University of Rhode Island

DigitalCommons@URI

Open Access Dissertations

2021

NORTH ATLANTIC ON THE EDGE: OCEAN CIRCULATION THROUGH GATEWAYS AT THE SUBPOLAR-SUBTROPICAL BOUNDARY

Afonso Gonçalves Neto

University of Rhode Island, afonsomgneto@gmail.com

Follow this and additional works at: https://digitalcommons.uri.edu/oa_diss

Recommended Citation

Gonçalves Neto, Afonso, "NORTH ATLANTIC ON THE EDGE: OCEAN CIRCULATION THROUGH GATEWAYS AT THE SUBPOLAR-SUBTROPICAL BOUNDARY" (2021). *Open Access Dissertations*. Paper 1233.
https://digitalcommons.uri.edu/oa_diss/1233

This Dissertation is brought to you for free and open access by DigitalCommons@URI. It has been accepted for inclusion in Open Access Dissertations by an authorized administrator of DigitalCommons@URI. For more information, please contact digitalcommons-group@uri.edu.

NORTH ATLANTIC ON THE EDGE: OCEAN CIRCULATION THROUGH
GATEWAYS AT THE SUBPOLAR-SUBTROPICAL BOUNDARY

BY

AFONSO GONÇALVES NETO

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
IN
OCEANOGRAPHY

UNIVERSITY OF RHODE ISLAND

2020

DOCTOR OF PHILOSOPHY DISSERTATION
OF
AFONSO GONÇALVES NETO

APPROVED:

Dissertation Committee:

Major Professor	Jaime B. Palter
	Kathleen Donohue
	Emily P. Diamond
	H. Thomas Rossby
	Brenton DeBoef
	DEAN OF THE GRADUATE SCHOOL

UNIVERSITY OF RHODE ISLAND

2020

ABSTRACT

Labrador Sea Water (LSW) is a major component of the deep limb of the Atlantic Meridional Overturning Circulation, yet LSW transport pathways and their variability lack a complete description. A portion of the LSW exported from the subpolar gyre is advected eastward along the North Atlantic Current and must contend with the Mid-Atlantic Ridge before reaching the eastern basins of the North Atlantic. In Chapter 1, we analyze observations from a mooring array and satellite altimetry, together with outputs from a hindcast ocean model simulation, to estimate the mean transport of LSW across the Charlie-Gibbs Fracture Zone (CGFZ), a primary gateway for the eastward transport of the water mass. The LSW transport estimated from the 25-year altimetry record is 5.3 ± 2.9 Sv, where the error represents the combination of observational variability and the uncertainty in the projection of the surface velocities to the LSW layer. Current velocities modulate the interannual to higher-frequency variability of the LSW transport at the CGFZ, while the LSW thickness becomes important on longer time scales. The modeled mean LSW transport for 1993–2012 is higher than the estimate from altimetry, at 8.2 ± 4.1 Sv. The modeled LSW thickness decreases substantially at the CGFZ between 1996 and 2009, consistent with an observed decline in LSW volume in the Labrador Sea after 1994. We suggest that satellite altimetry and continuous hydrographic measurements in the central Labrador Sea, supplemented by profiles from Argo floats, could be sufficient to quantify the LSW transport at the CGFZ.

A decade ago, abrupt warming of the Northwest Atlantic Shelf prompted an ecosystem shift with significant consequences for fisheries, and the cause of this event has remained unclear. In Chapter 2, we show that a heightened presence of the Gulf Stream at the Tail of the Grand Banks (TGB) reduced the westward connectivity of the Labrador Current that otherwise supplies cold, fresh,

oxygen-rich waters to the shelf. The appearance of anomalous properties at the TGB was followed by a southwestward progression of subsurface warming one year later. Thus, monitoring conditions at the TGB may offer predictability for shelf properties and ecosystem perturbations with substantial lead time. Historical observations suggest this recent shift was preceded by a similar phenomenon in the 1970s, indicating that just two events are responsible for much of the dramatic warming of the Northwest Atlantic Shelf over the past century. In Chapter 3, we performed a Lagrangian analysis in the high-resolution ocean circulation model HYCOM to investigate the control of the Gulf Stream on the continuity of the Labrador Current past the Grand Banks of Newfoundland. The Labrador Current can be completely cut off at or upstream of the Tail of the Grand Banks by Gulf Stream eddies and meanders that impinge on the shelf-break. In the simulation, this process became more common after 2007, leading to warm/salty anomalous conditions in the Northwest Atlantic.

ACKNOWLEDGMENTS

First, I would like to thank the Graduate School of Oceanography, the National Science Foundation and the National Oceanic and Atmospheric Administration for supporting my work as a graduate teaching and research assistant. Additionally, I would like to acknowledge the National Science Foundation, through SciWrite@URI, for supporting my writing internship with the Metcalf Institute and the Thomas and Kathy J. McNiff Graduate Student Endowment in Marine Sciences for the 2018 and 2020 awards.

I would like to thank my major professor, Dr. Jaime Palter, for the continuous support and amazing mentorship during the past years. I am specially thankful for her ability to understand my needs as a graduate student and guide me through the program. Beyond helping me become part of the conversation in the Physical Oceanography community, Dr. Palter has always supported my adventures in science communication and leadership roles. More importantly, she has taught me that, as scientists, our professional and personal lives are not in opposition and that we do not have to choose one or the other. This has really made a difference for me and has helped me shape my career goals.

My time at GSO would not have been the same without the continuous support of amazing faculty, staff and fellow students. I owe the GSO (and beyond) community a lot for helping me grow as a person and a scientist. In particular, I would like to acknowledge my committee, Dr. Kathy Donohue, Dr. Emily Diamond and Dr. Tom Rossby for the dedication to help me develop my work; my friends and lab mates, Don Rudnickas and Sarah Nickford for the daily support and friendship; my co-authors Xiaobiao Xu, Amy Bower, Heather Furey, Joe Langan and Paula Fratantoni for the terrific teamwork; faculty members Dr. David Smith, Dr. Kelton McMahon, Dr. Colleen Mouw, Dr. Kendall Moore, Dr. Sun-

shine Menezes and Dr. Nedra Reynolds for the lessons inside and outside of the classroom; staff members Cara Mitnick, April Pariseault, Chi Shen, Dr. Wilsage Sagesse, Cornelia Lineham and Meredith Clark for the amazing support.

To become a PhD student, I had massive support as an undergraduate at the Federal University of Santa Catarina (UFSC) and as a research technician at the Bermuda Institute of Ocean Sciences (BIOS). I would like to thank my mentors Regina Rodrigues and Alessandra Fonseca, at UFSC, and Rod Johnson, at BIOS, for teaching me so many valuable lessons along the way.

Last but certainly not least, I would like to thank the people who have been by my side through all of the good and bad moments from the beginning: my friends and family. I will not mention names, but each of them knows exactly how much they have helped me and how much I love them. As to my parents, Martha e José, they have to be mentioned. They have always prioritized my education and have provided me with all the love and resources to keep moving. This is for you! Amo vocês!

TABLE OF CONTENTS

ABSTRACT	ii
ACKNOWLEDGMENTS	iv
TABLE OF CONTENTS	vi
LIST OF FIGURES	ix
LIST OF TABLES	xix
MANUSCRIPT	
1 Labrador Sea Water Transport across the Charlie-Gibbs Fracture Zone	1
1.1 Introduction	1
1.2 Data and Methods	4
1.2.1 Moorings in the Western CGFZ	4
1.2.2 Satellite altimetry and other observational data	8
1.2.3 Numerical Simulation	10
1.3 Key Results	11
1.3.1 LSW Transport at the CGFZ Mooring Array	11
1.3.2 Circulation variability at the CGFZ from altimetry	13
1.3.3 Model-based LSW transport across the CGFZ	18
1.4 Discussion and Conclusions	22
List of References	26
2 Changes in the Gulf Stream Precede Rapid Warming of the Northwest Atlantic Shelf	33

	Page
List of References	88
BIBLIOGRAPHY	93

LIST OF FIGURES

Figure		Page
1	Circulation and bathymetry of the subpolar North Atlantic. (a) Schematic circulation diagram showing primary shallow (red) and deep (yellow and blue) currents in the subpolar North Atlantic. The LSW formation regions are shown as yellow circles, and its main pathways are indicated as yellow arrows. Abbreviations include Bight Fracture Zone (BFZ), Charlie-Gibbs Fracture Zone (CGFZ), Faraday Fracture Zone (FFZ) and Maxwell Fracture Zone (MFZ). (b) Detailed bathymetry at the Mid-Atlantic Ridge for the region that includes the Charlie-Gibbs Fracture Zone. The mooring array is indicated in black on the west side of the CGFZ.	2
2	Zonal velocity and salinity at the CGFZ mooring array. (a) Cross section of time-mean velocity from moored current meters. White triangles indicate the location of the current meters. Thick black line is the 0 cm s^{-1} isotach. Mooring labels are indicated on the top of the figure. (b) Similar to panel (a) but for salinity. White circles indicate the location of the MicroCats. Thick black line is the 34.94 isohaline. Dashed white lines are isopycnals. Mean locations of LSW, ISOW and Lower Deep Water (LDW) are indicated. Figure from [16].	3

- 3 LSW properties and transport at the mooring array. (a) 22-month Hovmöller diagram of the 30-day low-passed LSW thickness anomaly at the mooring array. The latitude of each mooring (A is the northernmost) is shown as dashed lines. Thick black line is the 0-m thickness anomaly contour. (b) 22-month mean LSW thickness across the mooring array. Dashed lines indicate the latitude of each mooring. (c) LSW zonal velocity anomaly through the mooring array, with same plotting conventions as in panel (a). Thick black line is the 0-cm s^{-1} zonal velocity anomaly contour. (d) 22-month mean LSW zonal velocity. (e) 22-month time series of the 30-day low-passed LSW thickness (blue), zonal velocity (green) and zonal transport (red) averaged across the eight moorings. The red lines, whose correlation coefficient R is equal to 0.33, are the transport as calculated from the moored instruments (solid) and the model output interpolated to the mooring positions (dashed). The dashed black line denotes LSW zonal velocity and transport equal to zero. 5
- 4 Mean surface geostrophic velocity (arrows) overlaying the mean dynamic topography (colors) in the CGFZ region. The red line to the left indicates the locations of the eight moorings. The black dashed line is the mean position of the 39.5 cm ADT isoline, which coincides with the maximum surface velocities associated with the northern branch of the NAC. A more ideal position for the quantification of LSW transport across the CGFZ is also shown in red, shifted 3° to the east of the moorings. This line is located at the mean position of the ridge crest at the latitude band of the CGFZ, as calculated from ETOP01 [40]. It is bounded by the 3,000-m isobath to both the north and south, and extends from 51.7°N to 53°N . The 3,000-m isobath is shown as a white contour. 11
- 5 Twenty-two month Hovmöller diagram of the 30-day low-passed zonal surface geostrophic velocity anomaly at the location of the mooring array. The solid black contour is the 0-cm s^{-1} velocity anomaly. Dashed lines indicate the latitude of each mooring. . . 14

Figure		Page
6	Eastward transport of LSW through the Mid-Atlantic Ridge. (a) Bathymetry map of the Mid-Atlantic Ridge between 40°N and 53°N, showing the 35°W line (red) and the center of the Mid-Atlantic Ridge rift valley (black). Dots represent the grid points at which the lines in panel b were calculated, with a 0.5° latitudinal resolution. (b) Total (light lines) and cumulative (dark lines) zonal transport of LSW at 35°W (red) and the Mid-Atlantic Ridge rift valley (black), as estimated from the CARS climatology (LSW thickness) and Argo/PALACE floats displacements (1,000-dbar velocity).	17
7	Time series of monthly (thin line) and 1-yr low-passed zonal surface geostrophic velocity at the CGFZ (32°W).	18
8	Maps of mean LSW thickness and velocity. (a) Observed climatological mean LSW thickness calculated from CARS2009 gridded temperature and salinity, and velocities derived from the Eulerian averaging of Argo and PALACE 1000 dbar float displacements. (b) Mean LSW thickness and velocities for 2010-2012 from the model output.	21
9	Model time series of LSW at CGFZ. (a) Zonal LSW velocity (green) and zonal LSW transport (red) at the CGFZ (32°W) in the model simulation from 1978 to 2012. The mean LSW velocity and transport for 1978-1994 (4.7 cm s ⁻¹ , 10.2 Sv) and 2003-2012 (4.6 cm s ⁻¹ , 8.7 Sv) are shown as straight lines. (b) Similar to a) except that the blue line is the LSW thickness. . .	24
10	Modeled velocity, salinity and temperature at CGFZ from 1978 to 2012. (a) Time-series of surface (red) and LSW (black) velocities. Thin lines are the monthly time series and thick lines are the low-passed time-series with one year cutoff. The correlation coefficient between the two time-series is given in the figure. (b) Hovmöller diagram of salinity. White lines indicate the upper and lower limits of the LSW layer. (c) Same as middle panel, except for temperature.	25

Figure		Page
11	Schematic of the circulation in the Northwest Atlantic Ocean and the water column average temperature difference between the period 2009–2018 and 2001–2007. The red (blue) shading indicates warming (cooling) of the vertically-averaged ocean temperature from the EN4 objective analysis to 2,000 m, or the seafloor if it is shallower than 2,000 m, in 0.5°C increments (change in the unshaded region is between -0.25°C and 0.25°C). Background in grayscale shows the bathymetry of the region, with darker shades representing shallower areas. The main surface circulation patterns associated with the Gulf Stream (red) and the Labrador Current (blue) systems are identified with arrows. Coastal and shelf areas of interest are indicated. TGB stands for Tail of the Grand Banks and SENR, for Southeastern Newfoundland Ridge.	56
12	Sea surface height shift at the Tail of the Grand Banks in 2008. (a) Mean Dynamic Topography (MDT) in the Northwest Atlantic between 1993 and 2018. The thick red contour is the 1,000-m isobath between 76°W and 48°W, along which the along-slope surface velocities displayed in Fig. 13a are calculated. The black dotted lines indicate the main cross-slope channels in the region: Great South Channel (GSCh), Northeast Channel (NECh) and Laurentian Channel (LCh). The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. (b) Sea surface height difference between 2009–2018 and 1993–2007. To emphasize spatial patterns, the sea surface height increase averaged over the entire region that is plotted (equal to 4.5 cm) has been subtracted. (c) Time-series of the monthly sea surface height (dark gray line) and seasonal ARIMA model (light gray line, estimated as described in the Methods section) at the Tail of the Grand Banks (i.e. averaged within the thick black contour in panel a). The blue and red horizontal lines indicate the averaged sea surface height of -10.5 cm and 0.3 cm before and after the shift in July of 2008. Colormaps in panels a and b are from the Cmocean package [53].	57

Figure		Page
13	<p>Along-slope velocity and temperature change west of the TGB. (a) The correlation coefficient between the deseasonalized along-slope surface velocity at the Tail of the Grand Banks of Newfoundland (along the 1,000-m isobath, at 42°51'N, 50°40'W) and the deseasonalized along-slope surface velocity downstream towards the Northwest Atlantic Shelf as a function of time lags and distance from the TGB. White contours represent the 95% significance level. The dashed black line represents a propagation speed of 130 km month⁻¹ (5 cm s⁻¹). (b) Map of change-point timing of the 149-m temperature in the Northwest Atlantic. The shades only show grid points whose single change point during 1993–2018 occurred between January/2009 and December/2011. The temperature changes, in °C, associated with these shifts are displayed in Fig. 17. Colors are displayed in 3-month intervals. The various hatch marks indicate the grid points used to calculate the regional lagged correlations shown in Table 1: Laurentian Channel (white circles), Scotia Shelf (black plus signs), Gulf of Maine (white triangles) and Mid-Atlantic Bight (black crosses). The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. The main cross-shelf features are identified. TGB = Tail of the Grand Banks; LCh = Laurentian Channel; NECh = Northeast Channel; GSCh = Great South Channel; CH = Cape Hatteras. Colormaps are from the Crameri package [54].</p>	58

- 14 Temperature-Salinity diagram at the TGB and time series of the 149-m temperature on the Scotian Shelf and Laurentian Channel. (a) Mean decadal profiles of temperature and salinity at the TGB between the 1930s and the 2010s are shown as a T-S diagram. The profiles are color-coded by decade of sampling. The blue and red solid lines indicate mean T-S profiles of the Labrador Current and the Gulf Stream, averaged over the blue and red boxes shown in the inset map, respectively. Thin, back lines are the potential density anomalies referenced to the sea surface, with the 27.68 kg m^{-3} isopycnal (i.e. potential density of $1027.68 \text{ kg m}^{-3}$) and 27.80 kg m^{-3} isopycnal indicating the upper and lower boundaries of the Labrador Sea Water. A robust locally estimated scatterplot smoothing is applied to the nine decadal profiles at the TGB to reduce the effect of outliers at poorly-sampled depths. The strategy used to create this diagram while minimizing the influence of variability in the location where the profiles were collected is described in the methods. (b) Monthly time series of the 149-m temperature averaged over the Scotian Shelf and Laurentian Channel (grid points marked with black dots in the inset map). Gray shades indicate the error estimate averaged over the respective grid points, calculated as shown in the Methods section. 59
- 15 Sample probability distribution and autocorrelation of the sea surface height time series at the TGB shown in Fig. 12c. (a) Sample probability distribution histogram of the pre-shift months (January 1993 – July 2008; blue bars) and post-shift months (August 2008 – December 2018; red bars). The transition in 2008 represented a 10.5 cm increase in the mean SSH and explains 24.1% of the variability in the time series. (b) Autocorrelation function, showing the highest values at 1-, 2-, 12- and 24-month lags, which indicates the strong month-to-month and seasonal signals. The temporal autocorrelation structure explains 45.8% of the variability in the time series. 60

Figure		Page
16	Along-slope velocity between Flemish Pass and Cape Hatteras. Black arrows indicate the 26-year mean surface velocity as interpolated onto the 1,000-m isobath. Red and blue arrows are the along- and across-slope velocity components as calculated from the angle between two consecutive grid points from Flemish Pass. Gray contours are the 100-m and 1,000-m isobaths. The monthly along-slope velocity is used to calculate the lagged correlations shown in Fig. 13a.	61
17	Magnitude of the 149-m temperature shift in the Northwest Atlantic. The temperature shift is calculated at each grid point as the mean temperature after the change-point timing minus the mean temperature before the change-point timing, as shown in Fig. 13b, for 1993–2018. The shades only show grid points whose single change point happened between January 2009 and December 2011. All of the changes on the shelf and slope are positive, i.e., represent a shift towards warmer waters. Colors are displayed in 0.5°C bins. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. The main cross-shelf features are identified. LCh = Laurentian Channel; NECh = Northeast Channel; GSCh = Great South Channel; CH = Cape Hatteras. Colormap from the Cmocean package [53].	62
18	Hydrographic and float profiles at the TGB from the EN4 database. (a-i) Each panel shows the locations of all profiles taken between April and June in each decade since the 1930s with pairs of temperature and salinity. Months are indicated by colors: blue for April, red for May and green for June. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray.	63
19	Post-2008 temperature shifts on the Northwest Atlantic Shelf. Timing (left; a,c,e,g) and magnitude (right; b,d,f,h) of the identified temperature shifts at the 122-m (a,b), 135-m (c,d), 166-m (e,f) and 185-m (g,h) layers from the EN4 objective analysis. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. Based on the results of the 149-m layer, analysis was limited to a polygon defined by the vertices (30°N, 80°W), (36°N, 65°W), (36°N, 52°W), (43°N, 48°W), (48°N, 42°W), (51°N, 47°W), (53°N, 54°W), and (48°N, 80°W). Colormaps in panels a and b are, respectively, from the Crameri [54] and Cmocean [53] packages.	64

Figure		Page
20	Ninety-year (1930-2019) time series of the normalized temperature anomalies at the TGB at 350 m—the mean depth of the 27.25 kg m ⁻³ isopycnal. The red line and shades are the three-year running mean and the standard error envelope, respectively. The gray circles are one-year averages. Plotting temperature at a fixed depth allows us to include the quality-controlled profiles of temperature that do not also have salinity (3,565 profiles that reach a depth of 350 m), and also combines the signal of the deepening thermocline (as in reference 14) as well as water mass changes on density surfaces. The method for making this time series is the same as that described in the methods for the T-S plot, where the bin-average was subtracted from each profile before averaging the anomalies. In this case, the anomaly was also normalized by the standard deviation in the bin, so that the plotted anomalies are the mean number of standard deviations from the mean temperature in the TGB box.	65
21	Hydrographic and float profiles from the EN4 database in the Northwest Atlantic. (a-i) The shades show the number of profiles by decade between the 1930s and the 2010s in 1° x 1° bins. Over 1.3 million profiles were used to build the objective analysis in this region, of which 57% were collected in the shelf regions that are used to construct Figs. 13b, 14b, 17 and Table 1. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. Colormaps are from the SciVisColor database [55].	66
22	(a) Schematic of general circulation in the Northwest Atlantic. The launch site is identified with a triangle. (b) Mean depth (colors) at the top of Layer 19 or the depth of the layer outcrop and Layer 19 mean velocity (arrows). Vectors whose speed is less 5 cm s ⁻¹ are shown in teal; between 5 and 10 cm s ⁻¹ , in dark gray; greater than 10 cm s ⁻¹ in red.	68
23	Mean sea surface height in the Northwest Atlantic Ocean for the period 1993-2017 (a) from altimetry and (b) in the Atlantic HYCOM output. The gray lines in both maps indicate the 1,000-m and the 4,000-m isobaths from ETOPO1.	71

Figure		Page
24	Cross-section of the ten-year mean (a) σ_2 , (b) temperature, (c) salinity and (d) velocity at the launch site LC, in the eastern Grand Banks. The velocity is normal to the section and has been rotated from the original zonal and meridional components, with positive values in the direction of the mean flow of the Labrador Current (i.e., southwestward). The boundaries between the model layers are shown in thin, gray contours, with Layer 19 highlighted with thick, black contours. The section is 68-km long (x-axis), with the continental shelf to its left and the open ocean to its right.	75
25	Probability distribution map of synthetic particles tracked for one year after deployment at launch site LC from 2003 to 2011, calculated at $1^\circ \times 1^\circ$ bins. The black lines represent the start and end points of one-year trajectories of 50 randomly-selected particles. The bathymetry is shown where the probability is equal to zero.	76
26	(a) Difference in the probability distribution of synthetic particles deployed between March and August (Spring/Summer) minus those deployed between September and February (Fall/Winter) and tracked for one year, calculated at $1^\circ \times 1^\circ$ bins. (b) Same as (a), but for particles deployed in 2008 minus particles deployed in 2004. The “Slope” and “NAC” boxes are identified in black. In panels (a) and (b) the grid points whose difference was between -1% and 1% were masked out. (c) One-year trajectories of “Slope” (blue) and “NAC Particles” (red) that were launched in 2004, defined as the particles that were sampled in the regions shown as black polygons in Fig. 26b. (d) Same as (c), but for those particles launched in 2008. The definition of “Slope” and “NAC Particles” was the same for both years. In 2008, only four particles met the criteria of entering the “Slope Region” less than three months after being deployed in the 10-day Lagrangian tracking output.	78

Figure		Page
27	(a) Mean potential thickness from Layer 19 to the bottom of the ocean. The “Slope Region” and the launch site are identified. The 2,300-m and the 3,000-m mean potential thickness contours are shown as thin black lines. (b) Monthly percentage of particles reaching the “Slope Region” within three months of launch (blue bars), plotted as a function of the date of launch between Jan 2003 and Dec 2011, and monthly potential thickness anomaly at the Tail of the Grand Banks (black line, averaged over box in Fig 30,a,b,c). The two time series have the highest correlation coefficient when the TGB potential thickness anomaly lags by one month ($R = 0.70$, $p\text{-value} < 0.001$).	81
28	Potential Thickness anomaly on three selected months with no “Slope Particles”, tracked for three consecutive months. Each row represents one event: (top) Dec 2007, (middle) Feb 2008 and (bottom) Jan 2010. The columns show the progress of each event from (left) the month of deployment to (right) two months after the deployment. In each map, the trajectories of 15 randomly-selected particles are shown. The trajectory of each particle in a given month is shown in dark red and the trajectory in the previous one (middle column) or two months (right column) are shown in salmon. The 1,000-m and 4,000-m isobaths are shown in thin black contours	84
29	Composite averages of (a, b) potential thickness below Layer 19, (c, d) salinity on Layer 19 and (e, f) velocity of Layer 19. The panels on the left show periods of strong Labrador Current connectivity: they are averaged for months in which 20% or more of the particles reached the “Slope Region” within 3 months. The right hand panels are for months in which no particles reached the “Slope Region”. The averages are for made during the second month after deployment.	85
30	Difference between the months with no “Slope Particles” and the months with 20% or more of “Slope Particles”. (a) SSH, (b) depth of Layer 19, (c) potential thickness and (d) thickness of Layer 19. The 200 m isobath is shown in black to show the approximate position of the shelf break.	87

LIST OF TABLES

Table		Page
1	SSH and temperature lagged correlations. Maximum correlation coefficients and time lags between the raw and prewhitened monthly SSH at the Tail of the Grand Banks and the monthly 149-m temperature averaged over each of the regions indicated in Fig 3b, for 1993–2018. Correlation coefficients in bold and italic fonts are significant at the 99.9% and 99% confidence levels, respectively.	48

Labrador Sea Water Transport across the Charlie-Gibbs Fracture Zone**1.1 Introduction**

Labrador Sea Water (LSW) is a major component of the deep limb of the Atlantic Meridional Overturning Circulation (AMOC). It forms in the central Labrador Sea and in the Irminger Sea through open-ocean deep convection and fills a large volume of the Atlantic Ocean [1, 2]. This water mass is also oxygen-rich and a major sink for anthropogenic carbon [3, 4, 5]. Thus, understanding the circulation and variability of LSW is important for a number of climate and biogeochemical questions. Although anomalies of the LSW water properties have been shown to be swiftly advected southward along the western boundary of the North Atlantic [6], the connection between variability in LSW formation and export rates to the subtropics was recently called into question [7, 8]. Such a disconnect could arise if the fraction of the LSW mixed and advected eastward were to vary, perhaps due to interactions with the North Atlantic Current. Ultimately, this eastward LSW transport must contend with the Mid-Atlantic Ridge (MAR), which acts as a barrier to zonal transport. Zonal currents tend to get funneled through deep fracture zones in the ridge, as schematized in Figure 1 [9].

