

The Fate of Atlantic Water Heat in the Arctic Ocean

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A thesis submitted for the degree of Doctor of Philosophy

Trinity 2022

Atlantic Water (AW) is the most significant source of oceanic heat in the Arctic Ocean, and has played a key role in recent Arctic sea ice loss. Better understanding of the spatial and temporal variability of AW heat and its future evolution is therefore key to understanding future Arctic change.

I use observations and model data from CESM-LE (Community Earth System Model Large Ensemble) to investigate 20th and 21st century AW layer changes. Observations show that AW heat has had an increased influence on the upper ocean in the eastern Arctic, but has become more isolated from the surface in the west. However, CESM-LE suggests that this is due to a regional lag in response, and that the AW will warm, freshen and shoal throughout the Arctic from the 2020s. These trends are anthropogenically forced. A shift in atmospheric circulation likely plays an important role in these changes, with the resultant atmospheric forcing from a weakened Beaufort High and expanded low over the eastern Arctic enhancing AW transport from the Atlantic, and causing more AW to be advected into the western Arctic.

Local processes such as mixing and cascading shelf flows also play an important role in modifying AW properties, and buffer against high frequency AW temperature variability.

In CESM-LE, despite AW warming and freshening, the stability of the upper water column increases in the future due to surface and halocline freshening. However, there is a shift to alpha ocean stratification in the summer Barents Sea towards the end of the 21st century which may have implications for AW heat loss and could occur in the Arctic basin itself in the longer term.

This thesis highlights the importance of AW heat in governing how the future Arctic may evolve in a changing climate, and the various processes that influence its fate.

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Contents

1	Introduction	1
1.1	Overview	1
1.2	Geography of the Arctic Ocean	3
1.3	Hydrography and circulation of the Arctic Ocean	4
1.3.1	Sea ice and the surface mixed layer	4
1.3.2	Atmospheric circulation and surface forcing	5
1.3.3	The halocline	7
1.3.4	Atlantic Water	9
1.3.4.1	AW advection pathways	12
1.3.4.2	AW propagation into sub-basin interiors	16
1.3.4.3	The relationship between AW in the Arctic and AW in the Nordic Seas	17
1.3.5	Deep waters	18
1.3.6	Fluxes between layers	18
1.4	A Changing Arctic	19
1.4.1	Sea ice loss and freshwater storage	20
1.4.2	Ocean heat content	21
1.4.3	Stratification	23
1.4.4	Forced trends and natural variability in the AW layer	24
1.4.5	Consequences of a changing Arctic	25
1.5	Arctic Ocean observational challenges and model performance	26
1.6	Thesis Aims	27
1.7	Thesis Structure	28

2	Spatial and temporal variability of Atlantic Water in the Arctic from forty years of observations	30
2.1	Introduction	30
2.2	Data and Methods	31
2.3	Results and Discussion	35
2.3.1	The Atlantic Water core across the Arctic	35
2.3.2	Regional differences in Atlantic Water properties and their temporal variability	39
2.3.3	AW variability at transects and moorings	45
2.3.3.1	Eurasian Basin	45
2.3.3.2	Canada Basin	48
2.3.4	Relationships between profile metrics	53
2.4	AW core properties and their relationship to AW temperature profile shape .	56
2.5	Conclusion	57
3	The evolution of the AW heat in the 21st century	59
3.1	Introduction	59
3.2	Data and methods	61
3.2.1	Model description	61
3.2.2	Model assessment	61
3.2.3	Statistics	63
3.3	Results and Discussion	66
3.3.1	Hydrographic properties of the AW layer within the Arctic in the CESM-LE and their natural variability	66
3.3.1.1	The evolution of AW layer properties	66
3.3.1.2	Forced trends and natural variability of the AW layer in the 20 th and 21 st centuries	70
3.3.1.2.1	AW temperature and heat content	71
3.3.1.2.2	AW salinity and density	75
3.3.1.2.3	AW depth and thickness	77
3.3.1.2.4	Seasonal cycle emergence	81
3.3.2	Gateway transports and AW advection	82

3.3.2.1	Hydrography and velocity at the gateways	82
3.3.2.2	Gateway transports	89
3.3.2.3	AW advection within the Arctic	96
3.3.3	The link between AW variability in the Arctic and gateway transports	99
3.3.3.1	Natural variability in AW transports through gateways . . .	100
3.3.3.2	30-year temperature trends	101
3.3.3.3	Dominant timescales of variability	104
3.4	Conclusion	108
4	The impact of AW evolution on the stability of the upper water column	110
4.1	Introduction	110
4.2	Data and methods	112
4.3	Results and discussion	114
4.3.1	Water column stability and its drivers	114
4.3.2	The role of temperature and salinity in governing stratification	127
4.3.3	Quantifying water column stability and its trends	133
4.4	Conclusion	137
5	Summary and conclusions	138
5.1	Thesis summary and discussion	139
5.2	Implications for the Arctic region and wider climate system	147
5.3	Further work	149
5.4	Final word	151
	Bibliography	153

List of Figures

1.1	Map of the Arctic Ocean with geographical features key to this thesis labelled.	3
1.2	Monthly average sea ice extent for March and September 2021. Data and figure from http://nsidc.org/arcticseaicenews/sea-ice-comparison-tool/ (National Snow and Ice Data Center 2022).	5
1.3	Illustration of the key circulation features of the Arctic Ocean. Diagram credit: Jack Cook, Woods Hole Oceanographic Institution.	6
1.4	Schematic of halocline source waters - Pacific Water, Atlantic Water, and river run-off, and their circulation throughout the Arctic Ocean. Figure from (Rudels et al. 2004).	8
1.5	Pan-Arctic sections of potential temperature and salinity created using data from the World Ocean Database 2018 (Boyer 2018). Figure from Timmermans et al. (2020).	10
1.6	Potential temperature, salinity, and depth of the AW core (the AW layer temperature maximum) in March and September of the MIMOC climatological dataset Schmidtke et al. (2013).	11
1.7	Schematic of the potential vorticity contours followed by the pan-Arctic AW boundary current and surface/halocline water local to the western Arctic via the wind-driven mechanism proposed by (Nøst et al. 2003). Typical sea level pressure is also shown. Figure from (Timmermans et al. 2020).	14
1.8	Arctic sea ice extent anomalies for March and September. Data and figure from (Perovich et al. 2018).	20
1.9	Annual normalised maximum temperature of the AW layer from a pan-Arctic observational dataset. Figure from (Polyakov et al. 2012b).	22

2.1	Arctic map showing relevant geographic features along with the spatial and temporal distribution of all AW core observations.	33
2.2	A demonstration of different smoothing lengths on a profile from the Canada Basin.	34
2.3	Maps showing the depth, potential temperature, salinity, and potential density anomaly of all AW core data points used in Chapter 2.	36
2.4	Un-smoothed potential temperature and salinity profiles at various locations and times from across the Arctic.	38
2.5	Annual normalised histograms of AW core potential temperature for the eastern and western Arctic.	41
2.6	As Figure 2.5 for AW core salinity	42
2.7	As Figure 2.5 for AW core potential density anomaly	43
2.8	As Figure 2.5 for AW core depth	44
2.9	Potential temperature and salinity along a repeated CTD transect in the eastern Eurasian Basin.	46
2.10	Water column properties across a repeated CTD transect in the eastern Eurasian Basin.	47
2.11	Water column properties across two repeated CTD transects in the Canada Basin.	49
2.12	Time series of monthly mean AW core properties from four moorings in the Canada Basin.	50
2.13	Hovmöller plots of potential temperature from three moorings in the Canada Basin.	52
2.14	Maps of total AW layer heat content from all profiles which sampled the entire AW layer.	53
2.15	Scatter plots between various AW core and water column properties in both the eastern and western Arctic, with R-squared values and regression lines. .	55
3.1	A comparison of observational and model water column profiles from both the eastern and western Arctic.	63
3.2	Map showing the key sections and regions used in Chapters 3 and 4.	65
3.3	Ensemble mean maps of depth-mean potential temperature of the AW layer.	67

