end of 2018. Furthermore, Tsubouchi et al. (2021) corroborated these results, attributing the warming to the observed 21 terawatt increase after 2001 in oceanic heat transport across the Greenland-Scotland Ridge, marking the southern limit of the Nordic Seas. Mork et al. (2019) have also observed a freshening trend in the upper 400-500 m of the Nordic Seas, attributed to a decrease in salinity of approximately 0.1 psu of the inflow of AW.

Using data from two oceanographic moorings in the western Fram Strait spanning the period from 2003 to 2019, De Steur et al. (2023) observed both a sea ice reduction and a warming of the summer PW in the Fram Strait. The sea ice extent in the Fram Strait reached its lowest level in August 2018, consistent with the documented minimum sea ice transport through the Fram Strait as outlined by Sumata et al. (2022). During winter, the most significant negative anomalies were recorded in January, characterized by a westward displacement of the sea ice edge by 65 to 100 km in the two lowest years, 2016 and 2018, relative to the average January sea ice extent from 2003 to 2015.

Over the same period, they observed an increase in summer temperatures of PW at 55 m depth. This increase was accompanied by greater variability after 2010, with higher temperatures in August and September and prolonged periods of elevated temperatures. The observed warming can be attributed to two factors: the increased fraction of open water, leading to enhanced heat absorption by the ocean, and the emergence of a warmer and saltier water mass associated with the recirculation of AW. For the latter, they identified a significant regime shift towards an AW-dominated regime in June 2015. After this date, salinity and temperature measurements within the Fram Strait displayed more frequent occurrences of AW than before. Notably, winter temperatures consistently rose above 0°C in every season, a condition not previously observed. In summary, their observations indicated a summer warming of the PW and a more frequent presence of AW during winter. They also demonstrated that the observed temperature increases in the Fram Strait had a consequential impact on downstream sea ice concentration, extending as far as the Denmark Strait.

Karpouzoglou et al. (2022) observed a decrease in the mean freshwater transport through the Fram Strait after 2015, with the single exception of 2017, compared to the 2003-2015 period. This reduction in freshwater transport has led to an increase in downstream salinity. They also noted an increased presence of AW in the Fram Strait and observed that the PW layer had become thinner by 40-50 m and more stratified. Additionally, they found that the boundary between PW and AW within the Fram Strait had shifted westwards in 2015. This boundary was previously

known to exhibit little seasonal variation, and AW was generally not found on the shelf break in the past (Holfort and Hansen, 2005). These results from Karpouzoglou et al. (2022) show that the Fram Strait is now experiencing the “*Atlantification*” a process which was first observed in the Arctic, as in the Barents Sea (Årthun et al., 2012; Lind et al., 2018) or the Eurasian basin (Polyakov et al., 2017).

Karpouzoglou et al. (2022) attributed the observed decrease in freshwater transport to a weakening of the overall transport mechanism, as well as an increase in salinity. They showed that this decrease in transport was caused by the slowdown of the EGC since 2015. Through the computation of the geostrophic current in Fram Strait, they found that this slowing down was attributable, in part, to a weakening of zonal density gradients.

We will see in this study that ORAS5 indeed reproduces a slowing down of the EGC, between 74°N and 79°N, and that it is attributable to a weakening of density gradient. We will investigate in detail the direct role of the sea ice melting on the observed increasing salinity on the Greenland continental shelf.

Chapter 2

Data used in this study

Our objective is to investigate the impact of the sea ice melting on the EGC velocity structure. To achieve this, we utilize ORAS5 from the European Centre for Medium-Range Weather Forecasts (ECMWF) in this study ([Copernicus Climate Change Service, Climate Data Store, 2021](#)). This reanalysis has been selected because it incorporates the coupling between sea ice and ocean dynamics, which is the focus of our investigation. Moreover, the assimilation of observational data in this reanalysis ensures that our results closely reflect the actual evolution of the current, a level of accuracy that would not be achievable with a purely model-based approach.

In this chapter, we present the ORAS5 dataset. We detail the variables natively present in the dataset and the ones we computed additionally. We also introduce the observational data that will be employed in the subsequent chapter to validate the ORAS5 dataset.

2.1 ORAS5

The ORAS5 dataset has been released in 2021 by the ECMWF. It provides a reconstruction of the global state of the ocean and sea ice from 1958 to present in two forms: a consolidated and a near real time product. The former extends from 1958 to 2015 and the latter from 2015 to present.

ORAS5 is a global eddy-permitting ocean reanalysis produced with the ECMWF OCEAN5 system. It relies on the Nucleus for European Modelling of the Ocean (NEMO V3.4.1) ocean model, coupled with the LIM2 sea ice model. The resolution is $1/4^\circ \times 1/4^\circ$ in the horizontal, with 75 levels in the vertical, spaced 1 m apart at the surface up to 200 m in the deep ocean. The reanalysis contains 5 ensemble members, opa0 to opa4, with opa0 the control member, without perturbation. The perturbations are generated on the initial conditions but also on the forcing and observations ([Zuo et al., 2019](#)).

ORAS5 has been forced by the atmospheric reanalysis ERA-40 before 1979, then ERA-Interim ([Dee et al., 2011](#)) until 2015 and ECMWF operational Numerical Weather Prediction (NWP) after. Beside the atmospheric forcing, four variables are

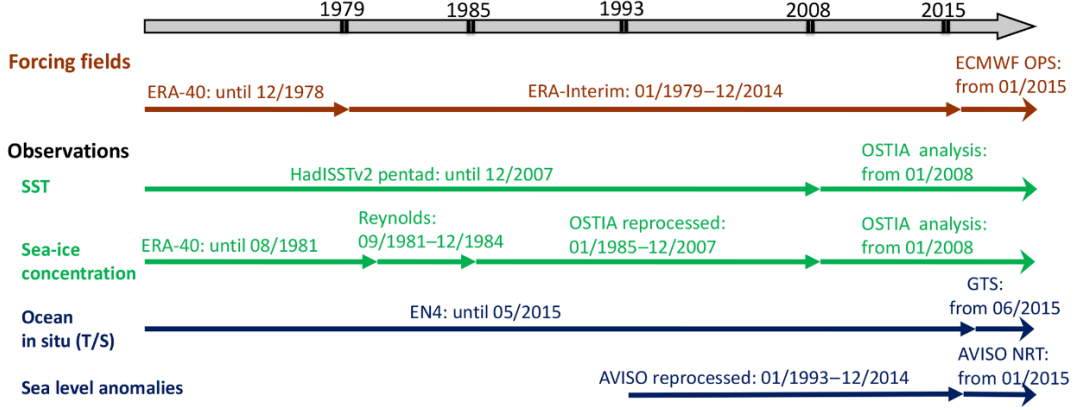


Figure 2.1: Time line of different forcings and assimilation data used in ORAS5, figure from (Zuo et al., 2019).

assimilated in ORAS5: sea surface temperature (SST), sea ice concentration (SIC), vertical salinity and temperature profiles, and sea level anomalies. SST and SIC come from the OSTIA analysis which use sea ice data from the Ocean and Sea Ice Satellite Application Facility (OSI SAF) and SST data from the Group for High-Resolution SST (GHRSSST) and in situ SST from the Global Telecommunication System (GTS) (Donlon et al., 2012). Vertical profiles of temperature and salinity come from in situ measurements, first from EN4 (Good et al., 2013) until July 2015 and then from GTS after this date. Finally, the sea level anomalies comes from AVISO reprocessed until the end of 2014 and from AVISO NRT after (Pujol et al., 2016). These information are synthesised on a timeline (Fig. 2.1) which can be found in Zuo et al. (2019).

In this study, we used the unperturbed control member opa0 from the ORAS5 dataset. We accessed this data through the Integrated Climate Data Centre (ICDC) at Hamburg University, where it is given monthly up to the end of 2018. The product has been interpolated on a grid with resolution $1^\circ \times 1^\circ$, which correspond to ~ 110 km in the latitude, ~ 50 km in longitude near the Denmark Strait and ~ 20 km in longitude near the Fram Strait.

2.1.1 Derived variables

The variables natively present in the ORAS5 dataset that we used in this study are summarized in Tab. 2.1. From these variables, we calculated the SIV, the sea ice extent (SIE), the sea ice melt (SIM), the kinetic energy (KE), and the eddy kinetic energy (EKE). The detailed calculations for these derived variables are provided below.

SIE and SIV

SIE is defined as the cumulative surface area of all grid cells with at least 15% of SIC. SIV is calculated as the sum, over all grid cells, of the product of sea ice

Table 2.1: ORAS5 native variables used in this study.

Name	Description	Dimension	Unit
ileadfra	SIC	2D	%
iicethic	SIT	2D	m
iicevelu	Sea ice zonal velocity	2D	m s ⁻¹
iicevelv	Sea ice meridional velocity	2D	m s ⁻¹
sozotaux	Wind stress along i-axis	2D	N m ⁻²
sometauy	Wind stress along j-axis	2D	N m ⁻²
vozocrtx	Water zonal velocity	3D	m s ⁻¹
vomecrtv	Water meridional velocity	3D	m s ⁻¹
votemper	Potential temperature	3D	C
vosaline	Salinity	3D	psu

thickness (SIT), SIC, and the grid cell surface area.

$$\begin{aligned}
 SIE &= \sum_{SIC > 0.15} \text{surface} \\
 SIV &= \sum_{all} SIT \times SIC \times \text{surface}
 \end{aligned} \tag{2.1}$$

SIE is given in km² and SIV in km³.

SIM

SIM represents the volume of sea ice that has melted in a given area over a specified time period. It is calculated as the difference between the net transport of sea ice (T), which is the difference between sea ice volume import and export in the area, and the sea ice volume balance (B), which is the difference between the SIV at the end and the beginning of the month:

$$SIM = T - B \text{ [km}^3 \text{ month}^{-1}\text{]}. \tag{2.2}$$

For a given area composed of multiple grid cells, its border can be decomposed into the smaller segments that separate the grid cells inside the area from those outside. These segments are defined by the edges of the grid cells at the perimeter of the area. To determine T within this area, we have to estimate the SIT, SIC, and sea ice velocity at these borders. This is achieved through linear interpolation between the given values in the two adjacent cells, i.e., the west and east cells for a border along meridians and the north and south cells for a border along the parallels. This interpolation ensures that the estimates of these variables at the cell boundaries accurately reflect the spatial gradients between grid cells. If the individual borders are indexed by i , the transport is calculated as:

$$T = \sum_i SID_i^\perp \times SIT_i \times SIC_i \times l_i, \quad (2.3)$$

where $SID_i^\perp$ is the sea ice drift vector perpendicular to the considered border, counted positive when directed towards the area, and l_i is the length of the border segment.

To calculate the sea ice volume balance (B), it is necessary to determine the SIV at both the beginning and end of the month. Given that we only have monthly data, we performed a linear interpolation as follows:

$$\begin{aligned} SIV_{\text{start}} &= \frac{SIV_{\text{previous month}} + SIV_{\text{current month}}}{2} \\ SIV_{\text{end}} &= \frac{SIV_{\text{current month}} + SIV_{\text{next month}}}{2}. \end{aligned} \quad (2.4)$$

The balance is then given by

$$B = SIV_{\text{end}} - SIV_{\text{start}}. \quad (2.5)$$

KE and EKE

In order to assess the characteristics of the current, we calculated two derived variables based on the surface current velocity: KE and EKE. If u and v denote the current velocity in the zonal and meridional directions, respectively, then we have,

$$\begin{aligned} KE &= \frac{1}{2}(u^2 + v^2) \\ EKE &= \frac{1}{2}(u'^2 + v'^2) \end{aligned} \quad (2.6)$$

where u' and v' are the difference between the instantaneous and the average velocity $u' = u - \bar{u}$ (Reynolds decomposition). It's worth noting that, in our case, the signal may not be stationary and could exhibit trends; therefore, u' will be computed based on the detrended signal.

2.2 Observational data

In the next chapter (Chapter III: ORAS5 validation), we aim to evaluate the accuracy of the ORAS5 reanalysis in depicting the EGC and sea ice phenomena. This requires the use of observational data distinct from those employed in the ORAS5 assimilation, in order to have meaningful comparisons. Here, we provide an overview of the various sources used in our validation. Specifically, sea ice thickness data are sourced from CryoSat-2 (CS2), while sea ice concentration and sea ice drift data are obtained from OSI SAF. Additionally, sea surface salinity (SSS) and SST data are acquired from Copernicus.

2.2.1 CS2

CS2 is a satellite launched on April 8, 2010 by the European Space Agency (ESA), replacing CryoSat-1, which faced a launch failure in 2005. This satellite is specifically designed for SIT measurements and monitoring changes in ice sheets. It is equipped with a synthetic aperture radar (SAR) Interferometric Radar Altimeter, measuring the surface elevation.

SIT remote sensing relies on sea ice freeboard altimetry, which measures the elevation of sea ice relative to sea level. Under the assumption of hydrostatic equilibrium, this measurement provides an estimate of the SIT. This technique has been in use for decades, and we have pan-Arctic winter SIT charts from 1993 to the present day (Laxon et al., 2003; Kwok and Cunningham, 2008; Laxon et al., 2013; Petty et al., 2020; Landy et al., 2022). Unfortunately, summer remote sensing of SIT, between May and September, poses significant challenges due to the presence of saturated water vapor near the sea ice surface, the formation of melt ponds and the lead detection (Min et al., 2019; Dawson et al., 2022). The saturated water vapor can absorb certain parts of the electromagnetic spectrum used by the satellite radar. However, this issue can be mitigated by identifying and avoiding the corresponding absorption lines (Zege et al., 2015).

Melt ponds form when snow on top of sea ice and sea ice melt, resulting in the accumulation of liquid water. During the initial 20 days of melting, the coverage of melt ponds can increase from 10% to 40% (Tschudi et al., 2008). These melt ponds can disrupt radar responses in two ways: by resembling open water channels and by biasing the measurement of sea ice elevation above the water level. Indeed, melt ponds level generally lies below the surrounding sea ice surface (Eicken et al., 2004), causing a positive electromagnetic range bias over ice floes and therefore an underestimation of the sea ice freeboard.

In winter, the radar response between open ocean and sea ice differs sufficiently to allow accurate classification, with leads returning a specular reflection, ice floes producing a lower reflection, and the open ocean yielding a smooth return form (Dawson et al., 2022). In summer, however, the high backscatter specular reflections from leads and melt ponds alter their responses, making them difficult to distinguish. While algorithms for classifying these different responses exist for winter months,

they were inefficient during the summer until recently (Dawson et al., 2022). To address this challenge, Dawson et al. (2022) trained a 1D convolutional neural network on 350 satellite images from summer months. Using this method, they were able to classify ice floes and leads with a precision greater than 80% for May-September between 2011 and 2020. However, their results remain negatively impacted by the presence of melt ponds.

To address the challenge of melt pond classification, Landy et al. (2022) modelled the CS2 radar echo over melt-pond-covered sea ice and performed simulations to characterize the associated electromagnetic bias. They used a Lagrangian snow evolution scheme, SnowModel-LG, to estimate snow cover and convert the measured freeboard into actual SIT. This innovative method resulted in the first year-round record of SIT, spanning from October 2010 to July 2020. The dataset is discussed in the paper Landy et al. (2022). It provides bi-weekly data (twice a month) with a spatial resolution of 80 km.

2.2.2 OSI SAF

OSI SAF is a project within the European Organisation for the Exploitation of Meteorological Satellites (EUMETSAT), which focuses on developing and providing satellite-derived products related to oceanography and sea ice. In this study, we utilize two variables sourced from OSI SAF: the sea ice drift (SID) and SIC. Note that the sea ice concentration data was integrated into the dataset provided by Landy et al. (2022), but it originates from OSI SAF (OSI-401-b).

SID

SID data product is the “OSI SAF Global Low Resolution Sea Ice Drift data record 1991-2020 (v1, 2022), OSI-455” (OSI SAF, 2022). This dataset provides daily sea ice motion vector fields with a spatial sampling of 75 km.

The retrieval methodology differs between winter and summer. In winter, SID vectors are computed using a continuous maximum cross-correlation (CMCC) algorithm on satellite images of sea ice. To do this, maps of daily-averaged satellite signals are first constructed, on which a Laplacian filter is applied to enhance the signal’s intensity pattern, and then a CMCC algorithm on pairs of processed images. Two polarization channels are used by the satellite, and the algorithm directly maximises the sum of the two cross correlations, resulting in one sea ice drift field available and not one per channel. At the end of the process, spurious vectors need to be either discarded or corrected.

For reasons similar to those discussed in the CS2 section, surface melting and atmospheric moisture hinder the retrieval of motion vectors during the summer season. Therefore, a free-drift model is employed, which describes the sea ice motion assuming negligible internal stresses within the sea ice. The parameters required for this model include the transfer coefficient of wind forcing, the deflection angle induced by Earth’s rotation, and the motion vectors caused by ocean currents. These parameters are pre-computed on a monthly basis from time-series data of sea ice drift

and wind vectors. Subsequently, daily sea ice drift vectors are calculated using daily wind forcing data provided by ERA5 and the free drift model with the corresponding monthly parameter values. To ensure that the two different retrieval methods did not produce inconsistencies between the summer and winter seasons, they used both sources during the transition months (May and October), with each method weighted by its error and the specific day of the month.

SIC

The SIC product used in this study is the “Global Sea Ice Concentration (SSMIS)” (OSI-401-b) from OSI SAF. These data are integrated into the dataset from [Landy et al. \(2022\)](#), on the same spatial and temporal resolution.

The detector, the Special Sensor Microwave Imager Sounder (SSMIS), measures the observed brightness temperature. The surface brightness temperature is derived from this measurement after applying corrections that account for air temperature, wind-induced agitation over open water, and atmospheric water vapor, using a radiative transfer model function and NWP data from ECMWF. Following these corrections, SIC is retrieved using a combination of the Bootstrap algorithm ([Comiso, 1986](#); [Comiso et al., 1997](#)) and the Bristol algorithm ([Smith, 1996](#)). This hybrid approach is discussed in [Andersen et al. \(2006\)](#), who demonstrated that the Bootstrap algorithm has the lowest sensitivity to atmospheric noise over open water, while the Bristol algorithm exhibits the lowest sensitivity to ice surface emissivity variability. The OSI SAF dataset is thus constructed using a hybrid algorithm that combines these two methods linearly, with the weighting depending on the SIC. Specifically, the Bootstrap algorithm is weighted more heavily at low SIC, whereas the Bristol algorithm is favored at high SIC.

2.2.3 Copernicus

Copernicus is the European Union’s Earth observation program, aimed at monitoring and providing detailed and accurate information about the Earth’s environment. It provides a broad range of remote sensing data from satellite observations. In this study, we used the “Multi Observation Global Ocean Sea Surface Salinity and Sea Surface Density” product (product ID: MULTIOBS_GLO_PHY_S_SURFACE_M_YNRT_015_013), developed by the Consiglio Nazionale delle Ricerche (CNR) ([Copernicus Marine Service Information, 2021](#)). It provides SSS and sea surface density (SSD) fields with a spatial resolution of $1/8^\circ \times 1/8^\circ$. This product includes four datasets, the “near real time” and “reprocessed” with both daily and monthly resolutions. Here we used the “cmems_obs-ob_glo_phy-sss_my_multi_P1M”, which is the reprocessed one with monthly resolution.

Remote sensing of salinity relies on the detection of conductivity changes caused by salinity variations, which imprint a distinct signature on microwave emissions. This technique represents a challenging measurement method that relies on the accuracy of highly sensitive radiometers. These radiometers are susceptible to various external factors, including surface temperature fluctuations and the presence of galactic background radiation.

The dataset combines data from five different sources. First, the background state is derived from the previous day’s analysis data over the open ocean. Along the coastal regions, it is obtained through a fusion of the SSS climatology from the preceding version of the product and a specifically enhanced Arctic climatology sourced from the Copernicus Marine Service platform (ARTIC_MULTI_YEAR_PHY_002_003). Second, we have the in situ data which are derived from surface measurements obtained from the Conductivity, Temperature, and Depth (CTD) profilers, as well as from Argo float measurements. Two satellite remote sensing sources contribute to the SSS fields: the first one is derived from the L3OS 2Q (Level 3 Ocean Salinity corrected for seasonal latitudinal bias and coastal bias) product generated from data obtained by the Soil Moisture and Ocean Salinity (SMOS) satellite, while the second one is from the Soil Moisture Active Passive (SMAP) satellite. Finally, the OSTIA analysis, developed by the Met Office, is also used.

These different datasets are then combined with the multidimensional optimal interpolation technique initially derived in [Nardelli \(2012\)](#) for in situ data. This method was improved by [Droghei et al. \(2016\)](#) to deliver dynamically consistent SSD fields, and further refined by [Buongiorno Nardelli et al. \(2016\)](#) and [Sammartino et al. \(2022\)](#) to incorporate satellite observations from SMOS and SMAP.

Chapter 3

Validation of ORAS5

Before utilizing the ORAS5 reanalysis, it is essential to verify its accuracy in replicating the structure of current velocity and sea ice distribution, as well as temperature and salinity fields. In this chapter, we compare the ORAS5 reanalysis data with the observational data described in the previous chapter.

First, we examine the representation of sea ice in ORAS5. We assess the ability of this reanalysis to reproduce the seasonal cycle of the sea ice and oceanic variables by comparing the spatial distributions of SIT, SIC, and SID with those obtained from CS2 and OSI SAF in March and September. Additionally, we analyze the interannual variability of SIV and SIE over our area of interest. Subsequently, we compare the sea ice transport through Fram Strait, an important variable for our study, as calculated using ORAS5 SID, SIC, and SIT, with values obtained in two recent studies.

In the second part, we validate the oceanic variables. The validation of the EGC representation in ORAS5 is only qualitative, because the remote sensing of sea surface velocity relies on sea surface elevation measurements and the current flows under sea ice for the majority of the year. Consequently, we evaluate whether ORAS5 accurately reproduces the EGC characteristics as observed in the literature and discussed in chapter II. The same qualitative approach is followed for the vertical profiles of temperature and salinity. Finally, we compare the SSS and SSD from ORAS5 with those obtained from Copernicus.

3.1 Sea ice

3.1.1 Thickness, concentration, volume, and extent

The mean seasonal cycle of SIT observed by CS2 is compared with that from ORAS5 in Fig. 3.1. The figure presents the mean over the eight-year period from 2011 to 2018 for the two extreme months, March and September. Additionally, maps of the mean SIT difference between ORAS5 and CS2 are displayed. Note that the ORAS5 data was interpolated onto the CS2 grid to produce this comparison. This choice to interpolate the ORAS5 grid over the CS2 grid, rather than the reverse, was made to ensure interpolation is performed over the coarser resolution, thereby avoiding the creation of artificial data that would result from interpolating in the

other way around. The same has been done for comparing SIC between ORAS5 and OSI SAF in Fig. 3.2. For the comparison, grid cells with SIC less than 15% and SIT less than 0.15 m have been discarded.

Looking at both figures (Figs. 3.1 and 3.2), we see that the reanalysis accurately represents the sea ice extent in both March and September. In March, sea ice extends as far as the Denmark Strait, while the eastern part of the Fram Strait remains ice-free. The eastward extent of sea ice reaches up to 300 km from the Greenland coast near the Fram Strait and narrows further south. In September, the southernmost extent of sea ice does not extend below 75°N , although the western part of the Fram Strait remains covered by sea ice.

Regarding the SIT values displayed in Fig. 3.1, it is obvious that there is significant disagreement between CS2 and ORAS5 near the coast in March, with CS2 showing sea ice that is one meter or more thicker. Additionally, in March, ORAS5 shows thicker SIT in the western part of Fram Strait and thinner SIT in the eastern part, as well as in the southernmost region. In other regions, the SIT values from ORAS5 and CS2 are relatively similar. The missing grid cells in the southernmost region of the “difference” map are due to interpolation issue arising from the thin distribution of data points. In September, there is generally good agreement between the two datasets, although ORAS5 tends to overestimate SIT compared to CS2. This overestimation is most pronounced in the Fram Strait, where it can reach up to half a meter.

If we look at the SIC maps in Fig. 3.2, there is generally good agreement between ORAS5 and OSI SAF in March, with minimal differences within the sea ice area. However, at the sea ice edge, ORAS5 shows a more eastward-extended sea ice cover, with differences of up to 10% compared to OSI SAF. Similar to the pattern observed in Fig. 3.1, the southernmost region exhibits interpolation issues, though it is evident that sea ice presence is less pronounced in ORAS5 than in OSI SAF in this area. In September, SIC from both datasets are comparable in the Fram Strait and at the southern extension of sea ice near 75°N , but ORAS5 exhibits significantly lower concentration in the central region.

Figure 3.3 illustrates the mean seasonal cycles between 2011 and the end of 2018 of the mean SIT and SIC averaged over the red area displayed in Fig. 3.3 and the SIV and SIE calculated over the same area. The standard deviation of the eight years for each month are also displayed, providing an overview of interannual variability. Note that for the SIV, both SIC and SIT were needed and, therefore, we needed to use data from both CS2 (SIT) and OSI SAF (SIC).