LSW is known to follow several pathways as it leaves the Labrador Sea. Some LSW remains in the subpolar gyre in a weak anticyclonic recirculation that enters the Irminger Sea [10]. Evidence for this pathway has been obtained from hydrographic measurements and the averaging of float trajectories [11, 12, 13]. Conversely, much of the LSW is exported within the Labrador Current [12] and flows southeastward towards Flemish Cap and the Grand Banks of Newfoundland following the 2000 m isobath in the continental slope. The fate of LSW downstream of Flemish Cap and the Tail of the Grand Banks depends upon its interactions

Figure 1. Circulation and bathymetry of the subpolar North Atlantic. (a) Schematic circulation diagram showing primary shallow (red) and deep (yellow and blue) currents in the subpolar North Atlantic. The LSW formation regions are shown as yellow circles, and its main pathways are indicated as yellow arrows. Abbreviations include Bight Fracture Zone (BFZ), Charlie-Gibbs Fracture Zone (CGFZ), Faraday Fracture Zone (FFZ) and Maxwell Fracture Zone (MFZ). (b) Detailed bathymetry at the Mid-Atlantic Ridge for the region that includes the Charlie-Gibbs Fracture Zone. The mooring array is indicated in black on the west side of the CGFZ.

with the northeastward flowing Gulf Stream: LSW may continue southward along the western boundary or be advected into the interior of the basin along with the North Atlantic Current (NAC), as shown by [14]. The eastward transport of LSW carries climatic signals from the Labrador Sea to the deep ocean and contributes to the lower limb of the Meridional Overturning Circulation [15].

The presence of LSW in the Eastern North Atlantic basin has been reported since early studies of mid-depth circulation in the North Atlantic [17, 18, 19, 20, 21]. Moreover, LSW has long been thought to preferably cross the MAR via the two deep valleys of the Charlie Gibbs Fracture Zone (CGFZ), the largest geological fault in the North Atlantic portion of the MAR [22, 23]. The CGFZ is, indeed, a major water mass crossroads, as it funnels both the eastward-flowing LSW and other water masses carried by the NAC at surface and intermediate depths, and the westward-flowing Iceland Scotland Overflow Water (ISOW) in the deepest layers of the water column [24, 16]. In addition, the LSW contributes to the dilution of the

Figure 2. Zonal velocity and salinity at the CGFZ mooring array. (a) Cross section of time-mean velocity from moored current meters. White triangles indicate the location of the current meters. Thick black line is the 0 cm s^{-1} isotach. Mooring labels are indicated on the top of the figure. (b) Similar to panel (a) but for salinity. White circles indicate the location of the MicroCats. Thick black line is the 34.94 isohaline. Dashed white lines are isopycnals. Mean locations of LSW, ISOW and Lower Deep Water (LDW) are indicated. Figure from [16].

ISOW in this region [25]. Early efforts to quantify the LSW geostrophic transport through the MAR used hydrographic data and provided estimates that range from 4 to 13 Sv [26, 27, 23]. Though these studies have revealed much about the mean LSW transport into the eastern North Atlantic basin, its temporal variability is yet to be explored.

Continuous observational efforts to understand the rates and variability of LSW formation and transport have been made in regions such as the central Labrador Sea [28], and along the perimeter of the North American continent:

the western Labrador Sea at 53°N [29, 30], the Tail of the Grand Banks [31] and south of New England [32]. However, we lack continuous measurements of the eastward transport of LSW across the MAR. We aim to fill this gap with a combination of 22 months of direct measurements made by moorings in the CGFZ, inferences from altimetry, and a regional ocean circulation model that has been widely validated for its faithful representation of circulation in this region [33]. Characterizing the variability of the LSW transport through the CGFZ will contribute to the understanding of the water-mass propagation downstream from the formation region.

In the following section, we describe the mooring array (Section 1.2.1), the altimetry data along with other observational data (Section 1.2.2) and the ocean circulation model used in this study (Section 1.2.3). The measurements recorded by the moored instruments are used to quantify the transport of LSW across the mooring array and characterize its variability over 22 months of direct observations (Section 1.3.1). Next, the satellite-derived surface geostrophic velocities (Section 1.3.2) and the model simulations (Section 1.3.3) reveal the variability of the LSW transport at the CGFZ over the past three decades. Finally, we summarize and conclude our work with an outlook for the future (Section 1.4).

1.2 Data and Methods

1.2.1 Moorings in the Western CGFZ

A mooring array was deployed in the CGFZ between 52°N and 53°N along a nominal longitude of 35.33°W to quantify the transport of ISOW [16]. The array, deployed for 677 days between August 2010 and June 2012, consisted of eight moorings spanning from south of the fault ridge to the southern flank of the Reykjanes Ridge. The four “tall” (seafloor to 500 m) and four “short” (seafloor to 1500 m) moorings were deployed as shown on the meridional section in Figure

2, with letters A-H denoting the 8 moorings at the top of panel a. Temperature, conductivity, and pressure were recorded from 36 SBE-37 MicroCATs, while direct measurements of the velocity components were made with 28 current meters (18 Aanderaa RCM-11 and 10 Nortek AquaDopp 6000 DW). The mooring configuration and further details of the deployments can be found in [34].

Figure 3. LSW properties and transport at the mooring array. (a) 22-month Hovmöller diagram of the 30-day low-passed LSW thickness anomaly at the mooring array. The latitude of each mooring (A is the northernmost) is shown as dashed lines. Thick black line is the 0-m thickness anomaly contour. (b) 22-month mean LSW thickness across the mooring array. Dashed lines indicate the latitude of each mooring. (c) LSW zonal velocity anomaly through the mooring array, with same plotting conventions as in panel (a). Thick black line is the 0-cm s⁻¹ zonal velocity anomaly contour. (d) 22-month mean LSW zonal velocity. (e) 22-month time series of the 30-day low-passed LSW thickness (blue), zonal velocity (green) and zonal transport (red) averaged across the eight moorings. The red lines, whose correlation coefficient R is equal to 0.33, are the transport as calculated from the moored instruments (solid) and the model output interpolated to the mooring positions (dashed). The dashed black line denotes LSW zonal velocity and transport equal to zero.

The methodology implemented to estimate the LSW transport across the mooring array follows the one described by [16], who used the same dataset to quantify the volume transport in the ISOW layer. In their work, ISOW was defined as the waters with salinity greater than 34.94 g kg^{-1} . Here, we define LSW as the waters whose salinity is lower than 34.94 g kg^{-1} [26] and σ_θ is between 27.68 and 27.80 kg m^{-3} [35, 36]. As in [16], we filtered all data with a low-pass third-order Butterworth filter with a 40 h cut-off period, run forward and backward in order to eliminate phase shifts.

Temperature and salinity were averaged daily and linearly-interpolated vertically onto 1-m bins between the top and bottom MicroCATs for each mooring before potential density was calculated. When the 27.68 kg m^{-3} isopycnal was shallower than the shallowest instruments in the tall moorings, the top interface of the LSW was assumed to be 500 m. This occurred in 53% of the measurements at the tall moorings, and, as such, the LSW thickness may be slightly underestimated. Hydrographic sections and salinity observed at the mooring array suggest that the halocline consistently extends below 500 m, indicating that the 34.94 g kg^{-1} threshold is likely not far above the top instrument depth on the tall moorings (see [16] Figures 3 and 5). Because the 27.68 kg m^{-3} isopycnal is always shallower than 1500 m, the top interface of the LSW in the short moorings was calculated as the average depth of the 27.68 kg m^{-3} isopycnal in the two neighboring moorings. For mooring H, the southernmost short mooring with only one neighbor to the north, the top interface was determined by mirroring mooring G.

Similarly to temperature and salinity, the velocity components were averaged to daily values and linearly interpolated vertically onto 1-m bins between the top and bottom current meters in each mooring. As the moorings were deployed nearly along a line of constant longitude (35.33°W), the zonal component of the velocity

field is approximately normal to the mooring array, and only this component is used further in the analysis. The daily LSW velocity across each mooring was calculated as the average of the zonal velocity between the top and bottom interfaces of the water mass. To deduce velocities at depths above the short moorings we rely on the deduction from an Empirical Orthogonal Function (EOF) analysis that most of the zonal velocity variability at the mooring array is explained by a vertical mode with very little shear [16]. For this reason, the zonal velocity above 1500 m was constructed by adding the time mean values of neighboring moorings and the fluctuations of the 1500 m current meter at moorings B, D and F. At mooring H, the top 1000 m zonal velocity was calculated as the fluctuations of the 1500 m current meter plus the time mean zonal velocity of mooring G above 1500 m.

The daily time series of LSW transport per unit width ($\text{m}^2 \text{s}^{-1}$) across each mooring was then calculated by multiplying the mean zonal velocity between the top and bottom interfaces of the LSW layer by the LSW thickness (i.e., the distance between the two interfaces). Finally, this transport was multiplied by the distance separating the mid-points between adjacent moorings to arrive at a volume transport in Sverdrups. For moorings A and H, this value was calculated as the distance between the moorings and their closest neighbor. Because the LSW always resided above the shallowest bathymetry of the CGFZ, no transport needed to be estimated between an instrument and adjacent bathymetry. The time series of LSW properties are filtered with a low-pass 3rd order Butterworth filter with a 30-day cut-off to remove the signals from high frequency (weekly or shorter) variability.

It is important to note that the original purpose of the mooring array was to quantify ISOW transport across the CGFZ. Because the CGFZ is a transform fault, the position of the ridge crest is shifted on either side of the fracture zone:

North of the CGFZ, the ridge crest aligns with 35°W , and south of the ridge crest it lies nearly along 30°W (Figure 4). As the moorings were deployed at 35°W , they are only situated in a true ridge gap in the north valley, where they are bounded by the Reykjanes Ridge to the north and the transform ridge to the south. In contrast, the moorings south of the transform ridge are about 3° to the west of the MAR and have no bounding bathymetric feature directly to the south. This placement of the mooring array was designed to investigate the westward transport of ISOW into the western basin [16]. Unfortunately, the placement is less ideal for quantifying the eastward flow of LSW across the CGFZ, given that the eastward currents can meander south of the mooring array and still cross the MAR through the CGFZ. Because of this limitation, and to consider variability in CGFZ transport at time scales longer than 2 years, we turn to satellite altimetry and an ocean model in Sections 1.3.2 and 1.3.3, respectively.

1.2.2 Satellite altimetry and other observational data

The Global Ocean Gridded L4 Sea Surface Height product was obtained from the Copernicus Marine Environment Monitoring Service (CMEMS). The $0.25^{\circ} \times 0.25^{\circ}$ daily gridded dataset is a multimission altimeter product that combines data from multiple satellites, including Topex/Poseidon, Jason-1 and Jason-2. Data released includes the daily surface geostrophic currents with the same $0.25^{\circ} \times 0.25^{\circ}$ resolution, derived from the lateral gradient of the absolute dynamic topography. The gridded absolute dynamic topography and the surface geostrophic velocities from January 1993 to December 2017 were used to determine the mean position of the northern branch of the North Atlantic Current in the region between 50°N and 55°N , 29°W and 38°W , which shed light on the importance of calculating the LSW transport farther east from the mooring array, at 32°W .

The zonal geostrophic velocity was interpolated onto two sections: the first at

the mooring array (35.33°W) and the second along 32°W . [16] showed that the leading vertical mode of the zonal velocity was equivalent-barotropic, explaining 68-74% of the variance. The next leading mode, explaining 21-27% of the remaining variance was the first baroclinic mode, which changed sign beneath the base of the LSW layer. Therefore, the mooring observations suggest that the LSW velocity should co-vary with the surface velocity, and we confirm this relationship by calculating a linear regression between the surface and the LSW velocities at the moorings. Our goal was to build a proxy for LSW transport based on altimetry, thus allowing for the investigation of LSW velocity variability over the 25-year altimetry era at the center of the fault ridge, 3° to the east of the mooring array.

We calculated the ratio of LSW-layer velocity to the surface geostrophic velocity using two observational products in addition to that calculated from the 22-month LSW velocity from the mooring array compared against the co-located surface geostrophic velocities. In the first estimate, the vertical zonal velocity shear between surface and LSW-layer depth (nominally 1000 m) was calculated by applying the thermal wind equation to a climatological density field. We used the Australia’s Commonwealth Scientific and Industrial Research Organisation (CSIRO) Atlas of Regional Seas (CARS2009), a product with a 0.5° horizontal resolution and 79 vertical levels spanning from surface to 5500 m [37]. The CARS2009 data product was created with a modified Loess filter to interpolate the irregularly spaced observations into a regular grid. This scheme uses a bathymetry-influenced weighting, which more accurately preserves gradients in areas with steep topography such as the Mid-Atlantic Ridge; as such, we believe it to be an appropriate climatology for our purposes.

In the second estimate, the 1000 dbar time-mean velocity field was calculated based on float displacements for the Argo era. The Argo dataset is freely available

by the International Argo Program (<http://www.argo.ucsd.edu>). The 1000 dbar float displacements were then compared to surface pseudo-displacements calculated from altimetric geostrophic velocities interpolated to the starting position of each float displacement with the method described by [38].

1.2.3 Numerical Simulation

A $1/12^\circ$ eddy simulation based on the HYbrid Coordinate Ocean Model (HYCOM) was used to explore LSW transport variability in 1978-2012 and the role of variable LSW thickness and velocity in the transport variability. The model configuration has been described in [33], who also provided several detailed model-data comparisons in the subpolar North Atlantic. The model results are shown to realistically simulate the observed time-mean structure of the western boundary current transport off the Labrador Coast, as well as the warming and sea surface height change in the central Labrador Sea since the early 1990s. Additionally, the model exhibits almost no salinity drift, with a slow increase rate of only 0.02 per century. More recently, [39] examined the westward ISOW transport through the CGFZ and showed that the simulation reproduced most of the intraseasonal to interannual variability observed in the moored current meter arrays during 1988-1989 [24] and 2010-2012 [16].

The model outputs include potential temperature, salinity, as well as the zonal and meridional velocities. LSW thickness was determined as the layer whose σ_θ is between 27.68 and 27.80 kg m^{-3} . A third-order Butterworth low-pass filter with a 30-day cut-off, run forward and backward in order to eliminate phase shifts, was applied to emphasize variability at time scales longer than one month.

1.3 Key Results

1.3.1 LSW Transport at the CGFZ Mooring Array

Temperature and salinity observations reveal the key water masses situated along the mooring array (Figure 2). The LSW lies on top of the ISOW in the north valley and on the southern flank of the zonally-oriented transform ridge. Further south, the lower layer is filled with cold, fresh Lower Deep Water (LDW), a water mass that is thought to be a mixture of Antarctic Bottom Water and North Atlantic Deep Waters [19, 24]. The surface layer is dominated by warmer waters that are generally advected and mixed to the east by the North Atlantic Current and its eddies [16].

Figure 4. Mean surface geostrophic velocity (arrows) overlaying the mean dynamic topography (colors) in the CGFZ region. The red line to the left indicates the locations of the eight moorings. The black dashed line is the mean position of the 39.5 cm ADT isoline, which coincides with the maximum surface velocities associated with the northern branch of the NAC. A more ideal position for the quantification of LSW transport across the CGFZ is also shown in red, shifted 3° to the east of the moorings. This line is located at the mean position of the ridge crest at the latitude band of the CGFZ, as calculated from ETOPO1 [40]. It is bounded by the 3,000-m isobath to both the north and south, and extends from 51.7°N to 53°N. The 3,000-m isobath is shown as a white contour.

The mean zonal LSW transport measured across the mooring array was 1.9 ± 2.3 Sv, where the reported uncertainty is one standard deviation. Transport measured at moorings F-H, located south of the transform ridge, contributed the majority of the transport 1.3 ± 2.6 Sv. In the north valley, moorings C-E also recorded eastward mean transport in the LSW layer and contributed 0.7 ± 1.6 Sv. The mean eastward LSW transport is not ubiquitous throughout the array, as transport at the northernmost moorings A and B, located on the southern flank of the Reykjanes Ridge, averages -0.1 ± 0.6 Sv. In addition to this spatial variability, the LSW thickness and velocity also have strong temporal variability at the mooring array.

During the mooring deployment, several periods of elevated eastward flow were observed, the strongest of which reached 7 Sv between the autumn of 2011 and the winter of 2012 (Figure 3e). The three periods of strong eastward transport, all surpassing 4 Sv, were associated with two distinct spatial distributions of the thickness and velocity anomaly fields across the mooring array, respectively shown in Figures 3a and 3c. During the winters of 2010/2011 and 2011/2012, positive thickness and eastward velocity anomalies are measured at all moorings, with peak velocities in the north valley. In contrast, in the summer of 2011, strong positive thickness and eastward velocity anomalies are measured in the south valley, while negative thickness and westward velocity anomalies are observed in the north valley.

In contrast, the transport time series reveals that there were also periods of weak westward LSW transport averaged over the whole array during four events, collectively lasting about one fifth of the deployment period. These events were each driven by slightly different spatial configurations of the velocity field. In Nov-Dec/2010, the longest westward event, strong westward velocity anomalies

are measured by the southern moorings, while eastward anomalies are observed in the north valley. In Apr/2011, Feb-Mar/2012 and May/2012, westward velocity anomalies are measured across most of the array. The anomalous westward velocities are associated with negative thickness anomalies during all of these events (Figure 3a,c).

The transport variability over the 22-month deployment period is almost entirely controlled by the zonal velocity in the LSW layer, which explains 98% of its variability. The 30-day low-passed LSW thickness and zonal velocity averaged over the mooring array are shown in blue and green, respectively, in Figure 3e. The mean thickness is $1,069 \pm 73$ m, and single daily mean values range from 900 m to 1230 m. On average, there is a meridional thickness gradient along the array, with thicker LSW at mooring H (1202 m), thinning northward to mooring A (867 m, Figure 3b). The zonal velocity during the 22 months averages 1.4 ± 1.8 cm s⁻¹, ranging from -3.0 cm s⁻¹ (westward) to 6.7 cm s⁻¹ (eastward).

As noted in Section 1.2.1 above, the mooring array does not capture the maximum eastward velocities that cross the MAR at the CGFZ. A map of the mean surface circulation from satellite altimetry (Figure 4) shows that the average position of the maximum surface eastward velocity is south of the mooring array. Therefore, the red line positioned further east in Figure 4, at 32°W, represents a more ideal location for the quantification of the eastward transport across the CGFZ. We next attempt to quantify the LSW transport variability across the CGFZ at this line, as described in the following sections.

1.3.2 Circulation variability at the CGFZ from altimetry

Sea surface geostrophic velocities calculated from satellite altimetry are significantly correlated with LSW velocities at the mooring array. The ratio between surface and LSW velocities at the mooring array is 0.27 and the correlation coef-

ficient between the two is 0.51. In the north valley, the satellite-observed surface velocities and the LSW velocities are more strongly connected, with correlation coefficient rising to 0.81, averaged over moorings A through E only. The north valley is bounded by steep topography to both the north and south, and thus has similar bathymetric constraints on the NAC as those the current experiences where it crosses the CGFZ at 32°W . A visual comparison between observed surface and LSW velocities shows that strong events of eastward LSW transport are associated with the swift eastward surface velocities (Figures 3c and 5). Likewise, the westward LSW transport anomalies correspond with westward anomalies in surface velocities. The strong relationship between the altimetric surface velocities and LSW-layer velocities, particularly in the bathymetrically-constrained north valley, suggests that the altimeter record can provide an indicator of temporal variability in the LSW layer velocity as it crosses the CGFZ.

Figure 5. Twenty-two month Hovmöller diagram of the 30-day low-passed zonal surface geostrophic velocity anomaly at the location of the mooring array. The solid black contour is the 0-cm s^{-1} velocity anomaly. Dashed lines indicate the latitude of each mooring.

Figure 6 shows the mean total (light lines) and cumulative (dark lines) LSW transport in 0.5° bins between 53°N and 40°N . In red, we show this transport at 35°W , and in black interpolated to the center of the Mid-Atlantic Ridge rift valley. This figure is made from observational data: LSW thickness estimated from the CARS climatology and the LSW-layer velocity from the objectively-mapped

ARGO displacements at 1000dbar. The LSW transport is clearly much higher at the CGFZ than elsewhere along the MAR, peaking at 3.2 Sv at 52°N. The latitudinal band of the CGFZ, between 51.5°N and 53°N, accounts for 66% of the total eastward transport of LSW over the Mid-Atlantic Ridge (7.8 Sv across the CGFZ versus 11.8 Sv total). If we include the Faraday Fracture Zone (extending the southern limit to 50°N), this value goes to 85%. We chose to estimate the LSW transport at the latitude band of 51.7°N to 53°N because we were limited by the location of the mooring array and extent at which the estimated vertical shear was relatively constant. Finally, it is important to note that over 40% of the eastward LSW transport that crosses 35°W is recirculated southward along the western flank of the MAR and does not cross into the eastern basin, as shown by comparing red and black lines in Figure 6.

To understand whether the surface geostrophic velocity may serve as a proxy for the LSW-layer velocity at 32°W, where the swiftest eastward velocity crosses the MAR (see Figure 4), we must first assess whether the vertical shear between the surface and LSW layer is similar to that observed at the mooring array at 35°W. To do so, we compare the vertical shear calculated at the moorings to the shear at 32°W using two different, independent, observation-based techniques, as well as checking the relationship in a model simulation.

First, we compared the subsurface displacements of Argo floats drifting at 1000 dbar, the approximate center of the LSW layer at the CGFZ, to surface pseudo-trajectories constructed by interpolating the altimetric geostrophic velocities to the time and location of the floats, as was done in [38] and [13]. All subsurface float trajectories were gathered in 1° bins and regressed against the surface pseudo-trajectories, yielding a regression coefficient that represents the ratio of the 1000 dbar velocity to the surface velocity. Between the bin encompassing the mooring

location (centered at 35°W, 52°N) and the bin encompassing the location of maximum cross-MAR surface velocities (32°W, 52°N), this ratio varies between 0.22 and 0.31. The number of trajectories in each bin used to calculate this regression range between 65 and 82.

Second, the geostrophic (i.e. thermal wind) shear from the climatological density field was used to calculate the velocity difference between the surface and 1000 m. We found that the thermal wind shear is relatively constant over a zonally-elongated band at the CGFZ, with the ratio of the 1000 m velocities to surface velocities ranging from 0.53 at 35°W to 0.57 at 32°W.

Finally, we also compare the surface velocity to the LSW-layer velocity in the model, as discussed in more detail in the next section. In brief, the model relationship is consistent with the Argo- and mooring-based estimates of the shear, with the ratio between LSW and surface zonal velocities of 0.19. It is notable that the geostrophic shear inferred from the climatological meridional density gradient is at least two times weaker than the float-, mooring- and model based shear estimates. We therefore suspect that the smoothed climatological product underestimates the meridional density gradients over the scale of interest. Yet, we incorporate the full range of observation-based estimates of the ratio between surface and LSW layer velocities (0.22 - 0.57) into a conservative metric of the uncertainty of the calculated LSW velocity. In practice, the total uncertainty of the LSW velocity at 32°W is calculated as $\sigma_{U_{lsw}} = \sqrt{\sigma_{shear}^2 + \sigma_{U_{geo}}^2}$, where σ_{shear} and $\sigma_{U_{geo}}$ are, respectively, the uncertainty of the vertical velocity shear, described here, and the standard deviation of the surface geostrophic velocities.

The mean surface zonal velocity at 32°W, shown in Figure 7, is about twice as fast as at the location of the mooring array ($7.9 \pm 4.7 \text{ cm s}^{-1}$ versus $3.9 \pm 2.7 \text{ cm s}^{-1}$), averaged over the approximately 25-year altimetric record. Surface currents

Figure 6. Eastward transport of LSW through the Mid-Atlantic Ridge. (a) Bathymetry map of the Mid-Atlantic Ridge between 40°N and 53°N, showing the 35°W line (red) and the center of the Mid-Atlantic Ridge rift valley (black). Dots represent the grid points at which the lines in panel b were calculated, with a 0.5° latitudinal resolution. (b) Total (light lines) and cumulative (dark lines) zonal transport of LSW at 35°W (red) and the Mid-Atlantic Ridge rift valley (black), as estimated from the CARS climatology (LSW thickness) and Argo/PALACE floats displacements (1,000-dbar velocity).

across the CGFZ show intense variability at time scales ranging from intraseasonal to decadal, as seen from the monthly (thin line) and 1-yr (thick line) low-passed time series. The maximum cross-CGFZ annual mean velocity exceeds 13 cm s^{-1} in 1995, after which there is a slow decline to a 5-year period of annual mean velocity of less than 5 cm s^{-1} between 2001 and 2005, with a record minimum of 3 cm s^{-1} in 2002. By 2006, annual mean velocities recovered to about 10 cm s^{-1} , and averaged 9 cm s^{-1} until the end of the delayed-time available at the time of writing.

Given that the moorings show that LSW velocity is correlated with the surface geostrophic velocity, as was also implied by the strongly barotropic nature of the flow [16], and having assessed the uncertainty on the vertical velocity shear between 32°W and 35°W, we can infer the variability of the LSW velocity through the

CGFZ. Strong eastward LSW velocity events likely occurred in the early 1990s and after 2006, while weak eastward velocity and possibly westward reversals were likely between 2001 and 2006. Using the shear calculated from the comparison of the surface geostrophic velocity and LSW-layer velocity at the mooring array, as well as the two other observational shear assessments, we estimate that the mean LSW velocity for the altimetric era is $3.4 \pm 1.8 \text{ cm s}^{-1}$. Assuming a constant LSW thickness of 1,069 m, which is the time-averaged LSW thickness calculated from the mooring measurements, we estimate that the LSW volume transport at the CGFZ averages $5.3 \pm 2.9 \text{ Sv}$.

Having observed that thickness variability contributes little to the LSW transport variability over the 22-month mooring array, it is tempting to assume that the surface geostrophic velocities can accurately represent the LSW volume transport, and that the time series in Figure 7 is a good multi-decadal record of LSW transport variability. However, we next show in Section 1.3.3 that this assumption may not hold on longer time scales.

Figure 7. Time series of monthly (thin line) and 1-yr low-passed zonal surface geostrophic velocity at the CGFZ (32°W).

1.3.3 Model-based LSW transport across the CGFZ

The modeled time mean LSW transport at the location of the mooring array between August 2010 and June 2012 is $1.4 \pm 3.4 \text{ Sv}$, compared to $1.9 \pm 2.3 \text{ Sv}$

estimated from the observations. The lower transport in the model is due to the fact the model maximum eastward flow is located slightly to the south of that observed. The variability of the modeled and observed LSW transport through the CGFZ is displayed in Fig. 3e. There is some similarity between the model and observations but the agreement is not as good as in the ISOW as shown in Xu et al. (2018). This may not be surprising, since the mooring array captured only the northern edge of the eastward flow of LSW but the entire westward flow of ISOW.