3.4	As in Figure 3.3 but for AW layer mean salinity.	68
3.5	As in Figure 3.3 but for AW top depth.	68
3.6	Hovmöller plots of ensemble mean potential temperature and salinity in the upper 1000 <i>m</i> of the water column, averaged over eastern and western regions of the Arctic Ocean.	69
3.7	Ensemble mean maps of sea ice cover in March and September.	69
3.8	Time series of hydrographic properties for each ensemble member averaged across the eastern and western Arctic. Ensemble mean time series are also shown.	72
3.9	Forced trend, internal variability, and ratio between forced trend and natural variability during 1920–2000 for AW layer mean potential temperature, AW layer mean salinity, AW layer mean potential density, and AW layer heat content.	73
3.10	As in Figure 3.9 but for the period 2020–2100.	74
3.11	As in Figure 3.9 but for the depth of the top of the AW layer and AW layer thickness.	79
3.12	As in Figure 3.11 but for the period 2020–2100.	80
3.13	Ensemble mean hydrographic properties across the Fram Strait.	83
3.14	Ensemble mean hydrographic properties across the Barents Sea Opening.	86
3.15	Ensemble mean hydrographic properties across the St. Anna Trough.	87
3.16	Time series of ensemble mean volume and temperature transport through the Fram Strait.	89
3.17	As in Figure 3.16 but for the Barents Sea Opening.	90
3.18	As in Figure 3.16 but for the St. Anna Trough.	90
3.19	Time series of normalised NAO index and its running 30-year mean, computed using ensemble mean data from a point-wise sea level pressure difference between Iceland and the Azores.	93
3.20	Maps of ensemble mean velocity at 500 <i>m</i> depth.	97
3.21	Map of change in ensemble mean velocity at 500 <i>m</i> depth between 1920–1949 and 2070–2099.	97
3.22	Maps of time-averaged ensemble mean sea level pressure.	98

3.23	Time series of ensemble member and ensemble mean volume and temperature transport across the Fram Strait. Net transports are shown along with inflow and outflow transports.	100
3.24	As for Figure 3.23 but for the St. Anna Trough.	101
3.25	Maps of maximum ensemble trend correlations between 30-year trends in Fram Strait AW inflow temperature and AW layer mean temperature at each grid point. Time lags corresponding to the maximum correlations are also plotted.	102
3.26	As for Figure 3.25 but for inflow through the St. Anna Trough.	104
3.27	Maps showing the two most dominant timescales of AW temperature variability over 1920–2000 and 2020–2100, along with their ensemble statistics.	105
3.28	Spectral plots of AW inflow temperature variability through the Fram Strait and St. Anna Trough for every ensemble member.	107
4.1	March ensemble mean pan-Arctic sections of potential temperature, salinity, and potential density, with AW top denoted.	115
4.2	March and September ensemble mean pan-Arctic sections of the square of the Brunt-Väisälä frequency (N^2), with AW top denoted.	116
4.3	March eastern Eurasian Basin profiles of Brunt-Väisälä frequency squared (N^2) and potential density from each ensemble member.	118
4.4	As for Figure 4.3 but for temperature and salinity.	119
4.5	As in Figure 4.3 but for a location in the western Eurasian Basin.	121
4.6	As in Figure 4.4 but for a location in the western Eurasian Basin.	122
4.7	As in Figure 4.3 but for a location in the Canada Basin.	124
4.8	As in Figure 4.4 but for a location in the Canada Basin.	125
4.9	TS diagrams of ensemble mean data from March and September at the same Canada Basin location as Figures 4.7 and 4.8.	126
4.10	Ensemble mean density difference from the surface density at a location in the eastern Eurasian Basin, and the temperature and salinity contributions to this density difference - all relative to the 1920–1949 average.	128
4.11	As for Figure 4.10 but for a location in the Canada Basin.	129
4.12	As for Figure 4.10 but for a location in the Barents Sea.	130

4.13	March and September ensemble mean profiles of potential temperature and salinity for the same location in the Barents Sea as in Figure 4.12.	131
4.14	Time series of March and September ensemble mean and ensemble member available potential energy anomaly, relative to 1920, averaged over the eastern and western Arctic.	134
4.15	March maps of forced trend, natural variability; and the ratio between forced trend and natural variability for available potential energy.	135

List of Tables

2.1 Trends in monthly mean AW core potential temperature, salinity, potential density anomaly, and depth from four Canada Basin moorings.	51
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Chapter 1

Introduction

1.1 Overview

The Arctic experiences the impact of global warming more than any other region on Earth, with surface air temperatures here rising faster than the global mean in an effect known as “Arctic Amplification” (Overland et al. 2019; Rantanen et al. 2022; Serreze et al. 2011). One of the most noticeable signatures of this so far has been the rapid decline in summer sea ice extent in the Arctic Ocean (Deser et al. 2008; Notz et al. 2016; Perovich et al. 2019). The thickness of sea ice at the end of the melt season reduced by two thirds between 1958 and 2018, and more than half of the multi-year ice in the Arctic has been lost over the past two decades (Kwok 2018). This decline is expected to continue, with the Arctic heading towards ice-free summers and thinner, more mobile winter ice (Haine et al. 2017). According to some models, the Arctic will have ice-free summers by the middle of the 21st century (Collins et al. 2013; Notz et al. 2020).

The impact of these stark changes is not only felt locally, but will likely affect the wider global climate system as well. The reduction in albedo due to sea ice loss has resulted in an increase in the amount of solar short-wave radiation absorbed by Earth. Pistone et al. (2014) found the darkening of the Arctic due to sea ice loss resulted in an increase in radiative heating equivalent to 25 % of the direct CO₂ forcing of the preceding 30 years. Furthermore, given that Arctic outflows are a significant input to the North Atlantic, any changes in their density could have important consequences for the Atlantic Meridional Overturning Circulation (AMOC), which itself is a key component of the global climate system. Jahn et al. (2013) found that sea ice decline can cause an increase in Arctic freshwater export

and hence a surface freshening in the North Atlantic. This has the potential to reduce the strength of deep convection in the North Atlantic and hence weaken overturning in the region (Jahn et al. 2013).

Although, as mentioned, many of these changes can be linked to warming surface air temperatures, in recent years the role of the Arctic Ocean itself in sea ice loss has become a key research topic. The most significant source of oceanic heat to the Arctic Ocean is water originating from the Atlantic, known as Atlantic Water (AW) (Carmack et al. 2015). The AW has warmed in recent years, both within the Arctic and upon inflow, with AW inflow temperatures increasing by 1–1.5 °C between 1980–2012 (Mulwijk et al. 2018; Polyakov et al. 2012b). Inflowing AW temperature is in fact the warmest it has been for the past 2000 years (Spielhagen et al. 2011). This AW temperature increase has been linked to a shoaling of AW in parts of the eastern Arctic and a reduction in upper ocean stratification, resulting in an increase in vertical heat flux toward the surface in a process known as “Atlantification” (Polyakov et al. 2017; Polyakov et al. 2010). Given the important local and global consequences of Arctic sea ice loss discussed above, improving our understanding of how the role played by AW in sea ice loss may evolve in the coming decades is an important part of predicting how our future climate may change.

Building a detailed observational picture of the Arctic Ocean to get an idea of its current state and recent changes is an important part of gaining this better understanding of AW in the Arctic. However, the challenges of an ice-covered polar environment mean that hydrographic measurements of the Arctic Ocean are sparse and inhomogeneous in both time and space. Furthermore, many observational studies of the Arctic Ocean focus on a specific region only (e.g. Li et al. (2020), Lind et al. (2018), and McLaughlin et al. (2009)). Collating these individual datasets and analysing them as one would not only give a useful pan-Arctic picture of changing ocean heat content and stratification, but, crucially, also be useful for benchmarking and validating forced ocean-sea ice and coupled climate models of the Arctic.

Models themselves are also a key tool in improving our understanding of AW - particularly pan-Arctic AW circulation, the drivers of AW changes, and the future evolution of AW heat in the Arctic - but come with their own challenges.

