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THE EFFECT OF A WEAKENING ATLANTIC MERIDIONAL OVERTURINING CIRCULATION ON OCEAN TEMPERATURES IN THE GULF OF ST. LAWRENCE OVER THE PAST 150 YEARS

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

Abstract

Ocean circulation is essential for transporting heat, dissolved CO_2, and nutrients all around the globe. The strength of the ocean circulation, therefore, has a major influence on climate and vice versa. Over time, ocean currents can change causing climate to vary on local, regional, and indeed global scales; in turn, changing climate can affect the strength and location of ocean currents. Recent studies suggest that freshwater addition to the western North Atlantic from melting Arctic and Greenland ice is causing the Atlantic Meridional Overturning Circulation (AMOC), which encompasses the northward flow of warm saline water of the Gulf Stream and the southward flow of colder deep water of the Labrador Current, to weaken. This results in surface ocean temperatures offshore of Atlantic Canada to rise faster than in the rest of the global ocean. Ocean warming, in turn, is detrimental to a wide variety of marine species, thus impacting both the environment and the local economy.

The objective of this study is to assess past changes in ocean currents in the western North Atlantic, based on sea surface temperature (SST) reconstructions from alkenones preserved in sediments on the Scotian Shelf. Alkenones are long-chain carbon molecules uniquely synthesized by a certain group of phytoplankton that change in composition depending on surface water temperatures. To improve the spatial and temporal resolution of previous studies investigating past SST variability in the open NW Atlantic, I analyzed a high-resolution sediment core from the Cabot Strait in the Gulf of the St. Lawrence that contains a detailed record of the last 150 years. From 1850 to present, the alkenone-derived SST estimates fluctuate between 6.5 – 7 °C with a local error of estimation of 1.4 °C. This absence of any significant trend in SST in the central Gulf of the St. Lawrence over the last 150 years implies that factors other than open ocean warming affect our study site. Three possible mechanisms are presented that may mask or offset any SST warming expected from the weakening AMOC: the hydrography of the Gulf of St. Lawrence, the distribution of water masses in the Cabot Strait, and the timing of the spring phytoplankton bloom. From such analyses, we can gain a better understanding of the changes we are seeing now and more reliably predict what is to come.

Keywords: alkenones, sea surface temperature, Scotian Shelf, AMOC, currents

LIST OF ABBREVIATIONS

AMOC	Atlantic Meridional Overturning Circulation
AZMP	Atlantic Zone Monitoring Program
CBSS	Cabot Strait Subsurface water
CIL	Cold Intermediate water
DWBC	Deep Water Boundary Current
GS	Gulf Stream
GSL	Gulf of St. Lawrence
LC	Labrador Current
LGM	Last Glacial Maximum
LIA	Little Ice Age
NADW	North Atlantic Deep Water
SST	Sea Surface Temperature
WOA	World Ocean Atlas
WSW	Warm Slope Water

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CHAPTER I: INTRODUCTION AND MOTIVATION

1.0 Introduction and Motivation

The ocean plays a large role in the lives of people living in coastal regions all over the globe and can also impact climate of further inland regions. In the Maritimes, the Atlantic Ocean provides work and pleasure while keeping the climate fairly temperate.

The global average ocean temperature has increased ~0.4 ºC between 1971 – 2010 (IPCC, 2013) as a consequence of global warming. The northwest Atlantic is warming at a faster rate, with an increase of 1.5 – 2 ºC over the same time period (Thibodeau *et al.*, 2018). Models have predicted that the rate of warming will continue to increase over the coming decades (Figure 1). Ocean warming is detrimental to the marine ecosystem, industries that rely on the ocean, and the global climate. Rising ocean temperatures have caused a decline in populations of fish and other marine species, especially in the northwest Atlantic (Wilson *et al.*, 2020). This in turn affects the fishing industry, a major part of the Nova Scotian economy (CBC News, 2019).

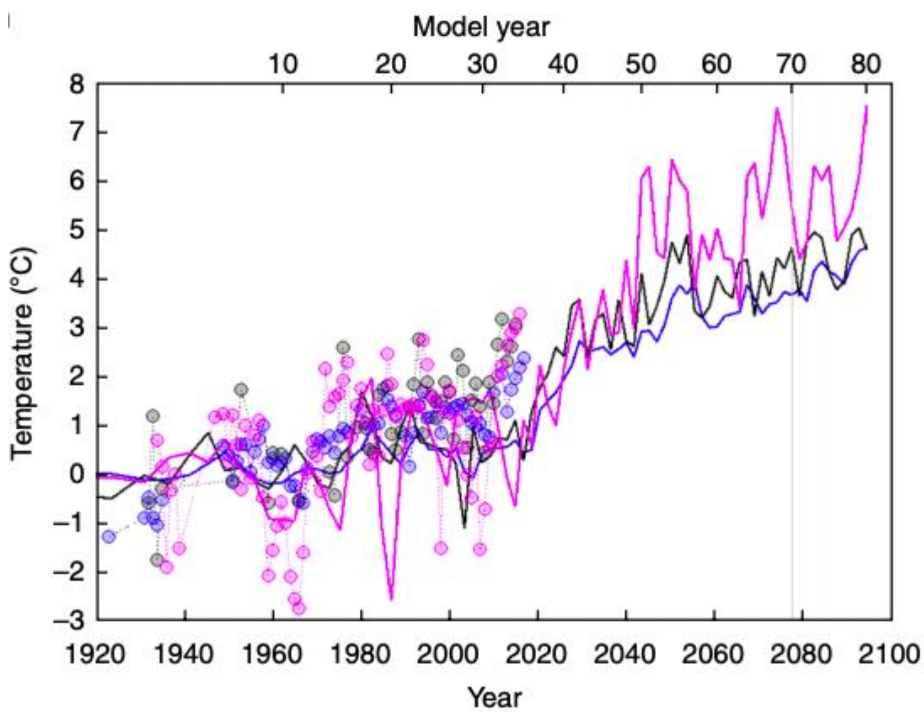

Figure 1: Model of change in potential temperature referenced to the surface in the coastal northwest Atlantic, circles are observations and solid lines are from models (from Claret *et al*., 2018)

The Atlantic Meridional Overturning Circulation (AMOC) encompasses the vertical and horizontal transport and admixture of water masses in the North Atlantic. It is responsible for the current heat distribution in the region, both air temperatures and sea surface. Warm saline shallow waters of the Gulf Stream (GS) are advected northwards where the waters cool, densify, and sink becoming North Atlantic Deep Water (NADW). The transport of surface waters north is balanced by the Labrador Current (LC) which transports cool surface waters south (Figure 2).

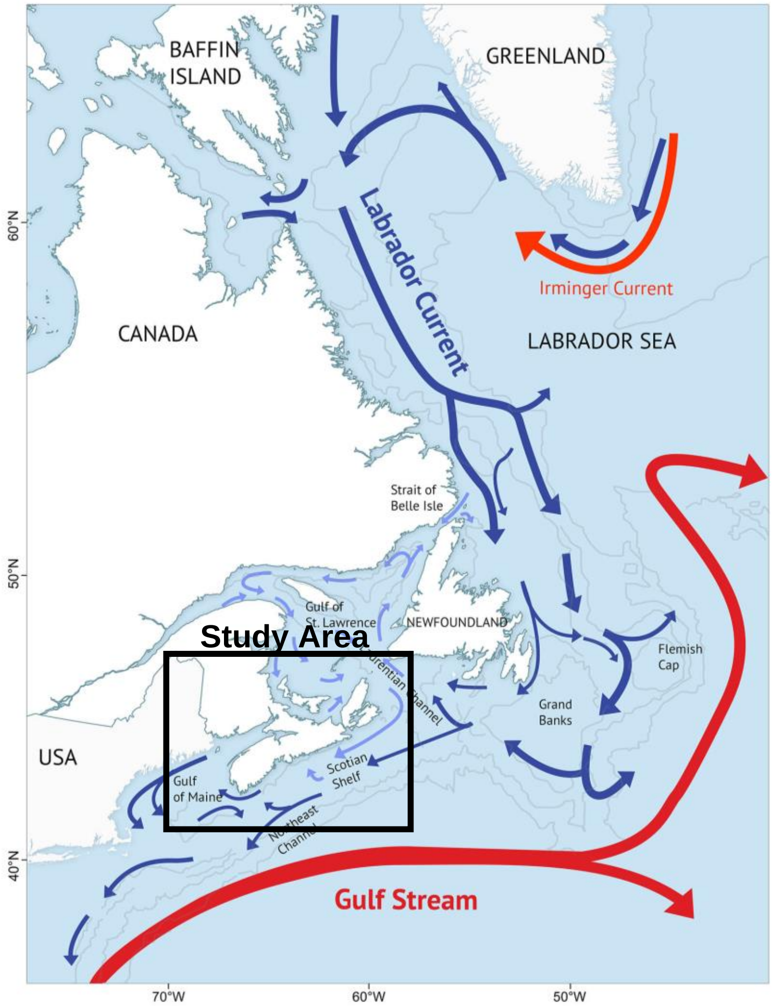

Figure 2: Ocean currents off the east coast of Canada, with cool currents in blue and warm currents in red (modified from Bernier *et al.*, 2018)