We observe in Fig. 3.3 that the seasonal cycle for the four variables exhibits similar patterns between ORAS5 and CS2/OSI SAF. The maximum values for all four variables are observed in April, while the minimum values occur in September for SIT and in August for the mean SIC, SIV, and SIE. The most significant decrease is noted between June and July for all variables, and the steepest increase occurs between September and October for SIE and SIC, and between October and November for

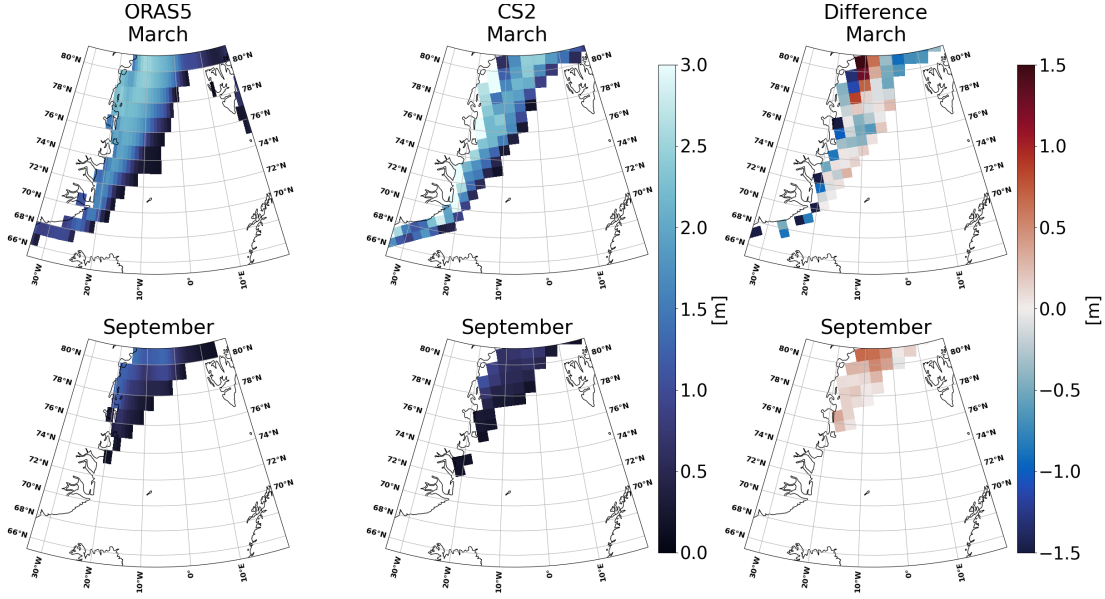


Figure 3.1: SIT averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and CS2. **Right panels:** the difference for the two months between ORAS5 and CS2.

SIT and SIV. The mean SIT reaches a maximum of approximately 0.7 m and a minimum of 0.1 m. The mean SIC has a maximum around 31% and a minimum near 5%. The greatest sea ice volume is noticed in April, with approximately 550 km^3 extending over $4 \times 10^5 \text{ km}^2$, while the minimum is in August, with less than 50 km^3 covering just over $5 \times 10^4 \text{ km}^2$.

Regarding the SIT, ORAS5 consistently overestimates CS2 by a few centimeters throughout the year. This discrepancy is most pronounced in summer but is barely noticeable in April and May. ORAS5’s mean SIT appears to exhibit greater interannual variability compared to CS2, particularly in summer, where CS2 displays low variability. On the mean SIC plot, a disagreement between ORAS5 and OSI SAF is observed in winter, with ORAS5 overestimating the mean SIC by 5%. This overestimation is most obvious in January, where the standard deviation bars do not overlap. However, this discrepancy diminishes in July and August and is generally less pronounced in summer. Additionally, the interannual variability is similar for both ORAS5 and OSI SAF. Regarding the SIV plot, there is good agreement between ORAS5 and the fusion of CS2 and OSI SAF, although, similar to SIT, the interannual spread is wider in ORAS5 during summer. SIE is slightly overestimated by ORAS5 in winter but shows good agreement in summer, with similar interannual variability. It is worth noting that, except for the mean SIC in January, the standard deviation of the different datasets always intersects, indicating consistency.

3.1.2 Transport

The average spatial distributions of SID for March and September as derived from ORAS5 and OSI SAF are compared in Fig. 3.4, following the same presentation as for SIT and SIC. Note that the “difference” map only indicates disagreement on the magnitude of the SID, not on its direction. The OSI SAF SID data covers a

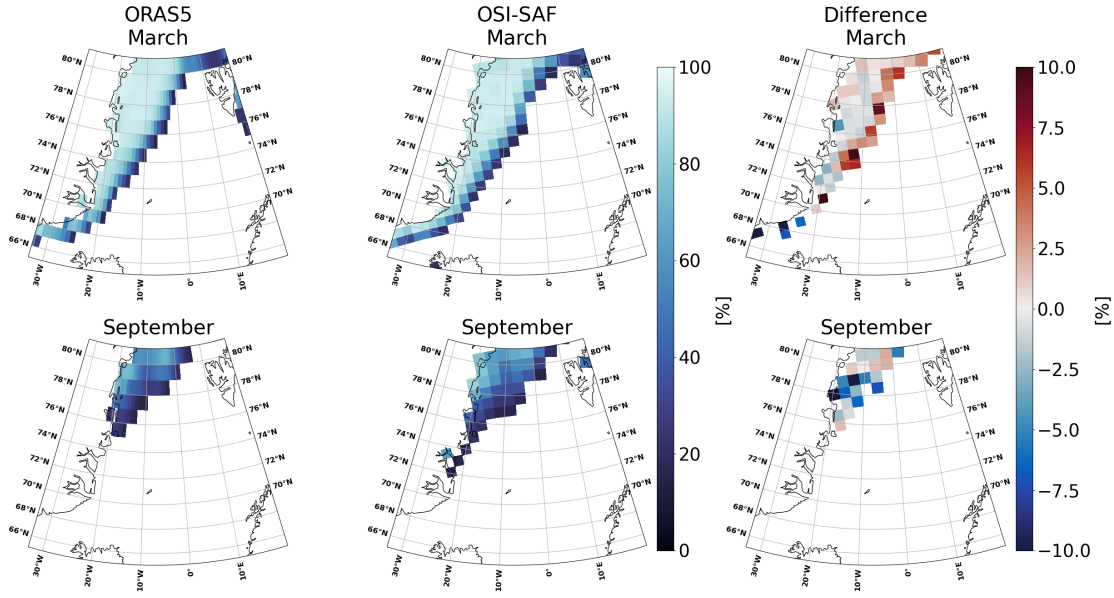


Figure 3.2: SIC averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and CS2. **Right panels:** the difference for the two months between ORAS5 and CS2.

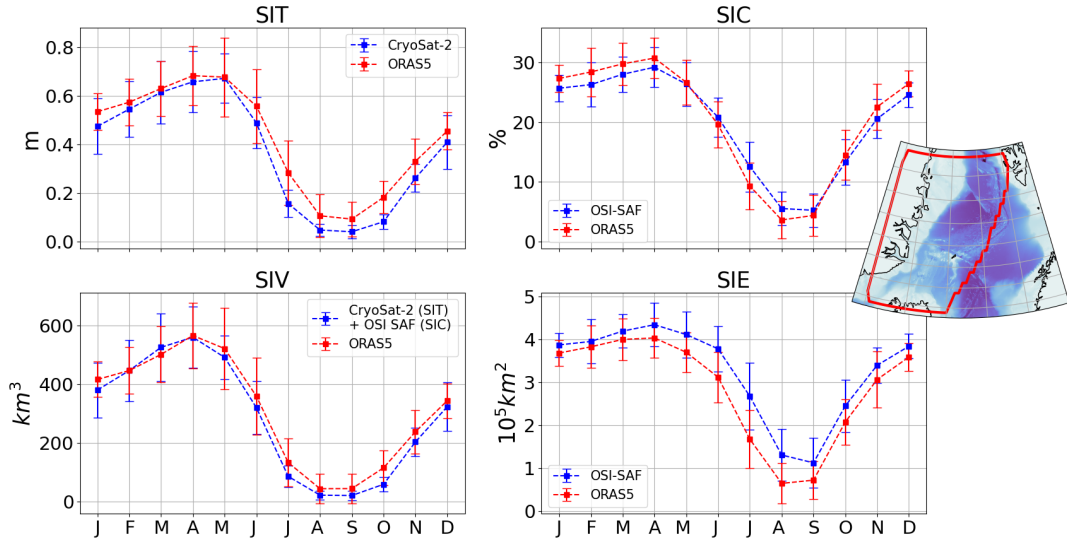


Figure 3.3: Mean SIT, mean SIC, SIV and SIE for both ORAS5 (red) and CS2/OSI SAF (blue). The dashed line represents the average over 8 years and the vertical bars the standard deviations. The area over which the SIT and SIC were averaged and SIE and SIV calculated is displayed in red on the inset map.

narrower area compared to the ORAS5 data, but both datasets are of the same order of magnitude and exhibit similar directional patterns. However, in March, the sea ice drift measured by OSI SAF is stronger in the central regions of the sea ice pack by approximately 5 cm s^{-1} and weaker at the edges compared to the ORAS5 reanalysis. The most significant difference is observed at the southernmost ice-covered region, where OSI SAF records slower sea ice movement, while ORAS5 indicates swift ice flow. In September, the SID is well-reproduced by the reanalysis in both direction and magnitude. The similarity in drift between OSI SAF and ORAS5 in September may partly result from the free drift model used to generate OSI SAF SID during summer, which is forced by ERA5, the same atmospheric analysis used to force ORAS5.

Finally, we assess the ability of ORAS5 to replicate the import of sea ice through the Fram Strait. Previous studies by [Ricker et al. \(2018\)](#) and [Min et al. \(2019\)](#) have quantified this sea ice import through a defined gate consisting of both a meridional segment along 82°N between 12°W and 20°E and a zonal segment between 80.5°N and 82.5°N along 20°E . This gate is highlighted in red in Fig. 3.5. The selection of a more northerly gate position than simply in the Fram Strait aims to mitigate errors and biases in low-resolution ice drift data, which tend to increase with higher ice velocities, typically observed south of 82°N ([Ricker et al., 2018](#)).

In [Min et al. \(2019\)](#), two distinct estimates (referred to as M1 and M2) of this sea ice transport between 2011 and 2017 are provided. The first estimate (M1) was computed using sea ice drift data from OSI SAF and SIT from the Combined Model and Satellite Thickness (CMST), which assimilated observations from CS2 and Soil Moisture and Ocean Salinity (SMOS) satellites. In contrast, the second estimate (M2) used both SID and SIT data from the CMST. On the other hand, [Ricker et al. \(2018\)](#) estimated this sea ice transport (R) employing sea ice drift data from the Polar Pathfinder Sea Ice Motion Vectors dataset, OSI SAF and IFREMER, along with SIT data from CS2. This information is summarized in Tab. 3.1. It is worth mentioning that only estimate M2 provides an annual coverage, while the other two are limited to the winter season.

We compared these findings with those derived from ORAS5 in Fig. 3.5. The total monthly series is presented alongside the mean annual cycle over the designated period. While the time series from ORAS5 extends beyond 2017, the mean annual cycle has been computed over the period for which the other estimates are available, spanning from 2011 to 2017, ensuring a consistent comparison. The error bars represent the standard deviation across different years for the respective months, offering insights into the interannual variability, as we have done in Fig. 3.3. The transport is positive when sea ice passes the gate southward and westward, towards the Nordic Seas.

Looking first at the monthly time series (Fig. 3.5), it is noteworthy that the lowest import occurs in February 2011, as indicated by ORAS5, displaying a negative sea ice import, implying movement towards the Arctic Ocean. Conversely, M1, M2, and R exhibit a slightly positive transport, around $20 \text{ km}^3 \text{ month}^{-1}$. The

Table 3.1: Data used for Fram Strait sea ice transport calculation in [Min et al. \(2019\)](#) (M1 and M2) and [Ricker et al. \(2018\)](#) (R).

Origin	SID	SIT
M1	OSI SAF	CMST
M2	CMST	CMST
R	Polar pathfinder + OSI SAF + Ifremer	CS2

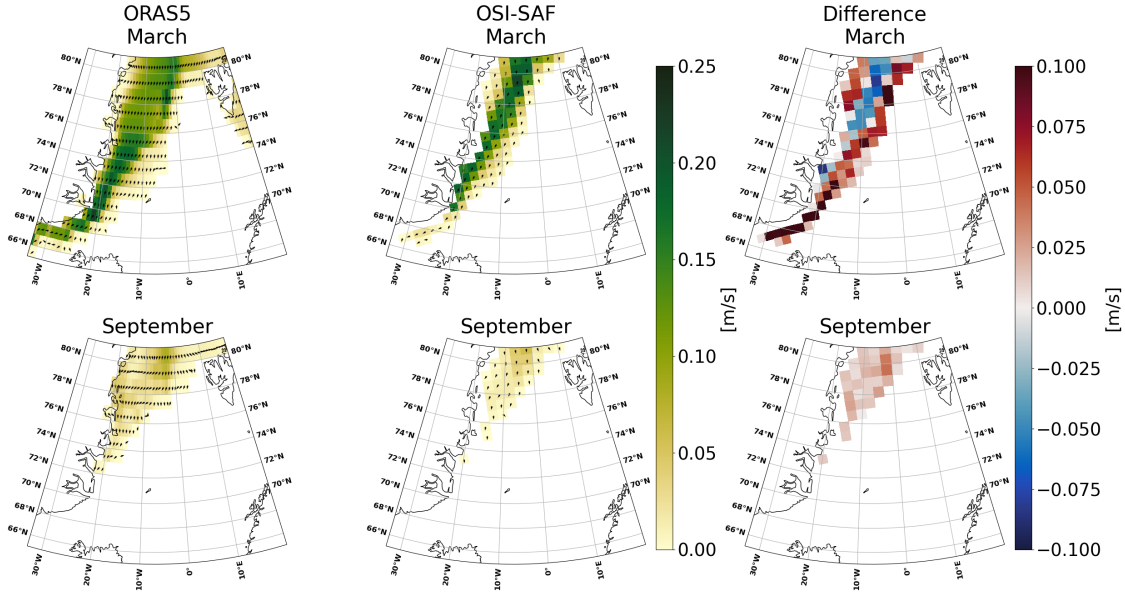


Figure 3.4: SID averaged between 2011 and the end of 2018 for the months of March and September from ORAS5 and OSI SAF. **Right panels:** the difference for the two months between ORAS5 and OSI SAF.

maximum ice transport during the period occurs in the subsequent month, specifically in March 2011, with approximately $450 \text{ km}^3 \text{ month}^{-1}$ of sea ice transport throughout the month, and reaching up to $500 \text{ km}^3 \text{ month}^{-1}$ according to the R estimation. Upon examination of the mean annual cycle plot (Fig. 3.5), we observe that the maximum transport occurs in March, ranging between $250 \text{ km}^3 \text{ month}^{-1}$ and $400 \text{ km}^3 \text{ month}^{-1}$, varying depending on the estimation, while the minimum is observed in September, with slightly under $100 \text{ km}^3 \text{ month}^{-1}$.

We find that ORAS5 consistently underestimates the sea ice transport but remains close to the other estimates. The largest discrepancy occurs between ORAS5 and R, reaching up to $150 \text{ km}^3 \text{ month}^{-1}$. However, even in this instance, the standard deviation bars intersect. There is notable similarity between ORAS5 and M2 during the summer months, with the exception of July, where ORAS5 slightly underestimates the transport. While the three estimates do not extend beyond 2017, it has been documented in [Sumata et al. \(2022\)](#) that 2018 witnessed an unprecedentedly low level of sea ice import through the Fram Strait, a phenomenon that is distinctly visible in our results.

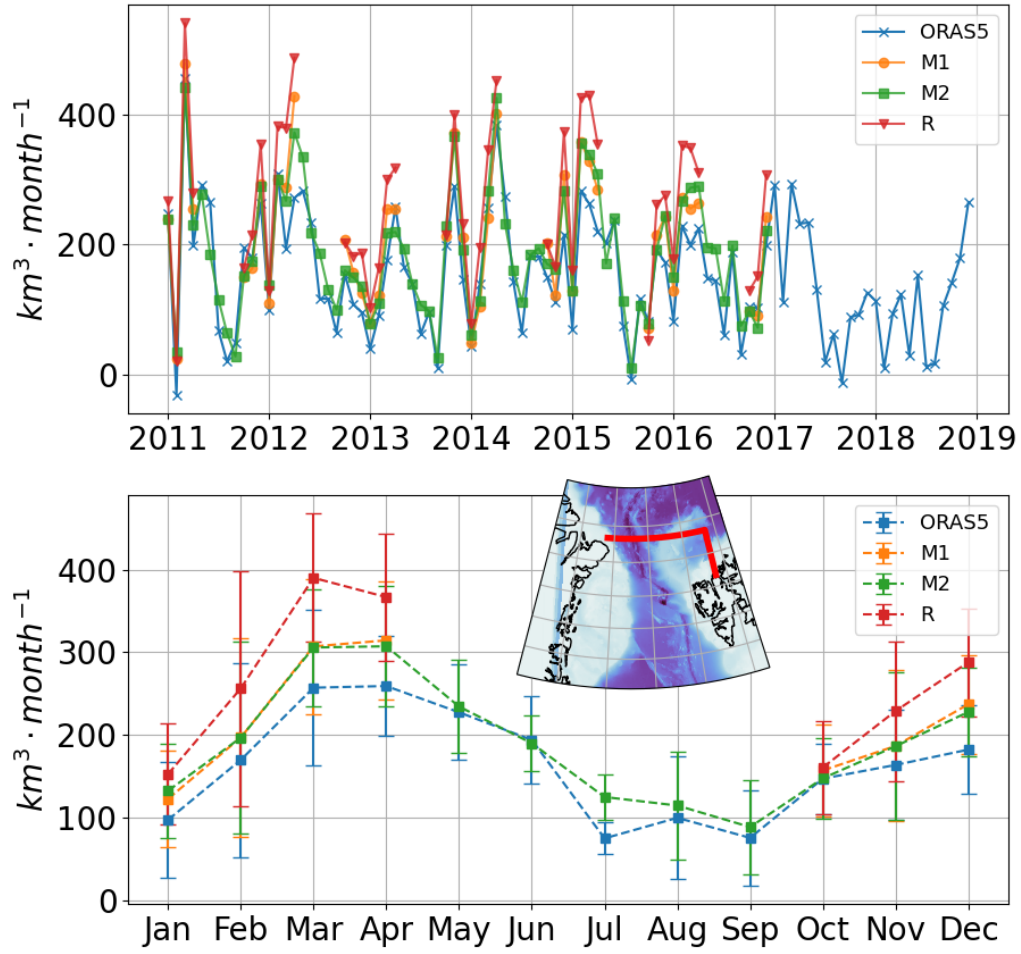


Figure 3.5: Comparison of Fram Strait sea ice transport between ORAS5 - [Min et al. \(2019\)](#) for M1 and M2 - and [Ricker et al. \(2018\)](#) for R. **Upper:** The monthly time series. **Below:** The seasonal cycle calculated by averaging each month between 2011 and 2017. Error bars indicate standard deviation over the 6 years. The gate used for the calculation is displayed in red in the inset map.

3.2 Ocean

3.2.1 Currents

Satellite remote sensing of surface ocean velocity depends on the measurement of sea level anomalies and, thus, cannot be conducted beneath sea ice. Given that the EGC primarily flows under sea ice, a quantitative year round validation of the ORAS5 surface ocean velocity field using satellite measurements is impractical. Consequently, we examine whether the representation of the EGC in ORAS5 aligns with the characteristics and features described in the literature, which were discussed in detail in chapter II.

Figure 3.6 displays the mean surface currents in March and September, averaged from 2011 to the end of 2018. The trajectory of the EGC is discernible between the Fram Strait and Denmark Strait, following the continental margin as shown in Fig. 3.10. The WSC is also visible, flowing northwards through the Fram Strait, with a portion recirculating within the EGC around 79°N. The seasonal variability is evident, with a stronger current in winter compared to summer, indicating that ORAS5 successfully captures this pattern. This variability is not observed in the Fram or Denmark Straits and is confined to the Greenland Sea gyre and the WSC, as noted in [Woodgate et al. \(1999\)](#).

We extracted the vertical distribution of the magnitude of the southward velocity component at four transects spanning the EGC path. We have plotted the average value between 2011 and the end of 2018, as well as the summer (July, August and September) and winter (January, February and March) in Figs. 3.7, 3.8 and 3.9.

Figure 3.7 shows that the main core of the current is centered over the continental slope in the two northern transects and in the southernmost one, while it is located over the continental margin in the third one. The strongest southward velocity component occurs at the surface, approximately 0.1 m s^{-1} , and decreases with depth, falling below 0.05 m s^{-1} between 100 and 150 m. Following this rapid decrease, the southward velocity component of the current remains nearly constant with increasing depth. Some minor northward flows are also observed over the continental margin, indicating the presence of an undercurrent. Additionally, the presence of the WSC flowing northwards is noted to the east of the Fram Strait transect. As expected, we observe that the WSC shows a strong barotropic mode, with velocity barely varying with depth.

A comparison between summer and winter, as illustrated in Figs. 3.8 and 3.9, reveals the seasonal cycle. In winter, the core of the current with a southward velocity component greater than 0.1 m s^{-1} is broader and extends deeper in the Fram Strait compared to summer, with the deep flow in this strait being stronger in winter. Conversely, in the second transect, the surface current is slightly stronger during summer but narrower, though this does not extend to greater depths. In the southernmost transects, the core of the current is both wider and stronger in winter than in summer. The observed undercurrents have a lower velocity compared to the surface southward flow but are slightly stronger during summer. Overall,

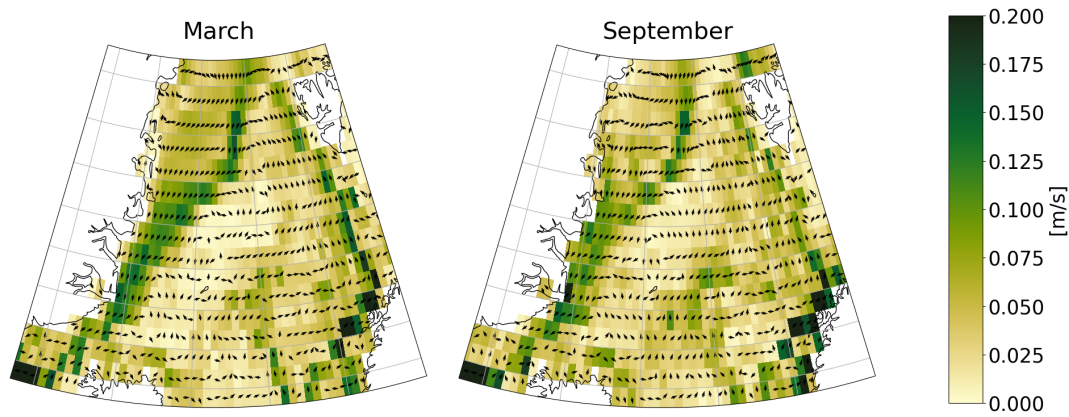


Figure 3.6: ORAS5 ocean surface velocity averaged between 2011 and the end of 2018. **Left:** March. **Right:** September.

baroclinicity is more pronounced in summer than in winter, consistent with what we have seen. The increased baroclinicity observed in summer is consistent with the presence of stronger undercurrents during this season. Examining the WSC in the Fram Strait, we observe that it also exhibits a seasonal cycle, peaking in winter. This observation corroborates the description of the EGC’s seasonal cycle driven by variations in the WSC, as provided by [Woodgate et al. \(1999\)](#) and discussed in Chapter I.

Our results can be compared with vertical sections measured by [Karpouzoglou et al. \(2022\)](#) in the Fram Strait over three periods: 2003-2009, 2009-2015, and 2015-2019. The ORAS5 data shows good agreement with this study, accurately reproducing both the velocity range and the depth extension of the main core, which is confined to depths above 150 m.

3.2.2 Temperature and salinity distribution

Observed vertical profiles of water salinity and temperature in the Greenland Sea are sparse, rendering it difficult to construct significant averages over the period of interest. Additionally, the majority of these profiles have been assimilated into ORAS5, thus they are not fully independent for validation purposes. However, as discussed in Chapter I, [Coachman \(1968b\)](#) provides a detailed description of the water mass vertical distribution near the Fram Strait. Here, we first compare the ORAS5 data with this description. [Rudels \(2002\)](#) analyzed vertical profiles along the EGC path measured by the Polarstern and the Valdivia in 1998, providing an extensive discussion and defining several water masses. The water masses of interest are transcribed in Tab. 3.2. Using these results, we will, in a second time, examine whether ORAS5 accurately reproduces the characteristics of these different water masses.