Figure 8 compares the mean LSW thickness derived from CARS [41] and the 1000 dbar velocity derived from Argo displacements [13] and the modeled LSW thickness and velocity. The LSW is overall thicker in the model than in observations. However, there is similarity in the spatial pattern between model results and observations: the LSW is thick in the Irminger Sea and very thin above the Reykjanes Ridge. South of the Reykjanes Ridge, in the western basin, the LSW flows eastward towards the MAR providing the thick layer that is observed at the CGFZ. Likewise, in both observations and the model, the LSW zonal velocity is notably higher at the latitude of the CGFZ than elsewhere along the MAR, with a maximum that resembles the mean meridional position of the northern branch of the NAC shown in Figure 4.

The modeled LSW transport at 32°W between 1978 and 2012 was 9.0 ± 4.4 Sv, about 70% larger than the 5.3 Sv estimated in observations for 1993-2017. This is due to both a thicker LSW layer and higher velocity in model results. However, comparing only the time period in which observations and the model overlap, the LSW transport is reduced to 8.2 ± 4.1 Sv, as the LSW thickness and velocity in the model are closer to those in the observed estimates. The model LSW thickness averages ($\pm$ one standard deviation) 1246 ± 32 m over 2010-2012 (compared to the mooring-based thickness of 1069 ± 73 m), and the simulated

velocity is $4.0 \pm 2.1 \text{ cm s}^{-1}$ after 1993 (compared to $3.4 \pm 1.8 \text{ cm s}^{-1}$ inferred from the altimetry-based surface velocity and the three observational estimates of vertical shear). The differences between the modeled and observed LSW thickness and velocity are greater when comparing the entire time period in the simulation, as the observational period saw a thinner LSW layer and slower velocity than earlier in the simulation.

As in the mooring observations, the LSW transport variability is mostly explained by changes in LSW velocity (96%) with a minor contribution from LSW thickness. Figure 9 shows the time series of LSW transport through the CGFZ as compared with the LSW velocity (top panel) and the LSW thickness (bottom panel). The monthly mean LSW velocity and thickness vary considerably during the entire period. Velocity fluctuations drive the monthly to interannual LSW transport variability. In the mid 1990s, both velocity and transport declined by nearly two thirds over the course of two years, during an anomalous period of reduced LSW transport through the CGFZ.

The results from the model output indicate that LSW thickness can be important on longer time scales. Prior to 1995, LSW transport across the CGFZ averaged 10.2 Sv. An abrupt decrease in LSW velocity in 1995 decreased the transport to 6.9 Sv between 1996 and 2003. After 2003, the velocity increased again to reach rates equal to the ones observed before 1995, but the transport did not recover to its pre-1995 average. The mean velocity during the last available decade (2003-2012) is 4.6 cm s^{-1} , statistically equal to the 4.7 cm s^{-1} simulated before 1995 (T-Test, 95% significance level). The transport, however, averaged 8.7 Sv after 2003, which is statistically lower than the mean 10.2 Sv before 1995, (T-Test, 95% significance level). This is explained by a continuous decrease in the LSW thickness between the mid 1990s and the mid 2000s that greatly impacts

its total transport. The LSW was consistently thicker before 1999, with annual averages above 1,500 m between 1991 and 1999, and dropped to less than 1,300 m by 2005. Therefore, shifts in LSW thickness have to be taken into consideration when inferring the cross-CGFZ LSW transport variability on decadal and longer time scales. Although the short term variability in the LSW transport across the CGFZ is modulated very little by the thickness, a long term thinning of the water mass reduces the volume of LSW crossing the CGFZ from 2003 to 2012.

The LSW velocity and thickness at the CGFZ vary on different time scales. Figure 10a reveals that the LSW velocity is highly correlated with the surface velocity ($R = 0.87$), similar to the relationship inferred from the mooring array and satellite altimetry. The LSW thickness is, by definition, controlled by the temperature and salinity stratification. These characteristics changed within the LSW layer after the mid 1990s. The fresh core of the LSW has become saltier, as the 34.92 isohaline vanished in 2007 after continuously thinning since 1999 (Figure 10b). The isotherms deepened and resulted in a warmer LSW, particularly after 2006 (Figure 10c).

Figure 8. Maps of mean LSW thickness and velocity. (a) Observed climatological mean LSW thickness calculated from CARS2009 gridded temperature and salinity, and velocities derived from the Eulerian averaging of Argo and PALACE 1000 dbar float displacements. (b) Mean LSW thickness and velocities for 2010-2012 from the model output.

1.4 Discussion and Conclusions

This study quantified the LSW transport across the CGFZ and explored its substantial temporal variability. Data from a mooring array at the western edge of the CGFZ was used to investigate the transport of LSW and its relationship with surface geostrophic currents in the region. However, the mean position of the northern branch of the NAC, the main driver of the LSW transport across the CGFZ, is located to the south of the mooring array. The NAC turns slightly northward east of the mooring array before crossing the MAR to the eastern basin. The mean surface velocity at the mooring array location is about half as strong as that at the center of the NAC. Therefore, we use altimetry to estimate the LSW transport where the swiftest cross-MAR eastward velocities are found. This approach depends on the assumption that the velocities in the LSW layer are strongly correlated to the surface layer, and that the vertical shear between the surface and LSW velocities can be deduced from existing data. The mooring array and numerical model simulation suggests that altimetric surface geostrophic velocities are well correlated with velocities on the LSW layer. Moreover, there is strong agreement in the vertical shear estimated from the mooring data and from the regression with hundreds of Argo float displacement velocities at 1000 m compared to the altimetric geostrophic surface velocities. The shear simulated in the model is also in close agreement with these two estimates. We therefore argue that surface geostrophic velocities can be used to create a proxy for LSW velocity.

The swift surface geostrophic velocities at the CGFZ at 32°W coincides with a thick layer of LSW (Figures 4 and 8), suggesting that this is the location of maximum eastward LSW transport across the MAR. At this ridge gap, the surface geostrophic velocity for the altimetric era varies strongly on intraseasonal to decadal time scales. The meridional excursion of the NAC at the MAR sets the

boundary between the North Atlantic subpolar and subtropical gyres and has been extensively investigated in a number of observational [42, 43, 44] and modeling [45] analyses. The LSW transport at CGFZ is locally controlled by changes in the NAC velocities on intraseasonal to interannual timescales. The meridional excursion of the NAC – and therefore its position relative to the CGFZ – is in turn associated with fluctuations in the zonal wind stress in the western European Basin east of the CGFZ [16, 39]. Assuming constant LSW thickness and using the calculated vertical shear of the zonal velocity, we estimated the altimetry-derived LSW transport through the CGFZ to be 5.6 ± 3.2 Sv.

Although fluctuations in the velocity field explain much of the LSW transport variability through the CGFZ on interannual and shorter time scales, the spatial and temporal variability of the LSW thickness must be accounted for when addressing the total LSW transport across the MAR. The model simulation indicates that the LSW thickness may play an important role in modulating LSW transport on decadal time scales. The simulation shows continuous thinning of the LSW layer between 1996 and 2009 at the CGFZ, during a time when velocities were recovering following a sharp slowdown in 1995. Thus, the LSW transport does not increase over the 13-year period as the velocity alone would imply (Figure 9).

LSW thickness at the CGFZ appears to be controlled by remote variations in LSW formation. Direct observations along the AR7W hydrographic section across the central Labrador Sea have shown that deep convection reached a depth of 2,400 m between 1987 and 1994, forming a thick layer of dense LSW [46, 47]. Following this maximum, the formation of denser LSW was suppressed for over a decade, [48, 47]. In the Labrador Sea, the shallower convection after 1994 was responsible for at least 1,000 m of thinning of the newly formed LSW after 1994 [43, 49, 50]. Subsequent thinning of the LSW was observed in mooring arrays and

Figure 9. Model time series of LSW at CGFZ. (a) Zonal LSW velocity (green) and zonal LSW transport (red) at the CGFZ (32°W) in the model simulation from 1978 to 2012. The mean LSW velocity and transport for 1978-1994 (4.7 cm s⁻¹, 10.2 Sv) and 2003-2012 (4.6 cm s⁻¹, 8.7 Sv) are shown as straight lines. (b) Similar to a) except that the blue line is the LSW thickness.

hydrographic sections downstream of the Labrador Sea along the DWBC between 56°N and 39°N [35, 6]. At CGFZ, our model accurately simulates the trend on the LSW layer thickness observed upstream.

The model used here also faithfully simulates the observed warming of the LSW layer along the AR7W section associated with the reduction in deep convection that started in 1994 (see Figure 4 in [33]). The transit time from the Labrador Sea to the CGFZ has been evaluated by tracing newly-formed LSW with temperature, salinity, and CFC concentration anomalies [21]. They found that newly-formed LSW reaches the Iceland Basin within 2 to 3.5 years. Consistent with that approximate timing, the simulated LSW thickness at the CGFZ peaks in 1996 (Figure 9), two years after its peak thickness in the Labrador Sea [28, 33]. Reduced convection and associated warming in the Labrador Sea is followed by

a deepening of the isotherms and salinification of the LSW core at the CGFZ beginning in the mid 1990s (Figure 10).

Figure 10. Modeled velocity, salinity and temperature at CGFZ from 1978 to 2012. (a) Time-series of surface (red) and LSW (black) velocities. Thin lines are the monthly time series and thick lines are the low-passed time-series with one year cutoff. The correlation coefficient between the two time-series is given in the figure. (b) Hovmöller diagram of salinity. White lines indicate the upper and lower limits of the LSW layer. (c) Same as middle panel, except for temperature.

The suite of observed and modeled evidence presented here suggests that the LSW transport at CGFZ varies in response to both fluctuations of the NAC velocity, which modulates the interannual to higher frequency variability, and remote changes in LSW formation in the central Labrador Sea, which can give rise to decadal variability of the transport. Exposing the substantial variability in the

eastward transport of LSW across the mid-Atlantic Ridge also points towards additional open questions about both the causes and consequences of such variability. For instance, it is interesting to consider whether large-scale wind patterns and convection in the LSW formation region co-vary, potentially causing changes in the volume and properties of exported LSW as well as its export pathways. Moreover, it remains unknown how variability in the LSW transport across the CGFZ may influence heat transport and stratification in the eastern basin, where ocean to atmosphere heat fluxes strongly influence regional climate. This work suggests that future efforts to quantify the eastward LSW transport across the CGFZ can take advantage of the relationship between surface and LSW-layer velocities, as long as a temporally-evolving estimate of LSW thickness can be resolved from hydrographic and Argo data.

List of References

- [1] R. S. Pickart, F. Straneo, and G. Moore, “Is Labrador Sea Water formed in the Irminger basin?” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 50, no. 1, pp. 23–52, Jan. 2003. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063702001346>
- [2] G. Gebbie and P. Huybers, “How is the ocean filled?: HOW IS THE OCEAN FILLED?” *Geophysical Research Letters*, vol. 38, no. 6, pp. n/a–n/a, Mar. 2011. [Online]. Available: <http://doi.wiley.com/10.1029/2011GL046769>
- [3] C. L. Sabine, “The Oceanic Sink for Anthropogenic CO₂,” *Science*, vol. 305, no. 5682, pp. 367–371, July 2004. [Online]. Available: <https://www.sciencemag.org/lookup/doi/10.1126/science.1097403>
- [4] S. Khatiwala, F. Primeau, and T. Hall, “Reconstruction of the history of anthropogenic CO₂ concentrations in the ocean,” *Nature*, vol. 462, no. 7271, pp. 346–349, Nov. 2009. [Online]. Available: <http://www.nature.com/articles/nature08526>
- [5] S. Khatiwala, T. Tanhua, S. Mikaloff Fletcher, M. Gerber, S. C. Doney, H. D. Graven, N. Gruber, G. A. McKinley, A. Murata, A. F. Ríos, and C. L. Sabine, “Global ocean storage of anthropogenic carbon,” *Biogeosciences*, vol. 10, no. 4, pp. 2169–2191, Apr. 2013. [Online]. Available: <https://www.biogeosciences.net/10/2169/2013/>

- [6] I. A. Le Bras, I. Yashayaev, and J. M. Toole, “Tracking Labrador Sea Water property signals along the Deep Western Boundary Current,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 7, pp. 5348–5366, July 2017. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017JC012921>
- [7] S. Zou and M. S. Lozier, “Breaking the Linkage Between Labrador Sea Water Production and Its Advective Export to the Subtropical Gyre,” *Journal of Physical Oceanography*, vol. 46, no. 7, pp. 2169–2182, July 2016. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO-D-15-0210.1>
- [8] S. Zou, M. S. Lozier, and M. Buckley, “How Is Meridional Coherence Maintained in the Lower Limb of the Atlantic Meridional Overturning Circulation?” *Geophysical Research Letters*, vol. 46, no. 1, pp. 244–252, Jan. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL080958>
- [9] A. S. Bower, B. Le Cann, T. Rossby, W. Zenk, J. Gould, K. Speer, P. L. Richardson, M. D. Prater, and H.-M. Zhang, “Directly measured mid-depth circulation in the northeastern North Atlantic Ocean,” *Nature*, vol. 419, no. 6907, pp. 603–607, Oct. 2002. [Online]. Available: <http://www.nature.com/articles/nature01078>
- [10] K. L. Lavender, R. E. Davis, and W. B. Owens, “Mid-depth recirculation observed in the interior Labrador and Irminger seas by direct velocity measurements,” *Nature*, vol. 407, no. 6800, pp. 66–69, Sept. 2000. [Online]. Available: <http://www.nature.com/articles/35024048>
- [11] K. L. Lavender, W. Brechner Owens, and R. E. Davis, “The mid-depth circulation of the subpolar North Atlantic Ocean as measured by subsurface floats,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 52, no. 5, pp. 767–785, May 2005. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063705000154>
- [12] J. B. Palter, M. S. Lozier, and K. L. Lavender, “How Does Labrador Sea Water Enter the Deep Western Boundary Current?” *Journal of Physical Oceanography*, vol. 38, no. 5, pp. 968–983, May 2008. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/2007JPO3807.1>
- [13] J. B. Palter, C. Caron, K. L. Law, J. K. Willis, D. S. Trossman, I. M. Yashayaev, and D. Gilbert, “Variability of the directly observed, middepth subpolar North Atlantic circulation,” *Geophysical Research Letters*, vol. 43, no. 6, pp. 2700–2708, Mar. 2016. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015GL067235>
- [14] A. Bower, S. Lozier, and S. Gary, “Export of Labrador Sea Water from the subpolar North Atlantic: A Lagrangian perspective,” *Deep*

- Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17-18, pp. 1798–1818, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000233>
- [15] D. Desbruyères, V. Thierry, and H. Mercier, “Simulated decadal variability of the meridional overturning circulation across the A25-Ovide section: DECADEAL VARIABILITY OF MERIDIONAL OVERTURNING CIRCULATION,” *Journal of Geophysical Research: Oceans*, vol. 118, no. 1, pp. 462–475, Jan. 2013. [Online]. Available: <http://doi.wiley.com/10.1029/2012JC008342>
 - [16] A. Bower and H. Furey, “Iceland-Scotland Overflow Water transport variability through the Charlie-Gibbs Fracture Zone and the impact of the North Atlantic Current: ISOW TRANSPORT THROUGH THE CGFZ,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 9, pp. 6989–7012, Sept. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2017JC012698>
 - [17] R. D. Pingree, “A COMPONENT OF LABRADOR SEA WATER IN THE BAY OF BISCAY: LABRADOR WATER IN BAY OF BISCAY,” *Limnology and Oceanography*, vol. 18, no. 5, pp. 711–718, Sept. 1973. [Online]. Available: <http://doi.wiley.com/10.4319/lo.1973.18.5.0711>
 - [18] L. D. Talley and M. S. McCartney, “Distribution and Circulation of Labrador Sea Water,” *Journal of Physical Oceanography*, vol. 12, pp. 1189–1205, 1982.
 - [19] M. McCartney, “Recirculating components to the deep boundary current of the northern North Atlantic,” *Progress in Oceanography*, vol. 29, no. 4, pp. 283–383, Jan. 1992. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/007966119290006L>
 - [20] S. A. Cunningham and T. W. Haine, “Labrador Sea Water in the Eastern North Atlantic. Part I: A Synoptic Circulation Inferred from a Minimum in Potential Vorticity,” *Journal of Physical Oceanography*, vol. 25, pp. 649–665, 1995.
 - [21] A. Sy, M. Rhein, J. R. N. Lazier, K. P. Koltermann, J. Meincke, A. Putzka, and M. Bersch, “Surprisingly rapid spreading of newly formed intermediate waters across the North Atlantic Ocean,” *Nature*, vol. 386, no. 6626, pp. 675–679, Apr. 1997. [Online]. Available: <http://www.nature.com/articles/386675a0>
 - [22] W. R. Wright and L. Worthington, “North Atlantic Ocean Atlas of Potential Temperature and Salinity in the Deep Water,” *Woods Hole Oceanographic Institution Atlas Series 2*, vol. 2, p. 24.
 - [23] J. Paillet, M. Arhan, and M. S. McCartney, “Spreading of Labrador Sea Water in the eastern North Atlantic,” *Journal of Geophysical Research:*

- Oceans*, vol. 103, no. C5, pp. 10 223–10 239, May 1998. [Online]. Available: <http://doi.wiley.com/10.1029/98JC00262>
- [24] P. M. Saunders, “The flux of overflow water through the Charlie-Gibbs Fracture Zone,” *Journal of Geophysical Research*, vol. 99, no. C6, p. 12343, 1994. [Online]. Available: <http://doi.wiley.com/10.1029/94JC00527>
 - [25] V. Racapé, V. Thierry, H. Mercier, and C. Cabanes, “ISOW Spreading and Mixing as Revealed by Deep-Argo Floats Launched in the Charlie-Gibbs Fracture Zone,” *Journal of Geophysical Research: Oceans*, vol. 124, no. 10, pp. 6787–6808, Oct. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019JC015040>
 - [26] L. Worthington, “On the North Atlantic Circulation,” *The Johns Hopkins Oceanographic Studies*, vol. 6, p. 110, 1976.
 - [27] W. J. Schmitz and M. S. McCartney, “On the North Atlantic Circulation,” *Reviews of Geophysics*, vol. 31, no. 1, pp. 29–49, Feb. 1993. [Online]. Available: <http://doi.wiley.com/10.1029/92RG02583>
 - [28] I. Yashayaev and J. W. Loder, “Recurrent replenishment of Labrador Sea Water and associated decadal-scale variability: 2015 Convection in Labrador Sea,” *Journal of Geophysical Research: Oceans*, vol. 121, no. 11, pp. 8095–8114, Nov. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012046>
 - [29] J. Fischer, M. Visbeck, R. Zantopp, and N. Nunes, “Interannual to decadal variability of outflow from the Labrador Sea: LABRADOR SEA OUTFLOW VARIABILITY,” *Geophysical Research Letters*, vol. 37, no. 24, pp. n/a–n/a, Dec. 2010. [Online]. Available: <http://doi.wiley.com/10.1029/2010GL045321>
 - [30] R. Zantopp, J. Fischer, M. Visbeck, and J. Karstensen, “From interannual to decadal: 17 years of boundary current transports at the exit of the Labrador Sea: INTERANNUAL LABRADOR SEA TRANSPORTS,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 3, pp. 1724–1748, Mar. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012271>
 - [31] F. A. Schott, J. Fischer, M. Dengler, and R. Zantopp, “Variability of the Deep Western Boundary Current east of the Grand Banks,” *Geophysical Research Letters*, vol. 33, no. 21, p. L21S07, Oct. 2006. [Online]. Available: <http://doi.wiley.com/10.1029/2006GL026563>
 - [32] J. M. Toole, M. Andres, I. A. Le Bras, T. M. Joyce, and M. S. McCartney, “Moored observations of the Deep Western Boundary Current in the NWAtlantic: 2004-2014: LINE W 2004-2014,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 9, pp. 7488–7505, Sept. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2017JC012984>

- [33] X. Xu, H. E. Hurlburt, W. J. Schmitz, R. Zantopp, J. Fischer, and P. J. Hogan, “On the currents and transports connected with the atlantic meridional overturning circulation in the subpolar North Atlantic: TRANSPORTS IN THE SPNA,” *Journal of Geophysical Research: Oceans*, vol. 118, no. 1, pp. 502–516, Jan. 2013. [Online]. Available: <http://doi.wiley.com/10.1002/jgrc.20065>
- [34] H. H. Furey, L. Trafford, and A. S. Bower, *A crossroads of the Atlantic Meridional Overturning Circulation : the Charlie-Gibbs Fracture Zone data report August 2010 June 2012*. Woods Hole, MA: Woods Hole Oceanographic Institution, 2014. [Online]. Available: <https://hdl.handle.net/1912/6846>
- [35] L. Stramma, D. Kieke, M. Rhein, F. Schott, I. Yashayaev, and K. P. Koltermann, “Deep water changes at the western boundary of the subpolar North Atlantic during 1996 to 2001,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 51, no. 8, pp. 1033–1056, Aug. 2004. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063704000640>
- [36] F. A. Schott, R. Zantopp, L. Stramma, M. Dengler, J. R. Fischer, and M. Wibaux, “Circulation and Deep-Water Export at the Western Exit of the Subpolar North Atlantic,” *JOURNAL OF PHYSICAL OCEANOGRAPHY*, vol. 34, p. 27, 2004.
- [37] J. Dunn and K. Ridgway, “Mapping ocean properties in regions of complex topography,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 49, no. 3, pp. 591–604, Mar. 2002. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063701000693>
- [38] J. K. Willis and L.-L. Fu, “Combining altimeter and subsurface float data to estimate the time-averaged circulation in the upper ocean: VELOCITY FROM ALTIMETER AND FLOAT DATA,” *Journal of Geophysical Research: Oceans*, vol. 113, no. C12, Dec. 2008. [Online]. Available: <http://doi.wiley.com/10.1029/2007JC004690>
- [39] X. Xu, A. Bower, H. Furey, and E. P. Chassignet, “Variability of the Iceland-Scotland Overflow Water Transport Through the Charlie-Gibbs Fracture Zone: Results From an Eddying Simulation and Observations,” *Journal of Geophysical Research: Oceans*, vol. 123, no. 8, pp. 5808–5823, Aug. 2018. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018JC013895>
- [40] B. E. Eakins and C. Amante, “ETOPO1 1 arc-minute global relief model: procedures, data sources and analysis,” National Geophysical Data Center, Boulder, Colorado, Tech. Rep. NESDIS NGDC-24, 2009.
- [41] K. R. Ridgway, J. R. Dunn, and J. L. Wilkin, “Ocean Interpolation by Four-Dimensional Weighted Least Squares—Application to the Waters around

Australasia,” *JOURNAL OF ATMOSPHERIC AND OCEANIC TECHNOLOGY*, vol. 19, p. 19, 2002.

- [42] A. S. Bower and W.-J. von Appen, “Interannual Variability in the Pathways of the North Atlantic Current over the Mid-Atlantic Ridge and the Impact of Topography,” *Journal of Physical Oceanography*, vol. 38, no. 1, pp. 104–120, Jan. 2008. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/2007JPO3686.1>
- [43] M. Rhein, D. Kieke, S. Hüttl-Kabus, A. Roessler, C. Mertens, R. Meissner, B. Klein, C. W. Böning, and I. Yashayaev, “Deep water formation, the subpolar gyre, and the meridional overturning circulation in the subpolar North Atlantic,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17-18, pp. 1819–1832, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000440>
- [44] A. Roessler, M. Rhein, D. Kieke, and C. Mertens, “Long-term observations of N orth A tantic C urrent transport at the gateway between western and eastern Atlantic,” *Journal of Geophysical Research: Oceans*, vol. 120, no. 6, pp. 4003–4027, June 2015. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014JC010662>
- [45] T. Breckenfelder, M. Rhein, A. Roessler, C. W. Böning, A. Biastoch, E. Behrens, and C. Mertens, “Flow paths and variability of the North Atlantic Current: A comparison of observations and a high-resolution model: NAC IN OBSERVATION AND MODEL,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 4, pp. 2686–2708, Apr. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012444>
- [46] R. G. Curry, M. S. McCartney, and T. M. Joyce, “Oceanic transport of subpolar climate signals to mid-depth subtropical waters,” *Nature*, vol. 391, no. 6667, pp. 575–577, Feb. 1998. [Online]. Available: <http://www.nature.com/articles/35356>
- [47] I. Yashayaev, M. Bersch, and H. M. van Aken, “Spreading of the Labrador Sea Water to the Irminger and Iceland basins,” *Geophysical Research Letters*, vol. 34, no. 10, p. L10602, May 2007. [Online]. Available: <http://doi.wiley.com/10.1029/2006GL028999>
- [48] K. Azetsu-Scott, “Time series study of CFC concentrations in the Labrador Sea during deep and shallow convection regimes (1991–2000),” *Journal of Geophysical Research*, vol. 108, no. C11, p. 3354, 2003. [Online]. Available: <http://doi.wiley.com/10.1029/2002JC001317>

- [49] D. Kieke and I. Yashayaev, “Studies of Labrador Sea Water formation and variability in the subpolar North Atlantic in the light of international partnership and collaboration,” *Progress in Oceanography*, vol. 132, pp. 220–232, Mar. 2015. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0079661114002201>
- [50] M. Rhein, R. Steinfeldt, D. Kieke, I. Stendardo, and I. Yashayaev, “Ventilation variability of Labrador Sea Water and its impact on oxygen and anthropogenic carbon: a review,” *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, vol. 375, no. 2102, p. 20160321, Sept. 2017. [Online]. Available: <https://royalsocietypublishing.org/doi/10.1098/rsta.2016.0321>

Changes in the Gulf Stream Precede Rapid Warming of the Northwest Atlantic Shelf

2.1 Introduction

The Northwest Atlantic Shelf is among the fastest warming regions in the global ocean [1] (Fig. 11). This region is home to ecologically- and economically-valuable marine species, including the American lobster and the Atlantic sea scallop—two of the most valuable single-species fisheries in the United States [2]. Ocean warming is expected to increasingly alter historically-exploited stocks in this region, demanding adaptation of fisheries risk assessments to maintain resilience in a changing climate [3, 4].

The Northwest Atlantic is also the locale where the swift, deep-reaching Gulf Stream and Labrador Current are in close proximity, as they transport warm, salty subtropical water poleward and cold, fresh Labrador Sea Water equatorward, respectively (see schematized currents in Fig. 11). In this region, some of the longest continuous records of ocean temperature and salinity have been collected, several dating to before the turn of the 20th century. More than 50 years ago, these records had already revealed that the sea surface temperature could fluctuate by up to 2°C on a decadal time scale over a region extending at least from the coast of New Jersey, US to Halifax, Canada [5]. Subsequent research linked these temperature fluctuations to ripples through the food web and fisheries [1, 6, 7, 8, 9, 10], and showed that their driver may be a modulation in the westward transport of the Labrador Current south of the shallow underwater plateau known as the Grand Banks of Newfoundland [11, 12, 13] (see Fig. 11 for currents and landmarks).