The rapid warming in the North Atlantic is linked to the weakening AMOC (Caesar *et al.*, 2020), specifically the pooling of the GS as the northwards pull on it lessens (Thibodeau *et al.*, 2018). As the admixture between the GS and LC changes, so do the boundaries and mixing of these water masses. Water masses have specific temperature and salinity signatures that are recorded by proxies in the sediment, meaning the location and boundaries of water masses can be tracked through time.

The western North Atlantic is an area of high sensitivity meaning changes in oceanographic conditions in this area are amplified. This makes the region more vulnerable to climate change, but also makes it the best place to detect changes. Studies and models have shown that the AMOC is weakening in response to anthropogenic forcing (Weijer *et al.*, 2019), with Smeed *et al.* (2018) reporting a weakening of 15% since 2004 based on RAPID array observations. A total collapse is thought to be unlikely but not impossible (McCarthy *et al.*, 2020).

While the northwest Atlantic off Nova Scotia is observed and projected to warm more strongly than the rest of the global ocean in response to a weakening AMOC, very little is known about how this region changed in response to past changes of the AMOC. Recent studies have inferred the AMOC to have been weakening over the last 150 years or so, (Thornalley *et al.*, 2018). Using alkenone paleothermometry, I have created a sea surface temperature (SST) reconstruction of the past 150 years for core MSM46 MC2 from the Gulf of St. Lawrence to determine if SST increased over this time period. Since this location is closely connected to the open ocean, I predict there would be an increase in SST.

In addition to core MSM46 MC2, I compiled alkenone-SST estimates of surface sediments from cores off the south shore of Nova Scotia and published data of SST in surface sediments on the shelf edge and offshore (Filippova *et al.*, 2016) to calculate a regional error of uncertainty of the proxy measure I use to infer past SSTs, alkenone paleothermometry. Finally, I used transects of modern shelf water collected from Atlantic Zone Monitoring Program (AZMP) cruises in 2014 to identify water masses present on the shelf and illustrate changes in their distribution through time and space (Figure 3).

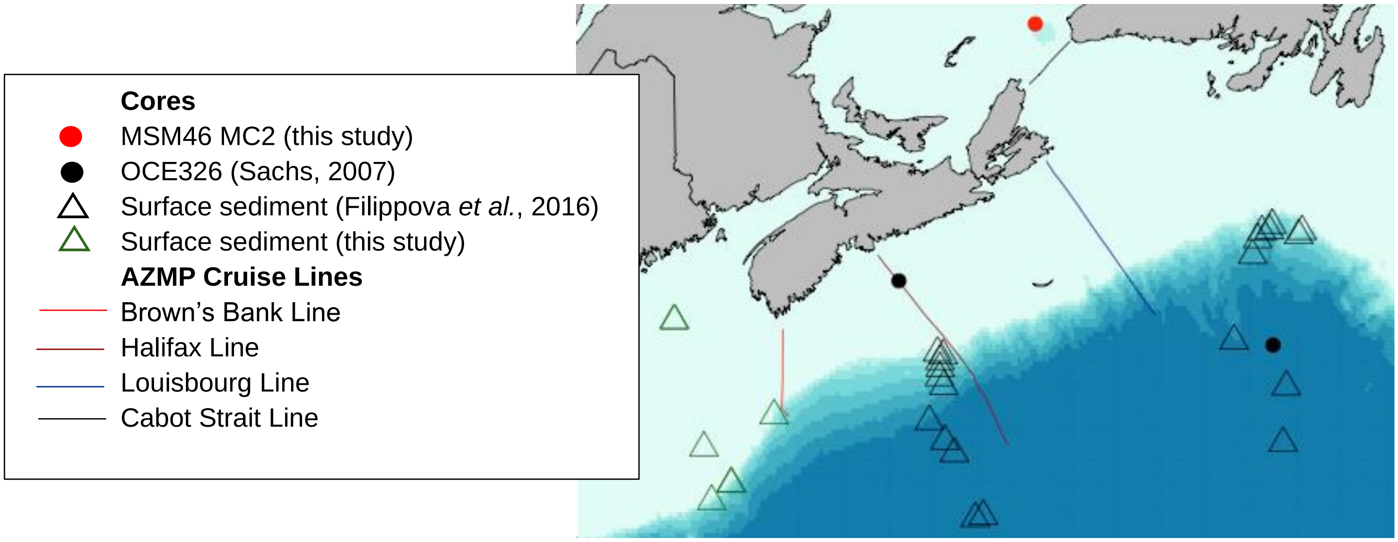

Figure 3: Locations of all data used in this study; cores, surface sediment, and AZMP cruises

CHAPTER II: BACKGROUND INFORMATION

2.1 Modern Oceanographic Setting

2.1.1 Global Ocean Circulation

Ocean currents transport huge amounts of water, heat, carbon, nutrients, and salt around the globe (Thibodeau *et al.*, 2018). Although ocean circulation is very complex, it can be simplified into the ocean conveyor belt, a schematic generalizing the path water takes around the globe. The ocean conveyor belt shows warm surface waters from low latitudes in the Atlantic flowing north to Greenland. Here, deep water is formed as the warm water cools and sinks to become bottom water which flows back south. Deep water eventually reaches the Pacific Ocean where it is upwelled to the surface, then makes the journey back to the North Atlantic as surface water to start the cycle again. In reality ocean circulation is much more nuanced. There are actually two sites of deep water formation in the North Atlantic, the Labrador Sea and the Nordic Seas (Bailey *et al.*, 2005), and one in the Southern Ocean (Broecker *et al.*, 1999). The Labrador Sea is the most dominant site of deep water formation, sourcing more than half of the waters in the southwards deep western boundary current (DWBC) which is a return flow of the northwards Gulf Stream at the surface (Bailey *et al.*, 2005). The cold air temperatures of the Labrador Sea densify surface waters to the point exceeding the density of the subsurface waters, causing an overturning where the surface water sinks. The Labrador Sea also has an excess of high salinity water due to salt extrusion, further increasing density and strengthening the overturning circulation.

The process of deep water formation is affected by melting polar ice. As the ice melts, freshwater is released into the region and reduces the density of the surface water making it harder to sink. This in turn weakens the entire AMOC and alters ocean heat transport, affecting the climate system.

2.1.2 Oceanographic conditions in western North Atlantic

The thermohaline circulation in the North Atlantic has exhibited significant interannual, decadal, and interdecadal variability in models, mostly due to changes in subpolar heat flux from the North Atlantic Oscillation (Bailey *et al.*, 2005). There have also been significant variations in circulation since the Last Glacial Maximum (LGM) due to meltwater events following glacial periods (Renssen *et al.*, 2015). During a warm period, such as the current interglacial, the AMOC is strong and deep (Lynch-Stieglitz, 2017) so heat is transported from the surface to the deep ocean (Weijer *et al.*, 2019). When enough freshwater is added to the system to shut down the AMOC completely, there is no longer any deep water formation and the ocean is highly stratified. During the last deglaciation (~18 ka – 11 ka) the intermittent AMOC shutdown caused abrupt climate changes, including large annual temperature variations in Greenland (Muschitiello *et al.*, 2019).