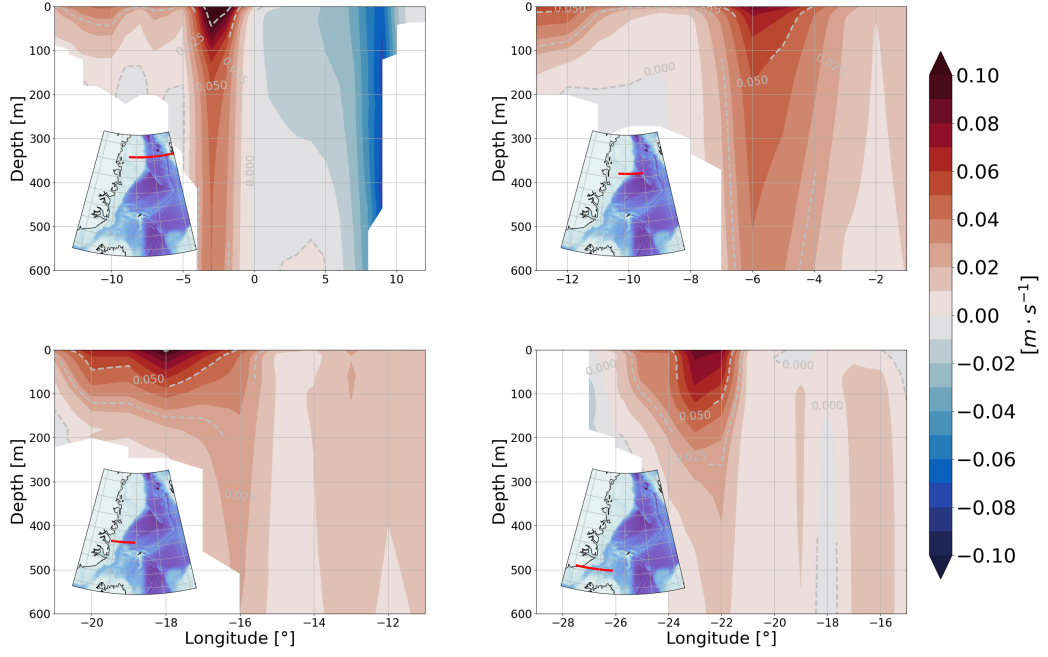


Figure 3.7: Four vertical transects of the mean southward velocity component magnitude. The transect location is indicated in the inset map.

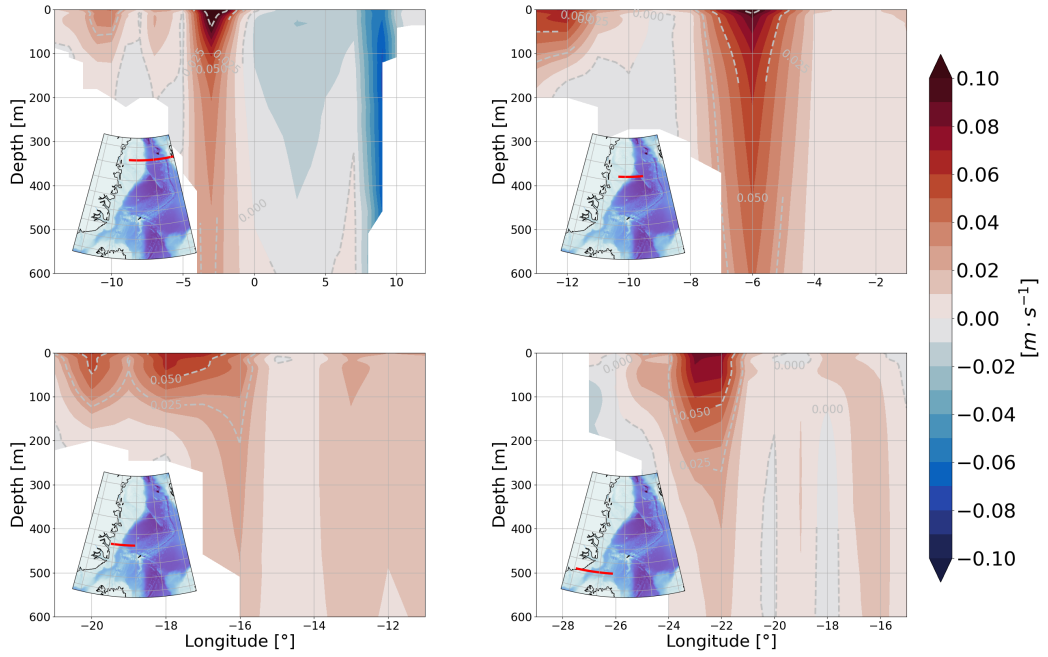


Figure 3.8: Four vertical transects of the summer (July, August and September) mean southward velocity component magnitude. The transect location is indicated in the inset map.

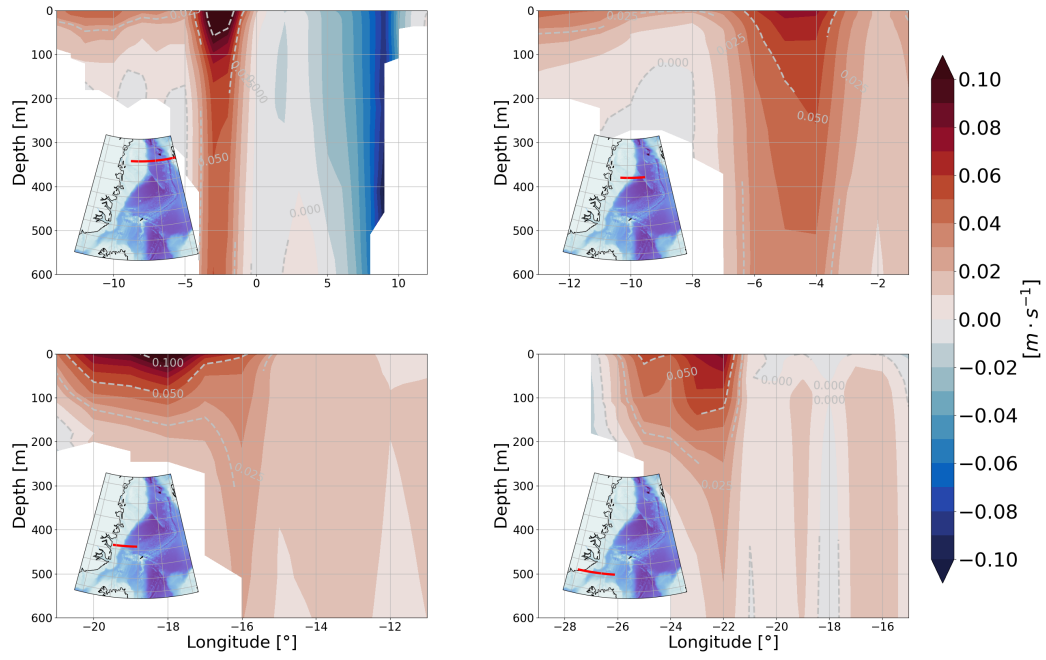


Figure 3.9: Four vertical transects of the winter (January, February and March) mean southward velocity component magnitude. The transect location is indicated in the inset map.

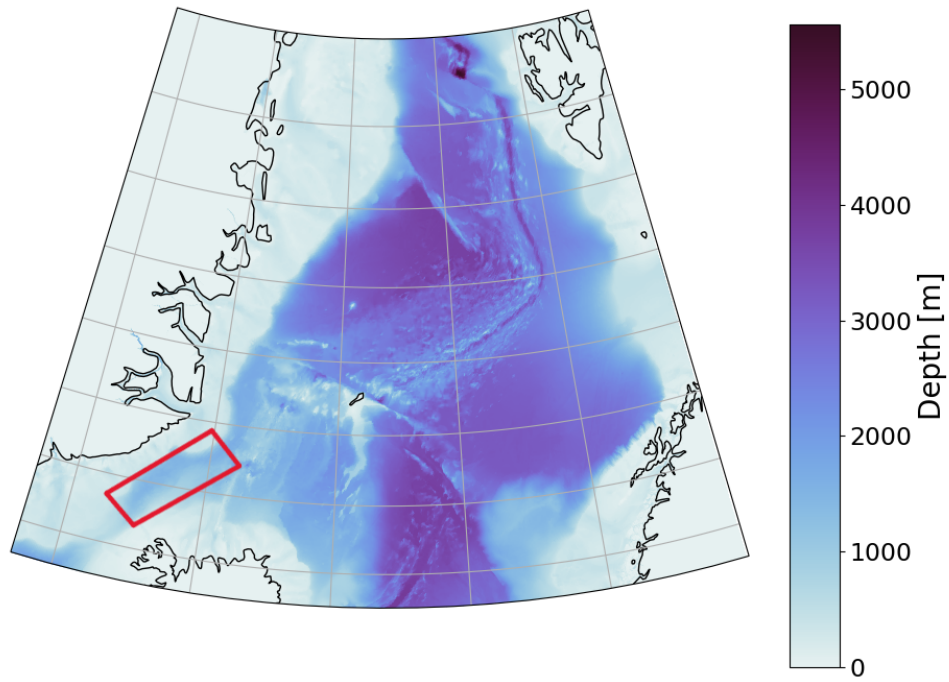


Figure 3.10: Bathymetry of the Nordic Seas. The Blosseville basin is delimited in red.

In Fig. 3.11, we present the T-S diagram obtained at 78.5°N and 3°W, averaged over the period spanning from 2011 to the end of 2018. The selected profile is situated south of the Fram Strait, where the AW has recirculated within the EGC. The T-S diagram is depicted for the entire year, as well as separately for the winter (January, February, and March) and the summer months (July, August, and September). Additionally, the boundaries of the water masses, as defined in Rudels (2002), are given in Tab. 3.2. A seasonal cycle is evident from the surface down to approximately 150 m depth, characterized by warmer and fresher water in summer, and colder and saltier water in winter.

As a reminder, Coachman (1968b) identified three distinct water masses within the EGC near the Fram Strait. The PW occupies the upper layer, extending from the surface to approximately 150 m depth, characterized by temperatures ranging from 0°C to the freezing point. Beneath the PW lies the AIW, found between 150 m and 800 m depth, featuring positive temperature and salinity values that stabilize below 400 m, typically ranging between 34.88 and 35 psu. The DW occupies depths below 800 m, where temperatures once again drop below 0°C. Examining Fig. 3.11, we observe a salinity increase with depth, reaching values between 34.88 and 35 psu generally above 400 m, along with the presence of PW exhibiting temperatures between 0°C and the freezing point in the uppermost 150 m. ORAS5 also captures the local temperature minimum reported by Coachman (1968b) around 50 m depth during summer. However, the depth at which the water temperature falls below 0°C is deeper than 800 m, approximately around 1000 m.

Applying the updated classification introduced by Rudels (2002), we observe in Fig. 3.11 that the upper layer comprises Polar Surface Water (PSW), with a minor presence of warm Polar Surface Water (PSWw) in summer, resulting from the melting of sea ice on warmer AW. This layer is followed by Arctic Atlantic Water (AAW), where the seasonal variability becomes less pronounced. Below 400 m, although the temperature decreases, the salinity remains relatively constant, indicating the presence of Recirculating Atlantic Water (RAW) originating from the WSC, which has descended beneath the PW layer within the Fram Strait. Beyond approximately 1000 m depth, the water temperature once again drops below 0°C, exhibiting salinity characteristics of the upper Polar Deep Water (uPDW) and the Nordic Deep Water (NDW).

Finally, we can juxtapose the SSS and SSD fields from ORAS5 with those derived from the Copernicus analysis. The spatial distributions of SSS and SSD, averaged over the period from 2011 to the end of 2018 for the months of March and September, are presented in Figs. 3.12 and 3.13, respectively. It is noteworthy that due to the coarser resolution of the ORAS5 dataset compared to the Copernicus product, the interpolation used to produce the “difference” map has been applied from the Copernicus grid to the ORAS5 one.

In Fig. 3.12, it is obvious that ORAS5 consistently overestimates SSS in March. This overestimation is relatively minor further offshore, but it increases to approximately 0.8 psu near the Greenland coast, particularly above 75°N. A slight underestimation is also noticeable in the Denmark Strait. The most substantial disparities

Table 3.2: Water masses definition from [Rudels \(2002\)](#).

Water mass	Characteristics	Origin
Polar Surface Water (PSW)	$\sigma_\theta \leq 27.7, \theta < 0$	Arctic Ocean
Polar Surface Water warm (PSWw)	$\sigma_\theta \leq 27.7$ $0 < \theta$	Sea ice melting on warmer Atlantic water
Recirculating Atlantic Water (RAW)	$27.97 < \sigma_\theta, \sigma_{0.5} \leq 30.444,$ $0 < \theta$	West Spitsbergen Current
Arctic Atlantic Water (AAW)	$27.7 < \sigma_\theta \leq 27.97, \theta < 2,$ If $\theta < 0$ then $S < 34.676 + 0.232\theta$	Arctic Ocean
Upper Polar Deep Water (uPDW)	$27.97 < \sigma_\theta, \sigma_{0.5} \leq 30.444,$ $\theta < 0$	Arctic Ocean
Nordic Deep Water (NDW)	$30.444 < \sigma_{0.5},$ $S < 34.915$	Includes the Greenland, Iceland and Norwegian Seas deep waters

between ORAS5 and the Copernicus dataset are observed in September. ORAS5 depicts fresher water in the Fram Strait and extending down to 75°N, with the greatest deviation occurring near the Greenland coast, reaching up to 1.5 psu difference. Conversely, as we move southwards, the discrepancy reverses, with ORAS5 now overestimating SSS, notably around 70°N. These patterns are closely mirrored in the SSD maps (Fig. 3.13), resulting in a more pronounced density gradient in the northern region in ORAS5 compared to the Copernicus product. It is important to consider this information given the central role of these density gradients in our analysis.

3.3 Summary and discussion

We have identified several differences between the datasets, particularly regarding the SIT along the east coast of Greenland in March (Fig. 3.1) and the position of the sea ice edge (Fig. 3.2). ORAS5 exhibits a smaller sea ice extent than OSI SAF but shows a slightly higher mean SIT compared to CS2, resulting in good agreement for the SIV. SID in ORAS5 shows significant differences compared to OSI SAF SID in March, with almost none in September (Fig. 3.4). The sea ice transport through Fram Strait, as computed from ORAS5 data, aligns well, albeit slightly lower, with the calculations by [Min et al. \(2019\)](#). This underestimation is more pronounced compared to the estimation by [Ricker et al. \(2018\)](#) (Fig. 3.5).

It is important to note that the SIC data from OSI SAF were assimilated into ORAS5 through the OSTIA analysis and, therefore, cannot be considered independent validation data. However, it is noteworthy that there is a discrepancy between the two in terms of the sea ice edge position (Fig. 3.2). Additionally, it should be remembered that the SID in summer (and thus September) was calculated using a free drift model forced with ERA5, which is the same atmospheric forcing used in ORAS5. This explains the good agreement observed between ORAS5 and OSI SAF

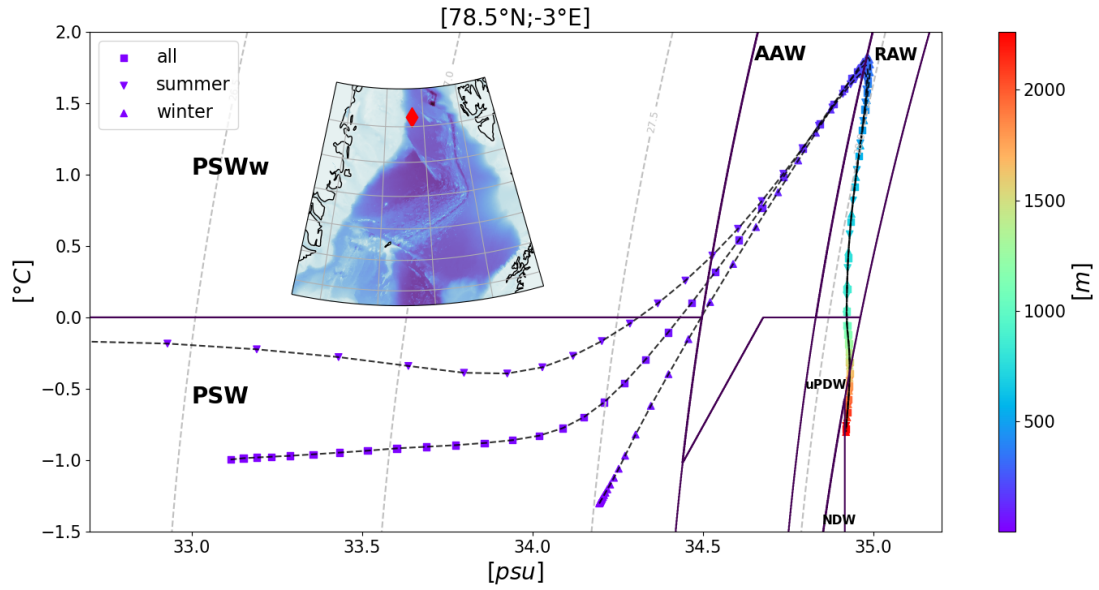


Figure 3.11: T-S diagram from ORAS5 at 78.5°N and 3°W. **All**: averaged over each month between 2011 and the end of 2018. **Winter**: January, February, and March. **Summer**: July, August, and September. The colour indicates the depth.

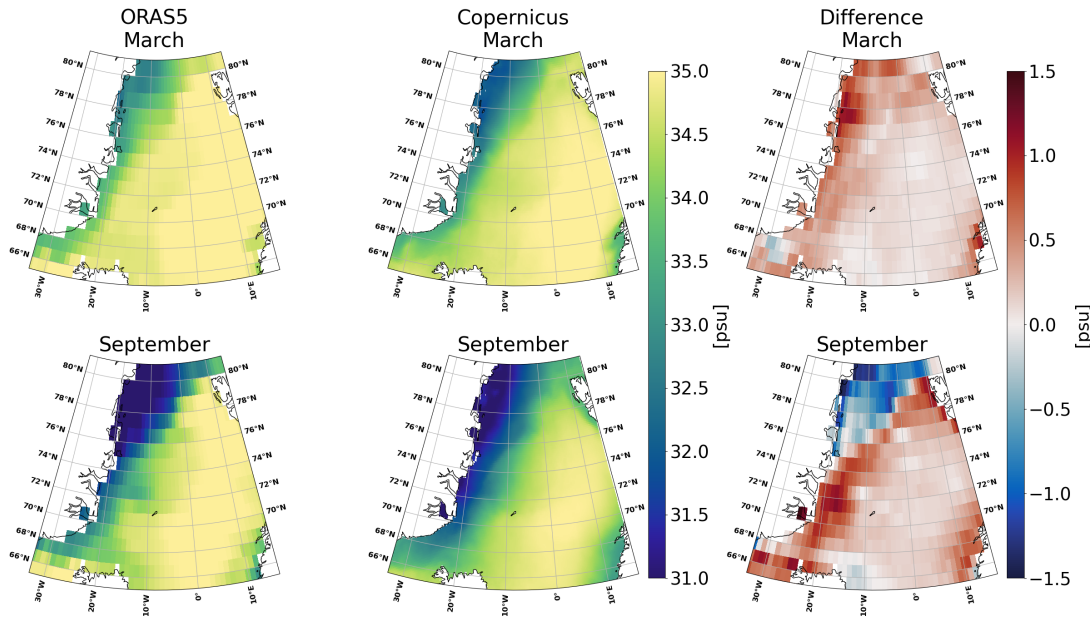


Figure 3.12: SSS climatological mean (between 2011 and the end of 2018) map for the months of March and September from ORAS5 and Copernicus. **Right panels**: the difference for the two months between ORAS5 and Copernicus.

SID in September, in contrast to the differences observed in March (Fig. 3.4).

The EGC as represented in ORAS5 exhibits characteristics consistent with those described in the literature, following the continental slope (Fig. 3.6). Seasonal variability is evident, with the current peaking in winter and displaying stronger baroclinicity in summer (Figs. 3.7, 3.8 and 3.9). The vertical distribution of differ-

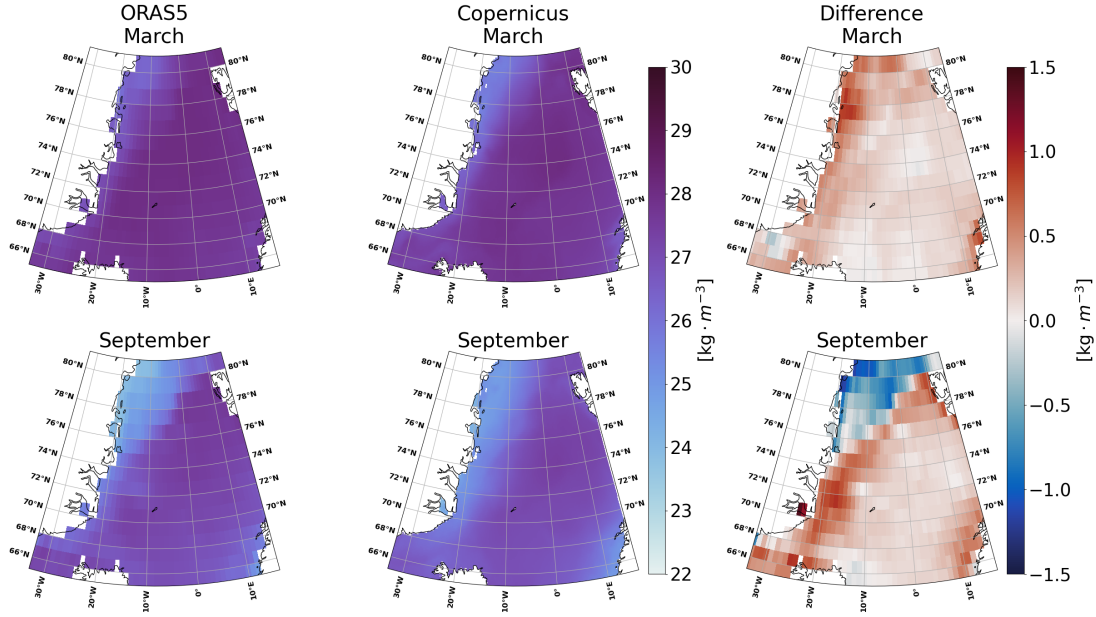


Figure 3.13: SSD anomaly climatological mean (between 2011 and the end of 2018) map for the months of March and September from ORAS5 and Copernicus. **Right panels:** the difference for the two months between ORAS5 and Copernicus.

ent water masses, where the WSC recirculates, aligns with what we should expect (Fig. 3.11). The SSS and SSD fields in ORAS5 differ from those in the Copernicus analysis, with ORAS5 showing saltier and denser surface water in the Nordic Seas and fresher, less dense water in the northwest. It is important to note that the Copernicus analysis, which relies partly on satellite remote sensing, is less accurate beneath sea ice.

Chapter 4

Results

In this chapter, we answer our research question: *“What is the impact of the sea ice accelerated melting on the velocity structure of the East Greenland Current?”*. Our hypothesis is that a reduction in the presence of sea ice in the EGC area would lead to a decrease in the release of low salinity water due to melting along the east coast of Greenland. This could consequently affect the zonal density gradient and alter the baroclinic component of the EGC. To explore this, we divide our main question into four sub-questions:

1. *“What are the observed changes in the EGC velocity structure over our study period?”*
2. *“Are the observed changes in the EGC velocity structure related to shifts in its baroclinic component due to perturbations in ocean density profiles?”*
3. *“Are these density changes related to the sea ice accelerated melting?”*
4. *“What future evolution of the sea ice melting in the EGC region can we expect?”*

This research is conducted using the ORAS5 dataset, whose accuracy and reliability between January 2011 to December 2018 were discussed in the preceding chapter.

Foremost, we will address our first question by examining the evolution of the EGC through its trends in surface KE and EKE, as well as vertical transects of the southward velocity component. Subsequently, we address our second question by investigating the evolution of zonal density gradients. Using the dynamical method, we quantify the impact of these density profile variations to determine their potential contribution to the observed current changes. We also examine the evolution of the wind forcing along the EGC path and qualitatively discuss the possibility that it completes the explanation for the observed changes. We then address our third question by exploring the connection between the alteration in baroclinicity and changes in sea ice. Firstly, by exploring the influence of salinity on these density changes, and secondly by calculating the SIM in the area and studying its potential impact on salinity distribution. Finally, we will answer our last question by exploring the underlying factors influencing change in SIM and explore its future evolution in the light of sea ice transport projections across the Fram Strait.

4.1 Changes in the EGC velocity structure

In this first section, we address the question “*What are the observed changes in the EGC velocity structure over our study period?*”. To do so, we examine the trends between 2011 and the end of 2018 in surface KE and EKE, as well as the vertical distribution of the southward velocity component.