More recently, a high-resolution model simulation suggested that the decreasing proportion of Labrador Current water on the Northwest Atlantic Shelf coincides

with a deepening thermocline at the crossroads of the subtropical and subpolar circulation at the Tail of the Grand Banks (TGB) [14]. Despite the long history of observations and these more recent model results, a description of the mechanisms involved in the rapid warming of the Northwest Atlantic Continental Shelf has been lacking, hindering our ability to predict these changes in advance. Here we connect abrupt migrations in the Gulf Stream position to the warming of the Northwest Atlantic Shelf and provide an observation-based assessment of the predictability of shelf property changes. Such predictability may ultimately improve forecasts of ecosystem changes in this region.

2.2 Sea Level Shift at the Tail of the Grand Banks

The strength and position of the Gulf Stream and Labrador Current can be tracked via satellite observations of sea surface height (SSH). The Gulf Stream Extension, with the 25 cm SSH contour at its center [15, 16, 17], approaches the TGB as a freely meandering jet at 50°W and, on average, aligns with the 4,000-m isobath south of the TGB (Fig. 12a). The Gulf Stream can impinge on the ridge, thus increasing the SSH inshore of the 4,000-m isobath, or meander away from it. Therefore, to evaluate the presence of the Gulf Stream at the TGB, we quantify the SSH variability in the area shoreward of the 4,000-m isobath (Fig. 12c). In the summer of 2008, a shift towards higher SSH suggests that the Gulf Stream migrated to a position closer to the TGB. This shift has persisted for more than a decade, with an increase in the mean SSH of 10.5 cm for 2009–2018 as compared to 1993–2007, detected beyond the 95% confidence level via change-point analysis (Fig. 15; see Methods). Accordingly, the Grand Banks has experienced anomalously high sea levels since 2009.

We searched for and quantified the SSH change point specifically at the TGB because the Labrador Current and Gulf Stream are known to interact at this

bathymetric feature [18]. The average SSH difference after the 2008 shift is a dipole-like structure, with large positive anomalies along and just shoreward of the 4,000-m isobath at the TGB, and negative anomalies in deeper waters (Fig. 12b). Although the dipole-like structure is strongest just offshore of the TGB, it extends as far west as the New England Seamounts along the all-time mean center of the Gulf Stream. This pattern coincides with an increase in the frequency of Gulf Stream deep cyclones after 2008 [17], and is consistent with evidence that the Gulf Stream’s path and speed have significantly changed to the east of the New England Seamounts during the altimetry era [19, 20]. In contrast, the SSH and water temperature north of the Grand Banks along the Labrador Current show no difference between the period before and after the shift. In other words, neither water column temperature (Fig. 11) nor SSH (Fig. 12b) indicate that changes at the TGB are driven by changes in the Labrador Current from the Labrador Sea to Flemish Cap.

2.3 Temperature anomalies on the Northwest Atlantic Shelf

The presence of the Gulf Stream at the TGB has consequences for the continuity of the Labrador Current west of the Grand Banks. The Labrador Current originates as a western boundary current at the edge of the Labrador shelf and flows southward along the Newfoundland shelf break and through the Flemish Pass before it reaches the TGB. Here, the current bifurcates and some of its volume is transported northeastward inshore of the North Atlantic Current [21], at an estimated rate as large as 2.6 Sv [22, 23]. The remainder continues to follow the continental slope west of the Grand Banks and can still be seen at the Northeast Channel of the Gulf of Maine [11, 24], as schematized in Fig. 11. At the ocean’s surface, the along-slope current speed can be estimated from the cross-slope gradient of the SSH field (Fig. 16), assuming geostrophic balance. Here, we use satellite

altimetry to estimate the variability of the Labrador Current along the 1,000 m isobath between Flemish Pass and Cape Hatteras. Anomalies in the along-slope velocity are significantly correlated with the velocity anomalies at the TGB over a distance of more than 2,000 km and at lag times consistent with the advective speed of the Labrador Current (Fig. 13a). The signal continuity persists to the southwest beyond the Laurentian Channel and the Northeast Channel of the Gulf of Maine, only breaking down at the Great South Channel. In contrast, Labrador Current anomalies north of the TGB are uncorrelated with velocity anomalies at the TGB and further to the southwest, further suggesting that the circulation variability along the Northwest Atlantic Shelf break originates at the TGB and not in the Labrador Sea.

The significant lagged correlations in Fig. 13a provide a means of quantifying the downstream propagation speed of anomalies originating at the TGB. An along-slope velocity anomaly takes, on average, nearly one year to reach the Great South Channel, which means that it propagates at about $130 \text{ km month}^{-1}$ (or 5 cm s^{-1}). Given that (1) higher SSH at the TGB is associated with lower along-slope velocities west of the Grand Banks of Newfoundland, and that (2) the SSH at the TGB has been at a higher state since July of 2008, we expect a decrease in the supply of relatively cold/fresh Labrador waters to the shelf and slope following this shift.

Indeed, temperatures on the Northwest Atlantic Shelf apparently responded to the abrupt sea level increase at the TGB in the summer of 2008, as indicated by strong column-integrated warming (color contours in Fig. 11; see also Fig. 17). The timing of the warming on the shelf depends on its proximity to the TGB, as expected from the propagation time scale of the velocity anomalies along the shelf break. Change-point analysis reveals that subsurface warming occurs at increasing

lag with distance from the TGB (Fig. 13b), a result that is robust for all depths between 100-m and 200-m depth (Fig. 18). In the spring of 2009, the Laurentian Channel experienced the onset of high temperature anomalies that have persisted through the end of our analysis (December/2018). By the summer of 2009, the warming reached the slope and shelf offshore of the Gulf of Maine. In subsequent months, the warm subsurface waters were swept into the Gulf of Maine through the continuous inflow on the northwest side of the Northeast Channel [25]. The magnitude of the subsurface warming reached 2.5°C in the Laurentian Channel and exceeded 1°C in most of the Gulf of Maine (Fig. 17). By the end of 2010, warmer subsurface waters enveloped the entire Northwest Atlantic Shelf between the Great South Channel and the Laurentian Channel.

South of the Great South Channel, the temperature shift is unlikely the direct result of the along-slope anomaly propagation from the TGB. In the Mid-Atlantic Bight, the significant warming shift shown in Fig. 13b occurred after a one-year lag following the warming on the Scotian Shelf, which is 8-10 months longer than if the anomaly propagated to this region at a speed of approximately 5 cm s^{-1} (the speed of the black dashed line in Fig. 13a). The breakdown of the coherent propagation downstream of the Great South Channel is consistent with water mass analysis[24], which showed a strong discontinuity in mean temperature and salinity, with a much warmer and saltier shelf break front in the Mid-Atlantic Bight likely associated with the close proximity of the Gulf Stream. This discontinuity suggests that other mechanisms influence subsurface temperature fluctuations here. For example, the warming after 2011 in this region may be linked to the increased frequency of warm core rings shed by the Gulf Stream and/or northward shifts in the Gulf Stream orientation downstream of the separation point near Cape Hatteras [26, 27, 28].

The association of the high SSH anomalies at the TGB and the rapid warming of the Northwest Atlantic Shelf after 2008 was not a one-off event. Between 1993 and 2018, the time series of SSH anomalies at the TGB was significantly correlated at the 99.9% confidence level with the subsurface temperature on the shelf (Table 1; temperatures are averaged over the four regions indicated in Fig. 13c), leading at timescales consistent with the propagation speed of the Labrador Current to the Great South Channel. The progressive lead-time of the correlations from the Laurentian Channel (11 months) to the Scotian Shelf (13 months), and Gulf of Maine (14 months) reinforces the westward propagation of temperature anomalies between the Grand Banks and the Northwest Atlantic Shelf. Moreover, the correlations remain significant beyond the 99% level, though smaller in magnitude, at similar lead-times after prewhitening (signal shown in Fig. 15b). The prewhitening procedure removes all autocorrelation as well as the change point. Thus, the robustness of the correlations to this procedure indicates that the association of SSH anomalies at the TGB and the shelf temperature anomalies is ubiquitous throughout the satellite altimetry era and is not tied only to the 2008 change point or to similarities in seasonal patterns. The association at a lag of about 2 years between the signal at the TGB and the temperature of the Mid-Atlantic Bight is also robust to prewhitening.

The advent of satellite altimetry and the surge in subsurface temperature/salinity measurements on the Northwest Atlantic Shelf in recent decades [29, 30] shows that the 2008 circulation shift at the TGB likely set off propagating velocity anomalies along the shelf break and associated shelf warming. Thus, monitoring sea level anomalies at the TGB may help predict impending shelf temperature anomalies with up to a year of lead time. Additionally, the long history of hydrographic surveys in the Northwest Atlantic allows us to place

this recent warming event in the context of the long-term variability of subsurface water masses before the satellite altimetry era.

2.4 Centennial-scale Shelf Warming

The TGB has been monitored for iceberg activity since the International Ice Patrol was formed in response to the sinking of the RMS Titanic in 1912 [31]. As such, it has one of the longest oceanographic records of temperature and salinity anywhere. Here, we look at the historical hydrographic records stretching back nearly a century, to put the 2008 shift in a broader context (Fig. 14). For this purpose, a temperature-salinity (T-S) diagram is useful, as these water properties are conserved beneath the ocean’s surface and change only through mixing. Thus, a T-S diagram reveals the provenance of different water masses (Fig. 14a; see Methods for how the 5,153 hydrographic and float profiles were analyzed to reduce the influence of temporal variability in sampling on this diagram). For instance, the cold, fresh Labrador Current north of the TGB contrasts strongly with the warm, salty Gulf Stream. All of the decadal-averaged T-S profiles at the TGB are within the envelope bounded by the Labrador Current and the Gulf Stream mean profiles in the layers shallower than the Labrador Sea Water.

The T-S diagram indicates that the last ten years are uniquely warm and salty compared to any time in the past 80 years. However, a shift to warmer and saltier water masses in the 1970s was of similar scale to this recent shift, relative to decades that preceded it. This warming and salinification in the 1970s may have been caused by a shift of the Gulf Stream toward the TGB, analogous with the more recent change evidenced from the 2008 rise in SSH (Fig. 12). While the shallower water masses of the thermocline have gotten warmer and saltier, consistent with more frequent incursions of the Gulf Stream onto the TGB, the temperature and salinity of the deep water masses, like the Labrador Sea Water,

have not changed dramatically or monotonically.

The coldest and freshest decades at the TGB occurred between the 1930s and 1960s, only returning to these conditions briefly in the 1990s. The 1990s were extraordinary in this region for a number of reasons. The wintertime deep convection in the Labrador Sea was the strongest since at least the end of the 1930s [32, 33], which resulted in the coldest, freshest and thickest Labrador Sea Water layer on record. Anomalously strong wintertime zonal winds in the subpolar North Atlantic in the early 1990s, expressed as a strong, positive North Atlantic Oscillation (NAO) index [34], helped drive this convection. This cold period lasted only for the first half of the decade; by the late 1990s, the temperature at the TGB returned to the warmer conditions of the post-1970s (Fig. 19).

On the Scotian Shelf and the Laurentian Channel, the subsurface temperatures have followed a similar signal as the TGB over the past nine decades (Fig. 14b, temperature averaged over the grid points marked with black dots in the inset of Fig. 14a). Between 1930 and 1970, the annual mean subsurface shelf temperatures fluctuated widely, with the 1930s and the 1960s being the coldest decades on record (average $\pm$ standard deviation of $7.3 \pm 0.9^\circ\text{C}$ and $7.7 \pm 0.9^\circ\text{C}$, respectively), separated by an intervening warmer period (1940s-1950s, averaging $8.5 \pm 0.9^\circ\text{C}$). At the end of the 1960s, rapid warming caused the mean annual shelf temperatures to exceed 9.5°C for the first time in 1968, a state that persisted with little variation for nearly 40 years (1970s-2000s, averaging $9.1 \pm 0.8^\circ\text{C}$). In 2009, a second warming event raised the shelf temperature by another 1.6°C (2010–2018, averaging $10.7 \pm 0.7^\circ\text{C}$). The subsurface shelf waters since 2012 were warmer than ever previously recorded.

The post-2008 dynamical connection established in Figures 2 and 3, which links SSH anomalies at the TGB to propagating downstream velocity anomalies

and shelf warming, is evident in the TGB T-S time series and Scotian Shelf temperatures (Fig. 14). We speculate that similar dynamics were operating earlier in the 20th century, when the appearance of warm and salty waters at the TGB in the 1970s coincides with warming all along the Northwest Atlantic Shelf. It is unclear if this earlier shift was part of a multidecadal oscillation, yet it is notable that only during the high NAO of the early 1990s did the Scotian Shelf or TGB experience a period nearly as cool or fresh as the 1930s-1960s and the warming after 2008 started from this warmer baseline.

2.5 Discussion

In this study, satellite-based and in situ observations show the influence of the Gulf Stream on the supply of cold, fresh waters from the Labrador Sea to the Northwest Atlantic Shelf. A heightened presence of the Gulf Stream at the TGB after 2008, revealed by a significant warming (Fig. 11), salinification (Fig. 14), and an increase in SSH (Fig. 12), was associated with subsurface warming along the continental shelf and slope between Nova Scotia and Cape Hatteras after 2009 (Figs. 13b, 17, 20 and Table 1). The more frequent impingements of the Gulf Stream at the Tail of the Grand Banks limited the advective connection of the Labrador Current along the edge of the Northwest Atlantic Shelf, thereby reducing the supply of cold, fresh and oxygen-rich waters to the shelf. This perturbation caused an anomaly to propagate along the slope and arrive at the Gulf of Maine nearly one year after the appearance of anomalous properties at the TGB.

Additionally, our analysis of nearly a century of hydrographic data suggests that a similar shift toward more subtropical water at the TGB was linked to shelf warming at the end of the 1960s, from which the system had never fully recovered. This long-term record may lend support to suppositions, based largely on climate modelling, of a 20th century slowdown in the Atlantic Meridional Overturning Cir-

ulation (AMOC), that is correlated with the warming of the Scotian Shelf [35, 36]. Idealized models predict a northward migration of the Gulf Stream accompanying AMOC slowdowns [37] in line with our observation of the Gulf Stream increasingly impinging on the TGB during shelf warming. In fact, the TGB has been called the “pacemaker” region for the AMOC, and simple dynamical arguments call for AMOC slowdowns to be accompanied by SSH increases at this boundary region [38], as observed following 2008 and likely inferred in the late 1960s from our water mass analysis (Fig. 14).

The recent subsurface warming of the Northwest Atlantic Shelf, associated with a dynamic change at the TGB, coincides with unprecedented surface warming [1], salinification [39], and severe marine heat waves [40, 41] that have likely contributed to long-noted trends in fisheries [1, 8, 42, 43, 44]. Our findings not only help to interpret the rapid temperature increase on the shelf, they also present an opportunity to enhance predictability of future warming. Accurately simulating Gulf Stream-Labrador Current interactions at the TGB appears to be crucial to reproducing the last century of warming on the shelf, and, therefore, will likely govern the future properties in this region. Furthermore, monitoring the impingement of the Gulf Stream at the TGB offers up to one year of lead time for warming events on the Northwest Atlantic Shelf, and these predictive capabilities may be valuable for forecasting ecosystem changes of consequence for fisheries management.

2.6 Methods

2.6.1 Satellite Altimetry

Altimetric data is derived from satellite observations with Topex/Poseidon (1992-2002), Jason I (2001-2012), and Jason II (2008-present) and made freely available through the Copernicus Marine Environment Monitoring Service

(CMEMS, <https://resources.marine.copernicus.eu/>). The monthly $0.25^\circ \times 0.25^\circ$ gridded absolute dynamic topography between January 1993 and December 2018 is used to calculate the mean SSH in the region bounded by 30°S , 60°N and 80°W , 40°W , as well as the SSH differences following the 2008 shift. The SSH time series at the TGB is calculated by averaging the absolute dynamic topography over the region highlighted with a thick black contour in Fig. 12a.

The monthly $0.25^\circ \times 0.25^\circ$ gridded surface geostrophic velocity, calculated from the gradient of the SSH between January 1993 and December 2018, provides a measure of the surface Labrador Current speed. The surface geostrophic velocity is interpolated onto the 1,000-m isobath between Flemish Pass and Cape Hatteras, as indicated by the red contour in Fig 2a, using a piecewise linear approximation. At each grid point along the 1,000-m isobath, the surface geostrophic velocity is decomposed into along-slope and across-slope components, with positive values pointing toward Cape Hatteras (along-slope) and inshore (across-slope). The direction of the along-slope component is estimated based on the angle between one grid point on the contoured isobath and the nearest point downstream (i.e., toward Cape Hatteras). Its magnitude is then calculated by projecting the surface geostrophic velocity vector onto the along-slope direction. Similarly, the magnitude of the across-slope component is calculated by projecting the velocity vector onto the across-slope direction. Fig. 16 illustrates the direction and magnitude of the all-time mean surface geostrophic velocity as projected onto these components. The along-slope velocity is considered the surface Labrador Current speed used to calculate the lagged correlations shown in Fig. 13a.

2.6.2 EN4 Profiles

Historical hydrographic and float profile data compiled and made freely available by the Met Office Hadley Centre

(<https://www.metoffice.gov.uk/hadobs/en4/>) are used to probe the multi-decadal variability of water mass composition at the TGB from the 1930s to the present [45]. A total of 5,153 profiles taken in April, May or June within the box 41°N-44°N, 40°W-53°W were analyzed (Fig. 20 shows the location of the profiles used for each decade). Pre-ARGO profiles have been historically biased towards these months, and this subset represents 51% of all profiles taken in this region. We avoid aliasing seasonal variability in our multidecadal time series by limiting our analysis to a single well-sampled season¹⁴. Poor data were removed based on EN4’s quality-control flag system, and only data points with accepted pairs of potential temperature and practical salinity were used. Each profile was linearly interpolated to a maximum of 55 vertical levels, with 5-m resolution in the top 100 m, 25-m resolution above 250 m, 50-m resolution above 1,550 m and 250-m resolution above 2,050 m. The maximum depth of the averaged profiles is 2,050 m, as less than 2% of the profiles in the region reach greater depths.

To avoid the aliasing of variability in the location where the profiles were collected in the TGB box, we subtract an appropriate gridded all-time mean T-S profile from each individual observation, as follows. The profiles were bin-averaged into thirty boxes of 0.5° latitude x 1° longitude with a terrain-following penalty, that sets the “effective distance” between the location of each profile and the center of each box, thereby clustering profiles collected at similar isobaths. The squared effective distance was calculated as

$$Def^2 = [(x_b - x_p)^2 + (y_b - y_p)^2] + [3\lambda(\frac{H_b - H_p}{H_b + H_p})]^2 \quad (1)$$

where λ is set to 1 km, and (x_b, y_b) and (x_p, y_p) are the location of the center of the box and the profile, respectively. H_b and H_p are the depths at those locations, estimated from ETOPO-1 [46]. This penalty lengthens bins in their along-isobath

direction and shortens them in the across-isobath direction, which is a more dynamic way to subdivide this region, given that the currents are steered by steep topography [47, 48].

Once each profile was assigned to one of these 30 terrain-following bins, we averaged the profiles taken in the same bin and month, to avoid giving disproportionate weight to correlated profiles collected very close to one another in space and time. Then, we calculated all-time, bin mean profiles:

$$\bar{T}_{x,y} = \frac{1}{N_{x,y}} \sum_{i=1}^{N_{x,y}} T_{i(x,y)}, \bar{S}_{x,y} = \frac{1}{N_{x,y}} \sum_{i=1}^{N_{x,y}} S_{i(x,y)} \quad (2)$$

where $T_{i(x,y)}$ and $S_{i(x,y)}$ are the temperature and salinity profiles (either individual profiles or the pre-averaged profiles collected in a given bin and month), numbering N quasi-independent profiles in each x,y bin. Anomaly profiles were then calculated by subtracting the all-time, bin mean $(\bar{T}_{x,y}, \bar{S}_{x,y})$ from each profile in the corresponding bin $(T_{i(x,y)}, S_{i(x,y)})$:

$$T'_{i(x,y)} = T_{i(x,y)} - \bar{T}_{x,y}, S'_{i(x,y)} = S_{i(x,y)} - \bar{S}_{x,y} \quad (3)$$

A decadal-mean anomaly profile $\bar{T}'_{decade}, \bar{S}'_{decade}$ was calculated by averaging the resulting anomalies from equation 3 over the entire TGB region for each of the nine decades. Finally, the all-time, spatially-averaged mean profile was added back to the decadal anomalies to calculate the mean decadal profiles:

$$\bar{T}_{decade} = \bar{T}'_{decade} + \frac{1}{N_{x,y}} \sum_{i=1}^{N_{x,y}} T_i, \bar{S}_{decade} = \bar{S}'_{decade} + \frac{1}{N_{x,y}} \sum_{i=1}^{N_{x,y}} S_i \quad (4)$$

We apply the same method to calculate the all-time mean profiles in the Labrador Current (44°N-47°N, 48°W-47°W) and Gulf Stream (40°N-41°N, 56°W-53°W) domains at a 10-m vertical resolution, as shown in solid lines in Fig. 14a.

2.6.3 EN4 Objective Analysis

The monthly $1^\circ \times 1^\circ$ objective analysis gridded product with 42 vertical levels [45, 49], made freely available by the Met Office Hadley Centre (<https://www.metoffice.gov.uk/hadobs/en4/>), is derived from the hydrographic and float dataset described above. The temperature field at 149 m in the region bounded by 33°S , 50°N and 77°W , 48°W is used in the change-point analysis described below. Here, the period analyzed is January 1993 to December 2018, coincident with the altimetric data. Twelve additional layers between 56 m and 235 m were also analyzed to determine the vertical extent of the changes observed at 149 m.

Decadal changes in the 149-m layer are analyzed in the time series extending back to 1930 (Fig. 14b). The authors of the EN4 objective analysis highlight that this dataset should be used with caution in the analysis of long-term trends, because, during periods with few observations, the analyses relax to climatology [45]. Fig. 21 shows that the Northwest Atlantic has been historically well-observed, as the number of profiles is plotted in a $1^\circ \times 1^\circ$ grid. Over 1.3 million profiles were used to build the objective analyses here, most of which were taken on the shelf and slope. The error estimate in Fig. 14b is calculated at each time-step t as

$$E = \text{sqr}t \frac{1}{N} \sum_{x=1}^N \sigma_x^2 + \frac{1}{N} \sum_{x=1}^N U_{x,t}^2 \quad (5)$$

where σ_x and $U_{x,t}$ are the standard deviation and objective analysis uncertainty estimate of the 149-m temperature at each of the N grid points, respectively.

2.6.4 Change-Point Analysis

The 2008 shift in the 1993–2018 SSH time series at the TGB is characterized using a seasonal AutoRegressive Integrated Moving Average (ARIMA) model, which explains the SSH based on its own past values (i.e. its own lags and lagged

observation errors). We jointly fit a seasonal ARIMA with each possible monthly change point, iterating over all months between January 1997 and December 2014. The timing of the SSH shift is selected as the month at which inserting a change point maximizes the model log-likelihood, and its inclusion is verified by comparing the Akaike information criterion (AIC) to a model without a change point. All change-point analyses are conducted in R [50] using the “forecast” package [51]. The orders of the resulting seasonal ARIMA, chosen via stepwise selection using the AIC, were $(2, 0, 0) \times (2, 0, 0)_{12}$. These orders indicate that the SSH time series is mean stationary, aside from the jointly-fit level shift, with significant autocorrelation at 1-, 2-, 12- and 24-month lags, as seen in Fig. 15b. These lags are consistent with strong month-to-month and seasonal signals. The temporal autocorrelation structure explains 45.8% of the variability in the time series, and the change point explains an additional 24.1%.

Similar to the analysis of the SSH at the TGB, temperature change points between January 1997 and December 2014 on the Northwest Atlantic Shelf are identified jointly with a seasonal ARIMA model, for each $1^\circ \times 1^\circ$ grid cell between 30° - 60° N and 40° - 80° W. The maximum orders we allowed for this seasonal ARIMA models are $(1, 0, 1) \times (1, 1, 1)_{12}$, chosen under the assumption that the temperature observations one month and one year prior to a measurement contain all available information for estimation. We also assume that the temperature time series are mean stationary after accounting for any change points, and therefore do not model a non-seasonal integrated process. However, variation in the magnitude of the seasonal cycle is permitted. Candidate temperature change points are selected as those that maximize the three-month running mean of the model log-likelihood, in order to avoid choosing isolated, sharp peaks in the likelihood function, and are retained if they both reduce the AIC over a model without a change point and

Table 1. SSH and temperature lagged correlations. Maximum correlation coefficients and time lags between the raw and prewhitened monthly SSH at the Tail of the Grand Banks and the monthly 149-m temperature averaged over each of the regions indicated in Fig 3b, for 1993–2018. Correlation coefficients in bold and italic fonts are significant at the 99.9% and 99% confidence levels, respectively.

		Mid-Atlantic Bight	Gulf of Maine	Scotian Shelf	Laurentian Channel
Raw Time Series	Correlation Coefficient	0.50	0.50	0.50	0.45
	Time Lag (months)	26	14	13	11
Prewhitened Time Series	Correlation Coefficient	<i>0.19</i>	0.20	0.21	0.20
	Time Lag (months)	24	14	14	10

occur between the SSH shift at the TGB (July, 2008) and December, 2011. We chose this window to identify only temperature change points that followed the SSH shift at the TGB, considering the time lags associated with the propagation speed of the Labrador Current described in Figure 3a.

To study the relationship between temperature change points detected in different regions of the Northwest Atlantic Shelf and the SSH shift at the TGB, mean temperature time series are calculated for areas of grid cells exhibiting similar change-point timing (hatch marks in Figure 3b). To prewhiten the SSH and regional temperature time series, we filter using the SSH seasonal ARIMA model fit, including the change point, using the R package “TSA” [52]. For both the raw and prewhitened time series, the strength and time lag of the correlations between the SSH at the TGB and the mean temperature of each identified region are evaluated (Table 1).