2.1.3 Labrador Current and Gulf Stream

This area off Nova Scotia is very sensitive because of the interplay between the warm Gulf Stream and cool Labrador Current, both part of the AMOC. The interaction between these currents affects the admixture of water masses on the Scotian Shelf, namely the Warm Slope Water (WSW), the Cold Intermediate Layer (CIL) and the Cabot Strait Subsurface (CBSS). The CBSS and CIL are both low salinity waters originating from the Gulf of St. Lawrence (Dever *et al.*, 2016), while the WSW is derived offshore from the Gulf Stream (Gatien, 1976). As a result of global warming, the Gulf Stream is shifting northwards and the Labrador Current is retreating, causing an increased proportion of WSW entering the Scotian Shelf leading to enhanced warming (Saba *et al.*, 2016).

2.1.4 Gulf of St. Lawrence

The Gulf of St. Lawrence (GSL) has two ocean inputs, through the Cabot Strait between Cape Breton and Newfoundland and through the Strait of Belle Isle between Newfoundland and Labrador, and a riverine input from the St. Lawrence River. It is highly stratified and has many forcings: 1) tides from the Atlantic, 2) meteorological seasonal and transient events, 3) freshwater runoff and heat flux, and 4) perturbations at the edge of the continental shelf (Koutitonsky and Bugden, 1991). Due to the input of both riverine and saltwater inputs, the GSL is a large estuary where eddy formation and current instabilities are expected, further complicating the temperature signature. Circulation in the GSL is difficult to study as it is highly variable both spatially and seasonally, though it has a broadly cyclonic tendency (Han *et al.*, 1999). The circulation is highly dependent on St. Lawrence River discharge, as it is responsible for over half the freshwater input in the GSL. Discharge varies seasonally, with minimum outflow in February and peak outflow in May (El-Sabh, 1976). The influence of different inputs is not uniform throughout the GSL, some sites are not affected by the St. Lawrence River and are reflective of the open ocean. This includes the core site from this study. Although SST in parts of the GSL are influenced by the admixture of the GS and LC, which in turn is driven by the AMOC, there are many other factors that can affect surface temperatures.

The Laurentian Channel extends from the continental slope to the GSL through the Cabot Strait, separating the eastern Scotian and southern Newfoundland shelves and providing a pathway for water exchange between the Atlantic Ocean and the GSL. The general structure of the Cabot Strait is two-layered, in which the upper layer flows out of the GSL and the lower layer flows in from the Atlantic (El-Sabh, 1977). However, outgoing transport is skewed so that the general direction of flow is out of the GSL in the northern Cabot Strait near Newfoundland and flow is into the GSL in the southern part near Cape Breton. Deep waters of the GSL are supplied continuously through the Laurentian Channel from Labrador- and Slope-derived waters from the Atlantic (Koutitonsky and Bugden, 1991).

2.2 Reconstructing Past SST

Past sea surface temperatures can be reconstructed using different proxies from sediment cores from a given location. In this study I will be using alkenone paleothermometry. Alkenones are long-chain primarily di- and tri-unsaturated ketones synthesized by living phytoplankton of the *Prymnesiophyceae* class (Prahl and Wakeham, 1987). The degree of unsaturation is a physiological response to growth temperature as it is a primary control of membrane fluidity (Prahl and Wakeham, 1987). Phytoplankton live in the surface water (approximately upper 100 m) in order to receive sunlight needed for photosynthesis (WHOI, 2021), meaning their internal chemistry reflects surface water living conditions such as temperature. When phytoplankton die, alkenone synthesis stops. The dead phytoplankton sink and are buried in the sediments, therefore preserving the alkenone proportions. The proportion is recorded as U^{K}_{37} which can then be converted to temperature using a calibration. One of the most widespread global calibrations and the one used in this study is the Müller calibration (Müller *et al.*, 1998), which is further explained in section 3.1. Paleo temperature can be calculated at set depths in a core with known ages to construct a time-series of SST for the given core location.

Though individual phytoplankton are very small, ranging from 2 – 200 μm (Acevedo-Trejos *et al.*, 2015), they appear in large communities known as blooms. Phytoplankton form blooms at particular times of the year that can last several weeks, though the lifespan of an individual organism rarely lasts more than a few days (Acevedo-Trejos *et al.*, 2015). The largest bloom in the GSL is in spring when the surface waters are replenished with nutrients from winter mixing (Bernier *et al.*, 2018). As a consequence, the temperature recorded by alkenones preserved in seafloor sediments only represents a short window of time and is not an annual average. However, most calibrations use annual average temperature on the x-axis of a calibration line, including in section 4.2 of this study.

2.3 Previous Paleo Work

Previous paleo studies focused on the Holocene on the Scotian Shelf and the northwest Atlantic mostly focused on the offshore region and over long time scales (2 – 12 kyrs), meaning they have low spatial and temporal resolution. Sea surface temperatures from past studies on the Scotian Shelf and offshore show a long-term cooling trend over the Holocene (Sachs, 2007; Keigwin *et al.*, 2003). A study by Sachs (2007) looked at cores OCE326-GGC26 and OCE326-GGC30, one on the Shelf and one past the continental margins (Figure 4). Both cores showed strong cooling trends in SST over the duration of the Holocene, with a 0.87 ºC/ky decrease in OCE326-GGC26 and a 0.68 ºC/ky decrease in OCE326-GGC30. However, due to the resolution of these records it is undetermined if the cooling trend has continued up to present day. The cooling of slope waters of the western North Atlantic during the Holocene is thought to be a result of declining insolation, increasing convection in the Labrador Sea, and equatorward shifting of the Gulf Stream (Sachs, 2007). The climate of the Holocene was relatively stable with a 1 – 3 ºC cooling trend in average global temperature, but because of the sensitivity of the northwest Atlantic, this cooling trend was amplified in the NW Atlantic (Sachs, 2007).

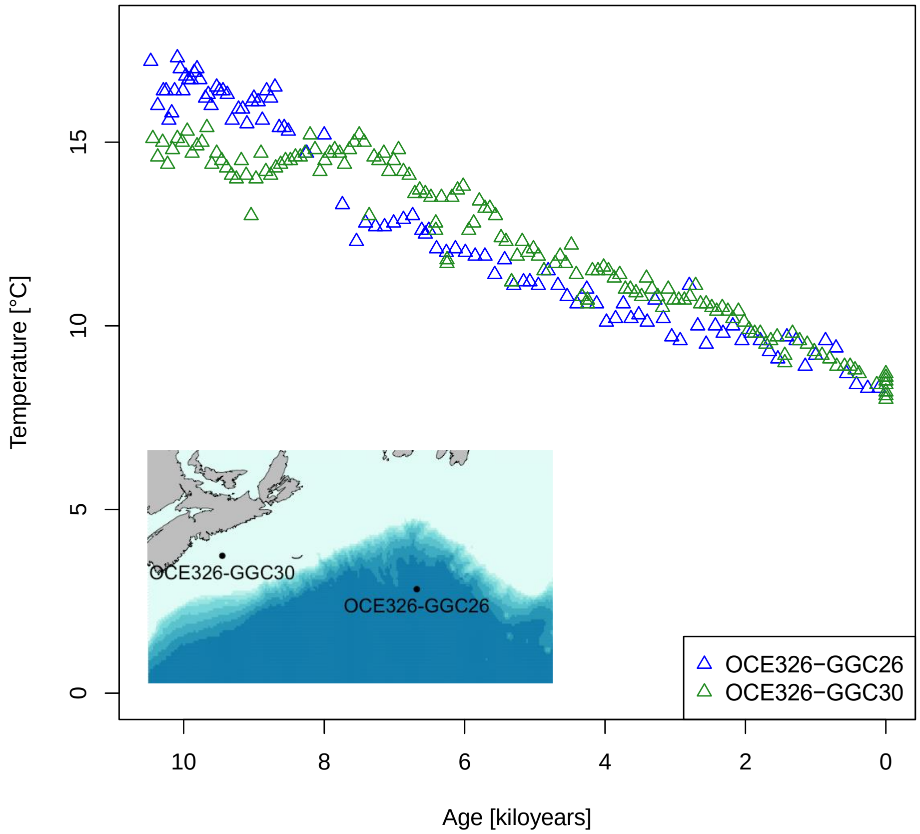

Figure 4: 10 kiloyear SST reconstruction from cores OCE326-GGC26 and OCE326-GGC30

Over the past 10 ka, there was a cooling trend in the region. However, there appears to be a slight uptick at the end of the records (present day) for cores OCE326-GGC26 and OCE326-GGC30 (Figure 4) though this cannot be confirmed without looking at a higher temporal resolution. The Holocene climate was relatively stable, especially compared to the period preceding the Holocene in which the climate was highly variable (Sachs, 2007). Though the climate was stable, air temperatures decreased over the last 8 ka due to decreasing summer insolation (McKay *et al.*, 2018). This cooling resulted in regrowth and advancement of glaciers in the Arctic as well as cooling in the surface ocean, shown by the decreasing SST. After 8 ka of decreasing, summer air temperatures in the Arctic have rapidly increased over the past century (McKay *et al.*, 2018).