We computed the surface KE and EKE for each grid point and for each month between 2011 and the end of 2018. This resulted in monthly time series for these variables for each $1^\circ \times 1^\circ$ cell. To mitigate seasonal variability, we applied a one-year running average to these time series. We then analyzed the trends, as depicted in Fig. 4.1. The mean path of the current is indicated by a black line, representing cells where the mean KE over our study period is at least 70% of the total mean KE across all grid cells. Additionally, a scatter plot shows all significant KE trends (marked by black dots on the maps) within the EGC against all EKE trends, whether significant or not.

Notably, a negative trend in KE is evident between 75°N and 79°N , reaching a magnitude as low as $-1.2 \text{ mJ kg}^{-1} \text{ year}^{-1}$. This decline translates into a reduction of approximately 20 – 30% in the current’s KE within this spatial domain over the entire study period. Conversely, a positive trend is observed southwards, particularly close to the Blosseville Basin (see Fig. 3.10) near the Denmark Strait. Although weak positive trends are identifiable offshore of the EGC, minor fluctuations are also observed between the EGC and the Greenland coast in its northern sector. We also note a slight decrease of the surface KE over the WSC region.

Regarding EKE, a slight increase is found along the entirety of the EGC’s trajectory, with occasional small decreases in areas where KE is also decreasing. Notably, there is a significant rise in EKE, approximately $+0.5 \text{ mJ kg}^{-1} \text{ year}^{-1}$, over the Blosseville Basin, coinciding with an increase in KE in the same region. This relationship between KE and EKE is corroborated by the scatter plot, which shows that positive KE trends are consistently associated with positive EKE trends, as observed in the Blosseville Basin. Conversely, negative KE trends do not exhibit a consistent pattern with EKE trends.

We also calculated the trends in the southward velocity component between 2011 and the end of 2018 along four transects spanning the EGC region. The results are plotted in Fig. 4.2. These vertical transects clearly indicate a marked weakening of the current in the northernmost region, with the southward velocity component decreasing by up to -0.01 m s^{-1} per year. This weakening is more pronounced in depths above 150 m and occurs above the continental shelf on the western side of the main core, as observed in Fig. 3.7. A smaller decrease in velocity can be seen between 3°W and 1°W , extending down to 600 m and peaking around 150 m. In the second transect along the 76.5°N parallel, we observe a decreasing southward velocity component above the continental slope, peaking at the surface and limited to the upper 200 m. Further east along this transect, there is a slightly uniform positive trend between 50 and 450 m depth, decreasing near the surface. On the easternmost

side, a negative trend is observed again, smaller than that on the continental slope, and constant with depth down to 450 m.

Along the third transect, corresponding to the 72.5°N parallel, we observe only minor trends, predominantly confined to the surface with the upstream signal of the current intensification reaching $+0.004 \text{ m s}^{-1}$. Lastly, near the Denmark Strait, along the southernmost transect, the primary core of the current (refer to Fig. 3.7) exhibits a strengthening trend, while its flanking sides display weakening tendencies. This suggests a narrowing of the current with an intensification at its center. These changes reach magnitudes of up to $0.008 \text{ m s}^{-1} \text{ year}^{-1}$, both positively and negatively.

4.1.1 Summary

In summary, the surface energy maps reveal contrasting signals: a weakening of the northern segment of the EGC, which is not necessarily accompanied by commensurate changes in EKE, juxtaposed with a reinforcement of both the mean and turbulent flow near the Denmark Strait. Upon examination of the depth distribution of the current, it is observed that the most significant decrease in the northern part of the current is confined to depths above 150 m over the continental shelf, while deeper slowing down is also observed above the continental slope. The upstream signal of the current intensification is already evident around 72.5°N in the third transect. Lastly, the strengthening in the southernmost part decreases with depth and appears to be associated with a narrowing of the current.

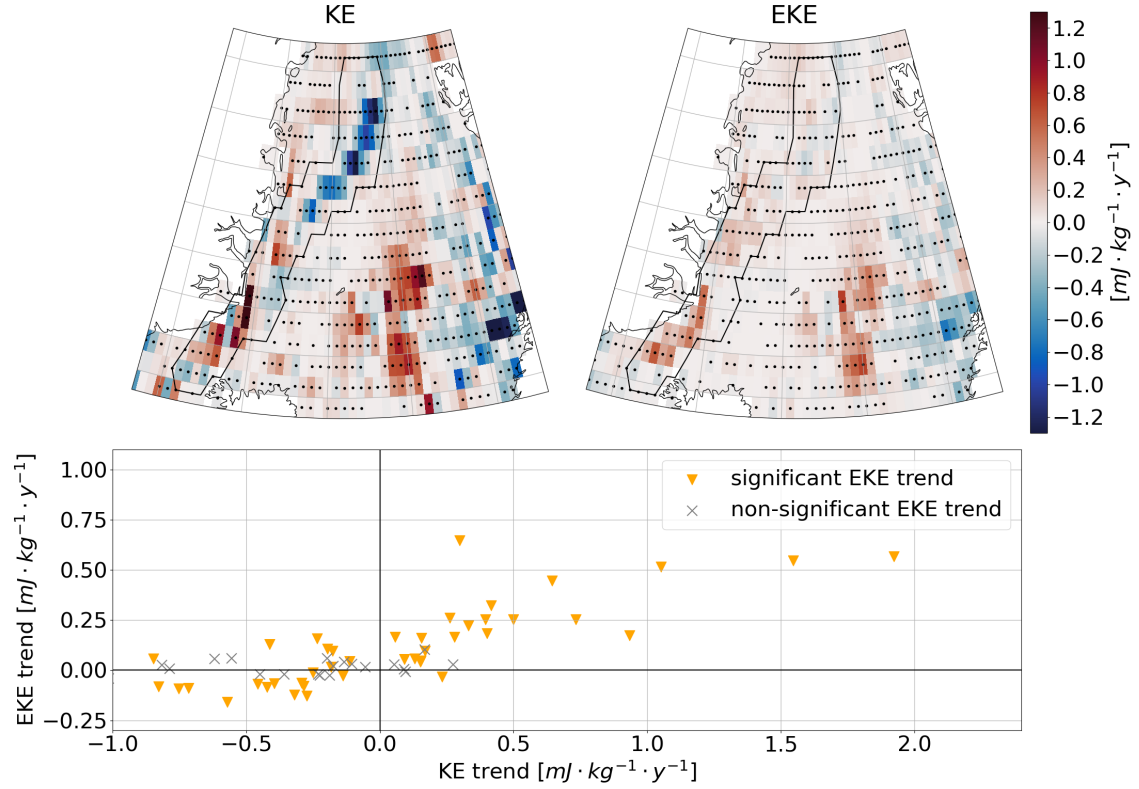


Figure 4.1: **Upper left:** Trend of KE. **Upper right:** Trend of EKE. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. **Bottom:** Scatter plot of the significant KE trends and EKE trends (both significant and non-significant).

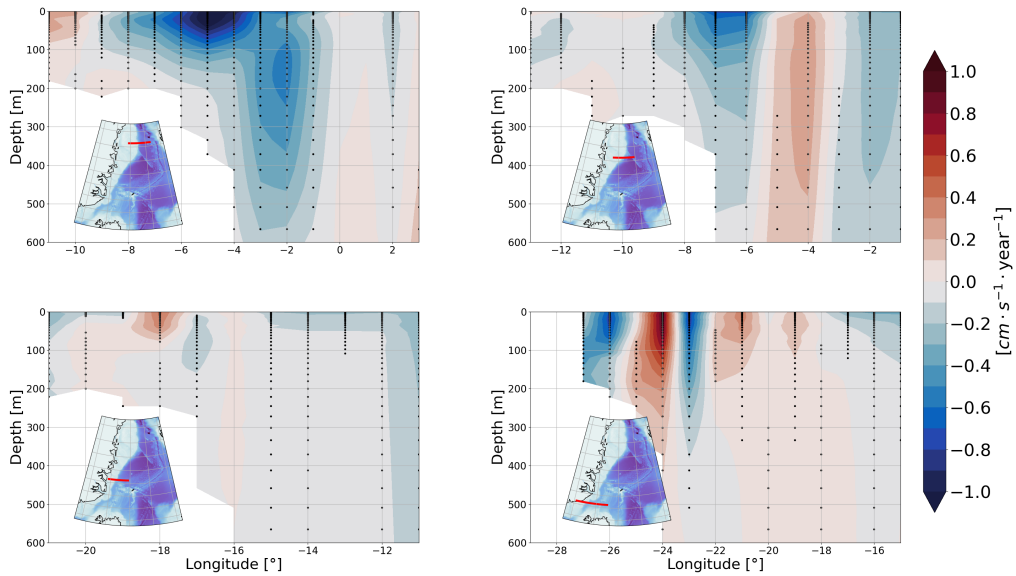


Figure 4.2: Trends of the southward velocity component between 2011 and the end of 2018 along four transects spanning the EGC path. Positive (negative) trends indicate a strengthening (weakening) of the southward flow. Statistically significant trends are marked with a dot.

4.2 Changes in the EGC drivers

In this section, we answer the question “*Are the observed changes in the EGC velocity structure related to shifts in its baroclinic component due to perturbations in ocean density profiles?*”. Consequently, we examine the evolution of zonal ocean density gradients and determine that they partially explain the observed changes in the EGC velocity structure. Additionally, through a qualitative analysis of wind evolution over the area, we explore its potential contribution to the remaining alterations in the EGC.

4.2.1 Zonal density gradients

The first step is to analyze the evolution of zonal density gradients in the ocean. To do this, we have established five transects perpendicular to the direction of the current, focusing on regions where significant changes in EGC KE have been observed: three in the weakening area and two in the strengthening area. For each transect, we have computed the zonal ocean density gradient between adjacent cells at every depth and every month, subsequently analyzing the trends of these density gradients over the study period.

The resulting trends are depicted in Fig. 4.3. We have plotted in this figure the KE trends over the EGC path and the position of the five transects as well as the trends of the zonal ocean density gradient for every depth above 200 m. Through this analysis, we observe a clear correlation between the weakening segments of the EGC and diminishing zonal ocean density gradients, and conversely, strengthening segments coincide with intensified density gradients (Fig. 4.3). This suggests that the trends in EGC KE can be elucidated by changes in its baroclinic component: a reduction (increase) in zonal ocean density gradients leads to a weakening (strengthening) of the baroclinic flow, subsequently manifesting as a weakening (strengthening) of the total current strength in the north (south).

4.2.2 Quantitative analysis with the dynamical method

We have observed that changes in the EGC velocity structure corresponded with variations in the zonal ocean density gradients. In the northernmost part of the domain, the weakening of the EGC is associated with a reduction in zonal ocean density gradient, while in the southernmost part, the strengthening of the EGC occurs simultaneously with an increase in density gradient. This suggests that the observed changes in the EGC may be attributable to variations in its baroclinic component. However, this assessment is qualitative; a quantitative investigation is necessary to determine the extent to which variations in density gradients can account for the changes in current strength.

To do so, we use the dynamical method, connecting the cross-section velocities to along-section density gradients. In this section, we first provide a succinct review of the dynamical method. Subsequently, we propose a hypothesis facilitating the computation of surface current evolution in response to observed ocean density gradient changes. We proceed with the implementation of the dynamical method

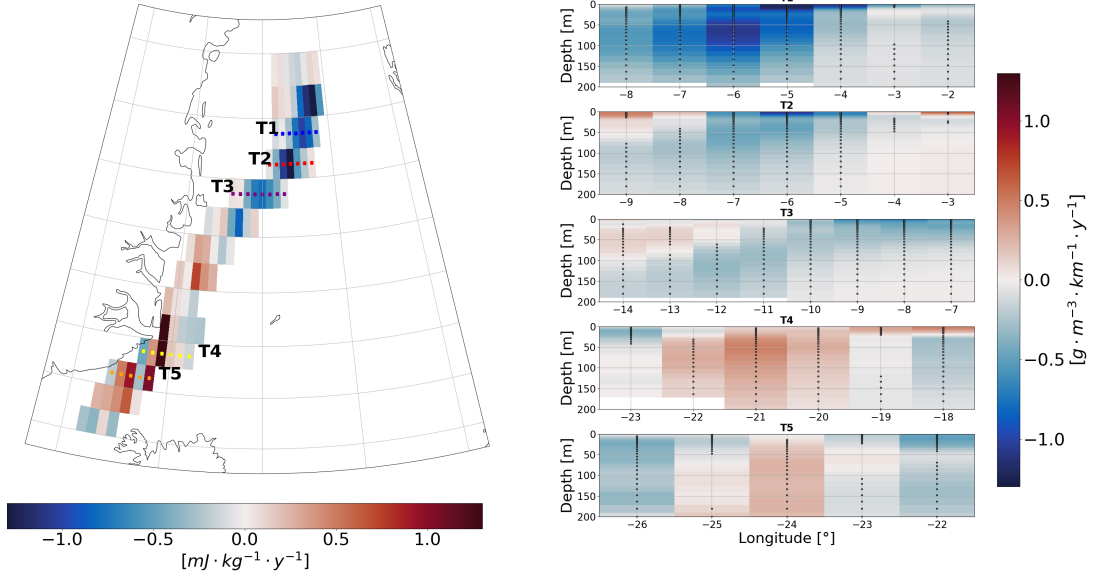


Figure 4.3: **Left:** Map of the KE trend between 2011 and the end of 2018 with the 5 transects (**T1**, **T2**, **T3**, **T4**, **T5**). **Right:** Trend of the zonal ocean density gradients between 2011 and the end of 2018 for each transect for each depth level above 200 m. Statistically significant trends are marked with a dot.

and calculate the surface velocity changes, taking into consideration the documented evolution of density gradient.

Theoretical framework

According to the geostrophic assumption, the current can be separated into two components: a barotropic component, which remains constant with depth, and a baroclinic component, which varies with depth and depends on zonal ocean density gradients. The dynamical method calculates the baroclinic component of the geostrophic current based on these density gradients (Fomin, 1964). In this section, we provide a concise overview of this method and derive a procedure for calculating surface velocity changes over time in response to density changes. We have the expression of the geostrophic current,

$$\mathbf{u}_g = \frac{1}{f\rho_{ref}} \mathbf{k} \times \nabla p \quad (4.1)$$

with f the Coriolis parameter, ρ_{ref} a density of reference, $\mathbf{k}$ the basis vector pointing upwards and p the pressure. With the hydrostatic assumption, we have the pressure in function of depth z given by

$$p(z) = p_s + \int_z^\eta g\rho dz' = p_s + g\langle\rho\rangle(\eta - z) \quad (4.2)$$

with p_s the atmospheric pressure, $\langle\rho\rangle = \frac{1}{\eta-z} \int_z^\eta \rho dz'$ and η the sea level anomaly. If we neglect atmospheric pressure variation, we have

$$\nabla p = g\langle\rho\rangle\nabla\eta + g(\eta - z)\nabla\langle\rho\rangle \quad (4.3)$$

and then the geostrophic current at a depth z is given by

$$\mathbf{u}_g(z) = \underbrace{\frac{g}{f} \mathbf{k} \times \nabla \eta}_{\text{barotropic comp.}} + \underbrace{\frac{g(\eta - z)}{f \rho_{ref}} \mathbf{k} \times \nabla \langle \rho \rangle}_{\text{baroclinic comp.}} \quad (4.4)$$

where we approximated $\langle \rho \rangle$ by ρ_{ref} in the first term. We see here that if the zonal ocean gradient of the vertically integrated density is not null, the geostrophic current will have a baroclinic component, characterized by a dependence in depth. We can use the thermal wind equation to compute the velocity difference between two depths, as follows:

$$\begin{cases} u_g(z) - u_g(z_1) = \frac{g}{f} \int_{z_1}^z \frac{1}{\rho_{ref}} \frac{\partial \sigma(z')}{\partial y} dz' = \frac{g}{f} \frac{\partial D}{\partial y} \\ v_g(z) - v_g(z_1) = -\frac{g}{f} \int_{z_1}^z \frac{1}{\rho_{ref}} \frac{\partial \sigma(z')}{\partial x} dz' = -\frac{g}{f} \frac{\partial D}{\partial x} \end{cases} \quad (4.5)$$

Where $D = \int_{z_1}^z \frac{\sigma}{\rho_{ref}} dz'$ is called the dynamical depth and $\sigma = \rho - \rho_{ref}$ is the density anomaly.

The conventional application of the dynamical method involves identifying a reference depth where the baroclinic current is zero and subsequently using the thermal wind equation to calculate the current at a specific depth. However, to determine how temporal changes in the density gradient affect the surface current, we can rely on the hypothesis that there exists a depth at which the current remains constant over time. We now consider the thermal wind equation for the meridional velocity between a given depth z_1 and the surface z_0 at two different moments, t_1 and t_2 ,

$$\begin{cases} v_g(z_0, t_1) - v_g(z_1, t_1) = -\frac{g}{f} \frac{\partial D(t_1)}{\partial x} \\ v_g(z_0, t_2) - v_g(z_1, t_2) = -\frac{g}{f} \frac{\partial D(t_2)}{\partial x} \end{cases} \quad (4.6)$$

and we take the difference between those two equations, with the assumption that the current at depth z_1 does not change between t_1 and t_2 , i.e. , $v_g(z_1, t_1) = v_g(z_1, t_2)$,

$$v_g(z_0, t_2) - v_g(z_0, t_1) = \frac{g}{f} \left[\frac{\partial D(t_1)}{\partial x} - \frac{\partial D(t_2)}{\partial x} \right]. \quad (4.7)$$

It is then possible to calculate the change in the surface baroclinic component associated with the change in underwater density profiles. The hypothesis that there is a depth z_1 where $v_g(z_1, t_1) = v_g(z_1, t_2)$ relies on the fact that, in deep regions, velocities do not change much, as can be seen in Fig. 4.2.

Application to the EGC

Using the dynamical method, as expressed by eq. 4.7, we conducted calculations to determine the meridional surface velocity change corresponding to observed density variations. This implies to calculate the dynamical height in each grid cell for

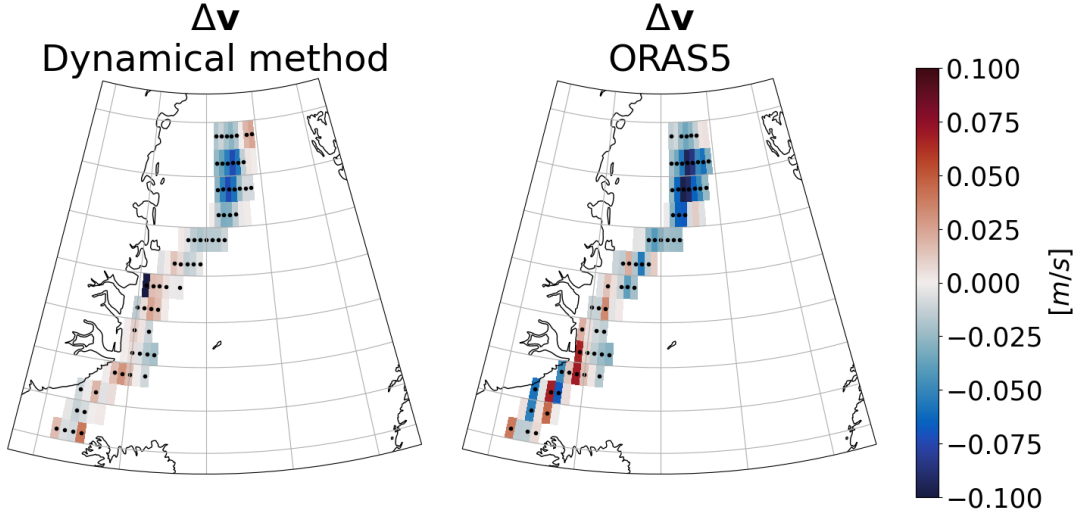


Figure 4.4: Change of the southward component of the current between 2011 and the end of 2018. **Right:** As computed by the dynamical method. **Left:** As given in ORAS5. Statistically significant changes are marked with a dot.

each month, as well as the zonal component of its horizontal gradient, $\frac{\partial D}{\partial x}$. We then computed the linear trends of these series for each cell and identified the two extreme values of these linear trends as $\frac{\partial D(t_1)}{\partial x}$ and $\frac{\partial D(t_2)}{\partial x}$ in eq. 4.7. Consequently, we obtained the two corresponding values, $v_g(z_0, t_2)$ and $v_g(z_0, t_1)$, and compared these to the extreme values of the linear trend of the actual meridional velocity present in ORAS5.

The resulting analysis is depicted in relation to the observed changes in ORAS5 in Fig. 4.4. Our findings reveal that the dynamical method successfully replicates the observed weakening of the EGC in the northern region, alongside a modest increase in the southern segment. From a comparison with observed trends, it becomes evident that the direction of change is largely consistent between the dynamical method and observations. However, it is notable that the dynamical method consistently underestimates the magnitude of change. This underestimation is particularly pronounced in the southern region compared to the northern one, where the disparity between calculated and observed values is relatively minimal.

4.2.3 Wind stress

As demonstrated, variations in the baroclinic component can elucidate a portion of the observed changes in the EGC, with more pronounced effects encountered in the northern region compared to the southern region. In this context, we aim to explore whether wind stress can clarify the remaining unexplained variations through a qualitative analysis of its temporal evolution.

The wind stress used to force the ocean surface in ORAS5 are natively presents in the dataset (see Tab. 2.1). We have computed the trends of its magnitude and direction, in the same way as described for KE and EKE. Note that the angle trends

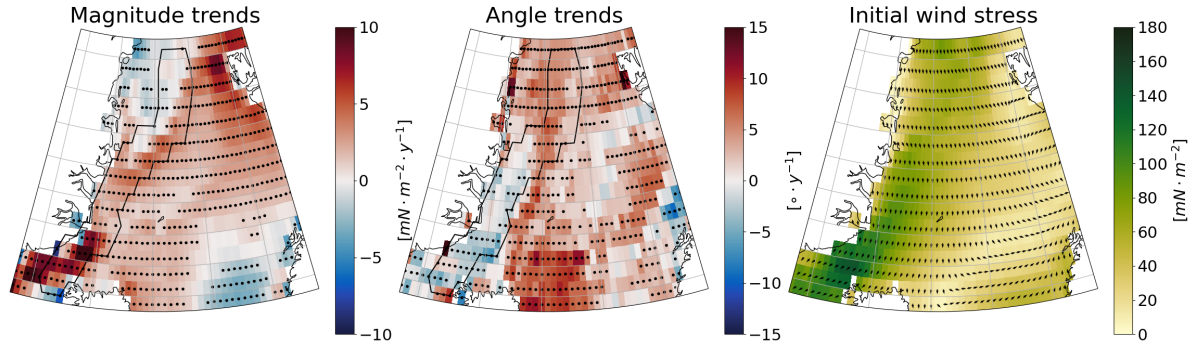


Figure 4.5: **Left:** Trends of surface wind stress magnitude between 2011 and the end of 2018. **Center:** Trends of angle between 2011 and the end of 2018, given in anti-clockwise degree per year. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. **Right:** Initial direction and magnitude of the wind stress.

are given in anti-clockwise degree of rotation per year. The initial wind stress is also displayed, calculated as the intercept of the trend analysis.

We see in Fig. 4.5 that the wind stress is initially strongest in the Denmark Strait and weakest in the northern region. The direction of the wind stress is about 45° to the left of the current direction in the north and align approximately with the EGC in the south. It is important to note that surface wind stress induces a surface current directed 45° to the right. This relationship is maintained between the wind stress and the EGC in the north, but the angle is insufficient in the south. Looking at the magnitude changes (Fig. 4.5), we see that the wind stress is intensifying across nearly the entire region, with the exception of the western Fram Strait, where slight negative or negligible trends are observed. A pronounced increase is obvious over the Blosseville Basin near the Fram Strait and in the eastern part of the Fram Strait. Additionally, there is a significant shift in wind stress direction: anticlockwise in the northern part of the EGC and clockwise in the southern part. Interestingly enough, no significant change in wind stress direction is observed over the central part of the EGC. These shifts cause the wind stress to significantly deviate from the expected 45° to the left of the current by opening the angle to the north and closing it to the south, resulting in an almost parallel alignment between the EGC and wind stress directions near the Denmark Strait.

Examining the trend in wind stress magnitude reveals that the previously observed intensification of the current near the Denmark Strait coincides with a concurrent wind stress strengthening in the same region, making the simultaneous increase in both KE and EKE reasonable. Conversely, the observed decline in EGC KE between 75°N and 79°N seems contradictory to the strengthening trend of the wind stress in this area. However, the shift in wind stress direction in the northern region results in a misalignment with the current, which may explain the weakening of the current. A similar misalignment due to a shift in wind direction is observed in the Denmark Strait. Nevertheless, the strengthening of the wind stress in this narrow passage would still contribute to an increase in EKE.

4.2.4 Summary

In summary, our analysis suggests that the observed changes in the EGC velocity structure can be partially explained by alterations in zonal ocean density gradients, resulting in weakening in the northern part and strengthening in the south. Using the dynamical method, we quantified that these changes accounted for a substantial portion of the weakening observed in the north, but only a minor portion of the strengthening in the south. Through a brief qualitative examination of the evolution of wind stress over the region, we suggest that it may elucidate the fraction of EGC changes not accounted for by the baroclinic alterations.