List of References

- [1] A. J. Pershing, M. A. Alexander, C. M. Hernandez, L. A. Kerr, A. Le Bris, K. E. Mills, J. A. Nye, N. R. Record, H. A. Scannell, J. D. Scott, G. D. Sherwood, and A. C. Thomas, “Slow adaptation in the face of rapid warming leads to collapse of the Gulf of Maine cod fishery,” *Science*, vol. 350, no. 6262, pp. 809–812, Nov. 2015. [Online]. Available: <https://www.sciencemag.org/lookup/doi/10.1126/science.aac9819>
- [2] N. M. F. S. , “Fisheries Economics of the United States 2016,” U.S. Dept. of Commerce, NOAA Tech. Memo. NMFS-F/SPO-187, U.S. Dept. of Commerce, NOAA Tech. Memo. NMFS-F/SPO-187, Tech. Rep., 2018.
- [3] S. K. Gaichas, J. S. Link, and J. A. Hare, “A risk-based approach to evaluating northeast US fish community vulnerability to climate change,” *ICES Journal of Marine Science*, vol. 71, no. 8, pp. 2323–2342, Oct. 2014. [Online]. Available: <https://academic.oup.com/icesjms/article/71/8/2323/758638>
- [4] J. A. Hare, W. E. Morrison, M. W. Nelson, M. M. Stachura, E. J. Teeters, R. B. Griffis, M. A. Alexander, J. D. Scott, L. Alade, R. J. Bell, A. S. Chute, K. L. Curti, T. H. Curtis, D. Kircheis, J. F. Kocik, S. M. Lucey, C. T. McCandless, L. M. Milke, D. E. Richardson, E. Robillard, H. J. Walsh, M. C. McManus, K. E. Marancik, and C. A. Griswold, “A Vulnerability Assessment of Fish and Invertebrates to Climate Change on the Northeast U.S. Continental Shelf,” *PLOS ONE*, vol. 11, no. 2, p. e0146756, Feb. 2016. [Online]. Available: <https://dx.plos.org/10.1371/journal.pone.0146756>
- [5] L. M. Lauzier, “Long-term temperature variations in the Scotian Shelf area,” *ICNAF Spec. Publ.*, vol. 6, pp. 807–816, 1965.
- [6] J. A. Nye, T. M. Joyce, Y.-O. Kwon, and J. S. Link, “Silver hake tracks changes in Northwest Atlantic circulation,” *Nature Communications*, vol. 2, no. 1, p. 412, Sept. 2011. [Online]. Available: <http://www.nature.com/articles/ncomms1420>
- [7] A. J. Pershing, C. H. Greene, C. Hannah, D. Sameoto, E. Head, D. J. Mountain, J. W. Jossi, M. C. Benfield, P. C. Reid, and T. G. Durbin, “Oceanographic Responses to Climate in the Northwest Atlantic,” *Oceanography*, vol. 14, no. 3, pp. 76–82, 2001.
- [8] X. J. Davis, T. M. Joyce, and Y.-O. Kwon, “Prediction of silver hake distribution on the Northeast U.S. shelf based on the Gulf Stream path index,” *Continental Shelf Research*, vol. 138, pp. 51–64, Apr. 2017. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S027843431630067X>
- [9] A. Le Bris, K. E. Mills, R. A. Wahle, Y. Chen, M. A. Alexander, A. J. Allyn, J. G. Schuetz, J. D. Scott, and A. J. Pershing, “Climate vulnerability

- and resilience in the most valuable North American fishery,” *Proceedings of the National Academy of Sciences*, vol. 115, no. 8, pp. 1831–1836, Feb. 2018. [Online]. Available: <http://www.pnas.org/lookup/doi/10.1073/pnas.1711122115>
- [10] A. T. Schartup, C. P. Thackray, A. Qureshi, C. Dassuncao, K. Gillespie, A. Hanke, and E. M. Sunderland, “Climate change and overfishing increase neurotoxicant in marine predators,” *Nature*, vol. 572, no. 7771, pp. 648–650, Aug. 2019. [Online]. Available: <http://www.nature.com/articles/s41586-019-1468-9>
 - [11] B. Petrie and K. Drinkwater, “Temperature and salinity variability on the Scotian Shelf and in the Gulf of Maine 1945–1990,” *Journal of Geophysical Research*, vol. 98, no. C11, p. 20079, 1993. [Online]. Available: <http://doi.wiley.com/10.1029/93JC02191>
 - [12] D. Gilbert, B. Sundby, C. Gobeil, A. Mucci, and G.-H. Tremblay, “A seventy-two-year record of diminishing deep-water oxygen in the St. Lawrence estuary: The northwest Atlantic connection,” *Limnology and Oceanography*, vol. 50, no. 5, pp. 1654–1666, Sept. 2005. [Online]. Available: <http://doi.wiley.com/10.4319/lo.2005.50.5.1654>
 - [13] D. Brickman, D. Hebert, and Z. Wang, “Mechanism for the recent ocean warming events on the Scotian Shelf of eastern Canada,” *Continental Shelf Research*, vol. 156, pp. 11–22, Mar. 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0278434317302650>
 - [14] M. Claret, E. D. Galbraith, J. B. Palter, D. Bianchi, K. Fennel, D. Gilbert, and J. P. Dunne, “Rapid coastal deoxygenation due to ocean circulation shift in the northwest Atlantic,” *Nature Climate Change*, vol. 8, no. 10, pp. 868–872, Oct. 2018. [Online]. Available: <http://www.nature.com/articles/s41558-018-0263-1>
 - [15] J. L. Lillibridge and A. J. Mariano, “A statistical analysis of Gulf Stream variability from 18+ years of altimetry data,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 85, pp. 127–146, Jan. 2013. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064512001208>
 - [16] T. Rossby, C. N. Flagg, K. Donohue, A. Sanchez-Franks, and J. Lillibridge, “On the long-term stability of Gulf Stream transport based on 20 years of direct measurements,” *Geophysical Research Letters*, vol. 41, no. 1, pp. 114–120, Jan. 2014. [Online]. Available: <http://doi.wiley.com/10.1002/2013GL058636>
 - [17] M. Andres, “On the recent destabilization of the Gulf Stream path downstream of Cape Hatteras: Gulf Stream Path Destabilization,”

- Geophysical Research Letters*, vol. 43, no. 18, pp. 9836–9842, Sept. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016GL069966>
- [18] T. Rossby, “On gyre interactions,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 46, no. 1-2, pp. 139–164, Jan. 1999. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S09670664598000952>
- [19] S. Dong, M. O. Baringer, and G. J. Goni, “Slow Down of the Gulf Stream during 1993–2016,” *Scientific Reports*, vol. 9, no. 1, p. 6672, Dec. 2019. [Online]. Available: <http://www.nature.com/articles/s41598-019-42820-8>
- [20] M. Andres, K. A. Donohue, and J. M. Toole, “The Gulf Stream’s path and time-averaged velocity structure and transport at 68.5W and 70.3W,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 156, p. 103179, Feb. 2020. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063719303255>
- [21] T. Rossby, “The North Atlantic Current and surrounding waters: At the crossroads,” *Reviews of Geophysics*, vol. 34, no. 4, pp. 463–481, Nov. 1996. [Online]. Available: <http://doi.wiley.com/10.1029/96RG02214>
- [22] J. W. Loder, B. Petrie, and G. G. Gawarkiewicz, “The coastal ocean off northeastern North America: A large-scale view,” in *The Sea 11*. New York: Wiley, 1998, pp. 105–133.
- [23] P. S. Fratantoni and M. S. McCartney, “Freshwater export from the Labrador Current to the North Atlantic Current at the Tail of the Grand Banks of Newfoundland,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 57, no. 2, pp. 258–283, Feb. 2010. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063709002222>
- [24] P. S. Fratantoni and R. S. Pickart, “The Western North Atlantic Shelfbreak Current System in Summer,” *Journal of Physical Oceanography*, vol. 37, no. 10, pp. 2509–2533, Oct. 2007. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO3123.1>
- [25] S. R. Ramp, R. J. Schlitz, and W. R. Wright, “The deep flow through the Northeast Channel, Gulf of Maine,” *Journal of Physical Oceanography*, vol. 15, pp. 1790–1808, 1985.
- [26] G. G. Gawarkiewicz, R. E. Todd, A. J. Plueddemann, M. Andres, and J. P. Manning, “Direct interaction between the Gulf Stream and the shelfbreak south of New England,” *Scientific Reports*, vol. 2, no. 1, p. 553, Dec. 2012. [Online]. Available: <http://www.nature.com/articles/srep00553>
- [27] J. S. T. Forsyth, M. Andres, and G. G. Gawarkiewicz, “Recent accelerated warming of the continental shelf off New Jersey: Observations from the

- CMV Oleander expendable bathythermograph line,” *Journal of Geophysical Research: Oceans*, vol. 120, no. 3, pp. 2370–2384, Mar. 2015. [Online]. Available: <http://doi.wiley.com/10.1002/2014JC010516>
- [28] A. Gangopadhyay, G. Gawarkiewicz, E. N. S. Silva, M. Monim, and J. Clark, “An Observed Regime Shift in the Formation of Warm Core Rings from the Gulf Stream,” *Scientific Reports*, vol. 9, no. 1, p. 12319, Dec. 2019. [Online]. Available: <http://www.nature.com/articles/s41598-019-48661-9>
- [29] J. M. Toole, M. Andres, I. A. Le Bras, T. M. Joyce, and M. S. McCartney, “Moored observations of the Deep Western Boundary Current in the NW Atlantic: 2004-2014: LINE W 2004-2014,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 9, pp. 7488–7505, Sept. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2017JC012984>
- [30] L. Smith, J. Barth, D. Kelley, A. Plueddemann, I. Rodero, G. Ulses, M. Vardaro, and R. Weller, “The Ocean Observatories Initiative,” *Oceanography*, vol. 31, no. 1, pp. 16–35, Mar. 2018. [Online]. Available: <https://tos.org/oceanography/article/the-ocean-observatories-initiative>
- [31] IIP, “Report of the International Ice Patrol in the North Atlantic (2019 Season),” CG-188-74, US Coast Guard Bulletin 105, 2019.
- [32] I. Yashayaev, “Hydrographic changes in the Labrador Sea, 1960–2005 [Prog. Oceanogr. 73 (2007) 242–276],” *Progress in Oceanography*, vol. 75, no. 4, pp. 857–859, Dec. 2007. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S007966110700184X>
- [33] I. Yashayaev and J. W. Loder, “Recurrent replenishment of Labrador Sea Water and associated decadal-scale variability: 2015 Convection in Labrador Sea,” *Journal of Geophysical Research: Oceans*, vol. 121, no. 11, pp. 8095–8114, Nov. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012046>
- [34] I. Yashayaev and J. W. Loder, “Further intensification of deep convection in the Labrador Sea in 2016,” *Geophysical Research Letters*, vol. 44, no. 3, pp. 1429–1438, Feb. 2017. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016GL071668>
- [35] V. S. Saba, S. M. Griffies, W. G. Anderson, M. Winton, M. A. Alexander, T. L. Delworth, J. A. Hare, M. J. Harrison, A. Rosati, G. A. Vecchi, and R. Zhang, “Enhanced warming of the Northwest Atlantic Ocean under climate change,” *Journal of Geophysical Research: Oceans*, vol. 121, no. 1, pp. 118–132, 2015. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015JC011346>

- [36] L. Caesar, S. Rahmstorf, A. Robinson, G. Feulner, and V. Saba, “Observed fingerprint of a weakening Atlantic Ocean overturning circulation,” *Nature*, vol. 556, no. 7700, pp. 191–196, Apr. 2018. [Online]. Available: <http://www.nature.com/articles/s41586-018-0006-5>
- [37] R. Zhang and G. K. Vallis, “The Role of Bottom Vortex Stretching on the Path of the North Atlantic Western Boundary Current and on the Northern Recirculation Gyre,” *Journal of Physical Oceanography*, vol. 37, no. 8, pp. 2053–2080, Aug. 2007. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO3102.1>
- [38] M. W. Buckley and J. Marshall, “Observations, inferences, and mechanisms of the Atlantic Meridional Overturning Circulation: A review,” *Reviews of Geophysics*, vol. 54, no. 1, pp. 5–63, Mar. 2016. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015RG000493>
- [39] S. A. Grodsky, N. Reul, B. Chapron, J. A. Carton, and F. O. Bryan, “Interannual surface salinity on Northwest Atlantic shelf: INTERANNUAL SSS IN NW ATLANTIC SHELF,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 5, pp. 3638–3659, May 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012580>
- [40] K. Mills, A. Pershing, C. Brown, Y. Chen, F.-S. Chiang, D. Holland, S. Lehuta, J. Nye, J. Sun, A. Thomas, and R. Wahle, “Fisheries Management in a Changing Climate: Lessons From the 2012 Ocean Heat Wave in the Northwest Atlantic,” *Oceanography*, vol. 26, no. 2, 2013. [Online]. Available: <https://tos.org/oceanography/article/fisheries-management-in-a-changing-climate-lessonsfrom-the-2012-ocean-heat->
- [41] A. Pershing, K. Mills, A. Dayton, B. Franklin, and B. Kennedy, “Evidence for Adaptation from the 2016 Marine Heatwave in the Northwest Atlantic Ocean,” *Oceanography*, vol. 31, no. 2, pp. 151–161, June 2018. [Online]. Available: <https://tos.org/oceanography/article/evidence-for-adaptation-from-the-2016-marine-heatwave-in-the-northwest-atl>
- [42] J. Nye, J. Link, J. Hare, and W. Overholtz, “Changing spatial distribution of fish stocks in relation to climate and population size on the Northeast United States continental shelf,” *Marine Ecology Progress Series*, vol. 393, pp. 111–129, Oct. 2009. [Online]. Available: <http://www.int-res.com/abstracts/meps/v393/p111-129/>
- [43] B. A. Dubik, E. C. Clark, T. Young, S. B. J. Zigler, M. M. Provost, M. L. Pinsky, and K. St. Martin, “Governing fisheries in the face of change: Social responses to long-term geographic shifts in a U.S. fishery,” *Marine Policy*, vol. 99, pp. 243–251, Jan. 2019. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0308597X18305463>

- [44] K. L. Oremus, “Climate variability reduces employment in New England fisheries,” *Proceedings of the National Academy of Sciences*, vol. 116, no. 52, pp. 26 444–26 449, Dec. 2019. [Online]. Available: <http://www.pnas.org/lookup/doi/10.1073/pnas.1820154116>
- [45] S. A. Good, M. J. Martin, and N. A. Rayner, “EN4: Quality controlled ocean temperature and salinity profiles and monthly objective analyses with uncertainty estimates: THE EN4 DATA SET,” *Journal of Geophysical Research: Oceans*, vol. 118, no. 12, pp. 6704–6716, Dec. 2013. [Online]. Available: <http://doi.wiley.com/10.1002/2013JC009067>
- [46] B. E. Eakins and C. Amante, “ETOPO1 1 arc-minute global relief model: procedures, data sources and analysis,” National Geophysical Data Center, Boulder, Colorado, Tech. Rep. NESDIS NGDC-24, 2009.
- [47] R. E. Davis, “Preliminary results from directly measuring middepth circulation in the tropical and South Pacific,” *Journal of Geophysical Research: Oceans*, vol. 103, no. C11, pp. 24 619–24 639, Oct. 1998. [Online]. Available: <http://doi.wiley.com/10.1029/98JC01913>
- [48] K. L. Lavender, W. Brechner Owens, and R. E. Davis, “The mid-depth circulation of the subpolar North Atlantic Ocean as measured by subsurface floats,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 52, no. 5, pp. 767–785, May 2005. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063705000154>
- [49] V. Gouretski and F. Reseghetti, “On depth and temperature biases in bathythermograph data: Development of a new correction scheme based on analysis of a global ocean database,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 57, no. 6, pp. 812–833, June 2010. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063710000671>
- [50] R. C. Team, “R: A language and environment for statistical computing,” Vienna, Austria, 2020. [Online]. Available: <https://www.R-project.org/>
- [51] R. Hyndman, G. Athanasopoulos, C. Bergmeir, G. Caceres, L. Chhay, M. O’Hara-Wild, F. Petropoulos, S. Razbash, E. Wang, and F. Yasmeen, “forecast: Forecasting functions for time series and linear models. R package version 8.10,” 2019. [Online]. Available: <https://pkg.robjhyndman.com/forecast/>
- [52] K.-s. Chan and B. Ripley, “TSA: Time series analysis. R package version 1.2,” 2018. [Online]. Available: <https://CRAN.R-project.org/package=TSA>
- [53] K. Thyng, C. Greene, R. Hetland, H. Zimmerle, and S. DiMarco, “True Colors of Oceanography: Guidelines for Effective and Accurate Colormap Selection,” *Oceanography*, vol. 29, no. 3, pp. 9–13,

- Sept. 2016. [Online]. Available: <https://tos.org/oceanography/article/true-colors-of-oceanography-guidelines-for-effective-and-accurate-colormap>
- [54] F. Crameri, “Scientific Colour-Maps: Perceptually uniform and colour-blind friendly,” 2018. [Online]. Available: <https://zenodo.org/record/3596401#.Xwfc0ZNKgWo>
- [55] S. Zeller and D. Rogers, “Visualizing Science: How Color Determines What We See,” *EOS*, vol. 101, 2020.

Figure 11. Schematic of the circulation in the Northwest Atlantic Ocean and the water column average temperature difference between the period 2009–2018 and 2001–2007. The red (blue) shading indicates warming (cooling) of the vertically-averaged ocean temperature from the EN4 objective analysis to 2,000 m, or the seafloor if it is shallower than 2,000 m, in 0.5°C increments (change in the unshaded region is between -0.25°C and 0.25°C). Background in grayscale shows the bathymetry of the region, with darker shades representing shallower areas. The main surface circulation patterns associated with the Gulf Stream (red) and the Labrador Current (blue) systems are identified with arrows. Coastal and shelf areas of interest are indicated. TGB stands for Tail of the Grand Banks and SENR, for Southeastern Newfoundland Ridge.

Figure 12. Sea surface height shift at the Tail of the Grand Banks in 2008. (a) Mean Dynamic Topography (MDT) in the Northwest Atlantic between 1993 and 2018. The thick red contour is the 1,000-m isobath between 76°W and 48°W, along which the along-slope surface velocities displayed in Fig. 13a are calculated. The black dotted lines indicate the main cross-slope channels in the region: Great South Channel (GSCh), Northeast Channel (NECh) and Laurentian Channel (LCh). The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. (b) Sea surface height difference between 2009–2018 and 1993–2007. To emphasize spatial patterns, the sea surface height increase averaged over the entire region that is plotted (equal to 4.5 cm) has been subtracted. (c) Time-series of the monthly sea surface height (dark gray line) and seasonal ARIMA model (light gray line, estimated as described in the Methods section) at the Tail of the Grand Banks (i.e. averaged within the thick black contour in panel a). The blue and red horizontal lines indicate the averaged sea surface height of -10.5 cm and 0.3 cm before and after the shift in July of 2008. Colormaps in panels a and b are from the Cmocean package [53].

Figure 13. Along-slope velocity and temperature change west of the TGB. (a) The correlation coefficient between the deseasonalized along-slope surface velocity at the Tail of the Grand Banks of Newfoundland (along the 1,000-m isobath, at 42°51'N, 50°40'W) and the deseasonalized along-slope surface velocity downstream towards the Northwest Atlantic Shelf as a function of time lags and distance from the TGB. White contours represent the 95% significance level. The dashed black line represents a propagation speed of 130 km month⁻¹ (5 cm s⁻¹). (b) Map of change-point timing of the 149-m temperature in the Northwest Atlantic. The shades only show grid points whose single change point during 1993–2018 occurred between January/2009 and December/2011. The temperature changes, in °C, associated with these shifts are displayed in Fig. 17. Colors are displayed in 3-month intervals. The various hatch marks indicate the grid points used to calculate the regional lagged correlations shown in Table 1: Laurentian Channel (white circles), Scotia Shelf (black plus signs), Gulf of Maine (white triangles) and Mid-Atlantic Bight (black crosses). The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. The main cross-shelf features are identified. TGB = Tail of the Grand Banks; LCh = Laurentian Channel; NECh = Northeast Channel; GSCh = Great South Channel; CH = Cape Hatteras. Colormaps are from the Crameri package [54].

Figure 14. Temperature-Salinity diagram at the TGB and time series of the 149-m temperature on the Scotian Shelf and Laurentian Channel. (a) Mean decadal profiles of temperature and salinity at the TGB between the 1930s and the 2010s are shown as a T-S diagram. The profiles are color-coded by decade of sampling. The blue and red solid lines indicate mean T-S profiles of the Labrador Current and the Gulf Stream, averaged over the blue and red boxes shown in the inset map, respectively. Thin, back lines are the potential density anomalies referenced to the sea surface, with the 27.68 kg m^{-3} isopycnal (i.e. potential density of $1027.68 \text{ kg m}^{-3}$) and 27.80 kg m^{-3} isopycnal indicating the upper and lower boundaries of the Labrador Sea Water. A robust locally estimated scatterplot smoothing is applied to the nine decadal profiles at the TGB to reduce the effect of outliers at poorly-sampled depths. The strategy used to create this diagram while minimizing the influence of variability in the location where the profiles were collected is described in the methods. (b) Monthly time series of the 149-m temperature averaged over the Scotian Shelf and Laurentian Channel (grid points marked with black dots in the inset map). Gray shades indicate the error estimate averaged over the respective grid points, calculated as shown in the Methods section.

Figure 15. Sample probability distribution and autocorrelation of the sea surface height time series at the TGB shown in Fig. 12c. (a) Sample probability distribution histogram of the pre-shift months (January 1993 – July 2008; blue bars) and post-shift months (August 2008 – December 2018; red bars). The transition in 2008 represented a 10.5 cm increase in the mean SSH and explains 24.1% of the variability in the time series. (b) Autocorrelation function, showing the highest values at 1-, 2-, 12- and 24-month lags, which indicates the strong month-to-month and seasonal signals. The temporal autocorrelation structure explains 45.8% of the variability in the time series.

Figure 16. Along-slope velocity between Flemish Pass and Cape Hatteras. Black arrows indicate the 26-year mean surface velocity as interpolated onto the 1,000-m isobath. Red and blue arrows are the along- and across-slope velocity components as calculated from the angle between two consecutive grid points from Flemish Pass. Gray contours are the 100-m and 1,000-m isobaths. The monthly along-slope velocity is used to calculate the lagged correlations shown in Fig. 13a.

Figure 17. Magnitude of the 149-m temperature shift in the Northwest Atlantic. The temperature shift is calculated at each grid point as the mean temperature after the change-point timing minus the mean temperature before the change-point timing, as shown in Fig. 13b, for 1993–2018. The shades only show grid points whose single change point happened between January 2009 and December 2011. All of the changes on the shelf and slope are positive, i.e., represent a shift towards warmer waters. Colors are displayed in 0.5°C bins. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. The main cross-shelf features are identified. LCh = Laurentian Channel; NECh = Northeast Channel; GSCh = Great South Channel; CH = Cape Hatteras. Colormap from the Cmocean package [53].

Figure 18. Hydrographic and float profiles at the TGB from the EN4 database. (a-i) Each panel shows the locations of all profiles taken between April and June in each decade since the 1930s with pairs of temperature and salinity. Months are indicated by colors: blue for April, red for May and green for June. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray.

Figure 19. Post-2008 temperature shifts on the Northwest Atlantic Shelf. Timing (left; a,c,e,g) and magnitude (right; b,d,f,h) of the identified temperature shifts at the 122-m (a,b), 135-m (c,d), 166-m (e,f) and 185-m (g,h) layers from the EN4 objective analysis. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. Based on the results of the 149-m layer, analysis was limited to a polygon defined by the vertices (30°N, 80°W), (36°N, 65°W), (36°N, 52°W), (43°N, 48°W), (48°N, 42°W), (51°N, 47°W), (53°N, 54°W), and (48°N, 80°W). Colormaps in panels a and b are, respectively, from the Crameri [54] and Cmocean [53] packages.

Figure 20. Ninety-year (1930-2019) time series of the normalized temperature anomalies at the TGB at 350 m—the mean depth of the 27.25 kg m^{-3} isopycnal. The red line and shades are the three-year running mean and the standard error envelope, respectively. The gray circles are one-year averages. Plotting temperature at a fixed depth allows us to include the quality-controlled profiles of temperature that do not also have salinity (3,565 profiles that reach a depth of 350 m), and also combines the signal of the deepening thermocline (as in reference 14) as well as water mass changes on density surfaces. The method for making this time series is the same as that described in the methods for the T-S plot, where the bin-average was subtracted from each profile before averaging the anomalies. In this case, the anomaly was also normalized by the standard deviation in the bin, so that the plotted anomalies are the mean number of standard deviations from the mean temperature in the TGB box.

Figure 21. Hydrographic and float profiles from the EN4 database in the Northwest Atlantic. (a-i) The shades show the number of profiles by decade between the 1930s and the 2010s in $1^\circ \times 1^\circ$ bins. Over 1.3 million profiles were used to build the objective analysis in this region, of which 57% were collected in the shelf regions that are used to construct Figs. 13b, 14b, 17 and Table 1. The 100-m, 1,000-m, 3,000-m and 4,000-m isobaths are contoured in gray. Colormaps are from the SciVisColor database [55].

The Warming of the Northwest Atlantic from a Lagrangian Perspective

3.1 Introduction

The Northwest Atlantic Shelf is one of the fastest warming regions in the ocean [1]. The accelerated warming experienced since 2008 has contributed to set the conditions for large marine heat waves and has challenged fisheries management [2, 3]. Numerical studies have linked water mass shifts on the Scotian Shelf and Gulf of St. Lawrence to changes in the westward transport of Labrador Current [4, 5]. Claret et al. (2018) further showed that a decrease in the connectivity of the Labrador Current past the shallow underwater plateau known as the Grand Banks of Newfoundland in a future warming scenario under a doubling of CO₂ would intensify the deoxygenation of the Gulf of St. Lawrence that has been underway for decades. Specifically, they pointed to possible circulation changes in the region south of the Grand Banks.

This region, commonly referred to as the Tail of the Grand Banks (TGB, Fig. 22), extends southeastward as a deep ridge. Here, the Gulf Stream interacts with both the complex bathymetry and the Labrador Current. An analysis of 26 years of satellite altimetry has revealed that positive sea surface height (SSH) anomalies at the Tail of the Grand Banks are associated with a slowing of the Labrador Current all along the Scotian and Northeast US Shelf (Gonçalves Neto et al., in review). The slowing of the Labrador Current is accompanied by progressive warming of the shelf waters. Specifically, an abrupt increase in sea surface height at the Tail of the Grand Banks in 2008 preceded the rapid subsurface warming that propagated from the Gulf of St. Lawrence to the Gulf of Maine in 2009 and 2010. Therefore, monitoring SSH at the the TGB may provide predictability for environmental conditions on the shelf. It remains unknown, however, why the sea

Figure 22. (a) Schematic of general circulation in the Northwest Atlantic. The launch site is identified with a triangle. (b) Mean depth (colors) at the top of Layer 19 or the depth of the layer outcrop and Layer 19 mean velocity (arrows). Vectors whose speed is less 5 cm s⁻¹ are shown in teal; between 5 and 10 cm s⁻¹, in dark gray; greater than 10 cm s⁻¹ in red.

surface height increase manifests as reduced connectivity of the Labrador Current.