A study of two cores from the Laurentian Fan indicate warming in slope waters during the Little Ice Age (LIA), approximately 600 – 150 years before present, and in one core there is no return to lower SST over the last century (Keigwin and Pickart, 1999). A core from the Emerald Basin showed sudden warming since the mid-19th century in five different proxies used to reconstruct SST, thought to be linked with the reduced transport of Labrador Current waters to the Scotian Shelf (Keigwin *et al.*, 2003). The $\delta\ ^{18}O$ records for two cores from the Laurentian Channel at the mouth of the St. Lawrence River show warming since 1850, synchronous with a sudden decrease in DWBC intensity from reduced Labrador Sea convection (Thibodeau *et al.*, 2018). Cores taken from the Gulf of Maine, the Scotian Shelf, and the Gulf of St. Lawrence (including the core investigated in this study) show an increase in anthropogenic CO_2 from the mid-20th century, though there is a lag between the atmosphere and the subsurface waters (Mellon *et al.*, 2019). This means there is collective evidence of anthropogenic perturbations, ocean warming, and a weakening AMOC in the past 150 years. For this project, the core analyzed is from the Gulf of St. Lawrence and shows a detailed record of the last 150 years, the time period with the biggest signal of change in AMOC strength from anthropogenic perturbations (Zhu and Liu, 2020).

CHAPTER III: METHODS

3.1 Alkenone Paleothermometry

Alkenone paleothermometry is a method for estimating past sea surface temperatures at the time of sediment deposition from alkenones extracted from sediment samples. The degree of unsaturation is dependent on water temperature at the time of synthesis. Unsaturation is measured by the ratio of di- to tri-unsaturated alkenones, C37:2 and C37:3 respectively, which is reported as U^{K}_{37}.

Alkenones were extracted from approximately 2 g of sediment per sample using the ASE Dionex 200 Accelerated Solvent Extractor at 1000 psi pressure and 100°C with 5 min static phases using a mixture of dichloromethane and methanol (93:7 v/v) to get total lipid extraction (TLE). After evaporating the TLE to complete dryness in the TurboVap LV at 40°C with gas pressure of 40 psi, the samples were saponified for 2 h at 80°C using approximately 15 mL of 0.5 M KOH/MeOH. I then used silica columns to separate the alkenones from the samples in a solution of dichloromethane and hexane (2:1 v/v), which was evaporated to dryness in a heated sand plate. I added hexane to re-dissolve the alkenones before gas chromatograph analysis.

The samples were analyzed by capillary gas chromatography, Agilent model. For each run, I additionally analyzed two blanks, an internal standard containing known concentrations of organic compounds with C19, C36 and C40, and a culture of *E. huxleyi*, as well as two samples from the eastern equatorial Pacific. The blanks are necessary to ensure there was no contamination from the extraction process. The internal standard was used to identify retention times for the C19, C36 and C40 peaks, and along with the blanks, used to quantify sample loss during the extraction process. Since the internal standard is added to every sample, including blanks, at the beginning of the extraction process, I was able to calibrate alkenone loss by comparing the peak area of the C36 in the blanks and the internal standard analyzed. The samples from the Pacific were used to ensure the method was working correctly, as they had been previously analyzed. Peak areas of C36, C37:2 and C37:3 in the samples were identified based on retention times of the *E. huxleyi* culture. The peak area of the alkenone peaks, C37:2 and C37:3, were used to calculate $U^{K}_{37}{}'$ in the equation:

$$U^{K}_{37}{}' = \frac{(C37{:}2 + C37{:}3)}{C37{:}2} \quad \text{Equation 1}$$

This ratio was used to estimate sea surface temperatures at the time of alkenone synthesis using the Müller calibration:

$$U^{K'}_{37} = 0.033\ SST + 0.044 \qquad \text{Equation 2 (Müller } et\ al., \text{ 1998)}$$

I extracted alkenones from various surface sediment samples in the general study area as well as the downcore record for MSM46 MC2 at every other centimetre, starting at 1 cm. I compiled the surface sample alkenone results in a plot against measured sea surface temperature at the same location from the World Ocean Atlas (2018) in order to determine the local error of uncertainty.

All sediment samples were provided as loose fine-grained sediments in containers labelled by cruise and depth. Core IDs and locations are in Table 1. MSM46 MC2 was collected in the Gulf of St. Lawrence in August 2015 from the German R/V Maria S. Merian (Mellon, 2018). The surface sediments I analyzed were the tops of cores taken from Powell-Munson Intercanyon, Nygren-Kinlan Intercanyon, Corsair Canyon, Jordan Basin, and Bedford Basin; all collected in June 2019.

Table 1: Information on cores investigated

Core ID	Latitude (N)	Longitude (W)	Location	Collection Date	Type
MSM46 MC2	47°50’	60°05	Gulf of St. Lawrence	August 2015	downcore samples
BB344	N/A	N/A	Bedford Basin	N/A	surface sediment
R2109-24 core 6	40°47.1731'	66°34.9348'	Nygren-Kinlan Intercanyon	June 2019	surface sediment
R2109-22 core 4	40°47.1741′	66°34.935′	Nygren-Kinlan Intercanyon	June 2019	surface sediment
R2113-3 core 3	41°53.0627′	65°39.6413′	NE Channel	June 2019	surface sediment
R2111-2 core 2	41°21.5038'	66°9.6026′	Corsair Canyon	June 2019	surface sediment
R2114-3 core 3	43°23.1858'	67°48.3726'	Jordan Basin	June 2019	surface sediment
R2108-17 core 7	40°30.1917′	67°0.1139′	Powell-Munson Intercanyon	June 2019	surface sediment

3.2 Creating Section Plots of Modern Scotian Shelf

Seasonal section plots of modern water column profiles act as a conceptual analogue to the predicted changes in distribution of water masses over time. I created section plots of water temperature data collected during the spring and fall Atlantic Zonal Monitoring Program

(AZMP) 2014 cruises for the Halifax, Brown's Bank, and Louisbourg lines on the Scotian Shelf. The plots were created in RStudio with the 'plot' command in the package 'oce', then I manually determined water mass boundaries based on temperature signals.

Data was provided for temperature, salinity, total alkalinity, dissolved inorganic carbon, and δ^{13}C. Each cruise line had multiple stations; Brown's Bank line had 7 stations, Halifax line had 12 stations, and Louisbourg line had 7 stations. At each station, data was collected every 10 m for the upper 60 m, then every 20 m down to 100 m, then every 50 m down to 200 m depth.

In order to plot the sections, I compiled the cruise ID, station ID, coordinates, depth at which data was collected, and all variables measured at that depth in a single Excel spreadsheet. Each cruise, identified by the line and the season (i.e. spring or fall), was assigned a number. To extract data for a specific cruise, I was able to create a subset from the spreadsheet based on the cruise number. For each subset, I used the 'as.section' command to compile the temperature, salinity, latitude, longitude, depth, and station objects into a section which I then plotted with distance on the x-axis. The 'oce' package in r is able to calculate distance between stations from latitude and longitude coordinates in the section. I limited the y-axis to not exceed 200 m water depth since the upper part of the water column is the most active biologically and this study focused on change in surface water temperatures. When plotting, I had to specify in the 'plot' command which variable I wanted to display. I created plots of temperature and salinity for each cruise, however, I focused on the temperature plots only.