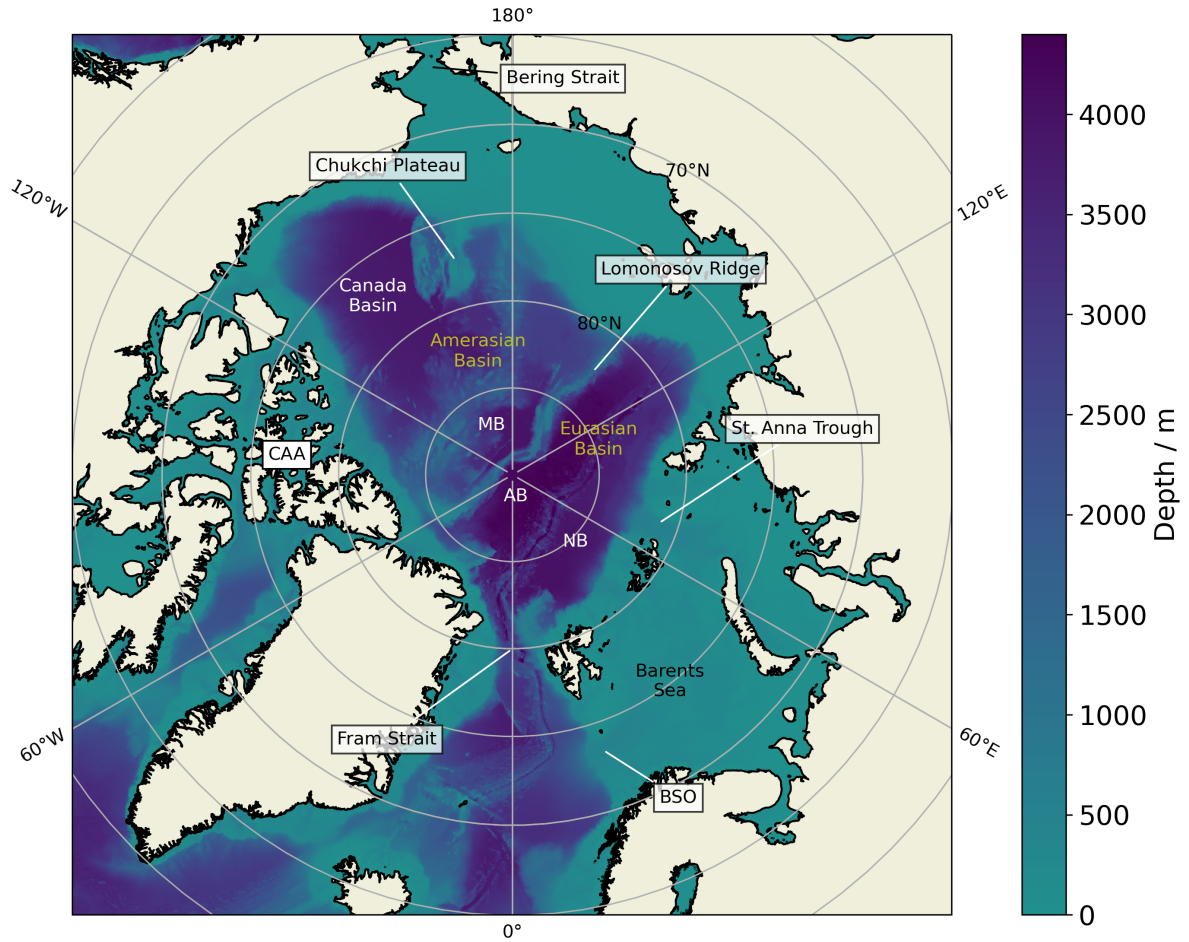


Figure 1.1: Map of the Arctic Ocean with geographical features key to this thesis labelled. Colours represent ocean bathymetry. The Arctic can be divided into two broad regions separated by the Lomonosov Ridge, with the Eurasian Basin (consisting of the Amundsen Basin (AB) and Nansen Basin (NB)) on the European side, and the Amerasian Basin (consisting of the Makarov Basin (MB) and Canada Basin) on the North American side. Key gateways for water mass transport into and out of the Arctic are the Bering Strait, St. Anna Trough, Barents Sea Opening (BSO), Fram Strait and Canadian Arctic Archipelago (CAA).

1.2 Geography of the Arctic Ocean

The Arctic Ocean, shown geographically in Figure 1.1, sits between the Eurasian and North American land masses, and is the smallest of the world's five oceans. It is dominated by shallow coastal shelves, although its three sub-basins - the Eurasian Basin, the Makarov Basin and the Canada Basin - are deep. It is linked to the Pacific via the shallow and narrow Bering Strait, and to the Nordic Seas and North Atlantic by the narrow and deep Fram Strait, the shallow and wide Barents Sea Opening (BSO), and several passages through the Canadian Arctic Archipelago (CAA). Perhaps the most stark bathymetric feature of the Arctic is the Lomonosov Ridge, which reaches depths of less than 1000 m in places. For hydrographic

reasons explained later, the Arctic Ocean is often divided into two broad regions either side of the Lomonosov Ridge - the Eurasian Basin (consisting of the Amundsen and Nansen Basins) and the Amerasian Basin (containing the Canada and Makarov Basins), which are sometimes referred to as the eastern and western Arctic, respectively. Despite holding less than 2 % of Earth’s ocean volume, the Arctic Ocean plays an important role in global heat and freshwater budgets, as will become clear in the following section.

1.3 Hydrography and circulation of the Arctic Ocean

1.3.1 Sea ice and the surface mixed layer

Most of the surface of the Arctic Ocean is covered by seasonal sea ice, with sea ice cover at a maximum in March and a minimum in September. Currently, about a third of the ocean surface is covered by perennial sea ice which, during winter, is around 2 *m* thick (Kwok 2018). Figure 1.2 shows this seasonal change in sea ice cover for the year 2021 (National Snow and Ice Data Center 2022).

The fresh surface mixed layer below the sea ice is around freezing temperature, although temperatures in this layer can rise to several degrees above freezing due to solar heating when not covered in sea ice, or in the presence of Atlantic or Pacific Waters (Timmermans et al. 2019). In the Nansen Basin, surface water derives from AW, which interacts with and melts sea ice upon its inflow (where it exists at the surface). The upper part of the AW here then becomes transformed into fresher, cooler, less dense surface water (Rudels 2016). In the rest of the deep Arctic basins, surface waters come from net precipitation (the atmosphere generally transports fresh water from the equator towards the poles), seasonal sea ice melt, and river run off (Rudels 2016; Timmermans et al. 2020). The extensive shelf seas surrounding the Arctic Ocean are a conduit for this river run off (which is greatly influenced by the terrestrial melt cycle), and seasonal sea ice melt here is also a key non-local source of the surface mixed layer of the deep Arctic basins. The thickness of the Arctic surface mixed layer varies temporally, with the mixed layer being deeper in winter than summer at around 25–50 *m* and 5–30 *m*, respectively (e.g. Peralta-Ferriz et al. (2015)).



Figure 1.2: Monthly average sea ice extent for March and September 2021. Data and figure from <http://nsidc.org/arcticseaicenews/sea-ice-comparison-tool/> (National Snow and Ice Data Center 2022).

1.3.2 Atmospheric circulation and surface forcing

Two main atmospheric wind patterns govern the advection of sea ice and the surface mixed layer of the Arctic. These are the Beaufort High, which tends to be centered over the Canada Basin, and the Icelandic Low (which we will return to later when discussing AW inflow). The Beaufort High drives an anticyclonic flow of sea ice and the surface mixed layer (and halocline) in the Canada Basin, as illustrated in the blue surface flow in Figure 1.3, in what is

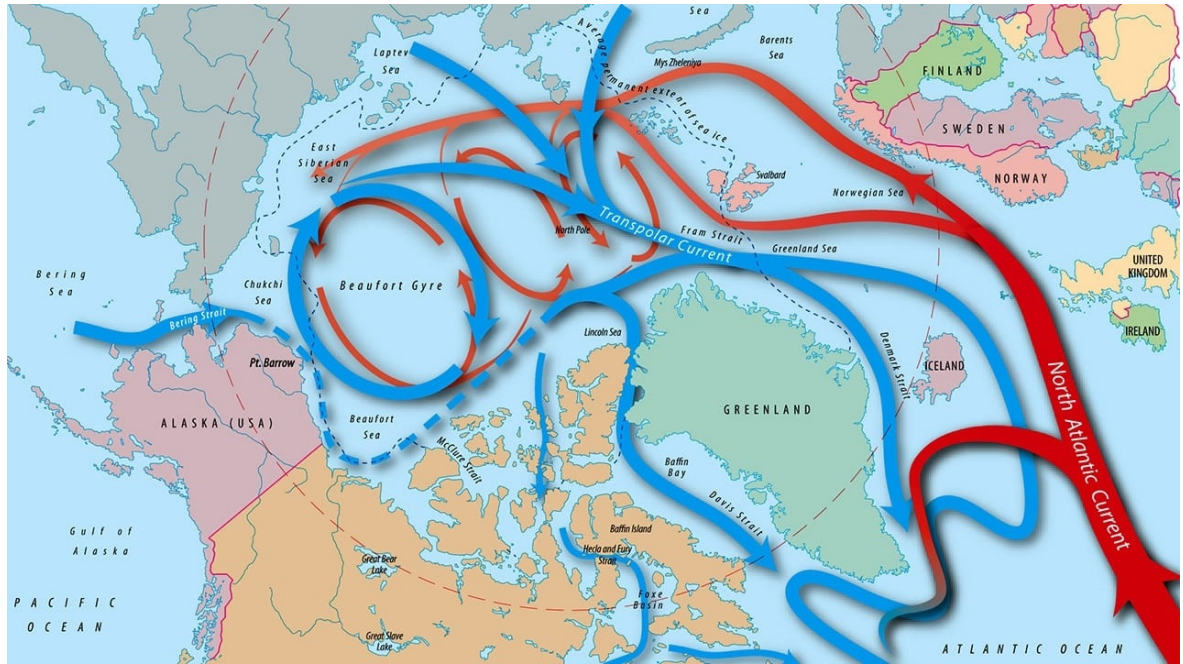


Figure 1.3: Illustration of the key circulation features of the Arctic Ocean. Blue arrows represent the advection of cool, fresh surface waters, with red arrows showing the advection of warm, saline Atlantic Water below. Diagram credit: Jack Cook, Woods Hole Oceanographic Institution.

known as the Beaufort Gyre. The anticyclonic nature of this flow results in the convergence of fresh surface waters and Ekman pumping/downwelling. As a result, the Beaufort Gyre is an important store of freshwater in the Arctic, with water of salinity $< 33 \text{ psu}$ typically extending down to around 250 m in recent decades (Timmermans et al. 2020), although how much freshwater is stored largely depends on the strength of the Beaufort High winds and the structural strength of the sea ice pack (as this mediates the transfer of wind forcing to the ocean surface (Doddridge et al. 2019; Meneghello et al. 2018)).