The complex circulation and bathymetry at the Tail of the Grand Banks, combined with the lack of substantial long-term subsurface observations, imposes difficulties to the understanding of the mechanisms that lead to the sharp shift in sea surface height here. Lagrangian analyses have been employed to reveal circulation features at and near the Tail of the Grand Banks during the past decade. For example, the discovery that Lagrangian floats within the Labrador Sea Water were exported to the ocean interior at Flemish Cap and the TGB [6, 7] challenged the long-standing hypothesis of continuity of the Deep Western Boundary Current [8]. The dense waters that “leak” continue to flow equatorward through interior pathways, greatly contributing to the AMOC [9]. Further analyses with synthetic Lagrangian floats in high-resolution ocean circulation models have corroborated the evidence that internal pathways greatly contribute to the total

equatorward transport in the deep North Atlantic [10, 11]. The curvature of the shelf-break create hot spots of leakiness, that can be observed in both Lagrangian and Eulerian frames of reference [12].

To investigate the circulation near the Tail of the Grand Banks in a high-resolution model, it is important to choose a model that performs well in this region. Ocean circulation models have historically failed to accurately simulate the Gulf Stream between Cape Hatteras and the Grand Banks. A recent model inter-comparison study has shown that the Hybrid Coordinate Ocean Model (HYCOM) outperforms other models in this region [13] and therefore was used in this study. The map of mean sea surface height from altimetry (Fig. 23a) reveals a meridional gradient along the entire extent of the Gulf Stream between Cape Hatteras and the Grand Banks, with the ocean surface being at a higher level to its south. South of the Grand Banks, at 50°W, the Gulf Stream flows seaward of the 4,000-m isobath and takes a sharp northeastward turn. The mean patterns in the Atlantic HYCOM output agrees with the altimetry sea surface height (Fig 23b), accurately simulating the mean angle of the Gulf Stream and its mean position at the Tail of the Grand Banks.

While past studies using Lagrangian analyses in the Northwest Atlantic have greatly contributed to the understanding of the AMOC and coastal processes, they have not resolved temporal changes. In this study, we use Lagrangian analyses in a high-resolution ocean circulation model to evaluate the circulation changes at the Tail of the Grand Banks that are associated with the rapid warming of the Northwest Atlantic Shelf and Slope after 2008. To do so, we deploy a suite of synthetic particles in the vicinity of the Labrador Current in the eastern Grand Banks, and track their dispersion in a coupled Lagrangian-Eulerian analysis.

3.2 Data and Methods

3.2.1 Numerical Model

This study is performed using output from a $1/12^\circ$ eddy-resolving ocean circulation model of the North Atlantic Ocean [14]. This product, that does not assimilate observational data, is based on HYCOM [15, 16, 17] and has a hybrid vertical configuration with 32 layers that combines aspects of z and sigma coordinates: beneath the surface mixed layer, the model layers are nearly isopycnal; in the surface mixed layer, the model layers are isobaric. This Atlantic simulation has been shown to accurately represent important aspects of the shallow and deep circulation of the North Atlantic [14, 18, 19]. In particular, this model output was shown to faithfully represent the mean full-depth transport structure observed with moored instruments in the eastern Labrador Sea at 53°N and at the Tail of the Grand Banks at 43°N [14, 20]. The original model output is available at daily averages for 1978-2017 and is forced by the European Center for Medium-Range Weather Forecasts (ECMWF) reanalysis ERA 40 for 1978-2001 [21] and by the Fleet Numerical Meteorology and Oceanography Center (FNMOC) Navy Operational Global Atmospheric Prediction System (NOGAPS) for 2002-present [22]. The Lagrangian analysis described below is performed on the period between January 2003 and December 2012.

Recent improvement in ocean reanalyses has allowed for a more accurate understanding of the Northwest Atlantic. In particular, the inaccuracy of ocean models to simulate the Gulf Stream separation has posed doubt on their ability to depict the path of the Gulf Stream between Cape Hatteras and the Tail of the Grand Banks [23]. A recent inter-comparison of the Gulf Stream in thirteen ocean reanalyses and unconstrained simulations showed that HYCOM has outperformed other eddy-resolving, eddy-permitting and coarse resolution models [13]. Specifically, HYCOM had the best simulation of the Gulf Stream path and a num-

Figure 23. Mean sea surface height in the Northwest Atlantic Ocean for the period 1993-2017 (a) from altimetry and (b) in the Atlantic HYCOM output. The gray lines in both maps indicate the 1,000-m and the 4,000-m isobaths from ETOPO1.

ber of metrics analyzed at the Oleander Line: maximum Gulf Stream velocity, 15°C isotherm position bias, mean Gulf Stream transport, as well as the highest correlation with observations in net transport and Gulf Stream latitude. It is important to note that the simulation used here is based on the HYCOM ocean model and refined for the North Atlantic, but was not the one used in the model inter-comparison described by Chi et al. (2018).

3.2.2 Lagrangian Analysis

The HYCOM velocity field is used to track the trajectories of synthetic particles launched from the launch site, shown in Fig. 22. The synthetic particles were deployed in the HYCOM reanalysis in offline mode. Among the advantages of this method, we highlight (1) the possibility of running the Lagrangian analysis forward

and backward in time and (2) the small computational cost compared to running new simulations with online synthetic particles. To deploy the particles, we used the open-source code Parcels (<https://oceanparcels.org/>) [24, 25] in Python.

HYCOM’s hybrid vertical configuration allows for a nearly-isopycnal 2-D Lagrangian analysis provided that the particles stay beneath the mixed layer during their entire path. This allows for a faithful representation of isopycnal flow while drastically reducing the computational cost that a fully 3-D analysis would require. We chose to perform the Lagrangian analysis on the model Layer 19 because this layer is shallow enough that the Labrador Sea Water is always beneath it, but deep enough that it is mostly isopycnal and does not outcrop in the region studied here. It was important that the chosen model layer was light enough to contribute to the Slope Water and ultimately influence the shelf properties, whose recent changes motivate this work. Most of the analyses presented here were repeated on Layer 20 and had similar results. While this work focuses on relatively shallow layers that may influence the Slope and Shelf water variability, similar analyses on the Labrador Sea Water (Layers 21-24) and the North Atlantic Deep Water (Layers 25 to bottom) layers could provide insights on the variability of the lower limb of the Atlantic Meridional Overturning Circulation in the Northwest Atlantic.

The typical isopycnal of Layer 19, as referenced to 2,000 dbar, is $\sigma_2 = 36.52 \text{ kg m}^{-3}$, with mean temperature and salinity equal to 6.46°C and 35.09, respectively, in the study region. The model works on σ_2 coordinates, and its typical value for Layer 19 is equivalent to $\sigma_\theta = 27.53 \text{ kg m}^{-3}$ at a mean depth of 570 m. This layer is purely isopycnal in 78.2% of the grid points over the study region showed in Fig. 23 and over all time analyzed (years 2003-2012). The layer outcrops in 19.2% of the grid points, which corresponds to the majority of the continental shelf in the region; here, the surface waters are less dense than those represented by Layer

19, so the layer essentially vanishes. Layer 19 behaves as an isobaric layer (i.e. is within the mixed layer) in only 2.5% of the grid points, mostly corresponding to the area northeastward of the Grand Banks. The spatial distribution of the mean depth and speed of Layer 19 are shown in colors and arrows, respectively, in Fig. 22b. Layer 19 is the shallowest along the continental slope, varying from about 300 m in the eastern Grand Banks to 500 m in the Mid-Atlantic Bight. Its thickness is the largest ($H \geq 80$ m) along the slope between the Tail of the Grand Banks and the Scotian Slope (50°W to 60°W), and farther offshore in the eastern side of the Grand Banks, between 50°W and 40°W .

The synthetic particles were deployed in the Labrador Current on the eastern flank of the Grand Banks (45.5°N , 48.5°W to 45°N , 48°W) between January 2003 and December 2011 and tracked forward in time for one year using the fourth-order Runge-Kutta scheme [26]. This method interpolates the velocity field between consecutive time-steps and therefore is suitable for an offline Lagrangian analyses performed at daily or longer temporal resolution [27]. Eighty particles were simultaneously deployed every 10 days, totaling more than 26,000 particles deployed over 9 years. Sensitivity analyses were performed with identical experiments with 20 and 40 simultaneously-deployed particles at the same location to assure that the run with 80 particles was robust [28], and the results in both cases were statistically indistinguishable from the ones presented below.

3.3 Results and Discussion

3.3.1 Properties of the Lagrangian launch site and mean trajectories

The Lagrangian particles were tracked from a release location in the Labrador Current upstream of the Tail of the Grand Banks. This location was chosen because it is immediately downstream of the region where, on average, the two branches of the Labrador Current (inshore and offshore of Flemish Cap) reassemble and there-

fore the fraction of particles that reach the Tail of the Grand Banks is maximized. Figure 24 shows the time-mean section where the particles were released. The Labrador Current skirts the shelf-break, with a surface-intensified core centered over the 500- and the 1,500-m isobaths. The mean southwestward flow extends to the seafloor with the Labrador Sea Water centered between 500 and 1,000 m. In the top two hundred meters, the Labrador Current is flanked farther offshore by the northeastward-flowing waters of the meandering North Atlantic Current (NAC). The characteristic low temperature and salinity of the Labrador Current are also evident in the mean view of the cross-section. Layer 19 is located at the bottom of the Labrador Current and has, on average, southwestward velocities spanning the entire section. Throughout the domain analyzed in this study, this layer is located immediately above the Labrador Sea Water, which is generally represented by Layers 21-24 [14].

The mean temperature, salinity, velocity and thickness of Layer 19 at the launch site cross-section are, respectively, $4.82 \pm 0.31^{\circ}\text{C}$, 34.88 ± 0.04 , $7.7 \pm 4.3 \text{ m s}^{-1}$ and $63.1 \pm 6.2 \text{ m}$, where the uncertainty is one standard deviation. There is considerable seasonality in all four variables, with temperature and salinity peaking in May ($5.02 \pm 0.25^{\circ}\text{C}$, 34.89 ± 0.04), velocity in November ($10.9 \pm 4.0 \text{ m s}^{-1}$) and thickness in March ($67.5 \pm 9.1 \text{ m}$). There is little interannual variability over the 10 years, with annual mean temperature and salinity ranging from ($4.40 \pm 0.29^{\circ}\text{C}$, 34.84 ± 0.03) to ($5.03 \pm 0.21^{\circ}\text{C}$, 34.92 ± 0.02) in 2009 and 2011, respectively.

Upon deployment, 89% of the particles flowed southward toward the Tail of the Grand Banks (i.e., the latitude ten days after deployment was below that at deployment). The remaining 11% of the particles, that initially flowed northward, were mostly deployed on the offshore edge of the cross-section and likely associated with the meandering of the NAC close to the shelf-break. Subsequently, the

Figure 24. Cross-section of the ten-year mean (a) σ_2 , (b) temperature, (c) salinity and (d) velocity at the launch site LC, in the eastern Grand Banks. The velocity is normal to the section and has been rotated from the original zonal and meridional components, with positive values in the direction of the mean flow of the Labrador Current (i.e., southwestward). The boundaries between the model layers are shown in thin, gray contours, with Layer 19 highlighted with thick, black contours. The section is 68-km long (x-axis), with the continental shelf to its left and the open ocean to its right.

one-year trajectories of the particles launched between 2003 and 2011 were highly dispersive and filled a great area of the Subtropical and Subpolar North Atlantic (Fig. 25). The mean probability map indicates that, within one year of deployment, some synthetic particles reached distant areas, such as the Iceland Basin and central subtropical gyre. The light shades of blue in Fig. 25 (probabilities $< 10\%$) indicate that the particles followed two typical paths: The westward propagation along southern edge of the Grand Banks and the northeastward propagation along the North Atlantic Current.

The black lines in Fig. 25 connect the start and end points of 50 randomly-selected particles that were tracked for one year after their deployment at launch site LC. Although the particles typically departed southward along the shelf-break upon deployment, the vast majority of them were exported near the Tail of the

Grand Banks and, after one year, completed their path to the northeast (45.6%, 12,028 particles) and southeast of the launch site (38.4%, 10,123 particles); only 3,759 particles (14.3%) were to the southwest of the launch site one year after upon deployment. The different lengths of these lines, along with the shape of the distribution map, suggest that the particles that flow northeastward tend to travel farther than those flowing southwestward. In fact, the particles that flowed northeastward, generally carried by the North Atlantic Current, traveled on average $3,923 \pm 798$ km versus only $2,618 \pm 585$ km traveled by the particles that completed their one-year path to the southwest of the launch site.

Figure 25. Probability distribution map of synthetic particles tracked for one year after deployment at launch site LC from 2003 to 2011, calculated at $1^\circ \times 1^\circ$ bins. The black lines represent the start and end points of one-year trajectories of 50 randomly-selected particles. The bathymetry is shown where the probability is equal to zero.

3.3.2 Seasonal and interannual variability in Lagrangian trajectories

The probability distribution maps have a strong seasonal cycle (Fig. 26a). The particles deployed in the spring and summer (March-August) are more likely

to turn toward the northeast upstream of the Tail of the Grand Banks and to travel farther distances, along the North Atlantic Current. Conversely, the probability map of the fall and winter (September-February) deployments indicates that more of these particles follow the Labrador Current westward along the southern edge of the Grand Banks in comparison to those deployed in the spring and summer. It is important to note that the seasonality in the probability distribution of particles is in agreement with the velocity of the Labrador Current at the launch site: higher Labrador Current velocities at the launch site in the Fall and Winter are coupled with higher fractions of particles flowing along the Labrador Current past the Tail of the Grand Banks. Similarly, lower velocities in the Spring and Summer are associated with higher export of particles to the North Atlantic Current.

These Lagrangian trajectories vary dramatically from year to year. Though the 10-year model output is not long enough to allow for inferences about trends, it reveals strong differences between years. The clearest difference occurred for the particles launched in 2004 compared to 2008. In 2004 there were relatively more particles moving westward along the shelf-break, versus more particles moving northeastward along the North Atlantic Current in 2008 (Fig. 26b). The distinct behavior of particles in 2008 as compared to 2004 is expected to result in a decrease of particles that reach the Scotian Slope. To identify the particles that flow along the shelf-break in the Scotian Slope, a terrain-following box is defined in Fig. 26b as the “Slope Region”. This box is defined between 52°W and 55°W , is bounded by the 1,000-m isobath to the north and extends 2° of latitude to the south. The particles that were sampled within this box at least once during the first three months upon deployment were considered “Slope Particles” (Blue lines, Fig. 26c,d). The timing is determined based on the early results that the particles that follow the Labrador Current along the shelf-break generally take two to three

Figure 26. (a) Difference in the probability distribution of synthetic particles deployed between March and August (Spring/Summer) minus those deployed between September and February (Fall/Winter) and tracked for one year, calculated at $1^\circ \times 1^\circ$ bins. (b) Same as (a), but for particles deployed in 2008 minus particles deployed in 2004. The “Slope” and “NAC” boxes are identified in black. In panels (a) and (b) the grid points whose difference was between -1% and 1% were masked out. (c) One-year trajectories of “Slope” (blue) and “NAC Particles” (red) that were launched in 2004, defined as the particles that were sampled in the regions shown as black polygons in Fig. 26b. (d) Same as (c), but for those particles launched in 2008. The definition of “Slope” and “NAC Particles” was the same for both years. In 2008, only four particles met the criteria of entering the “Slope Region” less than three months after being deployed in the 10-day Lagrangian tracking output.

months to round the Tail of the Grand Banks and enter the slope region. The longitude band (52°W to 55°W) was chosen to avoid accounting for particles that are exported to the Gulf Stream just to the west of the Tail of the Grand Banks, near 51°W .

During the 10-year output analyzed here, only 3.5% of the 26,351 particles are

“Slope Particles”. The fraction of particles that met this criteria varied considerably between years, ranging from 6.0% and 10.1% in 2003 and 2004, respectively, to 0.1% in both 2007 and 2008. In the latter years, only 4 out of 5,840 particles were “Slope Particles”. To track a common alternative pathway, we defined a second region south of Flemish Cap, that we call the “NAC Region” (Fig. 26b). The particles that were sampled within this box at least once are called “NAC Particles” (Red lines, Fig. 26c,d). Overall, 26.8% of the particles meet the “NAC Particle” criteria, ranging from 18.5% in 2004 to 31.9% in 2008.

The paths of the “NAC Particles” are visually indistinguishable between the 2004 and 2008. In both cases a small number of particles make a sharp eastward turn slightly west of the Tail of the Grand Banks, near 50°W , while the vast majority of the particles are swept eastward at the Tail of the Grand Banks or just to the north. In contrast, the behavior of the “Slope Particles”, is very different between these two years. In 2004, the “Slope Particles” occupied most of the Slope Sea between the Western Grand Banks and the Gulf of Maine, and invaded the Gulf of St. Lawrence. A small density of particles is observed farther offshore, possibly exported from the Slope Sea by mesoscale activity associated with Gulf Stream rings. In 2008, only three particles were classified as “Slope Particles,” and these never traveled west of 58°W , just offshore of the Laurentian Channel.

3.3.3 An Eulerian view of the interannual variability

Next we investigate the reason why the majority of the particles deployed in the Labrador Current in the eastern Grand Banks are exported at or upstream of the Tail of the Grand Banks. To do so, we use the concept of potential thickness (Fig. 27). Conservation of angular momentum constrains a water parcel of a given potential density to travel along contours of potential thickness of that density layer, in the absence of any external input of vorticity, like from wind stress

curl. The potential thickness is defined as the depth between two isopycnal layers, normalized to a reference latitude: $H = \frac{f \times (D_{lower} - D_{upper})}{f_{ref}}$ where f is the Coriolis parameter, D_{lower} and D_{upper} are the depths of the lower and upper bounding isopycnal, respectively, and f_{ref} is the Coriolis parameter at a reference latitude. The Labrador Current has a strong barotropic component (Figure 24d where the current is seen to extend to the sea floor). Layer 19 sits at the top of the less stratified/more barotropic part of the current. We propose a conceptual model of a two-layer ocean, with Layer 19 dividing a warm layer above from a cold layer below. The warm water layer above Layer 19 is thin north of the Gulf Stream and thick in and south of the Gulf Stream, reaching up to 1,000 m deep. Using the equation above, we define the potential thickness of the deeper layer as the depth between Layer 19 and the seafloor, normalized to 40°N. The deeper part of the Labrador Current, as calculated at the launch site (offshore, on the right-hand side of the sections in Fig. 24), is within that deeper layer with a mean potential thickness contour of 2,300 m.

At the Tail of the Grand Banks, the shallower bathymetry of the Southeast Newfoundland Ridge (SENR) protrudes from the TGB into the open ocean. As the Labrador Current approaches the TGB, the potential thickness contours spread widely to the east along the SENR (shades of brown/yellow in Fig. 27a). The shallowest potential thickness contour that, on average, extends offshore and continuously rounds the SENR is the 3,000 m, which is much deeper than the typical Labrador Current contour of 2,300 m. In contrast, the 2,300-m contour follows the shelf break bathymetry and does not extend out along the SENR. Thus, in the absence of any external input of vorticity, even the Labrador Current particles release in the deeper part of the launch site would be expected to follow the shelf-break past the Grand Banks in order to conserve potential vorticity.

Figure 27. (a) Mean potential thickness from Layer 19 to the bottom of the ocean. The “Slope Region” and the launch site are identified. The 2,300-m and the 3,000-m mean potential thickness contours are shown as thin black lines. (b) Monthly percentage of particles reaching the “Slope Region” within three months of launch (blue bars), plotted as a function of the date of launch between Jan 2003 and Dec 2011, and monthly potential thickness anomaly at the Tail of the Grand Banks (black line, averaged over box in Fig 30,a,b,c). The two time series have the highest correlation coefficient when the TGB potential thickness anomaly lags by one month ($R = 0.70$, $p\text{-value} < 0.001$).

To determine the fraction of particles that round the Tail of the Grand within the Labrador Current and reach the Western Grand Banks, we refer back to the “Slope Region” defined above. For each of the 108 months (Jan 2003 to Dec 2011) during which synthetic particles were deployed, the fraction of those that become “Slope Particles” relative to the total deployed in each month was calculated. Fig. 27b shows the time evolution of this fraction as a function of the month in

which the particles were deployed (blue bars). On average, nearly 3.5% of the particles were sampled in the “Slope Region” less than three months after being deployed, with the highest rates in 2003-2004 and the lowest in 2007-2008. The high rates in 2003-2004 are associated with positive potential thickness anomalies at the Tail of the Grand Banks (black line in Fig. 27b, calculated at the box in Fig 30a,b,c). Similarly, the low rates in 2007-2008 are coupled with negative potential thickness anomalies. The two time series are significantly correlated with the highest correlation coefficient when the percentage of “Slope Particles”, defined by the month of the particle release, leads by one month ($R = 0.70$, $p\text{-value} < 0.001$). This means that the potential thickness anomaly at the Tail of the Grand Banks one month after the release of the particles is tightly related to the to continuity of the Labrador Current.

In 27 out of the 108 months, there were more than 20% of “Slope Particles”, including for all fourteen months between October 2003 and November 2004. After 2006, only five months had more than 20% of “Slope particles”: Aug-Sep/2010 and Aug-Oct/2011. In both cases, there was a short period that lasted between two and three months in which the Gulf Stream meanders were farther offshore, allowing for the Labrador Current to contour the Tail of the Grand Banks. The conditions associated with these highest rates seem to be a function of both the speed of the Labrador Current and the presence/absence of Gulf Stream eddies at the TGB. For example, most of the particles released in November 2003 were exported offshore at the Tail of the Grand Banks in January 2004 by a large meander (not shown). The strong winter Labrador Current, however, still forced nearly 40% of the particles through to the Western Grand Banks.

In contrast, none of the particles released in 51 out of the 108 months reached the “Slope Region” less than three months after deployment. In Fig. 28, three

3-month sequences show examples of events where no particles reached the “Slope Region” for a given deployment month. These periods illustrate the various locations at which Labrador Current particles can be steered offshore by the Gulf Stream or its eddies. Here, the colors represent the potential thickness anomaly, i.e., the deviation of the potential thickness from the mean field shown in Fig. 27a. The tones of green/brown indicate the thinning of the bottom layer (i.e., expansion of the top layer) associated with anticyclonic Gulf Stream meanders and eddies. Alternatively, the tones of blue/cyan indicate the thickening of the bottom layer (i.e., contraction of the top layer), associated with the flow of the Labrador Current. With no exceptions, the particles swept off the shelf break and exported to the northeast at or upstream of the Tail of the Grand Banks during these months always do so on the eastern flank of an anticyclonic meander/eddy (i.e., to the left of the green/brown shades in Fig. 28).

We trace in more detail the three examples of periods of strong Labrador Current blocking shown in Fig 28 to understand the conditions leading to the offshore export of Labrador Current particles. In December 2007 (Fig. 28a), the only large GS meander impinging on the shelf-break was on the western flank of the Grand Banks. The particles deployed early in that month quickly reached the Tail of the Grand Banks. In January 2008 (Fig. 28b), the particles that were already at the Tail of the Grand Banks were steered offshore; simultaneously, a second meander approached the shelf-break in the eastern flank of the Grand Banks and blocked the particles that were deployed in late December. By February 2008 (Fig. 28c), all of the particles deployed in December 2007 had been exported in two locations by two different Gulf Stream meanders. The blocking of the Labrador Current and consequent export of particles to the Gulf Stream and/or the North Atlantic Current may occur at (Fig. 28d-f) or upstream of the Tail of the

Figure 28. Potential Thickness anomaly on three selected months with no “Slope Particles”, tracked for three consecutive months. Each row represents one event: (top) Dec 2007, (middle) Feb 2008 and (bottom) Jan 2010. The columns show the progress of each event from (left) the month of deployment to (right) two months after the deployment. In each map, the trajectories of 15 randomly-selected particles are shown. The trajectory of each particle in a given month is shown in dark red and the trajectory in the previous one (middle column) or two months (right column) are shown in salmon. The 1,000-m and 4,000-m isobaths are shown in thin black contours

Grand Banks (Fig. 28g-i). Despite the variable location at which the particles are exported, these events are always associated with one or more large Gulf Stream meanders and/or eddies impinging on the shelf-break.

The composites of the events with the highest rates of “Slope Particles” and those without “Slope Particles” reveal the spatial patterns associated with the causes and consequences of the Labrador Current cut-off. In Fig. 29, these composites are shown for the second month after the deployment for each case to represent the spatial patterns approximately at the time at which the particles

arrive at the Tail of the Grand Banks rather than at the time at which they are launched. We refer to months with more than 20% of “Slope Particles” as months with an “Open Valve” for the Labrador Current to reach the slope region, and months with zero “Slope Particles” as having a “Closed Valve.” The composites of Potential Thickness anomalies (Fig. 29a,b) show that, in fact, the “Closed Valve” is associated with the squashing of the bottom layer centered at the western flank of the Tail of the Grand Banks. Conversely, the “Open Valve” events are characterized by an anomalously thick bottom layer in the same region (Fig. 29a).

Figure 29. Composite averages of (a, b) potential thickness below Layer 19, (c, d) salinity on Layer 19 and (e, f) velocity of Layer 19. The panels on the left show periods of strong Labrador Current connectivity: they are averaged for months in which 20% or more of the particles reached the “Slope Region” within 3 months. The right hand panels are for months in which no particles reached the “Slope Region”. The averages are for made during the second month after deployment.

Changes in the velocity field help to identify how the altered position of the Gulf Stream and North Atlantic Current meanders/eddies influenced the connectivity of the Labrador Current (Fig. 29e,f). West of the Grand Banks, the Gulf

Stream had, on average, a large anticyclonic meander that reached the shelf-break in “Closed Valve” events, while during “Open Valve” events this region was characterized on average by a cyclonic meander that resulted from recirculation of the Labrador Current. The changes in temperature (not shown) salinity (Fig. 29c,d) and thickness of Layer 19 (Fig. 30d) are a consequence of the continuity of the Labrador Current: the cold and fresh tongue that extends from the western Grand Banks to the Scotian Slope following the “Open Valve” events vanish after “Closed Valve” events. It is important to note that the significant changes in temperature and salinity are not restricted to the shelf, as they extend farther offshore and encompass the Slope Water domain.

3.4 Conclusions

In this study we have performed a Lagrangian analysis in a high resolution ocean circulation model to identify the mechanism responsible for the substantial interannual variability in coherence of the Labrador Current at the Tail of the Grand Banks. The analysis of more than 26,000 synthetic particles revealed that the impingement of Gulf Stream meanders and eddies on the shelf-break at or near the Tail of the Grand Banks may completely cut off the Labrador Current, as they often did after 2007 in the simulation.