4.3 Role of the sea ice in the ocean density field perturbation

We will now look toward our third research question, “*Are these density changes related to the sea ice accelerated melting?*”. To address this question, we first demonstrate that the observed weakening in the north and strengthening in the south of the zonal ocean density gradients are attributable to a salinity increase along the east coast of Greenland. Secondly, we investigate whether the evolution of salinity in this area can be partly explained by the evolution of SIM.

4.3.1 Salinity as driver of the density field changes

Given that density is influenced by both salinity and temperature, the objective of this section is to ascertain the respective contributions of each of them to the observed variations in density. A linear approximation of the density anomaly in terms of deviations from reference temperature (T_0) and salinity (S_0), is expressed as

$$\sigma = \sigma_0 + \rho_{ref}(-\alpha_T[T - T_0] + \beta_S[S - S_0]), \quad (4.8)$$

with α_T the thermal expansion coefficient and β_S the haline one. If we consider the density anomaly variation between two points x_1 and x_2 , $\Delta\sigma = \sigma(x_2) - \sigma(x_1)$, we have its expression given in term of the temperature and salinity variation between these two points, $\Delta S = S(x_2) - S(x_1)$ and $\Delta T = T(x_2) - T(x_1)$,

$$\Delta\sigma = \rho_{ref}(\beta_S\Delta S - \alpha_T\Delta T). \quad (4.9)$$

To assess the relative impact of temperature and salinity changes on the density gradient, [Rudnick and Ferrari \(1999\)](#) introduces a dimensionless parameter denoted R , which is defined as:

$$R = \frac{\alpha_T\Delta T}{\beta_S\Delta S} \quad (4.10)$$

If R is less than 1, the density gradient is primarily influenced by the salinity gradient, whereas if R is greater than 1, the thermal gradient becomes the dominant contributor. We calculated this parameter along the same transects depicted in Fig. 4.3 and illustrated the results in Fig. 4.6. We observe that, overall, the salinity gradient exerts a primary influence on the density gradient along the EGC path. However, in the northernmost transects, particularly on the eastern side at depth, the influence of temperature becomes dominant. This phenomenon can be attributed to the recirculation of warm AW within the WSC in this region.

Therefore, we know that, in average during our study period, the predominant influence comes from the salinity. However, we still need to compare the time evolution of density to the one of salinity and temperature over our study period. We have calculated the trends of salinity, potential temperature and density, at the surface and at 50 m depth over the area and plotted the results in Fig. 4.7.

Firstly, focusing on the temperature maps, a nearly uniform warming trend of approximately one-fifth of a degree per year is discernible in the Nordic Seas. This

warming is present both at the surface and at a depth of 50 m. However, notable deviations from this trend include a comparatively weaker warming observed between the path of the EGC and the east Greenland coast north of 75°N , as well as cooling observed north of the Fram Strait at a depth of 50 m and a pronounced warming in the center of the seas at 50 m depth. This observed warming is coherent with what we have seen from [Mork et al. \(2019\)](#) and [Tsubouchi et al. \(2021\)](#).

Regarding salinity, its spatial distribution exhibits less uniformity, both in space and depth. At the surface, a slight negative trend is observed predominantly to the east of the EGC all over the Nordic Seas, as found by [Mork et al. \(2019\)](#). Conversely a positive trend, reaching up to 0.1 psu per year, is noted on the western part of the EGC in the northern part of the domain and on the eastern part in the southern part. Additionally, a significant freshening trend is noticed in the northern part of the Fram Strait, particularly pronounced at the surface compared to depths. Examining the salinity trends at a depth of 50 m, we observe that the positive trend location coincides with the region of weakening of the EGC, between 75°N and 79°N . Finally, the density map reveals trends that closely mirror those observed in salinity, exhibiting a consistent pattern across both surface and depth measurements.

The observed increase of salinity induces variations in the zonal ocean density gradient, albeit with differing spatial characteristics between the northern and southern regions. In the southern part of the domain, the salinity increase above the offshore side of the EGC contributes to an increase in the zonal ocean density gradient, thereby reinforcing the EGC, consistent with our calculations using the dynamical method. Conversely, in the northern EGC domain, the positive salinity trend is located on the western side of the EGC, resulting in a weakening of the density gradient.

An additional distinction lies in the depth penetration of these trends: in the southern part of the domain, the trends are prominently present at the surface but diminish rapidly with increasing depth, becoming indistinguishable at a depth of 50 m. In contrast, in the northern part of the domain, the positive trend persists and even intensifies at greater depths. We can also observe this in the transect T1 in Fig. 4.3, where the greatest trend in zonal ocean density gradient is found between 50 and 100 m depth. Bearing in mind that changes in surface baroclinic velocity are contingent upon the depth-integrated alterations in density gradient, we anticipate a more pronounced alteration of the baroclinic flow in the northern segment, which is coherent with our dynamical method calculations.

We have displayed in Fig. 4.8 the mean winter and summer geographical distribution of the SIT, SSS, and SSD anomaly. Consistent with our previous calculations, the density field mirrors the spatial pattern observed in the salinity field. The distribution of salinity (and consequently, density) exhibits seasonality, characterized by heightened baroclinicity during summer months, as outlined in Chapter III. Upon juxtaposing the salinity and sea ice maps, a notable observation emerges: regions of low salinity are predominantly confined beneath the sea ice cover, with this effect being more pronounced during summer compared to winter. The observed reduction in salinity during summer along the east Greenland coast can thus be attributed in

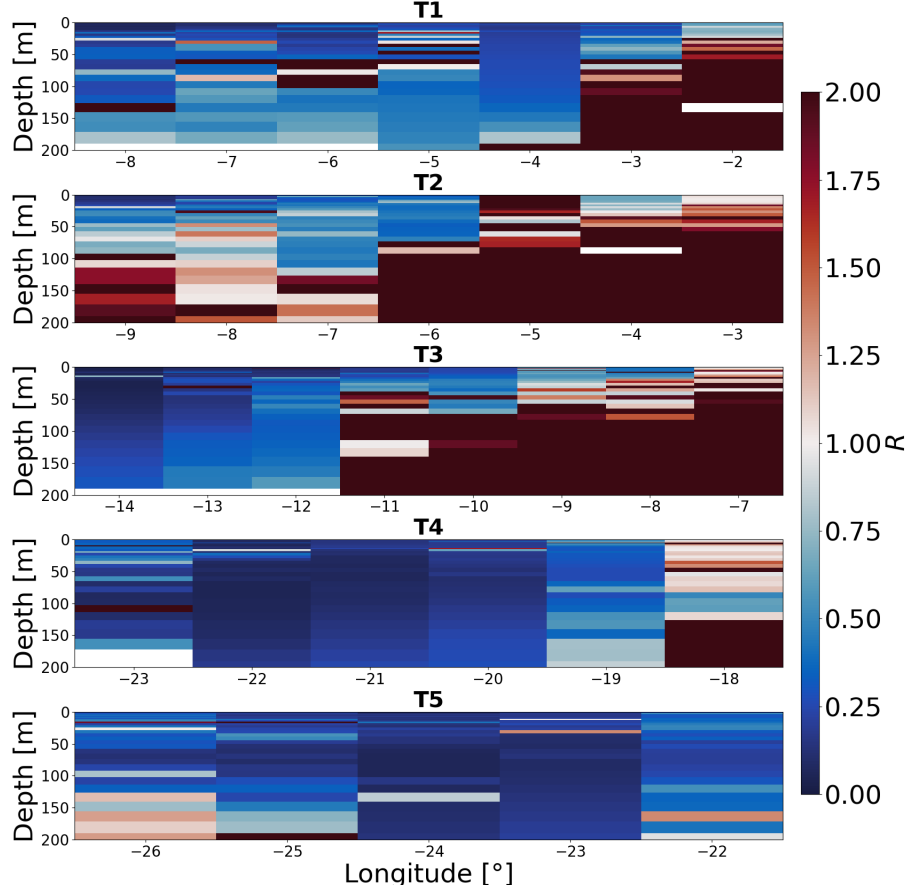


Figure 4.6: R factor as defined by eq. 4.10 averaged between 2011 and the end of 2018 for the five transects as defined in Fig. 4.3.

part to the melting of sea ice in the vicinity, thus releasing low salinity water into the surrounding oceanic environment. This observation motivates the investigation of a link between changes in sea ice cover and the observed salinity trends.

4.3.2 Sea ice trends

To investigate the potential impact of a change in the sea ice cover on the observed salinity increase discussed in the preceding section, we first examine the evolution of sea ice in the EGC region over our study period.

We have calculated the trends of SIT and SIC, and plotted the results in Fig. 4.9. Our analysis indicates a statistically significant negative trend for both SIT and SIC, spanning the entire area, consistent with the broader trend of decreasing sea ice observed across the Arctic region.

Specifically, concerning SIT, the most significant decrease is observed in the vicinity between the core of the EGC and the east Greenland coast, with negative trends reaching up to 12 cm year^{-1} . Conversely, the negative trend decreases when moving offshore or southwards. In contrast, for SIC, we observe the most pronounced negative trends near the sea ice edge, where negative trends of up to $4\% \text{ year}^{-1}$ are

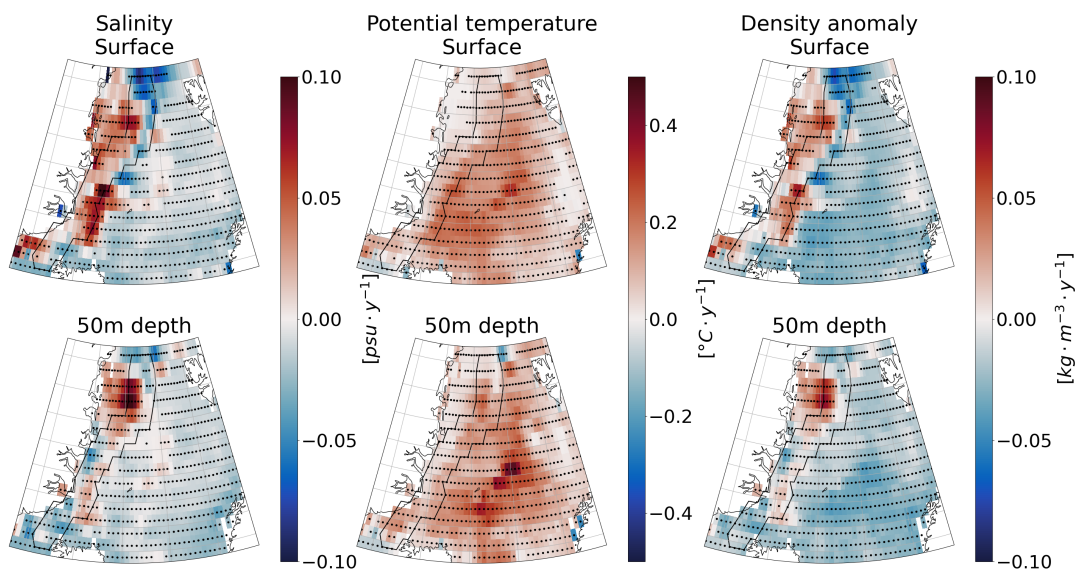


Figure 4.7: Trends of salinity, potential temperature and potential density anomaly between 2011 and the end of 2018 at the surface and at 50 m depth. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC.

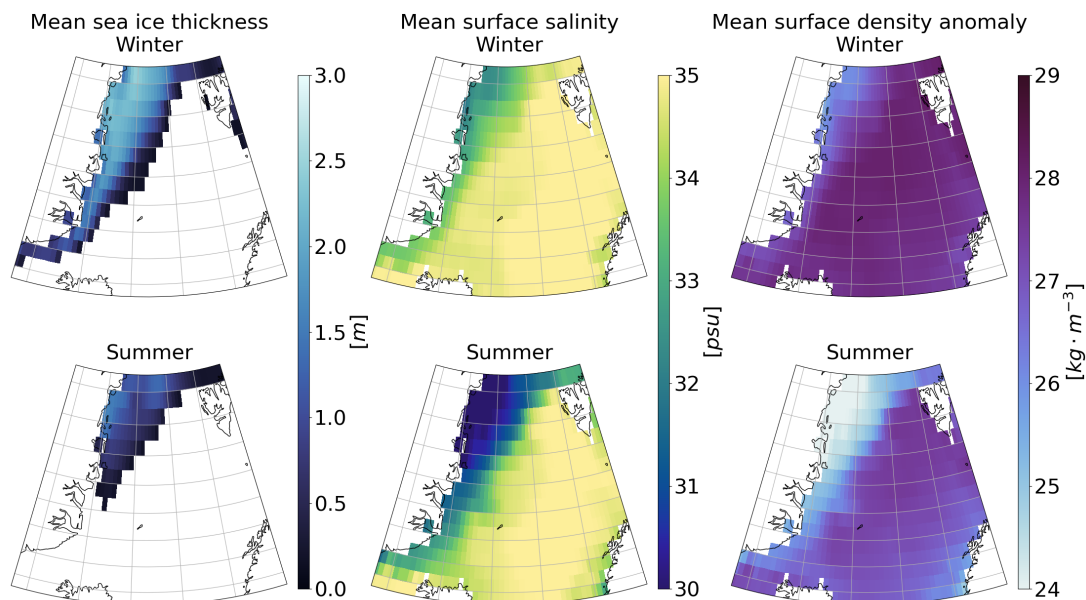


Figure 4.8: SIT, SSS and SSD anomaly averaged over winter (January, February and March) and Summer (July, August, September).

noted. However, within the sea ice pack, the decrease in SIC is less marked, exhibiting trends ranging between -2 and 0% year $^{-1}$. This difference between SIT and SIC trends is because the SIC map shows the retreat of the sea ice front, revealing a SIC reduction near the sea ice edge, hence the more pronounced negative trends in this region. In contrast, the trend in SIT is less conspicuous in these areas due to the thinner ice at the sea ice edge.

Additionally, we calculated the monthly sea ice transport through four meridional gates, spanning the EGC area, denoted as **G1**, **G2**, **G3**, and **G4**. A one year smoothing average has been applied to the time series to filter out the seasonal cycle and focus on the long term-trends. The results are also shown in Fig. 4.9.

The sea ice transport demonstrates a significant decline over the study period for all four gates considered. It is observed that the sea ice transport are similar for **G1** and **G2**, as well as for **G3** and **G4**. Specifically, the sea ice transport through the Fram Strait (**G1**) exhibits a significant decrease initially, reaching a minimum a few months before 2014, a trend also observable in **G2**. It is followed in **G1** by an increasing trend between 2014 and 2015 and then an almost constant decrease up to 2018. At **G2**, the transport increase is smoother and extend until 2016 and is also followed by a smooth decrease prior to 2018. After 2018, a steep decrease is observed at the four gates, coherent with the findings of [Sumata et al. \(2022\)](#), who reported an unprecedented low sea ice transport through the Fram Strait in 2018.

Conversely, **G3** and **G4** display a signal characterized by three distinct plateau: one centered around 2013, another between 2015 and 2016, and a third one between 2017 and 2018. Similarities in the shape of time series between **G1** and **G2** on one side, and **G3** and **G4** on the other side, suggest that the sea ice melting between **G1** and **G2**, as well as between **G3** and **G4**, appears to be relatively constant over time. However, there are indications of changes in sea ice melting rate between **G2** and **G3** over the period.

We computed the monthly SIV and SIE for the entire region. As usual, a one-year smoothing average was applied to highlight the interannual trends and hide the seasonal cycle. We have plotted the results in Fig. 4.10. The total SIV decreased from 400 km^3 in 2012 to slightly more than 200 km^3 in 2019, which corresponds to a decrease of almost half of the total volume, corresponding to a SIV reduction of $-27 \text{ km}^3 \text{ year}^{-1}$ over the period. Such a decline is also observed for the SIE, although to a lesser extent, with a reduction of around one fifth.

In light of these results, it is clear that the sea ice cover has undergone a dramatic decline throughout our study period. We could expect that this reduction would result in a decrease in the release of low salinity water associated with SIM, which could explain the observed increase in salinity. Therefore, we investigate the relationship between the SIM and salinity in the following sections.

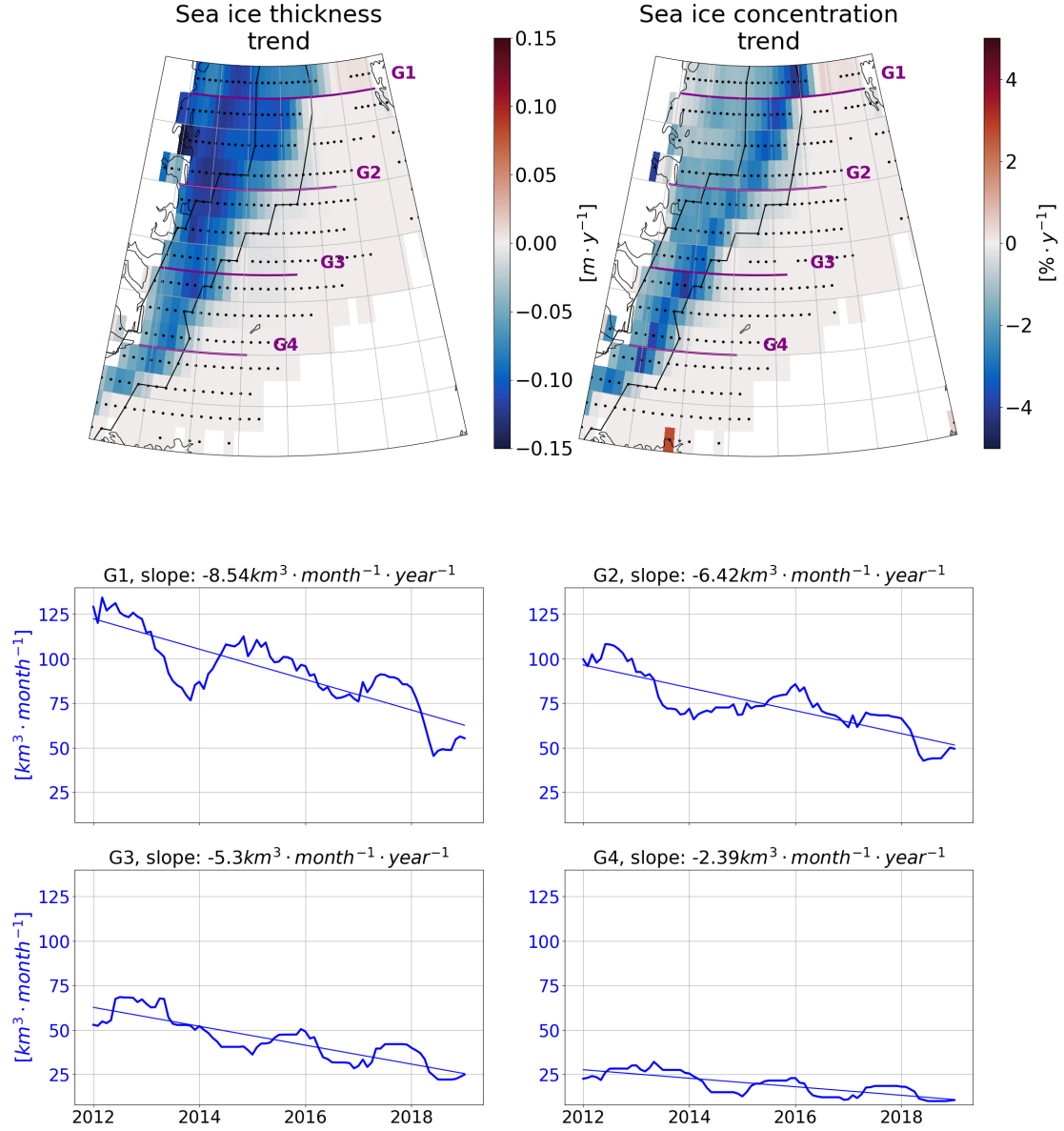


Figure 4.9: **Upper:** Trends of SIT and SIC between 2011 and the end of 2018. Statistically significant trends are marked with a dot and a solid black line delineates the mean path of the EGC. **Bottom:** Sea ice transport through the four gates displayed on the maps above. A one year smoothing average was applied to the time series and the linear trends are displayed, with the corresponding slopes indicated on the top of the figures.

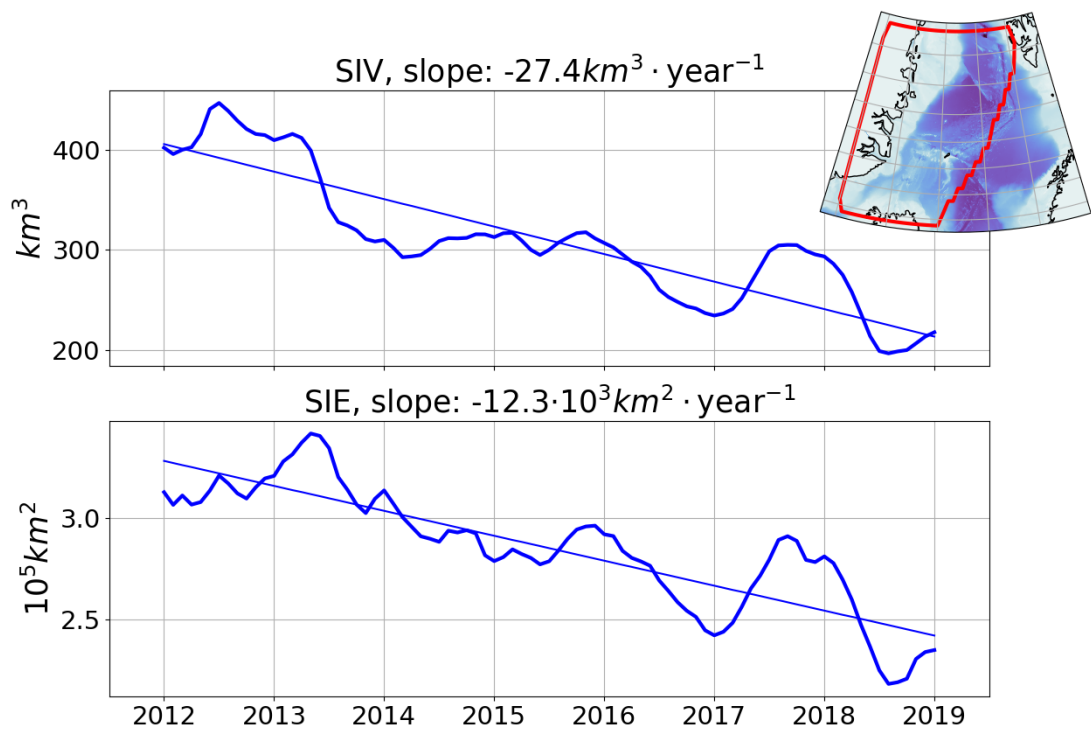


Figure 4.10: Evolution of SIV and SIE calculated over the area delineated in red between 2011 and the end of 2018. A one year smoothing average was applied to the time series. The linear trends are displayed, with the corresponding slopes indicated on the top of the figures.

4.3.3 Correlation between SIM and SSS

To explore the relationship between the observed increase in salinity and sea ice dynamics over the region, we have computed the thermodynamic melting of sea ice at each grid point. Subsequently, we calculated the trends and determined the correlation between SIM and SSS. The correlations were computed based on one-year averaged smoothed signals to mitigate any spurious correlations arising from the seasonal cycle. The trend of the time series has also been removed when computing correlation. The results are displayed in Fig. 4.11.

As can be seen from this figure the dominant correlation is negative, as anticipated: decreased SIM correlates with increased SSS. However, the map exhibits some degree of noise, with sporadic instances of positive correlation that may seem counter intuitive. Analyzing the trend map of SIM (still in Fig. 4.11), we observe even greater noise, although discernible patterns are still present. At the sea ice edge, there is a trend of decreasing SIM over the study period. Along the east coast of Greenland, where the EGC converges, both positive and negative trends in SIM are intermingled. Furthermore, in the northern region between the EGC core and the east coast of Greenland, trends are minimal and exhibit a mix of positive and negative values.

Indeed, while the reduction of sea ice at the sea ice edge straightforwardly leads to a decrease in SIM, complexities arise within the sea ice cover. As time progresses, less ice is available to melt. However, simultaneously, sea ice that initially crossed regions without melting may now undergo melting, consequently increasing SIM, especially in the northernmost regions. These competing effects give rise to the noisy signals observed on the SIM map.

To address this challenge, we extended the calculation of SIM over broader areas than individual grid cells and compared it to the mean SSS over these defined regions. To define these regions, we extended two cells in the meridional direction, with their eastern extension constrained to the offshore EGC boundary. The outcomes are illustrated in Fig. 4.12. We have six regions denoted as **A1**, **A2**, **A3**, **A4**, **A5**, and **A6**. We observe a significant negative correlation between SSS and SIM across all regions, except **A3**. Trend analysis reveals significant negative trends of SIM for each area, except **A6** (see Tab. 4.1).