CHAPTER IV: RESULTS

4.1 Modern Shelf Transects

The distinct temperatures signatures seen in Figure 5 represent different water masses. A water mass is a parcel of water with uniform characteristics throughout, such as temperature and salinity. I drew lines at the sharpest gradients to highlight the water mass boundaries. For all four cruise lines in the spring, there was a cold water mass at the surface close to the coast (0 km). Although the data is cut off due to bathymetry, it is evident this mass extended to about 80-100 m depth. The extent of this cold water mass increases moving north along the Scotian Shelf, and in the Cabot Strait spring line (Fig. 5g), the cold layer extends across the entire transect. I inferred this was the Cold Intermediate Layer (CIL). The CIL originates from the GSL and is characterized by low salinity waters between 50 – 120 m (Dever *et al.*, 2016).

Moving away from the coast closer to the shelf edge, there was a relatively sharp gradient in the Brown's Bank line in spring (Fig. 5a) where temperatures increase from ~2 ºC up to 10 ºC in the subsurface waters. An increase in surface temperatures offshore was also observed in the Louisbourg (Fig. 5e) and Halifax (Fig. 5c) spring lines, though in the Halifax line the gradient is less pronounced, and in the Louisbourg line the increase occurs much further from the coast. I inferred this warm offshore water mass to be the Warm Slope Water (WSW), a water mass derived offshore from the Gulf Stream (Gatien, 1976). In the Brown's Bank spring line (Fig. 5a), the WSW was about 40-60 m below the surface; in the Halifax spring line (Fig. 5c) the WSW was less constrained and reached the surface as well as extended further towards the coast; and in the Louisbourg spring line (Fig. 5e) the WSW was about 80 m below the surface. In the Cabot Strait spring line (Fig. 5g), the CIL dominated the entire top of the profile from the surface down to about 100 m depth. The water near the surface reached even colder temperatures than in the other lines, down to -1 ºC. There was a warmer water mass below about 200 m depth, though it was only 6 ºC compared to the 12 – 14 ºC WSW in the Brown's Bank (Fig. 5a) and Halifax (Fig. 5c) spring lines. However, in the Louisbourg spring line (Fig. 5e) the WSW was only about 8 ºC.

In the fall, the water masses and boundaries look considerably different from the spring conditions. In general, the Scotian Shelf waters were much more stratified in the fall due to surface warming during the summer months. The WSW was present at the surface in all cruise lines, most notably so in the Brown's Bank fall line (Fig. 5b) where it extended all the way to the coast. The warm water at the surface supresses the CIL; in the Halifax fall line (Fig. 5d) the CIL was present only between 40 and 80 m, while in the Brown's Bank fall line (Fig. 5b) the CIL was not present at all. The latter observation may mean that the CIL itself warmed in the summer. However, the CIL in the spring for the Brown's Bank line (Fig. 5a) is characterized by 2 – 3 ºC water, while the unidentified water mass in the fall profile below 60 m depth was approximately 10 – 12 ºC (Fig. 5b). This would be a large temperature increase for subsurface waters. The CIL was suppressed to below 50 m in the Louisbourg (Fig. 5f) and Cabot Strait (Fig. 5h) fall lines. Like the spring transects, the extent of the CIL off the coast, though now in the subsurface, diminished moving south along the Scotian Shelf. In the Cabot Strait spring line (Fig. 5h), the subsurface waters were colder than that of the other lines at < 2 ºC. Subsurface waters in the Louisbourg fall line (Fig. 5f) were ~2 ºC, increasing to ~ 4 ºC in the Halifax fall line (Fig. 5d) and reaching up to 10 – 12 ºC in the Brown's Bank fall line (Fig. 5b).

The influence of the WSW at the surface as well as the temperature increased moving south down the Scotian Shelf. As previously mentioned, the WSW reaches all the way to the coast in the Brown's Bank fall line (Fig. 5b), then moves further offshore in the north. However in the Cabot Strait fall line (Fig. 5h), the WSW was at the coast and only extended 40 km offshore. The Cabot Strait was very stratified in the fall, with warm surface waters of 10 – 14 ºC extending to less than 25 m depth, then immediately below the relatively warm surface waters, there was a sharp gradient to the CIL at < 2 ºC from about 50 – 150 m. Below the CIL the water actually warmed up again to about 6 ºC.

In the fall there was another water mass present in the Halifax (Fig. 5d), Louisbourg (Fig. 5f), and Cabot Strait (Fig. 5h) lines; the Cabot Strait Subsurface water (CBSS). By definition the CBSS is the layer that sits between 30 – 50 m (Dever *et al.*, 2014). The CBSS separated the CIL from the WSW and was approximately 8 – 10 ºC.

These transects display how water masses change position and shape both through space and time. As previously mentioned, the suppression of the WSW or CIL changes seasonally. In the Louisbourg line (Fig. 5e), WSW was observed on the shelf edge below 80-100 m depth. Moving south along the coast in the spring (Fig. 5c), the WSW shallowed and widened, effectively narrowing the CIL and pushing it closer to the surface. Further south (Fig. 5a), the WSW retreated back toward the shelf edge and deepened, allowing the CIL to extend further on the shelf though it stayed near the surface. In the fall, the WSW was only present on the shelf edge allowing the CBSS to extend across the shelf in a thin layer on top of the CIL (Fig. 5f). Moving south, the WSW shallowed and extended closer to shore (Fig. 5d). This pushed the CBSS and in turn the CIL slightly deeper, and also forced the warmer water from the CBSS to bend around the CIL making the CIL much smaller. Further south the WSW shallowed even more so that it was a wide wedge of warm water floating at the surface extending almost the entire length of the shelf (Fig. 5b). This pushed the warm waters surrounding the WSW even deeper so that the CIL was no longer present.