A recent observational study (Gonçalves Neto et al., in review) has shown that an abrupt shift to higher sea level at the Tail of the Grand Banks in 2008 preceded a rapid subsurface warming that propagated along the Northwest Atlantic Slope between the Gulf of St. Lawrence and the Gulf of Maine in 2009 and 2010. The simulated Lagrangian analysis described here provides a mechanistic explanation for this event: the strong positive sea surface height anomalies at the Tail of the Grand Banks, observed by satellite altimetry and corroborated by the HYCOM simulation (Fig. 30a) are associated with anticyclonic Gulf Stream meanders and

eddies that completely block the Labrador Current. In addition to the doming ocean surface, anticyclones are characterized by the deepening of the isopycnals, illustrated by the increase of the depth of Layer 19 (Fig. 30b). Picturing the water column as a two-layer system, the deepening of the interface between the two layers implies the squashing of the bottom layer (Fig. 30c). At the Tail of the Grand Banks, using the depth of the model Layer 19 as a proxy for the boundary between the two layers, the deepening isopycnals decrease the potential thickness of the bottom layer. To follow the potential thickness contours and conserve angular momentum, the Labrador Current is steered offshore when Gulf Stream meanders and eddies are present.

Figure 30. Difference between the months with no “Slope Particles” and the months with 20% or more of “Slope Particles”. (a) SSH, (b) depth of Layer 19, (c) potential thickness and (d) thickness of Layer 19. The 200 m isobath is shown in black to show the approximate position of the shelf break.

Although it is not our objective to determine the causes of the increased impingement of the Gulf Stream at the Tail of the Grand Banks, past theoretical and observational studies may provide useful connections to our findings. First, the Gulf Stream has undergone significant changes in recent years between the New England Sea Mounts (near 70°W) and the Grand Banks: It has slowed down, with a particular weakening of the cyclonic flank to its north, in the Slope Sea [29],

and its destabilization point has migrated westward, with a surge in Gulf Stream troughs and deep cyclones at Line W since 2008 [30]. The increased destabilization of the Gulf Stream and the weakening in its recirculation could have allowed for more eddies reaching the shelf-break at the TGB and consequently decreased the connectivity of the Labrador Current to the Northwest Atlantic.

The control of the Gulf Stream on the connectivity of the Labrador Current at the TGB likely plays a role on the strength and/or pathways of the cold limb of the AMOC at this latitude. Therefore, the continuity of the Labrador Current at the TGB could be associated with the coupling between the Northwest Atlantic Shelf warming and the AMOC slowing as is simulated in a high resolution model [31], and is expected to continue during this century under continuous greenhouse gas emissions [32]. Extending this Lagrangian analysis to deeper layers within the Labrador Sea Water and the overflow waters may provide insight on the connection between Northwest Atlantic Shelf warming and the AMOC and could help reveal the degree to which the offshore diversion of the Labrador Current leads to an increased importance of interior pathways for the cold branch of the AMOC [33].

List of References

- [1] A. J. Pershing, M. A. Alexander, C. M. Hernandez, L. A. Kerr, A. Le Bris, K. E. Mills, J. A. Nye, N. R. Record, H. A. Scannell, J. D. Scott, G. D. Sherwood, and A. C. Thomas, “Slow adaptation in the face of rapid warming leads to collapse of the Gulf of Maine cod fishery,” *Science*, vol. 350, no. 6262, pp. 809–812, Nov. 2015. [Online]. Available: <https://www.sciencemag.org/lookup/doi/10.1126/science.aac9819>
- [2] K. Mills, A. Pershing, C. Brown, Y. Chen, F.-S. Chiang, D. Holland, S. Lehuta, J. Nye, J. Sun, A. Thomas, and R. Wahle, “Fisheries Management in a Changing Climate: Lessons From the 2012 Ocean Heat Wave in the Northwest Atlantic,” *Oceanography*, vol. 26, no. 2, 2013. [Online]. Available: <https://tos.org/oceanography/article/fisheries-management-in-a-changing-climate-lessonsfrom-the-2012-ocean-heat->
- [3] A. Pershing, K. Mills, A. Dayton, B. Franklin, and B. Kennedy,

- “Evidence for Adaptation from the 2016 Marine Heatwave in the Northwest Atlantic Ocean,” *Oceanography*, vol. 31, no. 2, pp. 151–161, June 2018. [Online]. Available: <https://tos.org/oceanography/article/evidence-for-adaptation-from-the-2016-marine-heatwave-in-the-northwest-atl>
- [4] D. Brickman, D. Hebert, and Z. Wang, “Mechanism for the recent ocean warming events on the Scotian Shelf of eastern Canada,” *Continental Shelf Research*, vol. 156, pp. 11–22, Mar. 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0278434317302650>
- [5] M. Claret, E. D. Galbraith, J. B. Palter, D. Bianchi, K. Fennel, D. Gilbert, and J. P. Dunne, “Rapid coastal deoxygenation due to ocean circulation shift in the northwest Atlantic,” *Nature Climate Change*, vol. 8, no. 10, pp. 868–872, Oct. 2018. [Online]. Available: <http://www.nature.com/articles/s41558-018-0263-1>
- [6] A. S. Bower, M. S. Lozier, S. F. Gary, and C. W. Böning, “Interior pathways of the North Atlantic meridional overturning circulation,” *Nature*, vol. 459, no. 7244, pp. 243–247, May 2009. [Online]. Available: <http://www.nature.com/articles/nature07979>
- [7] A. Bower, S. Lozier, and S. Gary, “Export of Labrador Sea Water from the subpolar North Atlantic: A Lagrangian perspective,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17–18, pp. 1798–1818, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000233>
- [8] H. Stommel and A. Arons, “On the abyssal circulation of the world ocean—I. Stationary planetary flow patterns on a sphere,” *Deep Sea Research (1953)*, vol. 6, pp. 140–154, Jan. 1959. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/0146631359900656>
- [9] T. C. Biló and W. E. Johns, “Interior Pathways of Labrador Sea Water in the North Atlantic From the Argo Perspective,” *Geophysical Research Letters*, vol. 46, no. 6, pp. 3340–3348, Mar. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL081439>
- [10] S. F. Gary, M. Susan Lozier, C. W. Böning, and A. Biastoch, “Deciphering the pathways for the deep limb of the Meridional Overturning Circulation,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17–18, pp. 1781–1797, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000221>
- [11] M. S. Lozier, S. F. Gary, and A. S. Bower, “Simulated pathways of the overflow waters in the North Atlantic: Subpolar to subtropical export,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 85, pp. 147–153,

- Jan. 2013. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064512001233>
- [12] A. Solodoch, J. C. McWilliams, A. L. Stewart, J. Gula, and L. Renault, “Why Does the Deep Western Boundary Current “Leak” around Flemish Cap?” *Journal of Physical Oceanography*, vol. 50, no. 7, pp. 1989–2016, July 2020. [Online]. Available: <https://journals.ametsoc.org/jpo/article/50/7/1989/347277/Why-Does-the-Deep-Western-Boundary-Current-Leak>
 - [13] L. Chi, C. L. Wolfe, and S. Hameed, “Intercomparison of the Gulf Stream in ocean reanalyses: 1993–2010,” *Ocean Modelling*, vol. 125, pp. 1–21, May 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500318300568>
 - [14] X. Xu, H. E. Hurlburt, W. J. Schmitz, R. Zantopp, J. Fischer, and P. J. Hogan, “On the currents and transports connected with the atlantic meridional overturning circulation in the subpolar North Atlantic: TRANSPORTS IN THE SPNA,” *Journal of Geophysical Research: Oceans*, vol. 118, no. 1, pp. 502–516, Jan. 2013. [Online]. Available: <http://doi.wiley.com/10.1002/jgrc.20065>
 - [15] R. Bleck, “An oceanic general circulation model framed in hybrid isopycnic-Cartesian coordinates,” *Ocean Modelling*, vol. 4, no. 1, pp. 55–88, Jan. 2002. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500301000129>
 - [16] E. P. Chassignet, L. T. Smith, G. R. Halliwell, and R. Bleck, “North Atlantic Simulations with the Hybrid Coordinate Ocean Model (HYCOM): Impact of the Vertical Coordinate Choice, Reference Pressure, and Thermobaricity,” *JOURNAL OF PHYSICAL OCEANOGRAPHY*, vol. 33, p. 23, 2003.
 - [17] G. R. Halliwell, “Evaluation of vertical coordinate and vertical mixing algorithms in the HYbrid-Coordinate Ocean Model (HYCOM),” *Ocean Modelling*, vol. 7, no. 3–4, pp. 285–322, Jan. 2004. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500303000556>
 - [18] X. Xu, A. Bower, H. Furey, and E. P. Chassignet, “Variability of the Iceland-Scotland Overflow Water Transport Through the Charlie-Gibbs Fracture Zone: Results From an Eddying Simulation and Observations,” *Journal of Geophysical Research: Oceans*, vol. 123, no. 8, pp. 5808–5823, Aug. 2018. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018JC013895>
 - [19] A. Gonçalves Neto, J. B. Palter, A. Bower, H. Furey, and X. Xu, “Labrador Sea Water Transport Across the Charlie-Gibbs Fracture Zone,” *Journal of Geophysical Research: Oceans*, vol. 125, no. 8, Aug. 2020. [Online]. Available: <https://onlinelibrary.wiley.com/doi/10.1029/2020JC016068>

- [20] F. A. Schott, R. Zantopp, L. Stramma, M. Dengler, J. R. Fischer, and M. Wibaux, "Circulation and Deep-Water Export at the Western Exit of the Subpolar North Atlantic," *JOURNAL OF PHYSICAL OCEANOGRAPHY*, vol. 34, p. 27, 2004.
- [21] S. M. Uppala, P. W. Kllberg, A. J. Simmons, U. Andrae, V. D. C. Bechtold, M. Fiorino, J. K. Gibson, J. Haseler, A. Hernandez, G. A. Kelly, X. Li, K. Onogi, S. Saarinen, N. Sokka, R. P. Allan, E. Andersson, K. Arpe, M. A. Balmaseda, A. C. M. Beljaars, L. V. D. Berg, J. Bidlot, N. Bormann, S. Caires, F. Chevallier, A. Dethof, M. Dragosavac, M. Fisher, M. Fuentes, S. Hagemann, E. Hlm, B. J. Hoskins, L. Isaksen, P. A. E. M. Janssen, R. Jenne, A. P. McNally, J.-F. Mahfouf, J.-J. Morcrette, N. A. Rayner, R. W. Saunders, P. Simon, A. Sterl, K. E. Trenberth, A. Untch, D. Vasiljevic, P. Viterbo, and J. Woollen, "The ERA-40 re-analysis," *Quarterly Journal of the Royal Meteorological Society*, vol. 131, no. 612, pp. 2961–3012, Oct. 2005. [Online]. Available: <http://doi.wiley.com/10.1256/qj.04.176>
- [22] Rosmond, T, Teixeira, J, Peng, M, Hogan, T, and Pauley, R, "Navy operational global atmospheric prediction system (NOGAPS): Forcing for ocean models," *Oceanography*, vol. 15, no. 1, pp. 99–108, 2002. [Online]. Available: <https://doi.org/10.5670/oceanog.2002.40>
- [23] E. P. Chassignet and D. P. Marshall, "Gulf Stream separation in numerical ocean models," in *Geophysical Monograph Series*, M. W. Hecht and H. Hasumi, Eds. Washington, D. C.: American Geophysical Union, 2008, vol. 177, pp. 39–61. [Online]. Available: <http://doi.wiley.com/10.1029/177GM05>
- [24] P. Delandmeter and E. van Sebille, "The Parcels v2.0 Lagrangian framework: new field interpolation schemes," *Geoscientific Model Development*, vol. 12, no. 8, pp. 3571–3584, Aug. 2019. [Online]. Available: <https://gmd.copernicus.org/articles/12/3571/2019/>
- [25] M. Lange and E. van Sebille, "Parcels v0.9: prototyping a Lagrangian ocean analysis framework for the petascale age," *Geoscientific Model Development*, vol. 10, no. 11, pp. 4175–4186, Nov. 2017. [Online]. Available: <https://gmd.copernicus.org/articles/10/4175/2017/>
- [26] J. Dormand and P. Prince, "A family of embedded Runge-Kutta formulae," *Journal of Computational and Applied Mathematics*, vol. 6, no. 1, pp. 19–26, Mar. 1980. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/0771050X80900133>
- [27] E. van Sebille, S. M. Griffies, R. Abernathey, T. P. Adams, P. Berloff, A. Biastoch, B. Blanke, E. P. Chassignet, Y. Cheng, C. J. Cotter, E. Deleersnijder, K. Ds, H. F. Drake, S. Drijfhout, S. F. Gary, A. W. Heemink, J. Kjellsson, I. M. Koszalka, M. Lange, C. Lique,

- G. A. MacGilchrist, R. Marsh, C. G. Mayorga Adame, R. McAdam, F. Nencioli, C. B. Paris, M. D. Piggott, J. A. Polton, S. Rhs, S. H. Shah, M. D. Thomas, J. Wang, P. J. Wolfram, L. Zanna, and J. D. Zika, “Lagrangian ocean analysis: Fundamentals and practices,” *Ocean Modelling*, vol. 121, pp. 49–75, Jan. 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500317301853>
- [28] B. T. Jones, A. Solow, and R. Ji, “Resource Allocation for Lagrangian Tracking,” *Journal of Atmospheric and Oceanic Technology*, vol. 33, no. 6, pp. 1225–1235, June 2016. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JTECH-D-15-0115.1>
- [29] M. Andres, K. A. Donohue, and J. M. Toole, “The Gulf Stream’s path and time-averaged velocity structure and transport at 68.5W and 70.3W,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 156, p. 103179, Feb. 2020. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063719303255>
- [30] M. Andres, “On the recent destabilization of the Gulf Stream path downstream of Cape Hatteras: Gulf Stream Path Destabilization,” *Geophysical Research Letters*, vol. 43, no. 18, pp. 9836–9842, Sept. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016GL069966>
- [31] L. Caesar, S. Rahmstorf, A. Robinson, G. Feulner, and V. Saba, “Observed fingerprint of a weakening Atlantic Ocean overturning circulation,” *Nature*, vol. 556, no. 7700, pp. 191–196, Apr. 2018. [Online]. Available: <http://www.nature.com/articles/s41586-018-0006-5>
- [32] V. S. Saba, S. M. Griffies, W. G. Anderson, M. Winton, M. A. Alexander, T. L. Delworth, J. A. Hare, M. J. Harrison, A. Rosati, G. A. Vecchi, and R. Zhang, “Enhanced warming of the Northwest Atlantic Ocean under climate change,” *Journal of Geophysical Research: Oceans*, vol. 121, no. 1, pp. 118–132, 2015. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015JC011346>
- [33] J. Koelling, U. Send, and M. Lankhorst, “Decadal Strengthening of Interior Flow of North Atlantic Deep Water Observed by GRACE Satellites,” *Journal of Geophysical Research: Oceans*, vol. 125, no. 11, Nov. 2020. [Online]. Available: <https://onlinelibrary.wiley.com/doi/10.1029/2020JC016217>

BIBLIOGRAPHY

- , N. M. F. S., “Fisheries Economics of the United States 2016,” U.S. Dept. of Commerce, NOAA Tech. Memo. NMFS-F/SPO-187, U.S. Dept. of Commerce, NOAA Tech. Memo. NMFS-F/SPO-187, Tech. Rep., 2018.
- Andres, M., “On the recent destabilization of the Gulf Stream path downstream of Cape Hatteras: Gulf Stream Path Destabilization,” *Geophysical Research Letters*, vol. 43, no. 18, pp. 9836–9842, Sept. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016GL069966>
- Andres, M., Donohue, K. A., and Toole, J. M., “The Gulf Stream’s path and time-averaged velocity structure and transport at 68.5W and 70.3W,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 156, p. 103179, Feb. 2020. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063719303255>
- Azetsu-Scott, K., “Time series study of CFC concentrations in the Labrador Sea during deep and shallow convection regimes (1991–2000),” *Journal of Geophysical Research*, vol. 108, no. C11, p. 3354, 2003. [Online]. Available: <http://doi.wiley.com/10.1029/2002JC001317>
- Biló, T. C. and Johns, W. E., “Interior Pathways of Labrador Sea Water in the North Atlantic From the Argo Perspective,” *Geophysical Research Letters*, vol. 46, no. 6, pp. 3340–3348, Mar. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL081439>
- Bleck, R., “An oceanic general circulation model framed in hybrid isopycnic-Cartesian coordinates,” *Ocean Modelling*, vol. 4, no. 1, pp. 55–88, Jan. 2002. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500301000129>
- Bower, A. S., Le Cann, B., Rossby, T., Zenk, W., Gould, J., Speer, K., Richardson, P. L., Prater, M. D., and Zhang, H.-M., “Directly measured mid-depth circulation in the northeastern North Atlantic Ocean,” *Nature*, vol. 419, no. 6907, pp. 603–607, Oct. 2002. [Online]. Available: <http://www.nature.com/articles/nature01078>
- Bower, A. and Furey, H., “Iceland-Scotland Overflow Water transport variability through the Charlie-Gibbs Fracture Zone and the impact of the North Atlantic Current: ISOW TRANSPORT THROUGH THE CGFZ,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 9, pp. 6989–7012, Sept. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2017JC012698>

- Bower, A., Lozier, S., and Gary, S., “Export of Labrador Sea Water from the subpolar North Atlantic: A Lagrangian perspective,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17-18, pp. 1798–1818, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000233>
- Bower, A. S., Lozier, M. S., Gary, S. F., and Böning, C. W., “Interior pathways of the North Atlantic meridional overturning circulation,” *Nature*, vol. 459, no. 7244, pp. 243–247, May 2009. [Online]. Available: <http://www.nature.com/articles/nature07979>
- Bower, A. S. and von Appen, W.-J., “Interannual Variability in the Pathways of the North Atlantic Current over the Mid-Atlantic Ridge and the Impact of Topography,” *Journal of Physical Oceanography*, vol. 38, no. 1, pp. 104–120, Jan. 2008. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/2007JPO3686.1>
- Breckenfelder, T., Rhein, M., Roessler, A., Böning, C. W., Biastoch, A., Behrens, E., and Mertens, C., “Flow paths and variability of the North Atlantic Current: A comparison of observations and a high-resolution model: NAC IN OBSERVATION AND MODEL,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 4, pp. 2686–2708, Apr. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012444>
- Brickman, D., Hebert, D., and Wang, Z., “Mechanism for the recent ocean warming events on the Scotian Shelf of eastern Canada,” *Continental Shelf Research*, vol. 156, pp. 11–22, Mar. 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0278434317302650>
- Buckley, M. W. and Marshall, J., “Observations, inferences, and mechanisms of the Atlantic Meridional Overturning Circulation: A review,” *Reviews of Geophysics*, vol. 54, no. 1, pp. 5–63, Mar. 2016. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015RG000493>
- Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G., and Saba, V., “Observed fingerprint of a weakening Atlantic Ocean overturning circulation,” *Nature*, vol. 556, no. 7700, pp. 191–196, Apr. 2018. [Online]. Available: <http://www.nature.com/articles/s41586-018-0006-5>
- Chan, K.-s. and Ripley, B., “TSA: Time series analysis. R package version 1.2,” 2018. [Online]. Available: <https://CRAN.R-project.org/package=TSA>
- Chassignet, E. P. and Marshall, D. P., “Gulf Stream separation in numerical ocean models,” in *Geophysical Monograph Series*, Hecht, M. W. and Hasumi, H., Eds. Washington, D. C.: American Geophysical Union, 2008, vol. 177, pp. 39–61. [Online]. Available: <http://doi.wiley.com/10.1029/177GM05>

- Chassignet, E. P., Smith, L. T., Halliwell, G. R., and Bleck, R., “North Atlantic Simulations with the Hybrid Coordinate Ocean Model (HYCOM): Impact of the Vertical Coordinate Choice, Reference Pressure, and Thermobaricity,” *JOURNAL OF PHYSICAL OCEANOGRAPHY*, vol. 33, p. 23, 2003.
- Chi, L., Wolfe, C. L., and Hameed, S., “Intercomparison of the Gulf Stream in ocean reanalyses: 1993–2010,” *Ocean Modelling*, vol. 125, pp. 1–21, May 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500318300568>
- Claret, M., Galbraith, E. D., Palter, J. B., Bianchi, D., Fennel, K., Gilbert, D., and Dunne, J. P., “Rapid coastal deoxygenation due to ocean circulation shift in the northwest Atlantic,” *Nature Climate Change*, vol. 8, no. 10, pp. 868–872, Oct. 2018. [Online]. Available: <http://www.nature.com/articles/s41558-018-0263-1>
- Cramer, F., “Scientific Colour-Maps: Perceptually uniform and colour-blind friendly,” 2018. [Online]. Available: <https://zenodo.org/record/3596401#.Xwfc0ZKNKgWo>
- Cunningham, S. A. and Haine, T. W., “Labrador Sea Water in the Eastern North Atlantic. Part I: A Synoptic Circulation Inferred from a Minimum in Potential Vorticity,” *Journal of Physical Oceanography*, vol. 25, pp. 649–665, 1995.
- Curry, R. G., McCartney, M. S., and Joyce, T. M., “Oceanic transport of subpolar climate signals to mid-depth subtropical waters,” *Nature*, vol. 391, no. 6667, pp. 575–577, Feb. 1998. [Online]. Available: <http://www.nature.com/articles/35356>
- Davis, R. E., “Preliminary results from directly measuring middepth circulation in the tropical and South Pacific,” *Journal of Geophysical Research: Oceans*, vol. 103, no. C11, pp. 24 619–24 639, Oct. 1998. [Online]. Available: <http://doi.wiley.com/10.1029/98JC01913>
- Davis, X. J., Joyce, T. M., and Kwon, Y.-O., “Prediction of silver hake distribution on the Northeast U.S. shelf based on the Gulf Stream path index,” *Continental Shelf Research*, vol. 138, pp. 51–64, Apr. 2017. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S027843431630067X>
- Delandmeter, P. and van Sebille, E., “The Parcels v2.0 Lagrangian framework: new field interpolation schemes,” *Geoscientific Model Development*, vol. 12, no. 8, pp. 3571–3584, Aug. 2019. [Online]. Available: <https://gmd.copernicus.org/articles/12/3571/2019/>
- Desbruyères, D., Thierry, V., and Mercier, H., “Simulated decadal variability of the meridional overturning circulation across the A25–Ovide section: DECADAL VARIABILITY OF MERIDIONAL OVERTURNING CIRCULATION,”

- Journal of Geophysical Research: Oceans*, vol. 118, no. 1, pp. 462–475, Jan. 2013. [Online]. Available: <http://doi.wiley.com/10.1029/2012JC008342>
- Dong, S., Baringer, M. O., and Goni, G. J., “Slow Down of the Gulf Stream during 1993–2016,” *Scientific Reports*, vol. 9, no. 1, p. 6672, Dec. 2019. [Online]. Available: <http://www.nature.com/articles/s41598-019-42820-8>
- Dormand, J. and Prince, P., “A family of embedded Runge-Kutta formulae,” *Journal of Computational and Applied Mathematics*, vol. 6, no. 1, pp. 19–26, Mar. 1980. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/0771050X80900133>
- Dubik, B. A., Clark, E. C., Young, T., Zigler, S. B. J., Provost, M. M., Pinsky, M. L., and St. Martin, K., “Governing fisheries in the face of change: Social responses to long-term geographic shifts in a U.S. fishery,” *Marine Policy*, vol. 99, pp. 243–251, Jan. 2019. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0308597X18305463>
- Dunn, J. and Ridgway, K., “Mapping ocean properties in regions of complex topography,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 49, no. 3, pp. 591–604, Mar. 2002. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063701000693>
- Eakins, B. E. and Amante, C., “ETOPO1 1 arc-minute global relief model: procedures, data sources and analysis,” National Geophysical Data Center, Boulder, Colorado, Tech. Rep. NESDIS NGDC-24, 2009.
- Fischer, J., Visbeck, M., Zantopp, R., and Nunes, N., “Interannual to decadal variability of outflow from the Labrador Sea: LABRADOR SEA OUTFLOW VARIABILITY,” *Geophysical Research Letters*, vol. 37, no. 24, pp. n/a–n/a, Dec. 2010. [Online]. Available: <http://doi.wiley.com/10.1029/2010GL045321>
- Forsyth, J. S. T., Andres, M., and Gawarkiewicz, G. G., “Recent accelerated warming of the continental shelf off New Jersey: Observations from the CMVOleander expendable bathythermograph line,” *Journal of Geophysical Research: Oceans*, vol. 120, no. 3, pp. 2370–2384, Mar. 2015. [Online]. Available: <http://doi.wiley.com/10.1002/2014JC010516>
- Fratantoni, P. S. and McCartney, M. S., “Freshwater export from the Labrador Current to the North Atlantic Current at the Tail of the Grand Banks of Newfoundland,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 57, no. 2, pp. 258–283, Feb. 2010. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063709002222>
- Fratantoni, P. S. and Pickart, R. S., “The Western North Atlantic Shelfbreak Current System in Summer,” *Journal of Physical Oceanography*,

- vol. 37, no. 10, pp. 2509–2533, Oct. 2007. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO3123.1>
- Furey, H. H., Trafford, L., and Bower, A. S., *A crossroads of the Atlantic Meridional Overturning Circulation : the Charlie-Gibbs Fracture Zone data report August 2010 June 2012*. Woods Hole, MA: Woods Hole Oceanographic Institution, 2014. [Online]. Available: <https://hdl.handle.net/1912/6846>
- Gaichas, S. K., Link, J. S., and Hare, J. A., “A risk-based approach to evaluating northeast US fish community vulnerability to climate change,” *ICES Journal of Marine Science*, vol. 71, no. 8, pp. 2323–2342, Oct. 2014. [Online]. Available: <https://academic.oup.com/icesjms/article/71/8/2323/758638>
- Gangopadhyay, A., Gawarkiewicz, G., Silva, E. N. S., Monim, M., and Clark, J., “An Observed Regime Shift in the Formation of Warm Core Rings from the Gulf Stream,” *Scientific Reports*, vol. 9, no. 1, p. 12319, Dec. 2019. [Online]. Available: <http://www.nature.com/articles/s41598-019-48661-9>
- Gary, S. F., Susan Lozier, M., Böning, C. W., and Biastoch, A., “Deciphering the pathways for the deep limb of the Meridional Overturning Circulation,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17-18, pp. 1781–1797, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000221>
- Gawarkiewicz, G. G., Todd, R. E., Plueddemann, A. J., Andres, M., and Manning, J. P., “Direct interaction between the Gulf Stream and the shelfbreak south of New England,” *Scientific Reports*, vol. 2, no. 1, p. 553, Dec. 2012. [Online]. Available: <http://www.nature.com/articles/srep00553>
- Gebbie, G. and Huybers, P., “How is the ocean filled?: HOW IS THE OCEAN FILLED?” *Geophysical Research Letters*, vol. 38, no. 6, pp. n/a–n/a, Mar. 2011. [Online]. Available: <http://doi.wiley.com/10.1029/2011GL046769>
- Gilbert, D., Sundby, B., Gobeil, C., Mucci, A., and Tremblay, G.-H., “A seventy-two-year record of diminishing deep-water oxygen in the St. Lawrence estuary: The northwest Atlantic connection,” *Limnology and Oceanography*, vol. 50, no. 5, pp. 1654–1666, Sept. 2005. [Online]. Available: <http://doi.wiley.com/10.4319/lo.2005.50.5.1654>
- Gonçalves Neto, A., Palter, J. B., Bower, A., Furey, H., and Xu, X., “Labrador Sea Water Transport Across the Charlie-Gibbs Fracture Zone,” *Journal of Geophysical Research: Oceans*, vol. 125, no. 8, Aug. 2020. [Online]. Available: <https://onlinelibrary.wiley.com/doi/10.1029/2020JC016068>
- Good, S. A., Martin, M. J., and Rayner, N. A., “EN4: Quality controlled ocean temperature and salinity profiles and monthly objective analyses with uncertainty estimates: THE EN4 DATA SET,” *Journal of Geophysical*