Another notable feature of the cool, fresh surface flows shown in Figure 1.3 is the transpolar current (also known as the transpolar drift stream). This is the general advection of surface waters and sea ice from Siberia towards the Nordic Seas and Greenland. The general path of the transpolar drift stream is governed by the relative strength of the Beaufort High and Icelandic Low, with a weaker Beaufort High and stronger Icelandic Low resulting in a more cyclonic transpolar drift which flows from the Laptev Sea towards the Canada Basin before heading towards the Fram Strait (Kwok et al. 2013). A stronger Beaufort High and weaker Icelandic Low result in a more direct transpolar drift from the Laptev Sea towards the Fram Strait (Kwok et al. 2013). This relative strength of the Beaufort High to Icelandic Low

is essentially the Arctic Oscillation (AO) (Thompson et al. 1998), which is directly linked to the North Atlantic Oscillation (NAO) (e.g. Deser (2000)). A positive NAO and AO means that the sea level pressure in the Icelandic Low and Beaufort High is anomalously low, and vice versa. Variations in the AO can therefore impact the path of the transpolar drift and the transport of fresh shelf water to the surface layer of the deep Arctic basins.

1.3.3 The halocline

Another source of upper ocean water shown in Figure 1.3 is that which flows through the Bering Strait - i.e. Pacific Water. This inflow is driven by sea surface height differences between the Bering Sea and surrounding shelf seas, and also variations in local winds (e.g. (Danielson et al. 2014; Peralta-Ferriz et al. 2017)). Although Pacific Water forms part of the surface layer in the shelf seas surrounding the Canada Basin, it also enters the water column of the western Arctic just below the surface mixed layer via outcropping isopycnals at the edges of the Beaufort Gyre (e.g. Timmermans et al. (2017)) and via eddies shedding off the boundary current (e.g. Pickart et al. (2005)), forming part of the halocline layer and enhancing the stratification between the surface and deeper waters. This Pacific Water in the Canada Basin halocline can be classified into both winter (Pickart et al. 2005) and summer (Timmermans et al. 2014) varieties, depending on which time of year it was ventilated in the Bering Strait and local shelf seas. The warmer Pacific Summer Water has been shown to persist year-round in the Canada Basin halocline (Timmermans et al. 2017).

The halocline layer exists throughout the Arctic Ocean, and forms an important barrier between the surface layer above and the AW below. When defined as the potential density surface $\sigma = 27.4 \text{ kg m}^{-3}$ it exists at around 100 *m* depth in the eastern Arctic and 200-250 *m* in the west (due to the influence of the Beaufort Gyre) (Timmermans et al. 2020). The influence of Pacific Water on the halocline extends west until approximately the Lomonosov Ridge where the front between Pacific and Atlantic influence on the upper water column lies (Morison et al. 1998; Steele et al. 1998). The halocline waters to the east of this front predominantly derive from AW - either that which has entered the Arctic via the Fram Strait or that which has traveled via the Barents Sea. The halocline waters originating in the Barents Sea are formed through interaction with saline water produced during sea ice formation (Aagaard et al. 1981) and/or freshwater from sea ice melt (Steele et al. 1995),

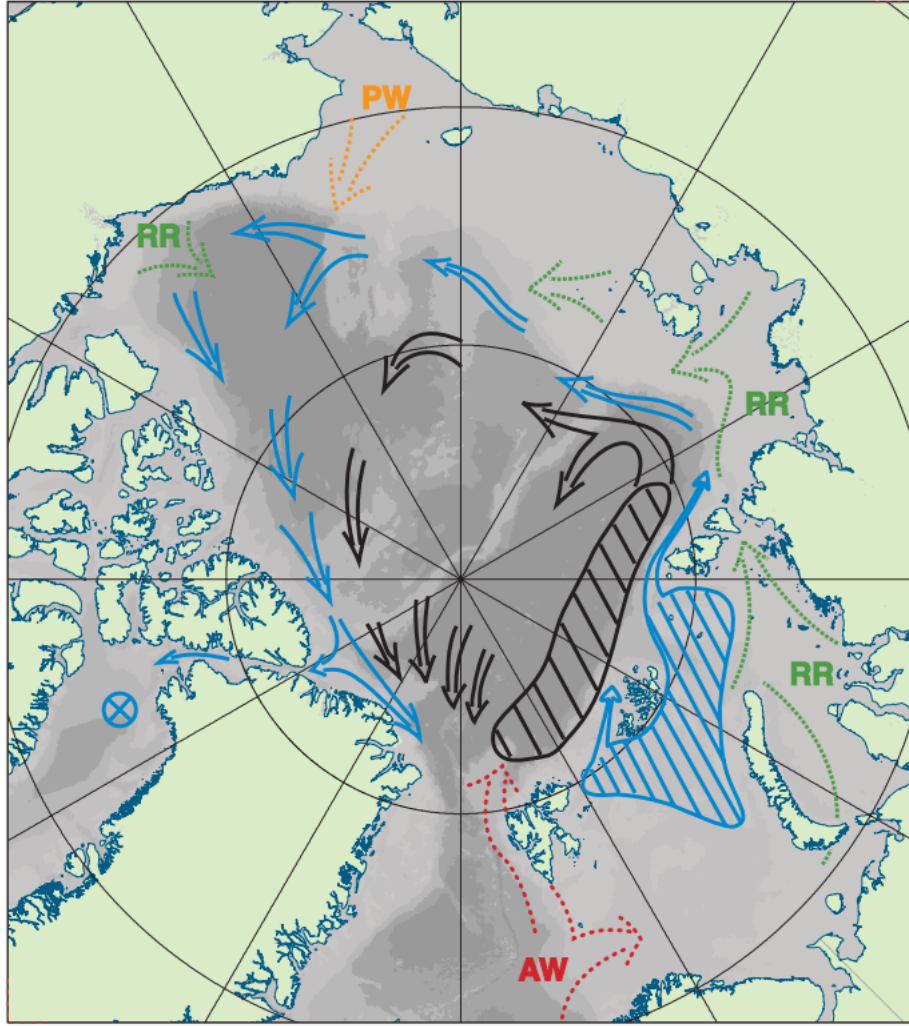


Figure 1.4: Schematic of halocline source waters - Pacific Water (PW), Atlantic Water (AW), and river run-off (RR), and their circulation throughout the Arctic Ocean. Figure from Rudels et al. (2004).

with Fram Strait halocline waters forming via the latter mechanism also (Rudels et al. 2004). Both these halocline waters are advected around the Arctic as shown in Figure 1.4. The advection of Barents Sea halocline waters into the western Arctic results in their interaction with inflowing Pacific Water in the Chukchi Sea, modifying the Pacific Water before it enters the Canada Basin halocline (Woodgate et al. 2005).

Freshwater from the shelves surrounding the Arctic plays a key role in the formation of lower halocline waters (i.e. the cold halocline) in the eastern Arctic. This freshwater is shown as river run-off (RR) in Figure 1.4. This formation mechanism has been termed the “advection-convection” process of cold halocline formation, and involves winter convection followed by the mixing of advected freshwater with the surface mixed layer, resulting in a cold

halocline sitting beneath a fresher surface layer (Steele et al. 1998). However, as mentioned, the advection pathway of fresh shelf waters often depends on the highly variable NAO/AO. In fact, a high (positive) NAO/AO phase during the 1990s was found to result in a “cold halocline retreat”, where the cold halocline in the Amundsen and Makarov Basins diminished in spatial extent due to the deflection of fresh shelf waters towards the Canada Basin (Schauer et al. 2002b; Steele et al. 1998).

Waters originating from the Arctic shelves can also directly impact deeper ocean layers. Cold, dense flows cascading from the shelves – which are formed from brine rejection during sea ice formation on the shelves – have been observed to interact with the lower halocline in the western Arctic (Ivanov et al. 2004; Luneva et al. 2020; Melling et al. 1995). Similar flows in the Eurasian Basin have been observed to penetrate even deeper into the water column and interact with local AW, modifying its properties (Ivanov et al. 2007).

The importance of the strongly-stratified halocline of the Arctic Ocean cannot be overstated. While subtropical oceans are alpha oceans, the Arctic Ocean is a beta ocean - with cool temperatures resulting in salinity being the dominant factor determining its density and stratification (Carmack 2007) (alpha and beta refer to the thermal expansion coefficient and saline contraction coefficient, respectively, in the equation of state of sea water). This not only allows sea ice to form (sea ice formation is only possible in beta oceans (Bulgakov 1962)), but also means that the warm AW that circulates throughout the Arctic Ocean is isolated from the surface layer and sea ice due to its relatively high salinity. The AW layer contains enough heat to melt all sea ice in the Arctic within a few years if brought to the surface in that time (Carmack et al. 2015; Turner 2010), and so the halocline is key in maintaining Arctic Sea ice and preventing it from being affected by this significant heat source. The importance of the halocline is clear in Figure 1.5, which emphasises the strong divide between the fresh, cool surface waters and the warmer more saline AW at depth.