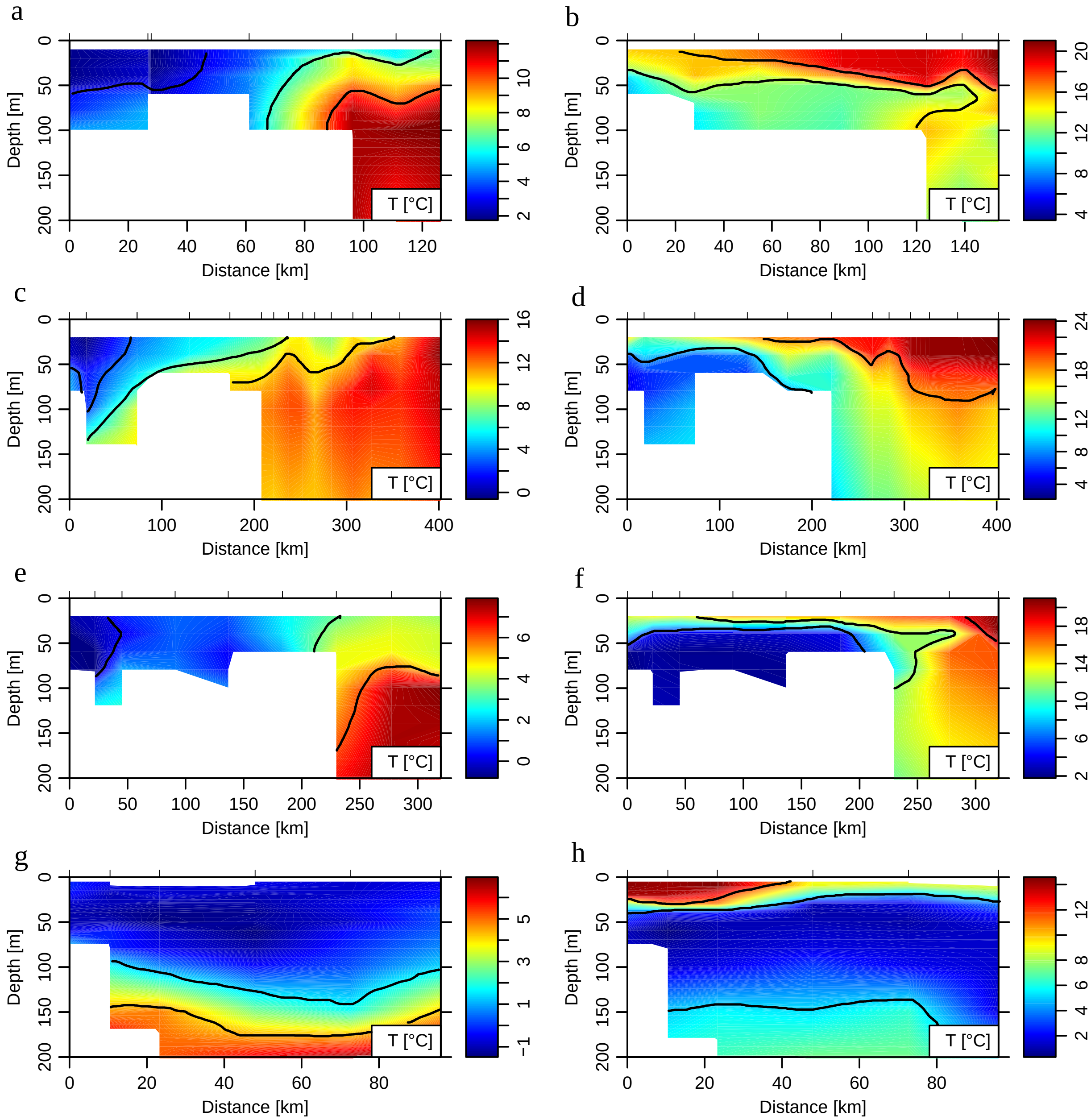

Figure 5: Temperature section plots of (a) Brown's Bank line spring and (b) fall, (c) Halifax line spring and (d) fall, (e) Louisbourg line spring and (f) fall, and (g) Cabot Strait line in spring and (h) fall from 2014 AZMP cruises

4.2 Surface Sediment Temperature Estimates

I looked at surface sediments samples from the Scotian Shelf from a previous study in this region (Filippova *et al.*, 2016) that had been analyzed for U^{K}_{37}. New surface sediment alkenone unsaturation data were merged with the published data from Filippova *et al.* (2016), as well as the core top sediment from MSM46 MC2 (Table 2). Since the sediment samples are from the surface, they are assumed to represent the current surface water conditions. As a result, I compared the U^{K}_{37} values with the 2018 World Ocean Atlas SST record for each sample location (Figure 6). The red line in figure 6 is the Müller Calibration (Equation 2). The sediment points are distributed relatively evenly on either side of the calibration. I excluded the sediment samples from the Bedford Basin and Corsair Canyon (Table 1) because they showed anomalously high and low U^{K}_{37} respectively. Anomalously high U^{K}_{37} values may be a result of contamination during the extraction process, while low values may indicate there were not enough alkenones in the sample to be detected. I calculated the root mean squared error (rmse) of the U^{K}_{37} values, using the 'rmse' command in r from the 'metrics' package, to be 0.0915 which I then converted to temperature using the Müller calibration to get an average local error of 1.4 ºC. This average local uncertainty is shown by the shaded red area. Some of the samples did not fall in this boundary of uncertainty because it is an average error of uncertainty for all the points. As mentioned previously in section 2.2, the Müller Calibration as well as the plot in figure 6 compare a mean annual temperature to the alkenone unsaturation ratio, which only records the surface temperature at the time of the bloom. Studies show that using temperatures from only one season, the time of the phytoplankton bloom, rather than mean annual temperatures has little effect on the overall U^{K}_{37}-SST relationship in regions with minor seasonal differences (Filippova *et al.*, 2016). Since there is strong seasonality in SSTs in the northwest Atlantic, it is possible the Müller Calibration has a seasonal bias for this region.

I used the Müller calibration (Müller *et al.*, 1998) to calculate SST. This is a widely used global calibration relating U^{K}_{37} to SST, based off samples from latitudes between 60 ºS and 60 ºN covering a 0 – 29 ºC range in SSTs (Filippova *et al.*, 2016). However, because it is a global calibration the uncertainty can change depending on region. This plot shows the uncertainty for the Scotian Shelf, and because the points are relatively close to the red line, this suggests the Müller calibration is a good calibration to use in this area. When estimating past SSTs for the downcore record of MSM46 MC2, I used the Müller calibration with an error of +/- 1.4 ºC.

Table 2: Surface sediment samples used to determine local error for Müller Calibration

Core ID	Latitude (N)	Longitude (W)	Depth (cm)	U^{K}_{37}	Mean Annual SST (WOA, 2018)	Source
2006_048_0029	44.77	55.08	0	0.23	8.648	Filippova *et al.*, 2016
99036_035	40.22	61.37	0	0.69	18.703	Filippova *et al.*, 2016

2002_046_079	44.85	55.02	0	0.23	8.648	Filippova *et al.*, 2016
2004_024_006	44.76	54.38	0	0.23	8.857	Filippova *et al.*, 2016
2000_036_016	42.88	62.17	0	0.32	10.612	Filippova *et al.*, 2016
98039_009	41.44	54.79	0	0.47	16.941	Filippova *et al.*, 2016
2006_048_0024	44.63	55.33	0	0.29	8.648	Filippova *et al.*, 2016
2006_048_0020	44.75	55.24	0	0.28	8.648	Filippova *et al.*, 2016
2000_036_030	42.35	62.04	0	0.53	10.612	Filippova *et al.*, 2016
98039_010	42.34	54.72	0	0.45	13.136	Filippova *et al.*, 2016
91020_014	41.79	62.35	0	0.54	13.838	Filippova *et al.*, 2016
2002_046_077	44.70	54.45	0	0.24	8.857	Filippova *et al.*, 2016
2004_030_003	42.83	62.05	0	0.34	10.612	Filippova *et al.*, 2016
9102_0008	41.47	62.02	0	0.71	13.838	Filippova *et al.*, 2016
96029_069	43.07	55.83	0	0.46	10.491	Filippova *et al.*, 2016
99036_052	42.74	62.12	0	0.34	10.612	Filippova *et al.*, 2016
99036_055	42.64	62.12	0	0.32	10.612	Filippova *et al.*, 2016
91020_012	41.27	61.81	0	0.57	15.159	Filippova *et al.*, 2016
99036_037	40.26	61.20	0	0.59	18.703	Filippova *et al.*, 2016
2002_046_074	44.39	55.44	0	0.28	8.648	Filippova *et al.*, 2016
R2109-24 core 6	40.78	66.58	1	0.57	13.003	This study
R2109-22 core 4	40.78	66.58	1	0.56	13.003	This study
R2113-3 core 3	41.88	65.67	1	0.53	11.056	This study
R2114-3 core 3	43.38	67.80	1	0.40	8.881	This study
R2114-3 core 3	43.38	67.80	3.5	0.47	8.881	This study
R2108-17 core 7	40.50	67.00	1	0.56	11.907	This study
MSM46 MC2	47.83	60.08	1.5	0.27	6.206	This study

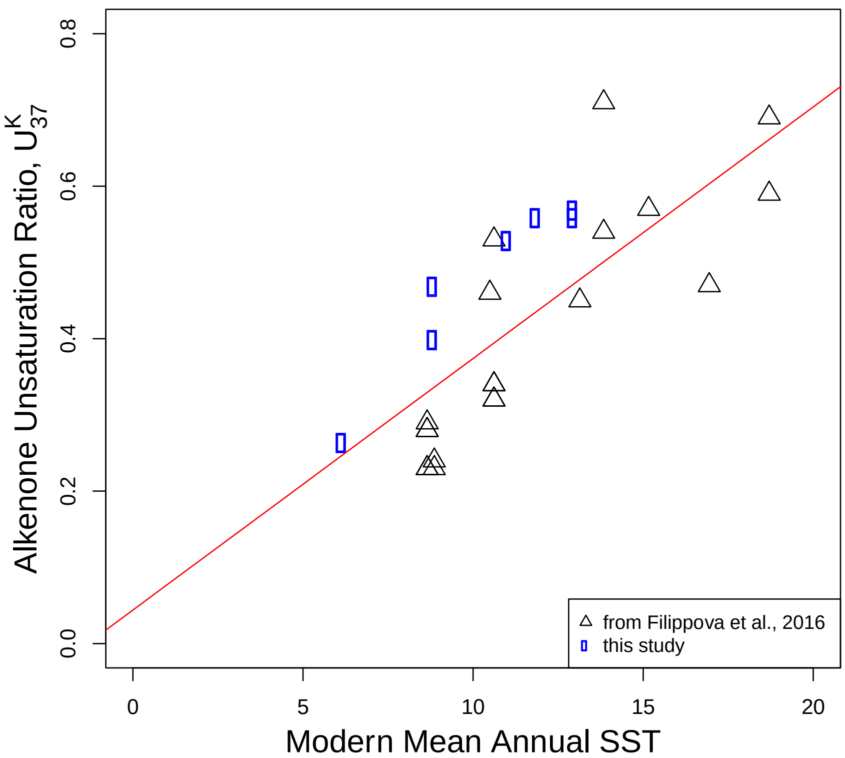

Figure 6: Surface sediment alkenone unsaturation ratio versus modern mean annual sea surface temperature from World Ocean Atlas (WOA, 2018)