The SSS signal is marked by abrupt changes occurring within a maximum period of one year, typically followed by a plateau. Regions **A1**, **A2**, and **A3** exhibit a similar temporal pattern, albeit with varying intensity. Initially, there is a gradual SSS decline, followed by a sudden rise around 2013. This increase is barely marked in **A1**, but reaches 0.3 psu in **A2**, and 0.6 psu in **A3**. Subsequently, a period of relative stability ensues, lasting approximately one year, followed by a steep decrease of 0.3 psu in **A1** and 0.75 psu in **A2** and **A3**. After another brief plateau of low salinity lasting less than a year, salinity undergoes another sharp increase, more pronounced in **A1**, while **A2** reaches a higher level than in the previous high plateau and **A3** a slightly lower level. The subsequent evolution entails a decline, followed by a sudden increase at the end of the study period.

Upon closer examination of **A4** and **A5** further southward, a steep decrease is observed between 2011 and 2012, amounting to approximately 0.5 psu in **A4** and 0.7 psu in **A5**. This is followed by a one-year period of low salinity plateau and a sudden increase of 0.7 psu for both regions. The subsequent evolution of surface salinity in **A4** and **A5** is characterized by greater stability, albeit with more pronounced oscillations in **A4**. Eventually, in the southernmost area **A6**, we notice a similar pattern as in the two above until 2014 followed by a smooth increase, reaching its maximum in 2015, and then a smooth decrease up to a minimum of 33.55 psu in 2018 before increasing again.

Examining the evolution of SIM reveals a notable negative correlation with SSS, particularly accentuated during periods of rapid SIM changes. Specifically, instances of increasing SSS observed prior to 2014 in regions **A1**, **A2**, and **A3** coincide with a reduction in SIM, with the trend being more pronounced in southernmost of these three regions, mirroring the behaviour of surface salinity. During this period, slight decreases of $10 \text{ km}^3 \text{ month}^{-1}$ in **A1** and $25 \text{ km}^3 \text{ month}^{-1}$ in both **A2** and **A3** are noted. Subsequent decreases in surface salinity align with an increase in SIM, amounting to $15 \text{ km}^3 \text{ month}^{-1}$ in **A1**, $20 \text{ km}^3 \text{ month}^{-1}$ in **A2**, and $15 \text{ km}^3 \text{ month}^{-1}$ in **A3**. However, following 2016, the negative correlation becomes less clear in **A2** and **A3**, where steep changes in SSS are no longer observed, though the correlation persists in **A1**. Conversely, in regions **A4**, **A5**, and **A6**, the pronounced decrease in SSS around 2013 aligns with an increase in SIM, albeit less discernible in **A5**. Subsequent changes in SIM exhibit a more gradual decline in **A4** and **A5**, while an upward trend in SIM is observed in **A6**, peaking in 2018 and coinciding with the SSS minimum. Following this peak, a steep decline in SIM is observed, accompanied by an increase in SSS.

In summary, our analysis shows a negative correlation between SIM and SSS, particularly strong during periods of steep variations. Notably, SIM exhibits a negative trend everywhere except in **A6**, where any SIM trend is discernible (see Tab. 4.1).

These findings suggest that the SIM is indeed negatively correlated with salinity, as anticipated. However, our analysis has thus far focused only on the evolution of SSS, which, as demonstrated in Fig. 4.7, exhibits notable disparities from salinity changes at depth. We explore the relationship between SIM and the depth-averaged distribution of salinity in the following section.

4.3.4 SIM and depth integrated salinity changes: quantitative assessment

While the correlation between SIM and SSS changes is obvious, the observed alterations in the EGC velocity structure are associated with changes in the vertical distribution of salinity. As depicted in Fig. 4.7, notable differences exist between the observed trends in surface and deeper salinities. The importance of the depth salinity is highlighted by the weakening of the EGC in its northernmost segment, coinciding with an increase in deep salinity, a phenomenon not observed elsewhere.

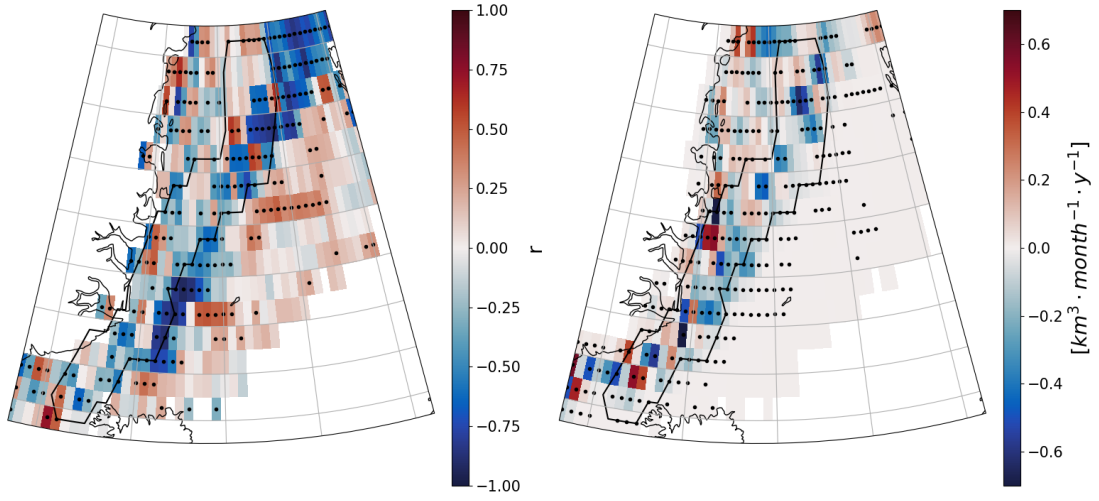


Figure 4.11: **Left:** Correlation between SIM and SSS. **Right:** Trends of SIM between 2011 and the end of 2018. The statistically significant correlations and trends are marked with a black dot. The mean path of the EGC is delineated with a solid black line.

Table 4.1: Linear trend (between 2011 and the end of 2018) slope and p-value for SIM over boxes **A1** to **A6** as defined in Fig. 4.12 and boxes **B1** and **B2** as defined in Fig. 4.13.

Area	Trend [$\text{km}^3 \text{ month}^{-1} \text{ year}^{-1}$]	p-value
A1	-2.17	6.2×10^{-9}
A2	-0.71	9.7×10^{-3}
A3	-0.61	3.4×10^{-2}
A4	-2.1	1.1×10^{-24}
A5	-2.28	6.6×10^{-30}
A6	0.04	0.78
B1	-2.21	5.7×10^{-7}
B2	-1.78	6.2×10^{-14}

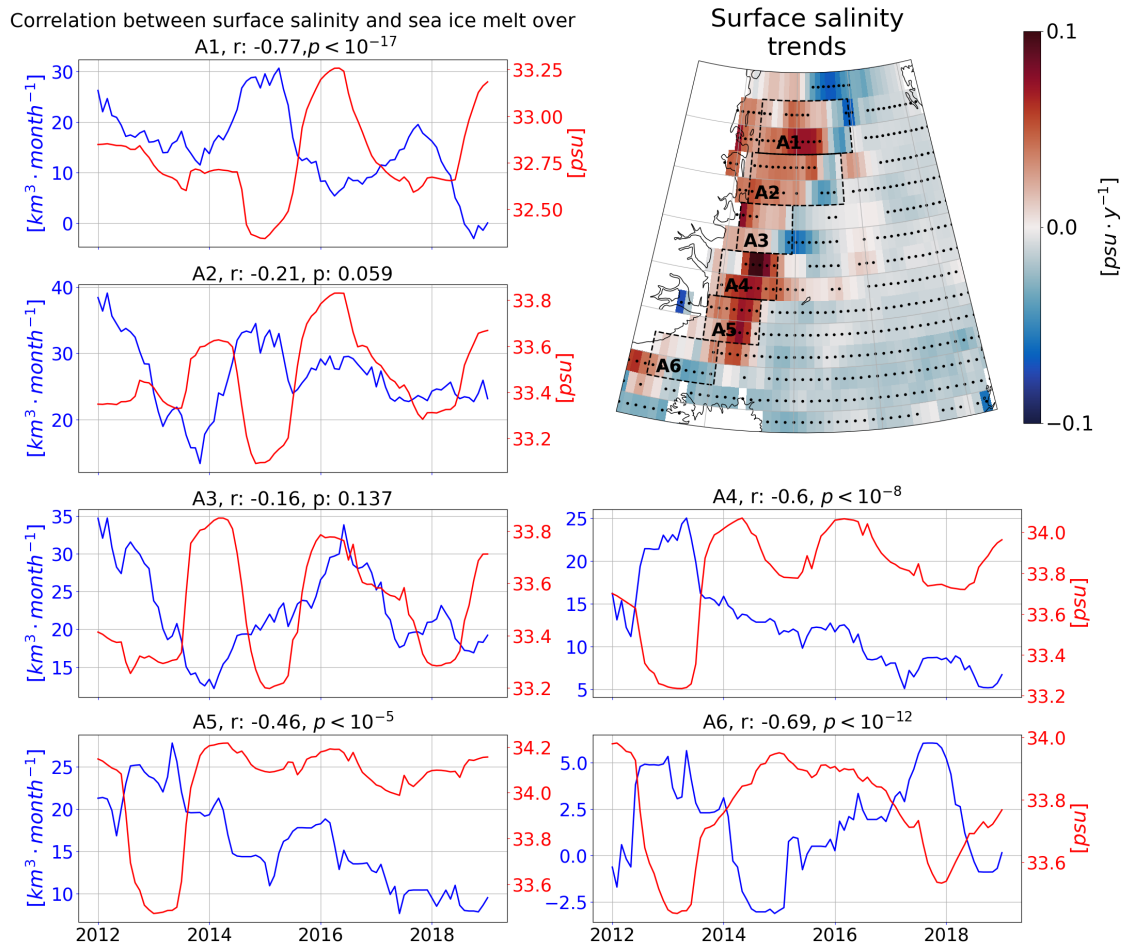


Figure 4.12: **The map:** SSS trends between 2011 and the end of 2018, with the statistically significant trends marked with a black dot and definition of the boxes **A1** to **A6**. **The time series:** SIM (in blue) and mean SSS (in red) for each of the 6 boxes. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.

Based on the results above, the following question arises: while SIM correlates with SSS changes, does this correlation persist when considering depth-averaged salinity? Moreover, can the volume of low salinity water resulting from SIM sufficiently explain the observed depth-averaged salinity change through a dilution mechanism? This inquiry is pivotal for understanding the intricate interplay between sea ice dynamics, ocean salinity variations, and the the EGC.

In ORAS5, the ocean salinity is given in psu for 75 depth levels with non constant spacing, the first ones spanning few meters, while the deeper ones span over hundred of meters. To compute the mean salinity between the surface and a given depth z_1 , we therefore calculate a weighted mean over all cells above the depth z_1 , indexed by i , as

$$\bar{S} = \frac{\sum_i S_i \cdot \text{height}_i}{z_1} \quad (4.11)$$

with height_i the height of the cell. Considering we have two instants, t_1 and t_2 , and a difference of SIM masses and mean salinity between these, $\Delta\text{SIM} = \text{SIM}(t_2) - \text{SIM}(t_1)$ (in kg) and $\Delta S = S(t_2) - S(t_1)$ (in psu). The expected change in salinity due to dilution can be calculated as,

$$\Delta S_{\Delta\text{SIM}} = \frac{S(t_1) \cdot \text{WM} + S_{SI} \cdot \Delta\text{SIM}}{\text{WM} + \Delta\text{SIM}} - S(t_1) \quad (4.12)$$

where WM is the weight of the water column, between surface and the considered depth, and S_{SI} the salinity of the sea ice, fixed at 5 psu. Therefore, when we have a change of depth averaged salinity which is negatively correlated with a change in SIM, we can calculate the ratio

$$R_S = \frac{\Delta S_{\Delta\text{SIM}}}{\Delta S} \quad (4.13)$$

which gives us the SIM contribution to the observed salinity change.

Figure 4.13 shows the trend in averaged salinity above 200 m. It can be seen that these patterns diverge from those observed in the SSS maps. A conspicuous positive trend is visible in the northernmost part of the domain, contributing to the weakening of the zonal ocean density gradient and the slowing down of the EGC. Additionally, a subtle increasing trend is found in the southernmost region, along the eastern flank of the EGC. This slight enhancement contributes to a mild strengthening of the zonal ocean density gradient and, consequently, the EGC. However, this increase is less pronounced at greater depths and therefore lacks significant impact on a 200 m depth-averaged map.

We aim here to focus on these two regions, because they are the ones responsible for the observed EGC changes. We therefore defined two new boxes, **B1** in the north and **B2** in the south, and compared in Fig. 4.13 the SIM and depth averaged salinity using these boxes. The correlation analysis indicates a statistically significant negative correlation between SIM and salinity in the two regions. This correlation is particularly significant during periods of sharp changes in salinity and SIM, but disappears when both variables exhibit gradual changes.

Indeed, the observed negative correlation appears less evident before 2015 in region **B1**. During this period, both the SIM and the depth averaged salinity exhibit gradual declines until 2014 and then an increase up to 2015. However, between 2015 and mid-2016, a notable shift occurs: SIM in region **B1** experiences a sudden decrease of nearly $30 \text{ km}^3 \text{ month}^{-1}$, followed by a subsequent, albeit smaller, increase until 2018, reaching a level lower than that of 2015. This change in SIM coincides with a corresponding shift in salinity, which undergoes a rapid increase of 0.25 psu between 2015 and mid-2016, followed by a subsequent decrease, albeit at a slightly slower rate, of approximately 0.1 psu.

In region **B2**, a robust negative correlation is observed until 2015. Between 2012 and 2013, the SIM undergoes a substantial increase of nearly $20 \text{ km}^3 \text{ month}^{-1}$, followed by a subsequent decrease of $14 \text{ km}^3 \text{ month}^{-1}$ between mid-2013 and 2015. Concurrently, the depth averaged salinity experiences a sharp decline of 0.1 psu, reaching its minimum at the same juncture when SIM peaks. Subsequently, the depth averaged salinity undergoes an increase of 0.05 psu, slowing down after 2014, shortly after the decrease in SIM. However, after 2015, the negative correlation appears to diminish, as both SIM and salinity exhibit a gradual increase until mid-2016, followed by a slow decrease.

To go further than correlation analysis, we aim to quantify the extent to which the observed changes in salinity can be attributed to variations in the release of low salinity water due to sea ice melting. To accomplish this, we delineated three distinct time points within each of the two regions, denoted as t_1^{B1} , t_2^{B1} , and t_3^{B1} for **B1**, and t_1^{B2} , t_2^{B2} , and t_3^{B2} for **B2**. These time points capture the two observed inverse steep fluctuations in SIM and salinity. Utilizing the previously described formula (eq. 4.12), we calculated the anticipated change in salinity, $\Delta S_{\Delta \text{SIM}}$, associated with the observed change in SIM, ΔSIM . By comparing this expected salinity change with the observed change in salinity, ΔS , we also calculate R_S (eq. 4.13) the relative contribution of ΔSIM to the measured ΔS . The closer R_S is to one, the more ΔSIM can explain the depth averaged salinity change. The results are summarized in Tab. 4.2.

As can be seen, ΔSIM explains a non-negligible part of the observed ΔS . In **B1**, the changes in SIM between t_1^{B1} and t_2^{B1} contribute to the changes in depth averaged salinity by 16% and by 24% between t_2^{B1} and t_3^{B1} . In **B2**, R_S amounts to about 28% between t_1^{B2} and t_2^{B2} and 42% between t_2^{B2} and t_3^{B2} .

4.3.5 Confounding variables?

We have observed that changes in salinity were inversely correlated with changes in SIM, and we have demonstrated that the alteration in low salinity water release can explain a portion of the observed salinity change through dilution. However, the observation of these correlations does not guarantee the existence of a causal relationship between SIM and salinity changes. Indeed, there may be another phenomenon, a confounding variable, which could explain the evolution of both salinity and SIM and thus negate the causality between them.

Table 4.2: Change SIM (ΔSIM), depth averaged salinity (ΔS), expected change in depth averaged salinity related to change in SIM ($\Delta S_{\Delta\text{SIM}}$) and relative importance of $\Delta S_{\Delta\text{SIM}}$ in ΔS (R_S). Results are for time periods $t_1^{B1} \rightarrow t_2^{B1}$, $t_2^{B1} \rightarrow t_3^{B1}$, $t_1^{B2} \rightarrow t_2^{B2}$, $t_2^{B2} \rightarrow t_3^{B2}$ and area **B1** and **B2** as defined in Fig. 4.13.

		ΔSIM [GT]	ΔS [psu]	$\Delta S_{\Delta\text{SIM}}$ [psu]	R_S [%]
B1	$t_1^{B1} \rightarrow t_2^{B1}$	-30.835	0.225	0.036	16.19
	$t_2^{B1} \rightarrow t_3^{B1}$	16.716	-0.081	-0.02	24.4
B2	$t_1^{B2} \rightarrow t_2^{B2}$	18.335	-0.089	-0.025	27.95
	$t_2^{B2} \rightarrow t_3^{B2}$	-14.74	0.048	0.02	41.54

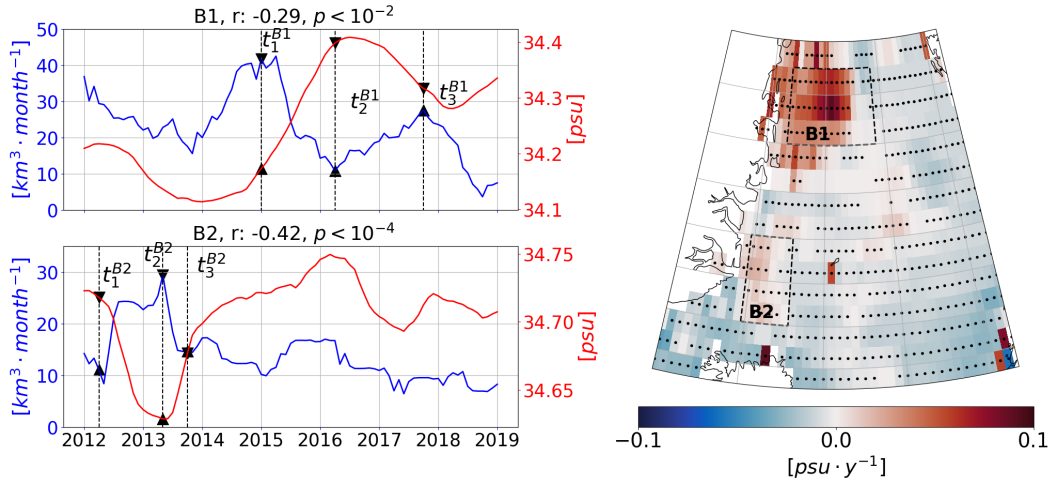


Figure 4.13: Correlation between SIM and depth averaged salinity. **Left:** Time series of SIM (blue) over boxes **B1** and **B2** and depth averaged salinity (red) above 200 m depth in the same boxes. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed. **Right:** Trends of salinity averaged above 200 m depth and boxes definition.

Another plausible explanation for the increasing salinity observed in the northern part of our domain could be an elevated transport of salt from the WSC or from the Arctic Ocean. Examination of the 200 m depth-averaged salinity trends depicted in Fig. 4.13 reveals no discernible positive salinity trend in the region of the WSC. Additionally, analysis of the surface KE trends in Fig. 4.1 indicates a slight weakening of the WSC at the surface.

These observations suggest that the salinity transport through the WSC may have slightly decreased during our study period. However, a more precise analysis can be conducted by calculating the salt transport (ST) using the following formula

$$ST(t) = \int_{L_0}^{L_1} dx \int_{Z_0}^{Z_1} dz V(t, x, z) S(t, x, z) \rho(t, x, z). \quad (4.14)$$

from [Karpouzoglou et al. \(2022\)](#). Here, L_0 and L_1 represent the two horizontal boundaries of a gate, while Z_0 and Z_1 denote the depth boundaries. V is the ocean velocity component perpendicular to the gate, S represents salinity, and ρ denotes water density. Therefore, ST is expressed in grams of salt per second.

We delineated two gates: one meridional at the initial segment of the EGC, where all the flow originates from the Arctic Ocean, and one zonal where the WSC recirculates within the EGC. We calculated the oceanic ST through these two gates from the surface to a depth of 300 m, applied a one-year smoothing average, and presented the results in Fig. 4.14.

We observe that the ST from the WSC is approximately twice as high as that from the Arctic Ocean, as expected. The ST from the Arctic Ocean remains around $30 \text{ kT}_{\text{salt}} \text{ s}^{-1}$ until 2014, after which it experiences a decline until 2016, stabilizing slightly below $20 \text{ kT}_{\text{salt}} \text{ s}^{-1}$. In contrast, the evolution of ST from the WSC is more variable, with a peak level of around $45 \text{ kT}_{\text{salt}} \text{ s}^{-1}$ in 2013, followed by a decrease to less than $20 \text{ kT}_{\text{salt}} \text{ s}^{-1}$ shortly before 2016. Subsequently, the ST rises again between 2016 and 2017 to approximately $40 \text{ kT}_{\text{salt}} \text{ s}^{-1}$, followed by a gradual decrease to approximately the same level as in 2016.

Comparing the ST (Fig. 4.14) and the depth-averaged salinity in **B1** (Fig. 4.13), we note that trends of the two are in opposite directions. While the ST from both the WSC and the Arctic Ocean decreases, an increase in salinity is observed in **B1**. Furthermore, the rise in salinity in **B1** between 2014 and 2016 coincides with a period of declining ST from both the WSC and the Arctic Ocean.

Therefore, the changes in ST cannot account for the observed increase in salinity; moreover, they would counteract the observed salinity trends. Two other potential confounding factors that merit consideration are a shift in the evaporation/precipitation balance and the melting of the Greenland ice sheet, both of which can influence salinity levels. It is noteworthy that any alteration in the evaporation/precipitation balance in the region would likely be linked to changes in the sea ice cover, which serves as a mediator between the ocean and atmosphere. Consequently, an insightful enhancement to our study would involve investigating this

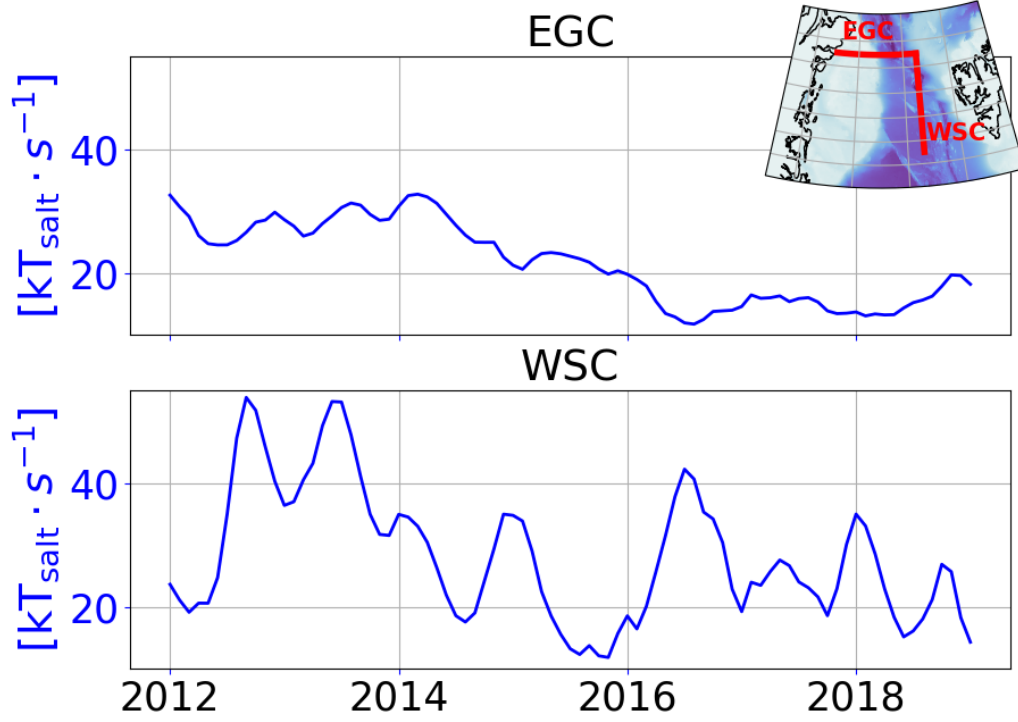


Figure 4.14: ST transport as calculated with eq. 4.14 through two transects displayed in red in the inset map. A one year smoothing average has been applied to the time series.

phenomenon using ERA5, the atmospheric analysis forcing ORAS5, in conjunction with an analysis of changes in sea ice cover.