1.3.4 Atlantic Water

The Atlantic Water layer is the main focus of this thesis. It is the most significant source of oceanic heat in the Arctic (Carmack et al. 2015), and contains enough heat to melt all sea ice in the Arctic within a few years if brought to the surface in that time (Carmack et al. 2015; Turner 2010). It sits beneath the halocline throughout the Arctic Ocean, and hence is

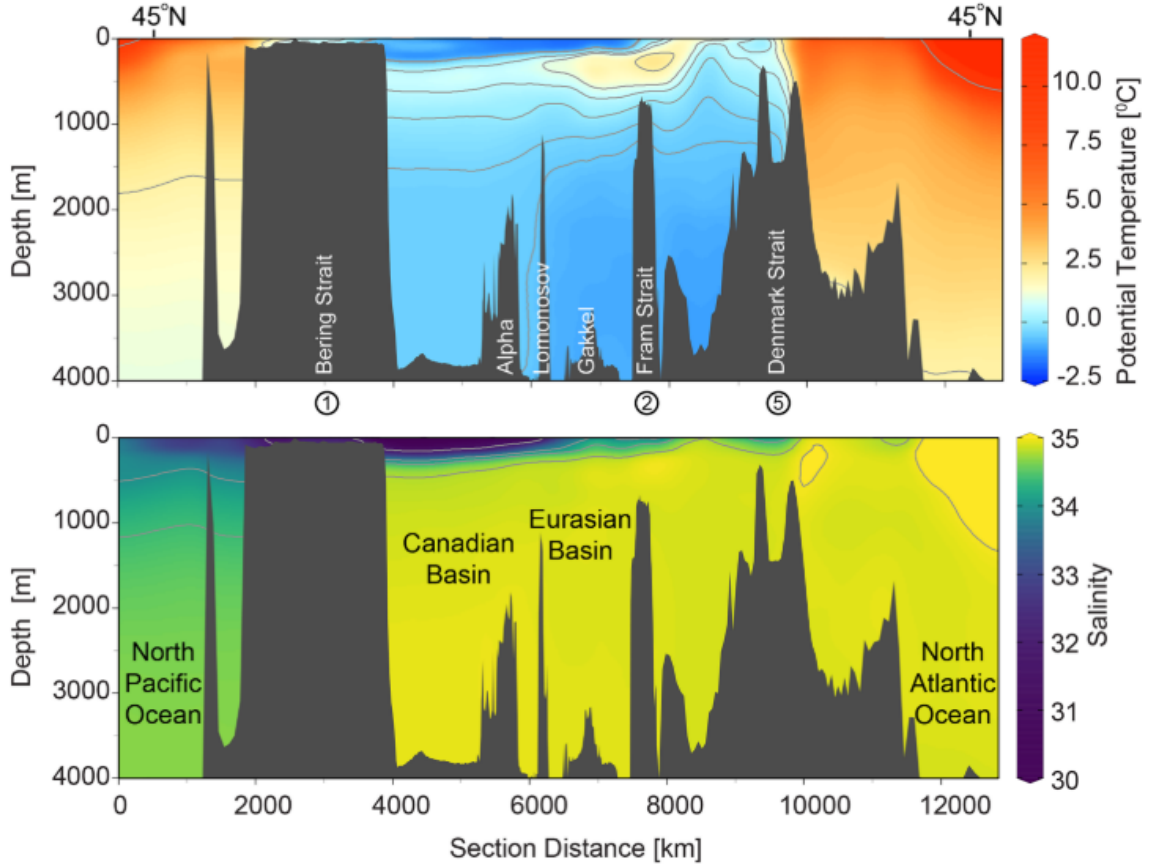


Figure 1.5: Pan-Arctic sections of potential temperature and salinity created using data from the World Ocean Database 2018 (Boyer 2018). Figure from Timmermans et al. (2020).

generally deeper in the western Arctic than it is in the eastern Arctic, as shown in Figure 1.5. This figure also emphasises the relatively high temperatures of the AW - something which is particularly prominent in the east of the potential temperature section. The AW layer is often defined by its warm temperature, with temperatures over 0°C commonly used in conjunction with salinity criteria to differentiate it from other warm surface/near-surface water masses.

Figure 1.6 gives an idea of the general properties of the pan-Arctic AW layer, showing the potential temperature, salinity and depth of the AW layer temperature maximum (also known as the AW core). Data are from the MIMOC climatology, which aimed to resemble the state of the ocean between 2007–2011 (Schmidtke et al. 2013). We define the AW core in this dataset as the warmest point in the water column with salinity greater than 34.7 psu, following Lique et al. (2012). A key feature of Figure 1.6 is the difference in AW core temperatures between the Nordic Seas (where AW core temperature reaches 10°C in September) and the Arctic basin itself, indicating strong AW heat loss upon entry to the Arctic basin. A second

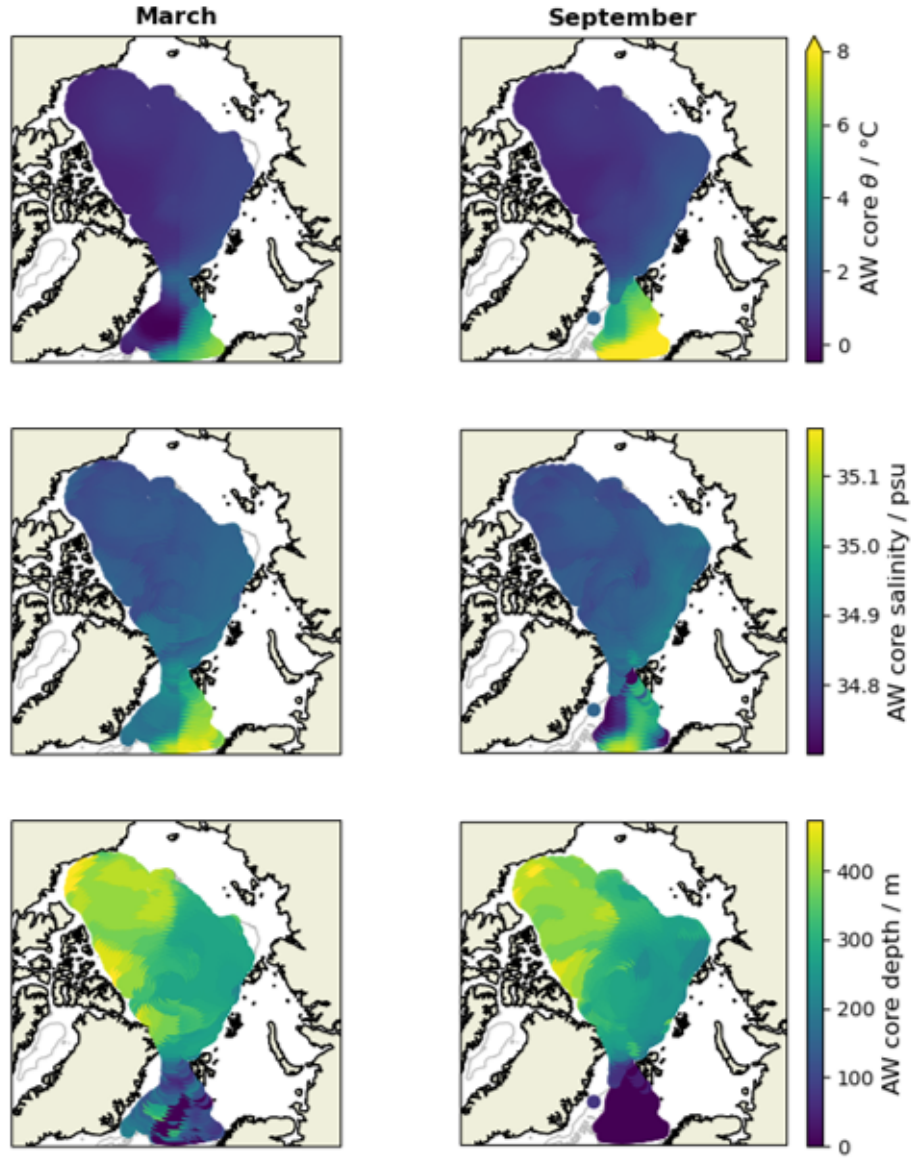


Figure 1.6: Potential temperature, salinity, and depth of the AW core (the AW layer temperature maximum) in March and September of the MIMOC climatological dataset Schmidtke et al. (2013).

notable feature is the change in depth of the AW core. The core exists at the surface south of the Fram Strait, before subducting underneath the fresher surface layer to around 200–300 *m* in the Eurasian Basin, and then deepening to over 400 *m* in the Canada Basin under the influence of the Beaufort Gyre (the AW advection pathway will be described in more detail below). A third key takeaway from the climatological data is the minimal seasonal cycle in AW core properties within the Arctic basin, as reported by Lique et al. (2012).