4.3 Downcore Record of Past SST

Figure 7 shows the reconstructed sea surface temperatures from 1844-2004 using the Müller calibration in core MSM46 MC2 (Table 3). There is no trend in the data, meaning there was no warming or cooling over the past 150 years; SST has not changed over the past 150 years within the error of uncertainty. This means that the uptick seen in Figure 4 was not actually the start of a warming trend, though it does corroborate that the cooling trend ended. I hypothesized that there would be a warming trend caused by the weakening of the AMOC. Previous studies on the Northwest Atlantic continental shelf and slope show there is an accelerated warming trend in both the surface and subsurface waters over recent decades, serving as evidence that the AMOC is weakening (Chen *et al.*, 2020). There are a few possibilities to explain why there was no change in SST over the study period in core MSM46 MC2 which are discussed in section 5.2.

Table 3: Summary of MSM46 MC2, ages and $\delta^{13}C$ from *N. Pachyderma* (from Mellon, 2018)

Depth (cm)	Age (year)	U^K_{37}	Temperature (ºC)	$\delta^{13}C$

1.5	2000	0.27	6.71	0.670
3.5	1992	0.27	6.90	0.610
5.5	1985	0.26	6.69	0.770
7.5	1977	0.27	6.78	0.820
9.5	1969	0.28	7.21	0.790
11.5	1961	0.28	7.10	0.840
13.5	1953	0.27	6.81	0.910
15.5	1944	0.26	6.62	0.930
17.5	1934	0.27	6.90	0.830
19.5	1925	0.28	7.02	0.890
21.5	1915	0.27	6.94	0.950
23.5	1905	0.26	6.56	0.960
25.5	1895	0.27	6.70	0.890
27.5	1885	0.26	6.59	0.850
29.5	1875	0.28	7.25	0.860
31.5	1865	0.27	6.99	0.850
33.5	1855	0.27	6.94	0.990
35.5	1844	0.28	7.09	0.920

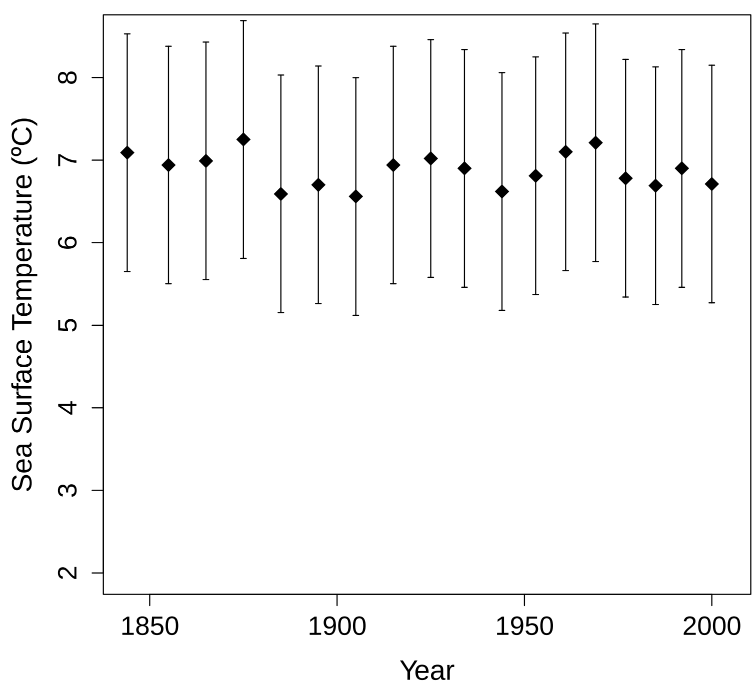

Figure 7: SST reconstructions for 150-year downcore record of MSM46 MC2

Published data of $\delta^{13}C$ from the same core shows a clear decrease in ^{13}C since the mid-20th century (Figure 8), the same isotopic signature as offshore cores. The core analyzed in this study is therefore representative of the open ocean conditions in the northwestern Atlantic. Anthropogenic CO_2 has an isotopically light signature since most fossil fuels area derived from plant material. This core location was clearly affected by anthropogenic perturbations.

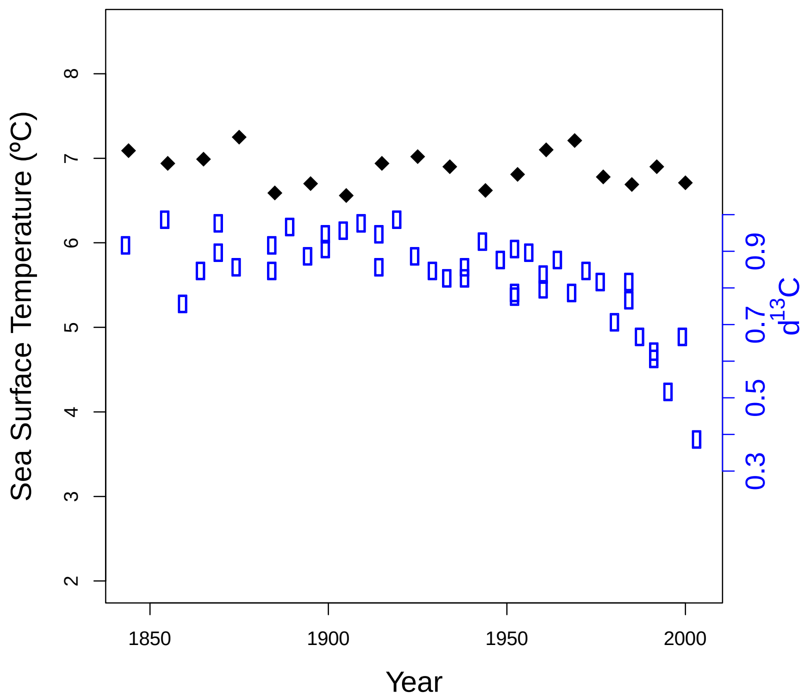

Figure 8: SST reconstructions for 150-year downcore record of MSM46 MC2 with $\delta^{13}C$

CHAPTER V: DISCUSSION

5.1 Effect of water mass distribution

I identified three water masses in the transects of modern Scotian Shelf waters (Figure 5): the Cold Intermediate Layer (CIL), the Warm Slope Water (WSW) and the Cabot Strait Subsurface water (CBSS). The influence of WSW and CIL on surface temperatures of the Scotian Shelf changed through space and time. The CIL was much more prevalent at the surface in the spring with a waning influence moving south from the GSL towards Brown’s Bank since the CIL is sourced from the GSL (Dever *et al.*, 2016). The Cabot Strait line runs between Cape Breton and Newfoundland instead of between the Nova Scotian coast and the edge of the Scotian Shelf, so there was a greater influence of the CIL in both spring (Fig. 5g) and fall (Fig. 5h).

Though water masses influence surface temperatures, there is also greater radiative forcing from the atmosphere on surface water in the summer months which causes warming (Richaud *et al.*, 2016). Surface warming affects the structure of water masses as it causes stronger stratification of Scotian Shelf waters, seen in the fall transects (Fig. 5).