We have demonstrated that the observed increase in depth averaged salinity was not associated with a change in oceanic ST from either the Arctic Ocean or the WSC. This finding strengthens our hypothesis regarding the influence of SIM changes on the salinity. However, we have shown (Fig. 4.13) that the magnitude of SIM changes was insufficient to explain the observed salinity changes, indicating the presence of additional processes operating in the region.

4.3.6 Summary

In summary, we have shown that between 2011 and the end of 2018 the zonal ocean density gradients along the EGC path were primarily due to zonal ocean salinity gradients (Fig. 4.6). We observed positive salinity trends along the east coast of Greenland (Fig. 4.13). In the northern part of the area, the increasing salinity trends are observed both in surface and depth, whereas in the southern part, they are primarily confined to the surface (Fig. 4.7).

These increasing salinity trends coincide with decreasing SIM (Tab. 4.1), and our calculations indicate a negative correlation between the two variables (Figs. 4.13 and 4.12). However, an examination of these time series reveals that the statistically

significant negative correlation primarily arises during periods of steep variations. When both SIM and salinity exhibit more gradual changes, the negative correlation is less pronounced. Indeed, the positive salinity trends should not obscure the fact that its evolution has not been gradual during our study period. As demonstrated in the temporal series of SSS and depth averaged salinity (Figs. 4.12 and 4.13), the evolution is characterized more by abrupt changes than by gradual trends.

The causal nature of the correlation between increasing salinity along the east coast of Greenland and decreasing SIM is not yet fully established. Confounding variables such as the melting of the Greenland ice sheet, the evaporation/precipitation balance, or oceanic ST could also contribute to the observed correlations. However, our study presents two arguments supporting a causal influence of SIM on the salinity changes.

The first argument pertains to our calculation of the expected salinity changes resulting from four observed sudden variations in SIM. We found that the alteration in low salinity water release from SIM could potentially explain between one quarter and 40% of the observed salinity changes (Fig. 4.13 and Tab. 4.2). However, we have not investigated the depth penetration of low salinity water resulting from SIM, neither the duration of salinity anomalies resulting from changes in SIM. Incorporating this analysis could contribute to determining whether the observed salinity changes are indeed attributable to variations in SIM.

The second argument concerns the oceanic ST towards the region. Our analysis indicates a decrease in ST, from both the Arctic Ocean and the WSC during our study period (Fig. 4.14), which contradicts the observed trend of increasing salinity. Therefore, this highlights the importance of other factors in the observed change such as SIM, Greenland ice sheet melting, or alterations in the evaporation/precipitation balance. It is noteworthy that changes in the evaporation/precipitation balance may also be influenced by the reduction of sea ice cover, which acts as an insulator between the ocean and atmosphere. A further step in elucidating the causes of the salinity increase would involve calculating this balance based on ERA5, the atmospheric analysis forcing ORAS5.

4.4 Future evolution of SIM in the EGC region

In this section, we will examine our final research question, namely, “*What future evolution of the sea ice melting in the EGC region can we expect?*”. To address this inquiry, we will delve into the factors contributing to the observed decline in SIM and demonstrate its strong correlation with ice transport through the Fram Strait. Subsequently, we will discuss the future trajectory of SIM in the EGC region in relation to the projected evolution of this transport.

First, we conducted a comparison between the evolution of the total SIV in the area of interest and the sea ice import from the Fram Strait, as depicted in Fig. 4.15, with a one year smoothing average. As anticipated, we observe a close relationship between the SIV in the EGC regions and the sea ice transport through the Fram Strait, both facing a significant decrease during our study period. The correlation analysis, performed on the time series with trends removed, reveals a notable positive correlation. This correlation is particularly significant during periods of rapid decrease in sea ice transport, immediately followed by steep declines in SIV. However, the SIV does not increase between 2014 and 2015 as the sea ice transport does.

We want to see how the SIM in our boxes **A1** to **A6** are related to the sea ice import from Fram Strait. The mean sea ice velocity in the region is around 0.1 m s^{-1} and a grid cell is about 110 km long in latitude. We can therefore expect that, in one month, sea ice travels around 250 km i.e., more or less two grid cells. Because of that, we calculate the correlation between the SIM and Fram Strait sea ice import by adding one month of lag time for each area when we look southwards. For example, the SIM in **A4** at a given month is compared with the Fram Strait sea ice import three months ago. A one year smoothing average was applied to the time series to hide the seasonal cycle and the correlation analysis was based on trend removed time series. The results are illustrated in Fig. 4.16.

A significant positive correlation is noticed everywhere, except in **A4** and **A5**. The pronounced decrease of SIM in **A2** and **A3** in 2014 mirrors closely the decreasing Fram Strait sea ice transport. The high SIM at the beginning of the time series in **A1**, **A2**, and **A3** coincides with a high sea ice transport. In **A4**, **A5**, and **A6**, the SIM is originally low but increases suddenly in the first months, reaching a maximum coinciding with the high value of sea ice transport once the lag time is taken into account.

Based on these findings, it appears that the SIM over the EGC path will align with the future trajectory of sea ice transport through the Fram Strait. [Zanowski et al. \(2021\)](#) analyzed seven climate models projections from the Coupled Model Intercomparison Project Phase 6 (CMIP6) under two future emissions scenarios, SSP1-2.6 and SSP5-8.5. They observed that the sea ice flux through the Fram Strait was projected to decline in both scenarios, with SSP5-8.5 approaching near-zero sea ice flux by the end of the century. Based on these projections, it is expected that the SIM will continue to decrease in the EGC region. This would lead to a

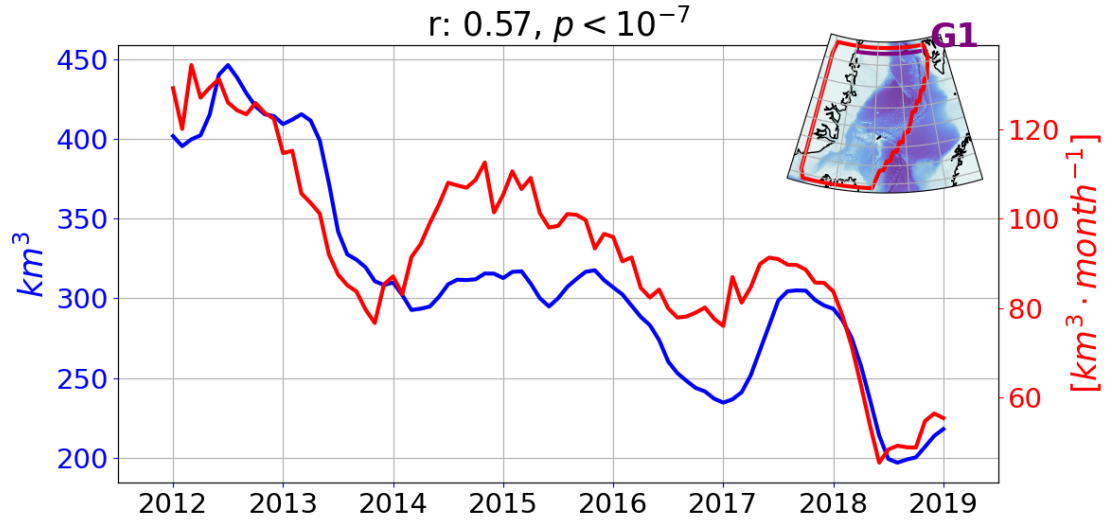


Figure 4.15: SIV (in blue) in the red area in the inset map and sea ice transport (red) through the gate G1 in the inset map. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.

decrease of low salinity water release from SIM and therefore an increase in salinity, which could impact the baroclinic component of the EGC.

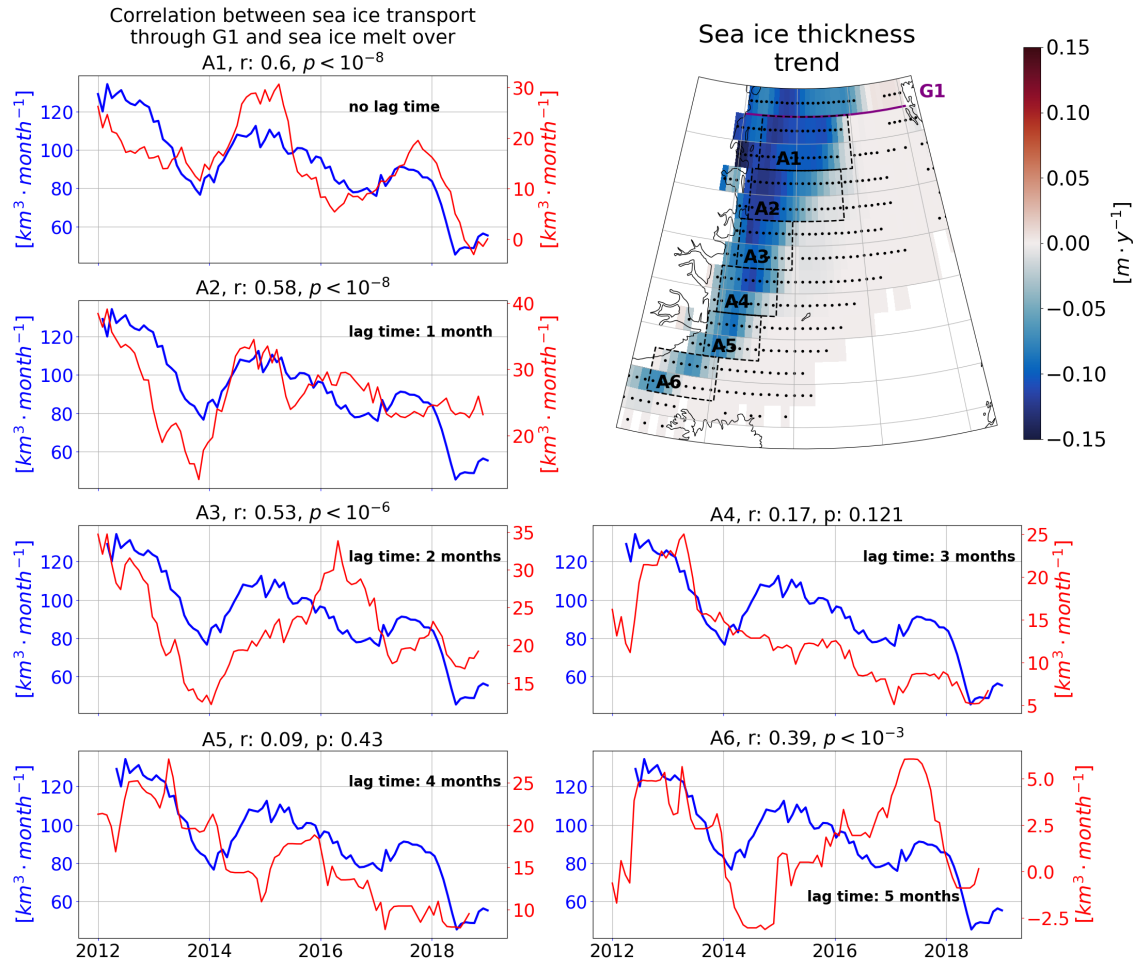


Figure 4.16: **The map:** SIT trends between 2011 and the end of 2018, with the statistically significant trends marked with a black dot, definition of the boxes **A1** to **A6** and the gate **G1**. **The time series:** SIM (in red) for each of the 6 boxes and sea ice transport through the gate **G1**. A one year smoothing average has been applied to all the time series. The correlation coefficient and the p-value of the correlation are displayed at the top of each plot. For the correlation analysis, the trends of the time series were removed.

Conclusion

Global warming has already affected numerous components of the climate system, with enhanced impact in the Arctic region. Consequently, it is crucial to improve our understanding of the climate system and its potential future perturbations. In this context, our study analyzed the link between the SIM and the velocity structure of the EGC, an ocean current connecting the Arctic to the Atlantic. The research question guiding this investigation was: *“What is the impact of the sea ice accelerated melting on the velocity structure of the East Greenland Current?”*

We conducted this research over the period January 2011 - December 2018 using the reanalysis ORAS5 ([Copernicus Climate Change Service, Climate Data Store, 2021](#)) from the ECMWF, which we validated against several observational data in Chapter III. We used the SIT from CS2 ([Landy et al., 2022](#)), the SIC and SID from OSI SAF ([OSI SAF, 2022](#)), the SSS and SSD from Copernicus ([Copernicus Marine Service Information, 2021](#)), the water masses definition from [Rudels \(2002\)](#) and estimates of the Fram Strait sea ice transport from [Ricker et al. \(2018\)](#) and [Min et al. \(2019\)](#). We found that some differences exist between those datasets and the reanalysis but the EGC, its water masses properties and the sea ice in the region are in general correctly represented in ORAS5.

In chapter IV, we answered the research question, which we have decomposed in four parts.

1) *“What are the observed changes in the EGC velocity structure over our study period?”*

Through the analysis of surface KE and EKE, as well as the vertical distribution of southward ocean velocity component, two notable changes in the EGC have been identified. In the northern region, a decrease in surface KE was observed above the continental shelf, without a concurrent change in EKE. This deceleration is confined near the surface and is not evident beyond 200 m depth, while another deceleration is observed at depth over the continental slope. In the southern part of the domain, an increase in both surface KE and EKE was observed. This strengthening of the current is limited to the upper 300 m and coincides with a narrowing of the current core.

2) *“Are the observed changes in the EGC velocity structure related to shifts in its baroclinic component due to perturbations in ocean density profiles?”*

We conducted an analysis to determine the trends of the zonal ocean density gradients and found that the variations in these gradients correspond to the observed changes in EGC velocity. Specifically, we observed a zonal density gradient weakening (strengthening) where the EGC was weakening (strengthening). This led to the conclusion that the observed changes in the EGC velocity were partly attributed to a modification in its baroclinic component.

Using the dynamical method, we quantified the extent to which the baroclinic modification contributed to the observed changes in the EGC and found that it explains a significant portion in the north but only a small fraction in the south. Additionally, we performed a qualitative analysis of the wind stress evolution over the area and determined that it could account for the remaining changes observed in the surface current.

3) “Are these ocean density changes related to the sea ice accelerated melting?”

We have demonstrated that the zonal ocean density gradient, along which the EGC flows, is primarily attributable to a zonal salinity gradient. Our findings indicate a salinity increase along the east coast of Greenland during the study period. This rise in salinity is observed along the western side of the EGC in the northern part of the domain, leading to a weakening of the zonal ocean density gradient. Conversely, in the southern region, the salinity increase is located on the eastern side of the EGC, resulting in a strengthening of the density gradient.

Our results showed a decrease in sea ice presence across the entire EGC region. Additionally, by calculating the SIM in various boxes spanning the path of the current, we observed decreasing trends of this quantity. We carried out comparisons between this SIM data and both surface salinity and depth-averaged salinity, identifying a negative correlation between the two variables. Notably, this negative correlation was most pronounced during periods characterized by steep changes. In examining four such instances of steep changes, we determined that the observed salinity fluctuations were partially attributable to variations in the release of low salinity water resulting from SIM.

These correlations do not constitute conclusive evidence of a causal relationship between SIM and salinity. It is possible that confounding variables, such as changes in the Greenland ice sheet melting, the evaporation/precipitation budget, or oceanic ST, may exist. To address this last influence, we conducted an analysis of ST from the Arctic Ocean and the WSC, and found that both were decreasing during our study period. Therefore, an alternative source of salinity must be identified to explain the observed increase. It is worth noting that, if this source is attributed to enhanced evaporation, it would also be linked to changes in sea ice cover. An analysis of this evaporation in conjunction with the evolution of sea ice cover would significantly enhance our study.

4) “What future evolution of the sea ice melting in the EGC region can we expect?”

We have observed a close correlation between the total SIV in the EGC region and the sea ice transport through the Fram Strait. Furthermore, our analysis has revealed that SIM in various parts of the EGC region correlates with this transport, considering a consistent lag time associated with the sea ice velocity. Given the projected decline in Fram Strait transport (Zanowski et al., 2021), it is reasonable to anticipate a corresponding decrease in SIM.

If we combine these four results, we can give the following answer to our research question.

“What is the impact of the sea ice accelerated melting on the velocity structure of the East Greenland Current?”

Between January 2011 and December 2018, the sea ice decrease within the EGC region could be partially responsible for the observed rise in salinity, attributed to a reduction in low salinity water release from SIM. This effect is notably pronounced during periods of sharp fluctuations in SIM. The increasing salinity has influenced the baroclinic structure of the EGC, leading to modifications in the zonal ocean density gradient. This alteration is particularly prominent in the northern sector of the region, where it has resulted in a noticeable weakening of the current.

If the reduction in sea ice presence in this region persists, as anticipated based on the projected decline in Fram Strait sea ice transport, we can anticipate a further increase in salinity along the east coast of Greenland. This would lead to a weakening of the zonal ocean density gradients and, consequently, a reduction in the baroclinic flow of the EGC. A summary of these observations and relationships is illustrated in Fig. 4.17.

In this study, our analysis primarily focused on assessing the impact of low salinity water release from SIM on salinity, with a brief discussion on oceanic ST. However, it is important to acknowledge that these factors do not represent the sole contributors to salinity variations. To comprehensively address this issue, further investigation should include an examination of the influence of Greenland ice sheet melting and the evaporation/precipitation balance. Additionally, enhancing our study would involve tracking the trajectory of low salinity water. While we conducted correlations and quantitative assessments to determine potential impacts on salinity, an in-depth analysis of factors such as the depth of penetration and residence time of low salinity anomalies remains unexplored.

Our study period of eight years provided a sufficient duration to investigate the physical interactions between changes in SIM, variations in salinity, and current dynamics. However, to examine the long-term evolution of the EGC, employing the ORAS5 dataset over an extended timeframe would be necessary. Nevertheless, such an extension would require additional validation, and measurements of summer SIT before 2011 are unavailable.

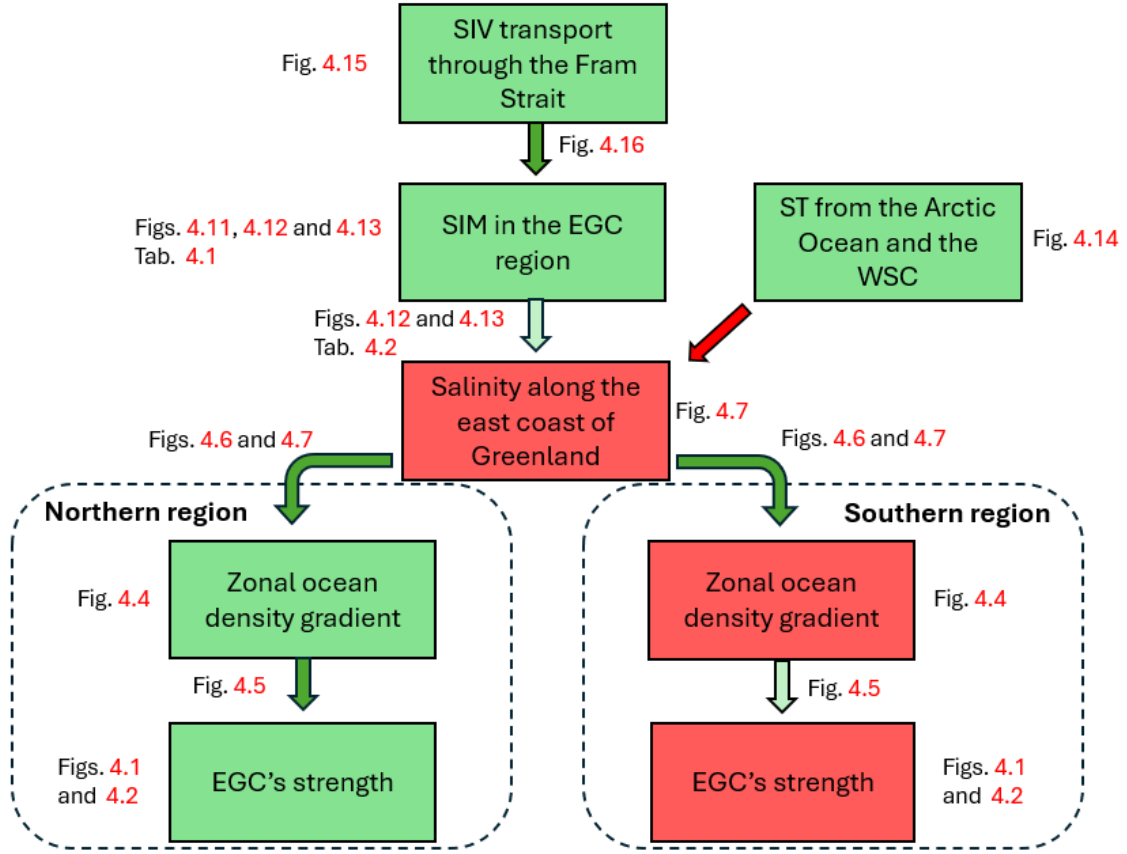


Figure 4.17: Summary of the main observations and relationships highlighted in this study. A green (red) box indicates a decrease (increase) in the given variable. A dark (light) green arrow indicates a strong (weak) relationship. The red arrow indicates a relationship at odds with the observed change. Figures and table references are displayed alongside the corresponding results.

Another improvement could be the use climate models to carry out sensitivity experiments aimed at evaluating the relative impacts of SIM, oceanic ST, Greenland ice sheet melting, and evaporation/precipitation balance on salinity in this region. For instance, we could decrease the SIM in the region while keeping the three other variables constant and observe the resulting salinity variation. We could also perform experiments involving freshwater releases in the area using a climate model to investigate the sensitivity of the EGC's baroclinic component.

Bibliography

- Aagaard, K. and Carmack, E. C. (1989). The role of sea ice and other fresh water in the Arctic circulation. *Journal of Geophysical Research: Oceans*, 94(C10):14485–14498. [_eprint: https://onlinelibrary.wiley.com/doi/pdf/10.1029/JC094iC10p14485](https://onlinelibrary.wiley.com/doi/pdf/10.1029/JC094iC10p14485).
- Aagaard, K. and Greisman, P. (1975). Toward new mass and heat budgets for the Arctic Ocean. *Journal of Geophysical Research*, 80(27):3821–3827.
- Andersen, S., Tonboe, R., Kern, S., and Schyberg, H. (2006). Improved retrieval of sea ice total concentration from spaceborne passive microwave observations using numerical weather prediction model fields: An intercomparison of nine algorithms. *Remote Sensing of Environment*, 104(4):374–392.
- Arrigo, K. R. (2014). Sea Ice Ecosystems. *Annual Review of Marine Science*, 6(1):439–467. [_eprint: https://doi.org/10.1146/annurev-marine-010213-135103](https://doi.org/10.1146/annurev-marine-010213-135103).
- Buongiorno Nardelli, B., Droghei, R., and Santoleri, R. (2016). Multi-dimensional interpolation of SMOS sea surface salinity with surface temperature and in situ salinity data. *Remote Sensing of Environment*, 180:392–402.
- Calvin, K., Dasgupta, D., Krinner, G., Mukherji, A., Thorne, P. W., Trisos, C., Romero, J., Aldunce, P., Barrett, K., Blanco, G., Cheung, W. W., Connors, S., Denton, F., Diongue-Niang, A., Dodman, D., Garschagen, M., Geden, O., Hayward, B., Jones, C., Jotzo, F., Krug, T., Lasco, R., Lee, Y.-Y., Masson-Delmotte, V., Meinshausen, M., Mintenbeck, K., Mokssit, A., Otto, F. E., Pathak, M., Pirani, A., Poloczanska, E., Pörtner, H.-O., Revi, A., Roberts, D. C., Roy, J., Ruane, A. C., Skea, J., Shukla, P. R., Slade, R., Slangen, A., Sokona, Y., Sörensson, A. A., Tignor, M., Van Vuuren, D., Wei, Y.-M., Winkler, H., Zhai, P., Zommers, Z., Hourcade, J.-C., Johnson, F. X., Pachauri, S., Simpson, N. P., Singh, C., Thomas, A., Totin, E., Arias, P., Bustamante, M., Elgizouli, I., Flato, G., Howden, M., Méndez-Vallejo, C., Pereira, J. J., Pichs-Madruga, R., Rose, S. K., Sahab, Y., Sánchez Rodríguez, R., Ürgen Vorsatz, D., Xiao, C., Yassaa, N., Alegría, A., Armour, K., Bednar-Friedl, B., Blok, K., Cissé, G., Dentener, F., Eriksen, S., Fischer, E., Garner, G., Guivarch, C., Haasnoot, M., Hansen, G., Hauser, M., Hawkins, E., Hermans, T., Kopp, R., Leprince-Ringuet, N., Lewis, J., Ley, D., Ludden, C., Niamir, L., Nicholls, Z., Some, S., Szopa, S., Trewin, B., Van Der Wijst, K.-I., Winter, G., Witting, M., Birt, A., Ha, M., Romero, J., Kim, J., Haites, E. F., Jung, Y., Stavins, R., Birt, A., Ha, M., Orendain, D. J. A., Ignon, L., Park, S., Park, Y., Reisinger, A., Cammaramo, D., Fischlin, A., Fuglestad, J. S., Hansen, G., Ludden, C., Masson-Delmotte, V., Matthews, J. R., Mintenbeck, K., Pirani, A., Poloczanska, E., Leprince-Ringuet, N., and Péan, C.