1.3.4.1 AW advection pathways

The circulation of AW in the Arctic is illustrated by the red arrows in Figure 1.3. AW enters the Arctic basin through the Fram Strait and via the Barents Sea through the St. Anna Trough. Approximately 7 *Sv* of AW flows in through the Fram Strait as Fram Strait Branch Water (FSBW) (Beszczynska-Moller et al. 2012), with some recirculated back to the south and about 2 *Sv* advected eastward along the southern Nansen basin boundary at around 150-600 *m* depth (Karcher et al. 2012; Pérez-Hernández et al. 2019). The branch of AW inflowing through the St. Anna Trough from the Barents Sea - referred to as Barents Sea Branch Water (BSBW) - is cooler and fresher than FSBW, with a volume transport of roughly 2 *Sv* (Schauer et al. 2002a; Schauer et al. 2002b). East of the St. Anna Trough these two branches meet, with the BSBW displacing the FSBW off the shelf so that the two branches are then advected parallel to each other along the continental slope of the eastern Eurasian Basin (Schauer et al. 2002b).

At the Kara Sea slope, the BSBW circulates along the upper slope, around 400 *m*, with its core temperature at 1.8 °C and core salinity at 34.85 *psu*. The FSBW core here is found further down the slope, above the 2000–3000 *m* isobaths, with temperature around 3 °C and salinity 34.95 *psu* (Dmitrenko et al. 2015). The cooler, fresher BSBW is a result of the transformation undergone by AW in the Barents Sea, where much AW heat is lost to the atmosphere, and interactions with freshwater and sea ice imported into the Barents Sea from the north result in a reduction in AW salinity (Smedsrud et al. 2013). However, the resultant temperature of these two branches is likely also affected by mixing between the two branches upon their meeting just downstream of the St. Anna Trough (Dmitrenko et al. 2015). This confluence can be turbulent, resulting in the formation of eddies (Schauer et al. 2002a), and mixing between the two branches continues along the slopes between the St. Anna Trough and the Lomonosov Ridge (Schauer et al. 2002a).

Upon meeting the junction between the Eurasian continental shelf and the ridge separating the Nansen and Amundsen Basin (the Gakkel Ridge), the main part of the FSBW leaves the continental slope and travels northward along this ridge back towards the Fram Strait (Rudels et al. 2013). The remaining FSBW and the BSBW continue along the continental slope toward the region where the Eurasian continental shelf meets the Lomonosov Ridge. This is an important junction for AW circulation. Here, some AW - predominantly the re-

maintaining FSBW - diverts northward along the Lomonosov Ridge to recirculate back towards the Fram Strait, whereas the BSBW branch continues eastward along the continental slope over the Lomonosov Ridge and into the western Arctic (Dmitrenko et al. 2008; Rudels et al. 2013; Woodgate et al. 2001).

Beyond the Lomonosov Ridge, the AW (BSBW) circulates cyclonically in both the Makarov and Canada Basins, before returning to the Fram Strait (Rudels et al. 2013). The rough topography of the Chukchi Plateau is a significant obstacle encountered by the AW upon its advection towards the Canada Basin. Here the AW splits into two branches, with one travelling around the north of the plateau and another travelling over it - the latter experiencing a significant degree of turbulent mixing because of this rough topography (Li et al. 2020; McLaughlin et al. 2009). The northern branch then enters the interior Canada Basin, with the other branch continuing along the continental slope of the basin (Li et al. 2020; McLaughlin et al. 2009).

Within the Canada Basin, AW advection is thought to be affected by the anticyclonic surface flow of the Beaufort Gyre, with measurements and models suggesting that a strengthening of the anticyclonic circulation of the gyre can reduce (or perhaps even reverse) the cyclonic advection of the AW below it (Karcher et al. 2012; Lique et al. 2015a). The strength of the Beaufort Gyre can also affect the depth of the AW layer. A stronger gyre circulation is associated with enhanced convergence of surface water and Ekman pumping, which can deepen the AW beneath to the point where AW in the interior CB underneath the gyre can exist deeper than denser AW near the continental slope, away from the gyre center (Zhong et al. 2014).

AW takes roughly 7.5 years to travel from the St. Anna Trough at the Eurasian Basin boundary to the southern Canada Basin, with current speeds in this boundary current measured to be around $2\text{--}4\text{ cm s}^{-1}$ (Mauldin et al. 2010; Woodgate et al. 2001).

The mechanisms forcing this cyclonic pan-Arctic AW boundary current are still an active area of research, but it is generally accepted that this flow is driven by a combination of wind forcing, eddy-topography interaction, and estuarine circulation. These will be briefly explained below.

Given the high-latitude of the Arctic Ocean, and the resulting small meridional gradient in the Coriolis parameter, f , the effect of Ekman upwelling and downwelling in the Arctic Ocean

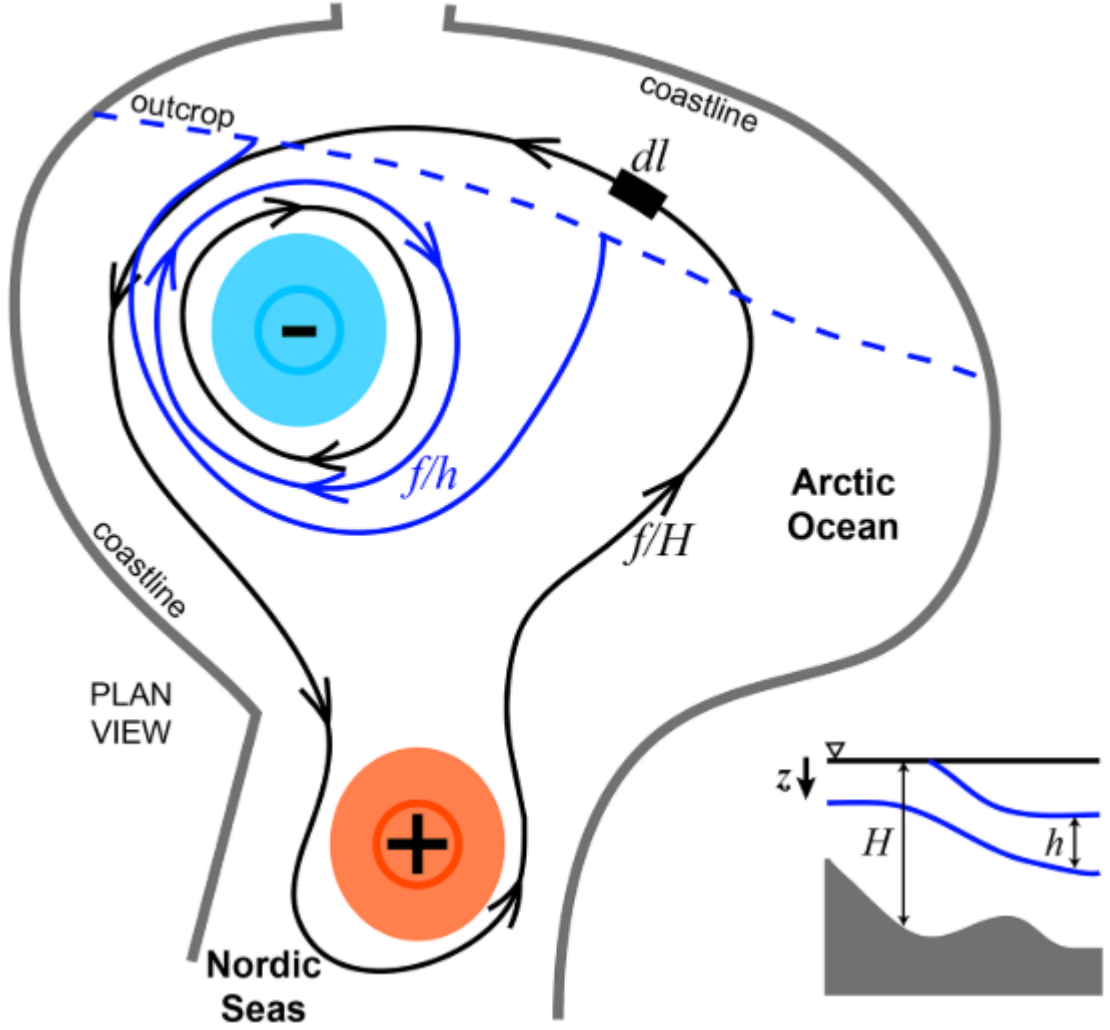


Figure 1.7: Schematic of the potential vorticity contours followed by the pan-Arctic AW boundary current (black) and surface/halocline water local to the western Arctic (blue) via the wind-driven mechanism proposed by (Nøst et al. 2003). Red and blue regions indicate typical sea level pressures and resulting winds driving the circulation. Figure from Timmermans et al. (2020).

is not compensated for by Sverdrup balance (i.e. depth-integrated meridional flow acting to balance vorticity input). Instead, wind-driven barotropic flows in the Arctic conserve potential vorticity (f/H) by following potential vorticity contours, which in the case of the Arctic (where, again, meridional gradient in the Coriolis parameter is small), results in flows following bathymetric contours. This was confirmed by Nøst et al. (2003) who found that the flow along a potential vorticity contour can be estimated using the wind stress curl within the contour, and that the sense of the AW flow is set by the cyclonic wind stress over the Nordic Seas. The cyclonic AW flow in the Canada Basin is a result of the relative weakness of the anticyclonic wind stress here compared to the cyclonic wind stress elsewhere in the

Arctic and Nordic Seas. This process is schematised in Figure 1.7. Bottom friction plays a key role in balancing wind stress curl, along with lateral momentum fluxes from baroclinic eddies (e.g. Dewar (1998)).