- Research: Oceans*, vol. 118, no. 12, pp. 6704–6716, Dec. 2013. [Online]. Available: <http://doi.wiley.com/10.1002/2013JC009067>
- Gouretski, V. and Reseghetti, F., “On depth and temperature biases in bathythermograph data: Development of a new correction scheme based on analysis of a global ocean database,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 57, no. 6, pp. 812–833, June 2010. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063710000671>
- Grodsky, S. A., Reul, N., Chapron, B., Carton, J. A., and Bryan, F. O., “Interannual surface salinity on Northwest Atlantic shelf: INTERANNUAL SSS IN NW ATLANTIC SHELF,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 5, pp. 3638–3659, May 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012580>
- Halliwel, G. R., “Evaluation of vertical coordinate and vertical mixing algorithms in the HYbrid-Coordinate Ocean Model (HYCOM),” *Ocean Modelling*, vol. 7, no. 3-4, pp. 285–322, Jan. 2004. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500303000556>
- Hare, J. A., Morrison, W. E., Nelson, M. W., Stachura, M. M., Teeters, E. J., Griffis, R. B., Alexander, M. A., Scott, J. D., Alade, L., Bell, R. J., Chute, A. S., Curti, K. L., Curtis, T. H., Kircheis, D., Kocik, J. F., Lucey, S. M., McCandless, C. T., Milke, L. M., Richardson, D. E., Robillard, E., Walsh, H. J., McManus, M. C., Marancik, K. E., and Griswold, C. A., “A Vulnerability Assessment of Fish and Invertebrates to Climate Change on the Northeast U.S. Continental Shelf,” *PLOS ONE*, vol. 11, no. 2, p. e0146756, Feb. 2016. [Online]. Available: <https://dx.plos.org/10.1371/journal.pone.0146756>
- Hyndman, R., Athanasopoulos, G., Bergmeir, C., Caceres, G., Chhay, L., O’Hara-Wild, M., Petropoulos, F., Razbash, S., Wang, E., and Yasmeen, F., “forecast: Forecasting functions for time series and linear models. R package version 8.10,” 2019. [Online]. Available: <https://pkg.robjhyndman.com/forecast/>
- IIP, “Report of the International Ice Patrol in the North Atlantic (2019 Season),” CG-188-74, US Coast Guard Bulletin 105, 2019.
- Jones, B. T., Solow, A., and Ji, R., “Resource Allocation for Lagrangian Tracking,” *Journal of Atmospheric and Oceanic Technology*, vol. 33, no. 6, pp. 1225–1235, June 2016. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JTECH-D-15-0115.1>
- Khatiwala, S., Primeau, F., and Hall, T., “Reconstruction of the history of anthropogenic CO₂ concentrations in the ocean,” *Nature*, vol. 462, no. 7271, pp. 346–349, Nov. 2009. [Online]. Available: <http://www.nature.com/articles/nature08526>

- Khatiwala, S., Tanhua, T., Mikaloff Fletcher, S., Gerber, M., Doney, S. C., Graven, H. D., Gruber, N., McKinley, G. A., Murata, A., Ríos, A. F., and Sabine, C. L., “Global ocean storage of anthropogenic carbon,” *Biogeosciences*, vol. 10, no. 4, pp. 2169–2191, Apr. 2013. [Online]. Available: <https://www.biogeosciences.net/10/2169/2013/>
- Kieke, D. and Yashayaev, I., “Studies of Labrador Sea Water formation and variability in the subpolar North Atlantic in the light of international partnership and collaboration,” *Progress in Oceanography*, vol. 132, pp. 220–232, Mar. 2015. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0079661114002201>
- Koelling, J., Send, U., and Lankhorst, M., “Decadal Strengthening of Interior Flow of North Atlantic Deep Water Observed by GRACE Satellites,” *Journal of Geophysical Research: Oceans*, vol. 125, no. 11, Nov. 2020. [Online]. Available: <https://onlinelibrary.wiley.com/doi/10.1029/2020JC016217>
- Lange, M. and van Sebille, E., “Parcels v0.9: prototyping a Lagrangian ocean analysis framework for the petascale age,” *Geoscientific Model Development*, vol. 10, no. 11, pp. 4175–4186, Nov. 2017. [Online]. Available: <https://gmd.copernicus.org/articles/10/4175/2017/>
- Lauzier, L. M., “Long-term temperature variations in the Scotian Shelf area,” *ICNAF Spec. Publ.*, vol. 6, pp. 807–816, 1965.
- Lavender, K. L., Brechner Owens, W., and Davis, R. E., “The mid-depth circulation of the subpolar North Atlantic Ocean as measured by subsurface floats,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 52, no. 5, pp. 767–785, May 2005. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063705000154>
- Lavender, K. L., Davis, R. E., and Owens, W. B., “Mid-depth recirculation observed in the interior Labrador and Irminger seas by direct velocity measurements,” *Nature*, vol. 407, no. 6800, pp. 66–69, Sept. 2000. [Online]. Available: <http://www.nature.com/articles/35024048>
- Le Bras, I. A., Yashayaev, I., and Toole, J. M., “Tracking Labrador Sea Water property signals along the Deep Western Boundary Current,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 7, pp. 5348–5366, July 2017. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017JC012921>
- Le Bris, A., Mills, K. E., Wahle, R. A., Chen, Y., Alexander, M. A., Allyn, A. J., Schuetz, J. G., Scott, J. D., and Pershing, A. J., “Climate vulnerability and resilience in the most valuable North American fishery,” *Proceedings of the National Academy of Sciences*, vol. 115, no. 8, pp. 1831–1836, Feb.

2018. [Online]. Available: <http://www.pnas.org/lookup/doi/10.1073/pnas.1711122115>
- Lillibridge, J. L. and Mariano, A. J., “A statistical analysis of Gulf Stream variability from 18+ years of altimetry data,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 85, pp. 127–146, Jan. 2013. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064512001208>
- Loder, J. W., Petrie, B., and Gawarkiewicz, G. G., “The coastal ocean off north-eastern North America: A large-scale view,” in *The Sea 11*. New York: Wiley, 1998, pp. 105–133.
- Lozier, M. S., Gary, S. F., and Bower, A. S., “Simulated pathways of the overflow waters in the North Atlantic: Subpolar to subtropical export,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 85, pp. 147–153, Jan. 2013. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064512001233>
- McCartney, M., “Recirculating components to the deep boundary current of the northern North Atlantic,” *Progress in Oceanography*, vol. 29, no. 4, pp. 283–383, Jan. 1992. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/007966119290006L>
- Mills, K., Pershing, A., Brown, C., Chen, Y., Chiang, F.-S., Holland, D., Lehuta, S., Nye, J., Sun, J., Thomas, A., and Wahle, R., “Fisheries Management in a Changing Climate: Lessons From the 2012 Ocean Heat Wave in the Northwest Atlantic,” *Oceanography*, vol. 26, no. 2, 2013. [Online]. Available: <https://tos.org/oceanography/article/fisheries-management-in-a-changing-climate-lessonsfrom-the-2012-ocean-heat->
- Nye, J., Link, J., Hare, J., and Overholtz, W., “Changing spatial distribution of fish stocks in relation to climate and population size on the Northeast United States continental shelf,” *Marine Ecology Progress Series*, vol. 393, pp. 111–129, Oct. 2009. [Online]. Available: <http://www.int-res.com/abstracts/meps/v393/p111-129/>
- Nye, J. A., Joyce, T. M., Kwon, Y.-O., and Link, J. S., “Silver hake tracks changes in Northwest Atlantic circulation,” *Nature Communications*, vol. 2, no. 1, p. 412, Sept. 2011. [Online]. Available: <http://www.nature.com/articles/ncomms1420>
- Oremus, K. L., “Climate variability reduces employment in New England fisheries,” *Proceedings of the National Academy of Sciences*, vol. 116, no. 52, pp. 26 444–26 449, Dec. 2019. [Online]. Available: <http://www.pnas.org/lookup/doi/10.1073/pnas.1820154116>

- Paillet, J., Arhan, M., and McCartney, M. S., "Spreading of Labrador Sea Water in the eastern North Atlantic," *Journal of Geophysical Research: Oceans*, vol. 103, no. C5, pp. 10 223–10 239, May 1998. [Online]. Available: <http://doi.wiley.com/10.1029/98JC00262>
- Palter, J. B., Caron, C., Law, K. L., Willis, J. K., Trossman, D. S., Yashayaev, I. M., and Gilbert, D., "Variability of the directly observed, middepth subpolar North Atlantic circulation," *Geophysical Research Letters*, vol. 43, no. 6, pp. 2700–2708, Mar. 2016. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015GL067235>
- Palter, J. B., Lozier, M. S., and Lavender, K. L., "How Does Labrador Sea Water Enter the Deep Western Boundary Current?" *Journal of Physical Oceanography*, vol. 38, no. 5, pp. 968–983, May 2008. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/2007JPO3807.1>
- Pershing, A., Mills, K., Dayton, A., Franklin, B., and Kennedy, B., "Evidence for Adaptation from the 2016 Marine Heatwave in the Northwest Atlantic Ocean," *Oceanography*, vol. 31, no. 2, pp. 151–161, June 2018. [Online]. Available: <https://tos.org/oceanography/article/evidence-for-adaptation-from-the-2016-marine-heatwave-in-the-northwest-atl>
- Pershing, A. J., Alexander, M. A., Hernandez, C. M., Kerr, L. A., Le Bris, A., Mills, K. E., Nye, J. A., Record, N. R., Scannell, H. A., Scott, J. D., Sherwood, G. D., and Thomas, A. C., "Slow adaptation in the face of rapid warming leads to collapse of the Gulf of Maine cod fishery," *Science*, vol. 350, no. 6262, pp. 809–812, Nov. 2015. [Online]. Available: <https://www.sciencemag.org/lookup/doi/10.1126/science.aac9819>
- Pershing, A. J., Greene, C. H., Hannah, C., Sameoto, D., Head, E., Mountain, D. J., Jossi, J. W., Benfield, M. C., Reid, P. C., and Durbin, T. G., "Oceanographic Responses to Climate in the Northwest Atlantic," *Oceanography*, vol. 14, no. 3, pp. 76–82, 2001.
- Petrie, B. and Drinkwater, K., "Temperature and salinity variability on the Scotian Shelf and in the Gulf of Maine 1945–1990," *Journal of Geophysical Research*, vol. 98, no. C11, p. 20079, 1993. [Online]. Available: <http://doi.wiley.com/10.1029/93JC02191>
- Pickart, R. S., Straneo, F., and Moore, G., "Is Labrador Sea Water formed in the Irminger basin?" *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 50, no. 1, pp. 23–52, Jan. 2003. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063702001346>
- Pingree, R. D., "A COMPONENT OF LABRADOR SEA WATER IN THE BAY OF BISCAY: LABRADOR WATER IN BAY OF BISCAY," *Limnology and*

- Oceanography*, vol. 18, no. 5, pp. 711–718, Sept. 1973. [Online]. Available: <http://doi.wiley.com/10.4319/lo.1973.18.5.0711>
- Racapé, V., Thierry, V., Mercier, H., and Cabanes, C., “ISOW Spreading and Mixing as Revealed by Deep-Argo Floats Launched in the Charlie-Gibbs Fracture Zone,” *Journal of Geophysical Research: Oceans*, vol. 124, no. 10, pp. 6787–6808, Oct. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019JC015040>
- Ramp, S. R., Schlitz, R. J., and Wright, W. R., “The deep flow through the Northeast Channel, Gulf of Maine,” *Journal of Physical Oceanography*, vol. 15, pp. 1790–1808, 1985.
- Rhein, M., Kieke, D., Hüttl-Kabus, S., Roessler, A., Mertens, C., Meissner, R., Klein, B., Böning, C. W., and Yashayaev, I., “Deep water formation, the subpolar gyre, and the meridional overturning circulation in the subpolar North Atlantic,” *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 58, no. 17-18, pp. 1819–1832, Sept. 2011. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064511000440>
- Rhein, M., Steinfeldt, R., Kieke, D., Stendardo, I., and Yashayaev, I., “Ventilation variability of Labrador Sea Water and its impact on oxygen and anthropogenic carbon: a review,” *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, vol. 375, no. 2102, p. 20160321, Sept. 2017. [Online]. Available: <https://royalsocietypublishing.org/doi/10.1098/rsta.2016.0321>
- Ridgway, K. R., Dunn, J. R., and Wilkin, J. L., “Ocean Interpolation by Four-Dimensional Weighted Least Squares—Application to the Waters around Australasia,” *JOURNAL OF ATMOSPHERIC AND OCEANIC TECHNOLOGY*, vol. 19, p. 19, 2002.
- Roessler, A., Rhein, M., Kieke, D., and Mertens, C., “Long-term observations of North Atlantic Current transport at the gateway between western and eastern Atlantic,” *Journal of Geophysical Research: Oceans*, vol. 120, no. 6, pp. 4003–4027, June 2015. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014JC010662>
- Rosmond, T., Teixeira, J., Peng, M., Hogan, T., and Pauley, R., “Navy operational global atmospheric prediction system (NOGAPS): Forcing for ocean models,” *Oceanography*, vol. 15, no. 1, pp. 99–108, 2002. [Online]. Available: <https://doi.org/10.5670/oceanog.2002.40>

- Rossby, T., "The North Atlantic Current and surrounding waters: At the crossroads," *Reviews of Geophysics*, vol. 34, no. 4, pp. 463–481, Nov. 1996. [Online]. Available: <http://doi.wiley.com/10.1029/96RG02214>
- Rossby, T., "On gyre interactions," *Deep Sea Research Part II: Topical Studies in Oceanography*, vol. 46, no. 1-2, pp. 139–164, Jan. 1999. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967064598000952>
- Rossby, T., Flagg, C. N., Donohue, K., Sanchez-Franks, A., and Lillibridge, J., "On the long-term stability of Gulf Stream transport based on 20 years of direct measurements," *Geophysical Research Letters*, vol. 41, no. 1, pp. 114–120, Jan. 2014. [Online]. Available: <http://doi.wiley.com/10.1002/2013GL058636>
- Saba, V. S., Griffies, S. M., Anderson, W. G., Winton, M., Alexander, M. A., Delworth, T. L., Hare, J. A., Harrison, M. J., Rosati, A., Vecchi, G. A., and Zhang, R., "Enhanced warming of the Northwest Atlantic Ocean under climate change," *Journal of Geophysical Research: Oceans*, vol. 121, no. 1, pp. 118–132, 2015. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015JC011346>
- Sabine, C. L., "The Oceanic Sink for Anthropogenic CO₂," *Science*, vol. 305, no. 5682, pp. 367–371, July 2004. [Online]. Available: <https://www.sciencemag.org/lookup/doi/10.1126/science.1097403>
- Saunders, P. M., "The flux of overflow water through the Charlie-Gibbs Fracture Zone," *Journal of Geophysical Research*, vol. 99, no. C6, p. 12343, 1994. [Online]. Available: <http://doi.wiley.com/10.1029/94JC00527>
- Schartup, A. T., Thackray, C. P., Qureshi, A., Dassuncao, C., Gillespie, K., Hanke, A., and Sunderland, E. M., "Climate change and overfishing increase neurotoxicant in marine predators," *Nature*, vol. 572, no. 7771, pp. 648–650, Aug. 2019. [Online]. Available: <http://www.nature.com/articles/s41586-019-1468-9>
- Schmitz, W. J. and McCartney, M. S., "On the North Atlantic Circulation," *Reviews of Geophysics*, vol. 31, no. 1, pp. 29–49, Feb. 1993. [Online]. Available: <http://doi.wiley.com/10.1029/92RG02583>
- Schott, F. A., Fischer, J., Dengler, M., and Zantopp, R., "Variability of the Deep Western Boundary Current east of the Grand Banks," *Geophysical Research Letters*, vol. 33, no. 21, p. L21S07, Oct. 2006. [Online]. Available: <http://doi.wiley.com/10.1029/2006GL026563>
- Schott, F. A., Zantopp, R., Stramma, L., Dengler, M., Fischer, J. R., and Wibaux, M., "Circulation and Deep-Water Export at the Western Exit of the Subpolar North Atlantic," *JOURNAL OF PHYSICAL OCEANOGRAPHY*, vol. 34, p. 27, 2004.

- Smith, L., Barth, J., Kelley, D., Plueddemann, A., Rodero, I., Ulses, G., Vardaro, M., and Weller, R., “The Ocean Observatories Initiative,” *Oceanography*, vol. 31, no. 1, pp. 16–35, Mar. 2018. [Online]. Available: <https://tos.org/oceanography/article/the-ocean-observatories-initiative>
- Solodoch, A., McWilliams, J. C., Stewart, A. L., Gula, J., and Renault, L., “Why Does the Deep Western Boundary Current “Leak” around Flemish Cap?” *Journal of Physical Oceanography*, vol. 50, no. 7, pp. 1989–2016, July 2020. [Online]. Available: <https://journals.ametsoc.org/jpo/article/50/7/1989/347277/Why-Does-the-Deep-Western-Boundary-Current-Leak>
- Stommel, H. and Arons, A., “On the abyssal circulation of the world ocean—I. Stationary planetary flow patterns on a sphere,” *Deep Sea Research (1953)*, vol. 6, pp. 140–154, Jan. 1959. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/0146631359900656>
- Stramma, L., Kieke, D., Rhein, M., Schott, F., Yashayaev, I., and Koltermann, K. P., “Deep water changes at the western boundary of the subpolar North Atlantic during 1996 to 2001,” *Deep Sea Research Part I: Oceanographic Research Papers*, vol. 51, no. 8, pp. 1033–1056, Aug. 2004. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S0967063704000640>
- Sy, A., Rhein, M., Lazier, J. R. N., Koltermann, K. P., Meincke, J., Putzka, A., and Bersch, M., “Surprisingly rapid spreading of newly formed intermediate waters across the North Atlantic Ocean,” *Nature*, vol. 386, no. 6626, pp. 675–679, Apr. 1997. [Online]. Available: <http://www.nature.com/articles/386675a0>
- Talley, L. D. and McCartney, M. S., “Distribution and Circulation of Labrador Sea Water,” *Journal of Physical Oceanography*, vol. 12, pp. 1189–1205, 1982.
- Team, R. C., “R: A language and environment for statistical computing,” Vienna, Austria, 2020. [Online]. Available: <https://www.R-project.org/>
- Thyng, K., Greene, C., Hetland, R., Zimmerle, H., and DiMarco, S., “True Colors of Oceanography: Guidelines for Effective and Accurate Colormap Selection,” *Oceanography*, vol. 29, no. 3, pp. 9–13, Sept. 2016. [Online]. Available: <https://tos.org/oceanography/article/true-colors-of-oceanography-guidelines-for-effective-and-accurate-colormap>
- Toole, J. M., Andres, M., Le Bras, I. A., Joyce, T. M., and McCartney, M. S., “Moored observations of the Deep Western Boundary Current in the NWAtlantic: 2004-2014: LINE W 2004-2014,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 9, pp. 7488–7505, Sept. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2017JC012984>
- Uppala, S. M., Kållberg, P. W., Simmons, A. J., Andrae, U., Bechtold, V. D. C., Fiorino, M., Gibson, J. K., Haseler, J., Hernandez, A., Kelly, G. A.,

- Li, X., Onogi, K., Saarinen, S., Sokka, N., Allan, R. P., Andersson, E., Arpe, K., Balmaseda, M. A., Beljaars, A. C. M., Berg, L. V. D., Bidlot, J., Bormann, N., Caires, S., Chevallier, F., Dethof, A., Dragosavac, M., Fisher, M., Fuentes, M., Hagemann, S., Hólm, E., Hoskins, B. J., Isaksen, I., Janssen, P. A. E. M., Jenne, R., McNally, A. P., Mahfouf, J.-F., Morcrette, J.-J., Rayner, N. A., Saunders, R. W., Simon, P., Sterl, A., Trenberth, K. E., Untch, A., Vasiljevic, D., Viterbo, P., and Woollen, J., “The ERA-40 re-analysis,” *Quarterly Journal of the Royal Meteorological Society*, vol. 131, no. 612, pp. 2961–3012, Oct. 2005. [Online]. Available: <http://doi.wiley.com/10.1256/qj.04.176>
- van Seville, E., Griffies, S. M., Abernathey, R., Adams, T. P., Berloff, P., Biastoch, A., Blanke, B., Chassignet, E. P., Cheng, Y., Cotter, C. J., Deleersnijder, E., Döös, K., Drake, H. F., Drijfhout, S., Gary, S. F., Heemink, A. W., Kjellsson, J., Koszalka, I. M., Lange, M., Lique, C., MacGilchrist, G. A., Marsh, R., Mayorga Adame, C. G., McAdam, R., Nencioli, F., Paris, C. B., Piggott, M. D., Polton, J. A., Rühls, S., Shah, S. H., Thomas, M. D., Wang, J., Wolfram, P. J., Zanna, L., and Zika, J. D., “Lagrangian ocean analysis: Fundamentals and practices,” *Ocean Modelling*, vol. 121, pp. 49–75, Jan. 2018. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S1463500317301853>
- Willis, J. K. and Fu, L.-L., “Combining altimeter and subsurface float data to estimate the time-averaged circulation in the upper ocean: VELOCITY FROM ALTIMETER AND FLOAT DATA,” *Journal of Geophysical Research: Oceans*, vol. 113, no. C12, Dec. 2008. [Online]. Available: <http://doi.wiley.com/10.1029/2007JC004690>
- Worthington, L., “On the North Atlantic Circulation,” *The Johns Hopkins Oceanographic Studies*, vol. 6, p. 110, 1976.
- Wright, W. R. and Worthington, L., “North Atlantic Ocean Atlas of Potential Temperature and Salinity in the Deep Water,” *Woods Hole Oceanographic Institution Atlas Series 2*, vol. 2, p. 24.
- Xu, X., Hurlburt, H. E., Schmitz, W. J., Zantopp, R., Fischer, J., and Hogan, P. J., “On the currents and transports connected with the atlantic meridional overturning circulation in the subpolar North Atlantic: TRANSPORTS IN THE SPNA,” *Journal of Geophysical Research: Oceans*, vol. 118, no. 1, pp. 502–516, Jan. 2013. [Online]. Available: <http://doi.wiley.com/10.1002/jgrc.20065>
- Xu, X., Bower, A., Furey, H., and Chassignet, E. P., “Variability of the Iceland-Scotland Overflow Water Transport Through the Charlie-Gibbs Fracture Zone: Results From an Eddying Simulation and Observations,” *Journal of Geophysical Research: Oceans*, vol. 123, no. 8, pp. 5808–5823,

- Aug. 2018. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018JC013895>
- Yashayaev, I., “Hydrographic changes in the Labrador Sea, 1960–2005 [Prog. Oceanogr. 73 (2007) 242–276],” *Progress in Oceanography*, vol. 75, no. 4, pp. 857–859, Dec. 2007. [Online]. Available: <https://linkinghub.elsevier.com/retrieve/pii/S007966110700184X>
- Yashayaev, I., Bersch, M., and van Aken, H. M., “Spreading of the Labrador Sea Water to the Irminger and Iceland basins,” *Geophysical Research Letters*, vol. 34, no. 10, p. L10602, May 2007. [Online]. Available: <http://doi.wiley.com/10.1029/2006GL028999>
- Yashayaev, I. and Loder, J. W., “Recurrent replenishment of Labrador Sea Water and associated decadal-scale variability: 2015 Convection in Labrador Sea,” *Journal of Geophysical Research: Oceans*, vol. 121, no. 11, pp. 8095–8114, Nov. 2016. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012046>
- Yashayaev, I. and Loder, J. W., “Further intensification of deep convection in the Labrador Sea in 2016,” *Geophysical Research Letters*, vol. 44, no. 3, pp. 1429–1438, Feb. 2017. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016GL071668>
- Zantopp, R., Fischer, J., Visbeck, M., and Karstensen, J., “From interannual to decadal: 17 years of boundary current transports at the exit of the Labrador Sea: INTERANNUAL LABRADOR SEA TRANSPORTS,” *Journal of Geophysical Research: Oceans*, vol. 122, no. 3, pp. 1724–1748, Mar. 2017. [Online]. Available: <http://doi.wiley.com/10.1002/2016JC012271>
- Zeller, S. and Rogers, D., “Visualizing Science: How Color Determines What We See,” *EOS*, vol. 101, 2020.
- Zhang, R. and Vallis, G. K., “The Role of Bottom Vortex Stretching on the Path of the North Atlantic Western Boundary Current and on the Northern Recirculation Gyre,” *Journal of Physical Oceanography*, vol. 37, no. 8, pp. 2053–2080, Aug. 2007. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO3102.1>
- Zou, S. and Lozier, M. S., “Breaking the Linkage Between Labrador Sea Water Production and Its Advective Export to the Subtropical Gyre,” *Journal of Physical Oceanography*, vol. 46, no. 7, pp. 2169–2182, July 2016. [Online]. Available: <http://journals.ametsoc.org/doi/10.1175/JPO-D-15-0210.1>
- Zou, S., Lozier, M. S., and Buckley, M., “How Is Meridional Coherence Maintained in the Lower Limb of the Atlantic Meridional Overturning Circulation?” *Geophysical Research Letters*, vol. 46, no. 1, pp. 244–252, Jan. 2019. [Online]. Available: <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL080958>