Looking at the seasonal differences in the transects illustrate how long-term changes in distribution or advection of water masses results in a cooling or warming trend at the surface. During a long-term warming trend, the water mass distribution would move more towards the fall-equivalent on annual average, while during a long-term cooling trend it would move towards a spring-equivalent distribution. In other words, greater influence of the CIL leads to colder surface waters, while greater influence of the WSW leads to warmer surface waters. The distribution of water masses on the Scotian Shelf is influenced by the interplay between the LC and GS, the strength of which is driven by the AMOC (Thibodeau *et al.*, 2018). Therefore, warming or cooling trends in surface water temperatures, which are recorded in alkenone unsaturation ratios in seafloor sediment, indicate changes in water mass distribution.

5.2 Mechanisms for temperature change

Based on previous paleo studies of SST and evidence of a weakening AMOC, it was predicted that the SST would show an increasing trend for core MSM46 MC2 over the past 150 years. However, after analysis I found there was no significant increase or decrease in SST at the core location over that time period. There are three possible, non-mutually exclusive mechanisms presented here to explain this temperature consistency: 1) the complex hydrodynamics of the Gulf of St. Lawrence, 2) the WSW not penetrating to the core site, and 3) changes in the timing of the phytoplankton bloom. We cannot, however, rule out the possibility that temperature increase is not spatially uniform and therefore does not occur over the entire region.

5.2.1 Hydrodynamics in the Gulf of St. Lawrence

The circulation of water in the Gulf is complex. It is possible that due to the complex circulation, the St. Lawrence River is influencing the sea surface temperature at the core location. However, the carbon isotopic signature shows this location experiences more ocean-like conditions. The core was from the northern part of the Cabot Strait close to Newfoundland where surface waters flow into the GSL from the Atlantic; however, the strength of the inward flow here is much weaker than the outward surface flow in the southern Cabot Strait (El-Sabh, 1977). It is possible the inflow from the Atlantic is too weak to bring in the warm surface waters, or the more dominant outflow to the south is diluting the incoming warm Atlantic surface waters. Waters derived from the GSL, both the CIL and CBSS, flow inshore down the Scotian Shelf (Dever *et al.*, 2016). To determine if the strength of outflow from the GSL has changed over time, one could look at a time series of the AZMP transects and determine if the extent of the CIL and CBSS changes.

Though it has been determined the general direction of flow at the core location is into the GSL, the strength and direction of geostrophic surface currents change seasonally. This leads to the formation and deconstruction of gyres which further complicate circulation (El-Sabh, 1976). Inflow at the core location is strongest in the summer and weakest in the winter (Han *et al.*, 1999), a result of the seasonal variations in St. Lawrence River discharge. The mean flushing time of the GSL is 3 months, meaning it takes on average 3 months for water at the mouth of the St. Lawrence River to exit the GSL through the Cabot Strait (El-Sabh, 1977). The timing of peak inflow through the northern Cabot Strait could change if the timing of peak St. Lawrence River discharge changes, which is dependent on highly variable factors such as snowmelt and precipitation. Sea ice is another factor that adds variability to circulation in the GSL, as it affects surface water salinity and therefore density. Historically, ice covers the southwestern and central parts of the GSL in the winter and is carried out through the Cape Breton side of the Cabot Strait (El-Sabh, 1977). However, there is high year-to-year variability in ice formation, ice cover, ice drift, and ice melting (Koutitonsky and Bugden, 1991). The relatively constant temperature signal seen in MSM46 MC2 (Figure 7) may be due to the complex hydrodynamics of the GSL, wherein the system is too variable to record any warming or cooling trends.

5.2.2 Water mass distribution in the Cabot Strait

Another possible reason why the temperature was relatively constant over the past 150 years is that there is little variability in water mass distribution at this location. Compare the distribution of temperature at Brown's Bank in spring (Fig. 5a) and fall (Fig. 5b) to that of the Cabot Strait in spring (Fig. 5g) and fall (Fig. 5h). In the spring, the CIL dominated the surface with water at 2 – 5 ºC (Fig. 5a) while in the fall the WSW extended the surface for the entire length of the transect with waters between 15 – 20 ºC (Fig. 5b). Even in the subsurface in the fall, the water was > 10 ºC. The fluctuation between spring and fall conditions were very strong here. In the Cabot Strait, the fluctuation was much smaller as there were less pronounced changes in the distribution the CIL and WSW between spring (Fig. 5g) and fall (Fig. 5h).

Transects of the Brown's Bank, Halifax, and Louisbourg AZMP lines show a distinct difference in distribution of the Warm Slope Water (WSW) and Cold Intermediate Layer (CIL) between fall and spring (Figure 5). If these transects are treated as analogues for water mass distribution during long term trends, it is clearly shown that during a warming trend from a weakening AMOC in which the average annual distribution would move towards fall-equivalent, the WSW would be much more prevalent at the surface waters along the Scotian Shelf. However, in the Cabot Strait line the WSW is much less prevalent (Fig. 5g & 5h). Even in fall conditions (Fig. 5h), it is only present in the southern end of the line while MSM46 MC2 was taken from the northern end of this line (Figure 3). That means that the surface water at the core location has little seasonal variability. Even during a period of a weakening AMOC moving towards a fall-equivalent distribution, the surface water would stay relatively cool; changes in the strength of the AMOC do not affect the surface water temperature at this location.

5.2.3 Timing of Phytoplankton Blooms in the Gulf of St. Lawrence

The third proposed explanation for the stable SST reconstruction at MSM46 MC2 is that the timing of the phytoplankton bloom is shifting earlier in the spring. If increasing surface water temperatures are detrimental to phytoplankton growth, the bloom would occur earlier in the year when the temperature is optimal to compensate. Phytoplankton blooms happen when there is sufficient sunlight and nutrients, which are brought to the surface by mixing. In summer months when the surface waters are much warmer than the subsurface, the water column becomes stratified and the surface does not get replenished with nutrients hence phytoplankton growth decreases. Warmer SST in spring can cause stratification that would inhibit phytoplankton growth, forcing the bloom to shift earlier in the year. Marine phytoplankton biomass and productivity has been shown to decrease with rising SST due to temperature-driven stratification (Huertas *et al.*, 2011).

Conditions in the GSL were very warm between 2010 – 2012, resulting in an earlier spring bloom (Bernier *et al.*, 2018). During this time period the chlorophyll *a* inventory in the top 100 m was below average, with a peak minimum in 2011, meaning there was decreased phytoplankton biomass (Bernier *et al.*, 2018). A study from the Gulf of Maine found that the phytoplankton blooms have been shifting later in the year for both the spring and fall blooms (Record *et al.*, 2019); however, a shift to later in the year would mean the blooms are happening in increasingly warmer waters and would not explain the SST record for MSM46 MC2.

CHAPTER VI: CONCLUSION

6.1 Summary

The AMOC plays a very important role in global ocean circulation and in the global climate system. Anthropogenic forcing is causing the AMOC to weaken, affecting the rate of heat transport and carbon sequestration as well as the transport of salt, nutrients and oxygen. The weakening AMOC is affecting the interplay between the warm Gulf Stream and cool Labrador Current causing ocean warming in the western North Atlantic at a faster rate than the global average ocean warming. It was expected that this warming would extent to the Scotian Shelf; however, I found there was no significant warming over the past 150 years in the northern Cabot Strait, the location of core MSM46 MC2. Looking at past ocean behavior and determining long-term trends in temperature helps researchers make models and predictions about future conditions. Warmer temperatures influence climate and are detrimental to fish populations, so identifying and predicting ocean warming is important in making management plans for fisheries (Wilson *et al.*, 2020).

6.2 Future Research

The next step is to reconstruct sea surface temperatures over the past 150 years for more cores from coastal waters of Nova Scotia to see if the temperature was statistically constant over that time period in other locations, or if the core from the Cabot Strait is an anomaly. By looking at more cores on the Scotian Shelf, we can determine if there is warming in other places and if so, determine which areas have warmed the most over the time period. Areas with the most warming are thought to be the most susceptible to changes in the AMOC. Policy makers can use this knowledge when creating management plans to prioritize the most vulnerable areas. Understanding changes to the AMOC and how that is affecting ocean warming is key to predicting and planning for what is to come.

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