Another way to view the circulation of AW in the Arctic is through an estuarine framework. This involves the mixing and entrainment of salty AW up into the fresher surface layer, which drives an inflow of AW from the Nordic Seas into the Arctic to compensate for the outflow of fresh polar water out of the Arctic. Atlantic Water flow into the Arctic has been modeled using an estuary framework (Stigebrandt 1981) and also using a double estuary framework (where the cooling in the Nordic Seas acts as a negative estuary (i.e. induces negative buoyancy forcing) and the freshening of AW within the Arctic acts as a positive estuary (e.g. Lambert et al. (2016))). The estuarine view of AW circulation suggests that mixing levels are important in driving AW advection. Mixing will be discussed in more detail later.

Eddy-topography interaction also appears to be important for AW boundary current circulation in what is known as the Neptune effect, where a correlation between the seafloor slope and stresses from eddy pressure anomalies drives flow along continental slopes (Holloway 1992; Nazarenko et al. 1998). The injection of positive potential vorticity in the Arctic (as inflowing isopycnal layers are thinner than those exiting the Arctic) may also play a key role in setting the cyclonic sense of the AW boundary current circulation (Karcher et al. 2007; Yang 2005).

The circulation of the AW boundary current described above is not constant, however. As is the case in the surface layer, AW advection is greatly influenced by large-scale atmospheric wind patterns - in particular the NAO and AO. Mulwijk et al. (2019) found that the cyclonicity of winds over the Greenland Sea can influence AW advection throughout the Arctic Ocean basin. This was also proposed by Karcher et al. (2012) who found that the amount of AW crossing the Lomonosov Ridge (which can act as a potential vorticity barrier) into the western Arctic increased during the high NAO phase of the 1990s. This is in agreement with the proposed influence of wind stress on the AW boundary current discussed above, with a strengthening cyclonic wind stress near the AW inflow and a weaker Beaufort High (both associated with a more positive NAO) resulting in a stronger cyclonic pan-Arctic boundary current. When the NAO index reduces, a stronger AW current is directed along

the Lomonosov Ridge, with less AW crossing into the western Arctic (Karcher et al. 2012).

1.3.4.2 AW propagation into sub-basin interiors

So far this section has focussed on the drivers and pathways of the cyclonic, topographically-steered AW boundary current. However, the AW layer is present throughout the Arctic Ocean, not just at the boundaries.

AW primarily propagates into the sub-basin interiors through thermohaline intrusions. These intrusions are the result of along-isopycnal gradients in temperature and salinity (Ruddick 1992), which exist between the warm inflowing AW boundary current and the cooler abyssal waters of the interior sub-basins, and travel laterally from the AW core (Bebieva et al. 2017; Rudels et al. 2009). Intrusions begin as small lateral perturbations of the warm, saline AW boundary current, which then form intrusions due to the divergence of double-diffusive vertical buoyancy fluxes driving further lateral advection (Ruddick 1992; Ruddick et al. 2003). Salt fingering (which transports salt more effectively than heat) results in a flux of salt across the base of the warm, saline intrusion. Heat and salt are lost through the top of the intrusion, first through reduced salt fingering and then through diffusive convection (which transports heat more effectively than salt) when the intrusion becomes more established. This results in the intrusion becoming less dense as it is advected, hence the lateral advection of the intrusion is self-driven (Ruddick 1992). Thermohaline intrusions carrying AW heat into the interior have been observed in both the Eurasian and Canada Basins (e.g. Kuzmina et al. (2011), McLaughlin et al. (2009), and Rudels et al. (2009)). They form around the AW core depth, as this allows for the differing types of double-diffusive convection on either side of the intrusion.

Eddies also contribute to the transport of AW from the boundary into the interior. Schauer et al. (2002a), for example, found evidence that eddies can detach from the front at the confluence of the FSBW and BSBW just downstream of the St. Anna Trough and migrate into the interior Eurasian Basin as a warm water lens. Similarly, eddies transport AW into the interior of the Canada Basin from the boundary current in the southwestern corner of the basin, where changes in topography result in eddy formation (McLaughlin et al. 2009). Eddies also play a key role in counter-acting isopycnal deepening in the Beaufort Gyre, driving a salt flux from the boundary towards the interior (e.g. (Doddridge et al. 2019)).

1.3.4.3 The relationship between AW in the Arctic and AW in the Nordic Seas

As mentioned above, the NAO and resultant wind stress over the Nordic Seas influences the advection of AW within the Arctic Ocean (Karcher et al. 2012; Muilwijk et al. 2019). Wind stress and atmospheric temperature over the Nordic Seas also influence the properties and volume of AW advected into the Arctic via both the Fram Strait and Barents Sea. Muilwijk et al. (2018) found a high correlation between AW volume transport into the Arctic Ocean and wind stress curl in the Nordic Seas and North Atlantic. Similar findings were also reported by Muilwijk et al. (2019), who found that Greenland Sea wind forcing can explain 50 % of the AW inflow variance, with strong cyclonic wind stress over the Greenland Sea (i.e. a positive NAO) resulting in greater AW inflow.

Muilwijk et al. (2018) also found that thermodynamic forcing from the Nordic Seas atmosphere damps/amplifies AW heat anomalies and accounts for 45 % of the temperature variability of AW inflowing into the Arctic. Wind forcing itself can also affect the temperature of AW flowing north through the Fram Strait. A stronger Greenland Sea gyre circulation (a result of a stronger cyclonic wind stress) was found to increase AW flow speed toward the Fram Strait and the northward transport of AW heat, resulting in higher AW inflow temperatures into the Arctic and less AW heat in the Nordic Seas (Chatterjee et al. 2018; Muilwijk et al. 2019). An increased Greenland Sea gyre circulation also results in greater import of cold polar water into the Nordic Seas via the East Greenland Current (which runs south through the western side of the Fram Strait and towards the southern tip of Greenland), further reducing the Nordic Seas heat content (Muilwijk et al. 2019).

After transit through the Arctic, the Atlantic-origin water eventually exits back into the Atlantic Ocean via the Canadian Arctic Archipelago (CAA) and the Fram Strait (as shown in Figure 1.3), fresher and cooler than it came in, having taken about 20-30 years to complete its journey (Karcher et al. 2002; Lique et al. 2010; Rudels 2015; Wefing et al. 2020). This outflow through the Fram Strait can influence the Nordic Seas and the broader North Atlantic. If the density of AW in the Arctic is low, i.e. it is particularly warm and/or fresh upon inflow or experiences reduced cooling and/or enhanced freshening within the Arctic, it is thought that this could be maintained throughout the AW circumnavigation of the Arctic Ocean, and hence reduce the density of outflows into the North Atlantic (Karcher et al. 2011). This is in addition to any increase in sea ice melt or freshwater export from the Arctic acting to reduce

the density of waters in the North Atlantic (Jahn et al. 2013).

1.3.5 Deep waters

The Arctic Ocean deep and bottom waters, below the AW layer, are dense waters thought to be formed through brine rejection on the continental shelves (Jones et al. 1995). Based on sparse observations, it has been suggested that these waters are advected slowly and cyclonically in both the Eurasian and Canada Basins, with residence times of several decades (Aagaard et al. 1981; Rudels 2015). Although largely divided by the Lomonosov Ridge, there is some exchange of deep waters between these two basins, as reported by Woodgate et al. (2001) who inferred that this exchange occurs through a gap in the Lomonosov Ridge at around 1700 *m* depth. Similarly to AW, deep Arctic waters also impact the Nordic Seas, as highlighted by Langehaug et al. (2012) who found an increase in the fraction of Eurasian Basin deep water observed in the Fram Strait between 1997–2008. The direct impact of these deep waters on the North Atlantic via the Denmark Strait is less well known, although deep waters from both the Canada and Eurasian basins have been observed just north of the Denmark Strait (Buch et al. 1996).