- (2023). IPCC, 2023: Climate Change 2023: Synthesis Report. Contribution of Working Groups I, II and III to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change [Core Writing Team, H. Lee and J. Romero (eds.)]. IPCC, Geneva, Switzerland. Technical report, Intergovernmental Panel on Climate Change (IPCC). Edition: First.
- Chylek, P., Folland, C. K., Lesins, G., Dubey, M. K., and Wang, M. (2009). Arctic air temperature change amplification and the Atlantic Multidecadal Oscillation. *Geophysical Research Letters*, 36(14). eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2009GL038777>.
- Coachman, K. A. a. L. K. (1968a). View of The East Greenland Current North of Denmark Strait: Part I.
- Coachman, K. A. a. L. K. (1968b). View of The East Greenland Current North of Denmark Strait: Part II.
- Comiso, J. C. (1986). Characteristics of Arctic winter sea ice from satellite multispectral microwave observations. *Journal of Geophysical Research: Oceans*, 91(C1):975–994. eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/JC091iC01p00975>.
- Comiso, J. C., Cavalieri, D. J., Parkinson, C. L., and Gloersen, P. (1997). Passive microwave algorithms for sea ice concentration: A comparison of two techniques. *Remote Sensing of Environment*, 60(3):357–384.
- Copernicus Climate Change Service, Climate Data Store (2021). ORAS5 global ocean reanalysis monthly data from 1958 to present. Copernicus Climate Change Service (C3S) Climate Data Store (CDS). DOI: 10.24381/cds.67e8eeb7 (Accessed on 20-Oct-2023).
- Copernicus Marine Service Information (2021). Global Ocean Gridded L 4 Sea Surface Heights And Derived Variables Reprocessed Copernicus Climate Service. E.U. Marine Data Store (MDS). DOI: 10.48670/moi-00145 (Accessed on 15-Feb-2024).
- Curry, J. A., Schramm, J. L., and Ebert, E. E. (1995). Sea Ice-Albedo Climate Feedback Mechanism. *Journal of Climate*, 8(2):240–247. Publisher: American Meteorological Society Section: Journal of Climate.
- Dawson, G., Landy, J., Tsamados, M., Komarov, A. S., Howell, S., Heorton, H., and Krumpen, T. (2022). A 10-year record of Arctic summer sea ice freeboard from CryoSat-2. *Remote Sensing of Environment*, 268:112744.
- De Steur, L., Hansen, E., Mauritzen, C., Beszczynska-Möller, A., and Fahrbach, E. (2014). Impact of recirculation on the East Greenland Current in Fram Strait: Results from moored current meter measurements between 1997 and 2009. *Deep Sea Research Part I: Oceanographic Research Papers*, 92:26–40.
- De Steur, L., Sumata, H., Divine, D. V., Granskog, M. A., and Pavlova, O. (2023). Upper ocean warming and sea ice reduction in the East Greenland Current from 2003 to 2019. *Communications Earth & Environment*, 4(1):261.

- Dee, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M. A., Balsamo, G., Bauer, P., Bechtold, P., Beljaars, A. C. M., van de Berg, L., Bidlot, J., Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A. J., Haimberger, L., Healy, S. B., Hersbach, H., Hólm, E. V., Isaksen, I., Kållberg, P., Köhler, M., Matricardi, M., McNally, A. P., Monge-Sanz, B. M., Morcrette, J.-J., Park, B.-K., Peubey, C., de Rosnay, P., Tavolato, C., Thépaut, J.-N., and Vitart, F. (2011). The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Quarterly Journal of the Royal Meteorological Society*, 137(656):553–597. [_eprint: https://onlinelibrary.wiley.com/doi/pdf/10.1002/qj.828](https://onlinelibrary.wiley.com/doi/pdf/10.1002/qj.828).
- Deutsche Nordpolar-Expedition (2nd : 1869-1870) (1874). *The German Arctic expedition of 1869-70, and narrative of the wreck of the "Hansa" in the ice*. London : Sampson Low, Marston, Low & Searle, Crown Buildings, 188, Fleet Street.
- Dickson, B., Dye, S., Jónsson, S., Köhl, A., Macrander, A., Marnela, M., Meincke, J., Olsen, S., Rudels, B., Valdimarsson, H., and Voet, G. (2008). The Overflow Flux West of Iceland: Variability, Origins and Forcing. In Dickson, R. R., Meincke, J., and Rhines, P., editors, *Arctic-Subarctic Ocean Fluxes*, pages 443–474. Springer Netherlands, Dordrecht.
- Donlon, C. J., Martin, M., Stark, J., Roberts-Jones, J., Fiedler, E., and Wimmer, W. (2012). The Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA) system. *Remote Sensing of Environment*, 116:140–158.
- Droghei, R., Buongiorno Nardelli, B., and Santoleri, R. (2016). Combining In Situ and Satellite Observations to Retrieve Salinity and Density at the Ocean Surface. *Journal of Atmospheric and Oceanic Technology*, 33.
- Eicken, H., Grenfell, T. C., Perovich, D. K., Richter-Menge, J. A., and Frey, K. (2004). Hydraulic controls of summer Arctic pack ice albedo. *Journal of Geophysical Research: Oceans*, 109(C8). [_eprint: https://onlinelibrary.wiley.com/doi/pdf/10.1029/2003JC001989](https://onlinelibrary.wiley.com/doi/pdf/10.1029/2003JC001989).
- Eldevik, T., Nilsen, J. E., Iovino, D., Anders Olsson, K., Sandø, A. B., and Drange, H. (2009). Observed sources and variability of Nordic seas overflow. *Nature Geoscience*, 2(6):406–410.
- Foldvik, A., Aagaard, K., and Tørresen, T. (1988). On the velocity field of the East Greenland Current.
- Fomin, L. M. (1964). *The dynamic method in oceanography*. Number 2 in Elsevier oceanography series. Elsevier Pub. Co, Amsterdam New York.
- Good, S. A., Martin, M. J., and Rayner, N. A. (2013). EN4: Quality controlled ocean temperature and salinity profiles and monthly objective analyses with uncertainty estimates. *Journal of Geophysical Research: Oceans*, 118(12):6704–6716.
- Hansen, B. and Østerhus, S. (2000). North Atlantic–Nordic Seas exchanges. *Progress in Oceanography*, 45(2):109–208.

- Hansen, B., Østerhus, S., Turrell, W. R., Jónsson, S., Valdimarsson, H., Hátún, H., and Olsen, S. M. (2008). The Inflow of Atlantic Water, Heat, and Salt to the Nordic Seas Across the Greenland–Scotland Ridge. In Dickson, R. R., Meincke, J., and Rhines, P., editors, *Arctic–Subarctic Ocean Fluxes: Defining the Role of the Northern Seas in Climate*, pages 15–43. Springer Netherlands, Dordrecht.
- Helland-Hansen, Koefoed, E. B. (1907). Hydrographie, Croisière océanographique accomplie à bord de la Belgica dans la Mer du Grönland 1905. in: Duc d’Orléan. pages pp. 275–343.
- Holfort, J. and Hansen, E. (2005). Timeseries of Polar Water properties in Fram Strait. *Geophysical Research Letters*, 32(19). _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2005GL022957>.
- Håvik, L., Våge, K., Pickart, R. S., Harden, B., Appen, W.-J. V., Jónsson, S., and Østerhus, S. (2017). Structure and Variability of the Shelfbreak East Greenland Current North of Denmark Strait. *Journal of Physical Oceanography*, 47(10):2631–2646.
- Intergovernmental Panel On Climate Change (2022). *The Ocean and Cryosphere in a Changing Climate: Special Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, 1 edition.
- Isachsen, P. E., Sørli, S. R., Mauritzen, C., Lydersen, C., Dodd, P., and Kovacs, K. M. (2014). Upper-ocean hydrography of the Nordic Seas during the International Polar Year (2007–2008) as observed by instrumented seals and Argo floats. *Deep Sea Research Part I: Oceanographic Research Papers*, 93:41–59.
- Jahn, A., Holland, M. M., and Kay, J. E. (2024). Projections of an ice-free Arctic Ocean. *Nature Reviews Earth & Environment*, pages 1–13. Publisher: Nature Publishing Group.
- Karpouzoglou, T., de Steur, L., Smedsrud, L. H., and Sumata, H. (2022). Observed Changes in the Arctic Freshwater Outflow in Fram Strait. *Journal of Geophysical Research: Oceans*, 127(3):e2021JC018122. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021JC018122>.
- Killworth, P. D. (1983). Deep convection in the World Ocean. *Reviews of Geophysics*, 21(1):1–26. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/RG021i001p00001>.
- Kwok, R. and Cunningham, G. F. (2008). ICESat over Arctic sea ice: Estimation of snow depth and ice thickness. *Journal of Geophysical Research: Oceans*, 113(C8). _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2008JC004753>.
- Kwok, R., Cunningham, G. F., and Pang, S. S. (2004). Fram Strait sea ice outflow. *Journal of Geophysical Research: Oceans*, 109(C1). _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2003JC001785>.
- Landy, J. C., Dawson, G. J., Tsamados, M., Bushuk, M., Stroeve, J. C., Howell, S. E. L., Krumpen, T., Babb, D. G., Komarov, A. S., Heorton, H. D. B. S., Belter, H. J., and Aksenov, Y. (2022). A year-round satellite sea-ice thickness record from CryoSat-2. *Nature*, 609(7927):517–522.

- Larson, L. M. (1917). *The king's mirror (Speculum regale-Konungs skuggsjá) translated from the old Norwegian by Laurence Marcellus Larson*. New York American-Scandinavian Foundation.
- Laxon, S., Peacock, N., and Smith, D. (2003). High interannual variability of sea ice thickness in the Arctic region. *Nature*, 425(6961):947–950. Number: 6961 Publisher: Nature Publishing Group.
- Laxon, S. W., Giles, K. A., Ridout, A. L., Wingham, D. J., Willatt, R., Cullen, R., Kwok, R., Schweiger, A., Zhang, J., Haas, C., Hendricks, S., Krishfield, R., Kurtz, N., Farrell, S., and Davidson, M. (2013). CryoSat-2 estimates of Arctic sea ice thickness and volume. *Geophysical Research Letters*, 40(4):732–737. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/grl.50193>.
- Li, H., Fedorov, A., and Liu, W. (2021). AMOC stability and diverging response to Arctic sea ice decline in two climate models. *Journal of Climate*, pages 1–47.
- Lind, S., Ingvaldsen, R. B., and Furevik, T. (2018). Arctic warming hotspot in the northern Barents Sea linked to declining sea-ice import. *Nature Climate Change*, 8(7):634–639. Publisher: Nature Publishing Group.
- Lindsay, R. and Schweiger, A. (2015). Arctic sea ice thickness loss determined using subsurface, aircraft, and satellite observations. *The Cryosphere*, 9(1):269–283. Publisher: Copernicus GmbH.
- Liu, W. and Fedorov, A. V. (2019). Global Impacts of Arctic Sea Ice Loss Mediated by the Atlantic Meridional Overturning Circulation. *Geophysical Research Letters*, 46(2):944–952.
- Manabe, S. and Stouffer, R. J. (1994). Multiple-Century Response of a Coupled Ocean-Atmosphere Model to an Increase of Atmospheric Carbon Dioxide. *Journal of Climate*, 7(1):5–23.
- Mann, M. E., Bradley, R. S., and Hughes, M. K. (1999). Northern hemisphere temperatures during the past millennium: Inferences, uncertainties, and limitations. *Geophysical Research Letters*, 26(6):759–762. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/1999GL900070>.
- Marcott, S. A., Shakun, J. D., Clark, P. U., and Mix, A. C. (2013). A Reconstruction of Regional and Global Temperature for the Past 11,300 Years. *Science*, 339(6124):1198–1201. Publisher: American Association for the Advancement of Science.
- Miller, G. H., Alley, R. B., Brigham-Grette, J., Fitzpatrick, J. J., Polyak, L., Serreze, M. C., and White, J. W. C. (2010). Arctic amplification: can the past constrain the future? *Quaternary Science Reviews*, 29(15):1779–1790.
- Min, C., Mu, L., Yang, Q., Ricker, R., Shi, Q., Han, B., Wu, R., and Liu, J. (2019). Sea ice export through the Fram Strait derived from a combined model and satellite data set. preprint, Sea ice/Sea Ice.
- Mohn (1887). *Den norskenordhavs-expedition, 1876-1878. Nordhavets dybder, temperatur og strømninger*.

- Mork, K. A., Skagseth, , and S iland, H. (2019). Recent Warming and Freshening of the Norwegian Sea Observed by Argo Data. *Journal of Climate*, 32(12):3695–3705. Publisher: American Meteorological Society Section: Journal of Climate.
- Nardelli, B. B. (2012). A Novel Approach for the High-Resolution Interpolation of In Situ Sea Surface Salinity. *Journal of Atmospheric and Oceanic Technology*, 29(6):867–879. Publisher: American Meteorological Society Section: Journal of Atmospheric and Oceanic Technology.
- Notz, D. and Community, S. (2020). Arctic Sea Ice in CMIP6. *Geophysical Research Letters*, 47(10):e2019GL086749. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2019GL086749>.
- OSI SAF (2022). Global Low Resolution Sea Ice Drift data record 1991-2020, OSI-455, doi: 10.15770/EUM_SAF_OSI-0012. EUMETSAT Ocean and Sea Ice Satellite Application Facility. Data extracted from OSI SAF FTP server: (2011-2019) accessed 18-08-2023.
- Paquette, R. G., Bourke, R. H., Newton, J. F., and Perdue, W. F. (1985). The East Greenland Polar Front in autumn. *Journal of Geophysical Research: Oceans*, 90(C3):4866–4882.
- Perovich, D. K., Light, B., Eicken, H., Jones, K. F., Runciman, K., and Nghiem, S. V. (2007). Increasing solar heating of the Arctic Ocean and adjacent seas, 1979–2005: Attribution and role in the ice-albedo feedback. *Geophysical Research Letters*, 34(19):2007GL031480.
- Petty, A. A., Kurtz, N. T., Kwok, R., Markus, T., and Neumann, T. A. (2020). Winter Arctic Sea Ice Thickness From ICESat-2 Freeboards. *Journal of Geophysical Research: Oceans*, 125(5):e2019JC015764. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2019JC015764>.
- Polyakov, I. V., Pnyushkov, A. V., Alkire, M. B., Ashik, I. M., Baumann, T. M., Carmack, E. C., Goszczko, I., Guthrie, J., Ivanov, V. V., Kanzow, T., Krishfield, R., Kwok, R., Sundfjord, A., Morison, J., Rember, R., and Yulin, A. (2017). Greater role for Atlantic inflows on sea-ice loss in the Eurasian Basin of the Arctic Ocean. *Science*, 356(6335):285–291. Publisher: American Association for the Advancement of Science.
- Polyakov, I. V., Rippeth, T. P., Fer, I., Alkire, M. B., Baumann, T. M., Carmack, E. C., Ingvaldsen, R., Ivanov, V. V., Janout, M., Lind, S., Padman, L., Pnyushkov, A. V., and Rember, R. (2020). Weakening of Cold Halocline Layer Exposes Sea Ice to Oceanic Heat in the Eastern Arctic Ocean. *Journal of Climate*, 33(18):8107–8123. Publisher: American Meteorological Society Section: Journal of Climate.
- Post, E., Bhatt, U. S., Bitz, C. M., Brodie, J. F., Fulton, T. L., Hebblewhite, M., Kerby, J., Kutz, S. J., Stirling, I., and Walker, D. A. (2013). Ecological Consequences of Sea-Ice Decline. *Science*, 341(6145):519–524. Publisher: American Association for the Advancement of Science.

- Pujol, M.-I., Faugère, Y., Taburet, G., Dupuy, S., Pelloquin, C., Ablain, M., and Picot, N. (2016). DUACS DT2014: the new multi-mission altimeter data set reprocessed over 20 years. *Ocean Science*, 12(5):1067–1090. Publisher: Copernicus GmbH.
- Ricker, R., Girard-Ardhuin, F., Krumpen, T., and Lique, C. (2018). Satellite-derived sea ice export and its impact on Arctic ice mass balance. *The Cryosphere*, 12(9):3017–3032.
- Rudels, B. (2002). The East Greenland Current and its contribution to the Denmark Strait overflow. *ICES Journal of Marine Science*, 59(6):1133–1154.
- Rudels, B., Korhonen, M., Budéus, G., Beszczynska-Möller, A., Schauer, U., Nummelin, A., Quadfasel, D., and Valdimarsson, H. (2012). The East Greenland Current and its impacts on the Nordic Seas: observed trends in the past decade. *ICES Journal of Marine Science*, 69(5):841–851.
- Rudnick, D. L. and Ferrari, R. (1999). Compensation of Horizontal Temperature and Salinity Gradients in the Ocean Mixed Layer. *Science*, 283(5401):526–529. Publisher: American Association for the Advancement of Science.
- Ryder, C. (1895). *Den østgrønlandske Expedition, udført i Aarene 1891-92 under Ledelse af C. Ryder*. Comm. Reitzel. Google-Books-ID: 7CGnzwEACAAJ.
- Sammartino, M., Aronica, S., Santoleri, R., and Buongiorno Nardelli, B. (2022). Retrieving Mediterranean Sea Surface Salinity Distribution and Interannual Trends from Multi-Sensor Satellite and In Situ Data. *Remote Sensing*, 14(10):2502. Number: 10 Publisher: Multidisciplinary Digital Publishing Institute.
- Schauer, U., Beszczynska-Möller, A., Walczowski, W., Fahrbach, E., Piechura, J., and Hansen, E. (2008). Variation of Measured Heat Flow Through the Fram Strait Between 1997 and 2006. In Dickson, R. R., Meincke, J., and Rhines, P., editors, *Arctic-Subarctic Ocean Fluxes*, pages 65–85. Springer Netherlands, Dordrecht.
- Schweiger, A., Lindsay, R., Zhang, J., Steele, M., Stern, H., and Kwok, R. (2011). Uncertainty in modeled Arctic sea ice volume. *Journal of Geophysical Research*, 116:C00D06.
- Scoresby (1820). *An account of the Arctic regions, with a history and description of the northern whale-fisher*.
- Selyuzhenok, V., Bashmachnikov, I., Ricker, R., Vesman, A., and Bobylev, L. (2020). Sea ice volume variability and water temperature in the Greenland Sea. *The Cryosphere*, 14(2):477–495. Publisher: Copernicus GmbH.
- Serreze, M. C., Barrett, A. P., Slater, A. G., Steele, M., Zhang, J., and Trenberth, K. E. (2007). The large-scale energy budget of the Arctic. *Journal of Geophysical Research: Atmospheres*, 112(D11). eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2006JD008230>.
- Serreze, M. C., Barrett, A. P., Stroeve, J. C., Kindig, D. N., and Holland, M. M. (2009). The emergence of surface-based Arctic amplification. *The Cryosphere*, 3(1):11–19. Publisher: Copernicus GmbH.

- Smith, D. M. (1996). Extraction of winter total sea-ice concentration in the Greenland and Barents Seas from SSM/I data. *International Journal of Remote Sensing*, 17(13):2625–2646.
- Smith, D. M., Screen, J. A., Deser, C., Cohen, J., Fyfe, J. C., García-Serrano, J., Jung, T., Kattsov, V., Matei, D., Msadek, R., Peings, Y., Sigmond, M., Ukita, J., Yoon, J.-H., and Zhang, X. (2019). The Polar Amplification Model Intercomparison Project (PAMIP) contribution to CMIP6: investigating the causes and consequences of polar amplification. *Geoscientific Model Development*, 12(3):1139–1164. Publisher: Copernicus GmbH.
- Smith, R. S. and Gregory, J. M. (2009). A study of the sensitivity of ocean overturning circulation and climate to freshwater input in different regions of the North Atlantic. *Geophysical Research Letters*, 36(15). eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2009GL038607>.
- Spreen, G., de Steur, L., Divine, D., Gerland, S., Hansen, E., and Kwok, R. (2020). Arctic Sea Ice Volume Export Through Fram Strait From 1992 to 2014. *Journal of Geophysical Research: Oceans*, 125.
- Sumata, H., de Steur, L., Gerland, S., Divine, D. V., and Pavlova, O. (2022). Unprecedented decline of Arctic sea ice outflow in 2018. *Nature Communications*, 13(1):1747.
- Tammiksaar, E., Sukhova, N., and Stone, I. (1999). Hypothesis Versus Fact: August Petermann and Polar Research. *Arctic -Montreal-*, 52:237–244.
- Taylor, P. C., Cai, M., Hu, A., Meehl, J., Washington, W., and Zhang, G. J. (2013). A Decomposition of Feedback Contributions to Polar Warming Amplification. *Journal of Climate*, 26(18):7023–7043. Publisher: American Meteorological Society Section: Journal of Climate.
- Tschudi, M. A., Maslanik, J. A., and Perovich, D. K. (2008). Derivation of melt pond coverage on Arctic sea ice using MODIS observations. *Remote Sensing of Environment*, 112(5):2605–2614.
- Tsubouchi, T., Våge, K., Hansen, B., Larsen, K., Østerhus, S., Johnson, C., Jonsson, S., and Valdimarsson, H. (2021). Increased ocean heat transport into the Nordic Seas and Arctic Ocean over the period 1993–2016. *Nature Climate Change*, 11:1–6.
- Untersteiner, N. (1988). On the ice and heat balance in Fram Strait. *Journal of Geophysical Research: Oceans*, 93(C1):527–531.
- Vermassen, F., O’Regan, M., De Boer, A., Schenk, F., Razmjooei, M., West, G., Cronin, T. M., Jakobsson, M., and Coxall, H. K. (2023). A seasonally ice-free Arctic Ocean during the Last Interglacial. *Nature Geoscience*, 16(8):723–729.
- Vihma, T. (2014). Effects of Arctic Sea Ice Decline on Weather and Climate: A Review. *Surveys in Geophysics*, 35(5):1175–1214.
- Våge, K., Pickart, R. S., Spall, M. A., Moore, G. W. K., Valdimarsson, H., Torres, D. J., Erofeeva, S. Y., and Nilsen, J. E. (2013). Revised circulation scheme north

- of the Denmark Strait. *Deep Sea Research Part I: Oceanographic Research Papers*, 79:20–39.
- Wadhams, P. (1983). Sea ice thickness distribution in Fram Strait. *Nature*, 305(5930):108–111.
- Walsh, J. E. and Chapman, W. L. (1990). Arctic Contribution to Upper-Ocean Variability in the North Atlantic. *Journal of Climate*, 3(12):1462–1473. Publisher: American Meteorological Society Section: Journal of Climate.
- Woodgate, R. A., Fahrbach, E., and Rohardt, G. (1999). Structure and transports of the East Greenland Current at 75°N from moored current meters. *Journal of Geophysical Research: Oceans*, 104(C8):18059–18072.
- Worthington, L. (1970). The Norwegian Sea as a mediterranean basin. *Deep Sea Research and Oceanographic Abstracts*, 17(1):77–84.
- Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K. (2001). Trends, Rhythms, and Aberrations in Global Climate 65 Ma to Present. *Science*, 292(5517):686–693.
- Zanowski, H., Jahn, A., and Holland, M. M. (2021). Arctic Ocean Freshwater in CMIP6 Ensembles: Declining Sea Ice, Increasing Ocean Storage and Export. *Journal of Geophysical Research: Oceans*, 126(4):e2020JC016930. _eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JC016930>.
- Zege, E., Malinka, A., Katsev, I., Prikhach, A., Heygster, G., Istomina, L., Birnbaum, G., and Schwarz, P. (2015). Algorithm to retrieve the melt pond fraction and the spectral albedo of Arctic summer ice from satellite optical data. *Remote Sensing of Environment*, 163:153–164.
- Zhang, Z., Bi, H., Sun, K., Huang, H., Liu, Y., and Yan, L. (2017). Arctic sea ice volume export through the Fram Strait from combined satellite and model data: 1979–2012. *Acta Oceanologica Sinica*, 36(1):44–55.
- Zuo, H., Balmaseda, M. A., Tietsche, S., Mogensen, K., and Mayer, M. (2019). The ECMWF operational ensemble reanalysis–analysis system for ocean and sea ice: a description of the system and assessment. *Ocean Science*, 15(3):779–808.
- Årthun, M., Eldevik, T., Smedsrud, L. H., Skagseth, , and Ingvaldsen, R. B. (2012). Quantifying the Influence of Atlantic Heat on Barents Sea Ice Variability and Retreat. *Journal of Climate*, 25(13):4736–4743. Publisher: American Meteorological Society Section: Journal of Climate.

UNIVERSITE CATHOLIQUE DE LOUVAIN

Faculté des sciences

Place des sciences, 2 bte L6.06.01, 1348 Louvain-la-Neuve, Belgique — www.uclouvain.be/sc