1.3.6 Fluxes between layers

In general, the Arctic has low mixing levels compared to the mid-latitude oceans. Given the high latitude of the Arctic, open ocean tidal forcing is weak, and so tidal flows over rough topography and bathymetric slopes are a key source of internal wave energy within the ocean (e.g. Rippeth et al. (2015)). Sea ice cover plays a role in the low-energy wave field by dissipating any internal waves that do exist (Morison et al. 1985) and also acting as a barrier to wind-driven momentum input to the ocean (Rainville et al. 2011). Turbulent kinetic energy dissipation in the Arctic halocline lies between 5×10^{-10} and $2 \times 10^{-9} \text{ W kg}^{-1}$ in the basin interiors, increasing by two orders of magnitude over the sloping bathymetry of the continental shelves (Fer 2009; Rippeth et al. 2015). Diapycnal diffusivity in the Arctic - an important part of the estuarine framework of AW circulation described earlier - ranges from 10^{-4} at the mixed layer base to $10^{-6} \text{ m}^2 \text{ s}^{-1}$ at the halocline of the deep basins (D’Asaro et al. 1991; Fer 2009). This low diapycnal diffusivity provided by the strongly stratified halocline is crucial in keeping sea ice isolated from the warm AW layer. Diapycnal diffusivity values of

around $10^{-6} \text{ m}^2\text{s}^{-1}$ were found to give the most realistic AW circulation and properties in an ice-ocean model (Zhang et al. 2007).

The temperature and salinity structure above the AW core, where both temperature and salinity increase with depth, together with the low levels of mixing in the sub-basin interiors, allows for the formation and maintenance of double-diffusive staircases (also known as diffusive convective staircases). These staircases have been observed in the temperature and salinity profiles of the upper AW layer in both the eastern and western Arctic (e.g. Polyakov et al. (2012a), Shibley et al. (2017), Sirevaag et al. (2012), and Timmermans et al. (2008)). Although they are the primary way through which heat is transported upwards from the AW layer toward the halocline and surface, typical double-diffusive heat flux values in basin interiors are of the order $10^{-2} - 10^{-1} \text{ Wm}^{-2}$, which is one to two orders of magnitude smaller than the average heat flux from the ocean to the ice e.g. Krishfield et al. (2005) to which other heat sources, such as solar heating and advected water masses such as the Pacific Water, contribute (Krishfield et al. 2005; Perovich et al. 2015). Where turbulence and mixing levels are high enough to prevent the formation and maintenance of these staircases, heat fluxes from the ocean to sea ice can be of the order 100 Wm^{-2} (Peterson et al. 2017).

Double-diffusive staircases are rarely found at basin boundaries where the topography of the continental shelves results in higher levels of turbulence and mixing (Rippeth et al. 2015; Shibley et al. 2017). Where they *are* found at boundaries, vertical heat fluxes from the AW are higher than those from staircases in the basin interiors, and are of the order 1 Wm^{-2} (e.g. Polyakov et al. (2012a)).

1.4 A Changing Arctic

As mentioned in the overview section of this chapter, the impacts of climate change on the Arctic region have been stark as “Arctic Amplification” causes temperatures to rise faster here than any other region on the planet (Overland et al. 2019; Rantanen et al. 2022; Serreze et al. 2011). Below, we will build on the description of the Arctic given in section 1.3 and give a brief overview of the drastic changes currently underway in the Arctic Ocean.

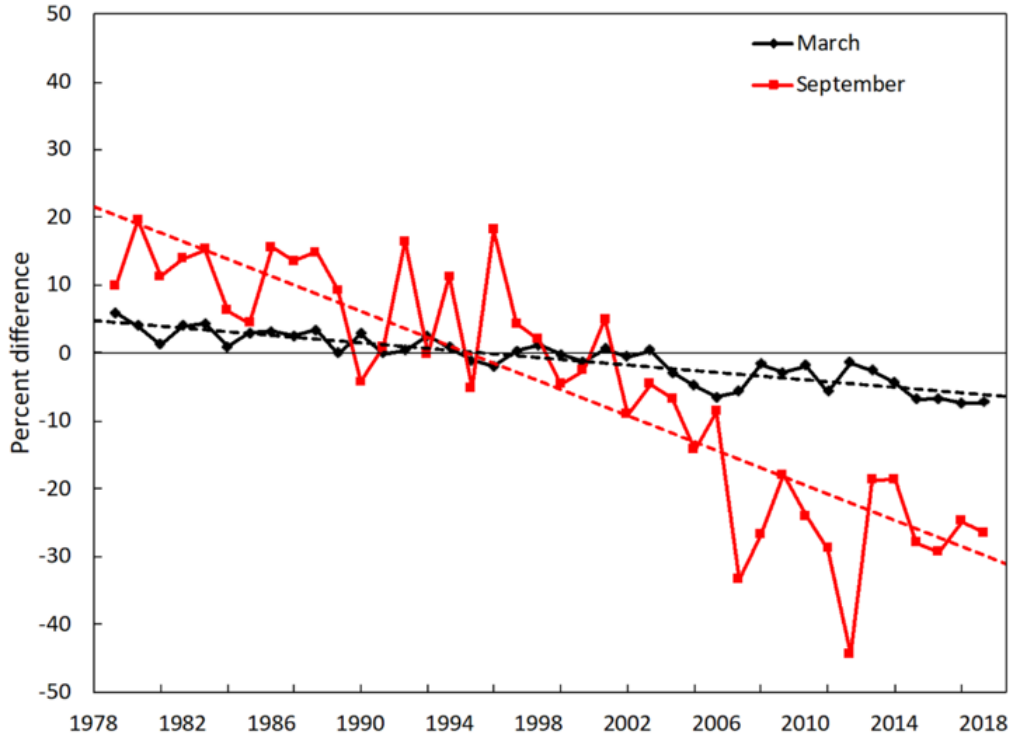


Figure 1.8: Arctic sea ice extent anomalies represented by the percentage difference in ice extent relative to mean values over 1981-2010, for March (black) and September (red). Dashed lines show the least squares linear regression. Data and figure from (Perovich et al. 2018).

1.4.1 Sea ice loss and freshwater storage

A decline in sea ice extent has perhaps been one of the most noticeable changes in the Arctic in recent decades. This has affected both the summer minimum (September) and winter maximum (March) extents, although reductions have been highest in September (see Figure 1.8) (Deser et al. 2008; Notz et al. 2016; Perovich et al. 2018; Perovich et al. 2019). Arctic sea ice has also been getting thinner on average, both due to increased melt and a decrease in the average age of sea ice, resulting in the Arctic losing more than half of its sea ice volume over the past 40 years (Carmack et al. 2015; Kwok 2018). As mentioned earlier, sea ice reductions are expected to continue, with some models predicting that the Arctic Ocean will be ice-free during summer as early as the middle of the 21st century (Collins et al. 2013; Notz et al. 2020), with winter sea ice continuing to become thinner and more mobile (Haine et al. 2017).

As the extent and volume of summer sea ice reduces more quickly than that in winter (Figure 1.8), the amount of seasonal sea ice melt (and formation) increases each year. This has driven an increase in the seasonal changes in Beaufort Gyre freshwater storage (Proshutinsky et al. 2019). The year-round freshwater storage in the Beaufort Gyre has also increased

in recent decades, which is important given that the gyre is the most significant store of freshwater in the Arctic Ocean. This is mainly due to an increase in freshwater input from local rivers and Pacific Water, and, crucially, an increase in the Ekman transport of freshwater towards the gyre center due to changing sea ice conditions and an increase in the intensity of the Beaufort High (Krishfield et al. 2014; Proshutinsky et al. 2019). As a result, the Beaufort Gyre accumulated 6400 km^3 of freshwater between 2003 and 2018 (Proshutinsky et al. 2019). The total freshwater content of the Arctic Ocean increased by approximately $600 \pm 300 \text{ km}^3$ per year between 1992 and 2012 (Rabe et al. 2014).

As alluded to above, increased melt is not the only way in which sea ice trends contribute to the increase in freshwater content of the Arctic and Beaufort Gyre. The thinning of the local sea ice pack, and the resultant decrease in its ability to withstand internal stresses, could reduce the role that sea ice plays in mediating the transfer of wind momentum to the surface ocean. This would result in a deeper gyre which, combined with greater runoff to the shelves and an increased transfer of freshwater from the shelves to the interior, stores more freshwater, with wind forcing balanced only by eddy fluxes (Doddridge et al. 2019; Meneghello et al. 2018).

Another potential impact of reduced sea ice cover in the Arctic is a more turbulent ocean with higher mixing levels (Rainville et al. 2011), which could have important consequences for heat fluxes between AW and sea ice. As discussed in section 1.3.6, these fluxes have been observed to be low - especially in basin interiors, away from boundary-induced turbulence (e.g. D’Asaro et al. (1991), Fer (2009), and Rippeth et al. (2015)). However, there have been no observed increases in mixing levels in the Arctic Ocean to date. Reduced stratification - a potential consequence of sea ice reduction - may also increase mixing levels. Stratification changes will be considered later, along with changes in vertical heat flux from the AW layer. First, though, changes in AW heat content itself will be briefly discussed.

1.4.2 Ocean heat content

Reductions in sea ice coverage and the resultant increase in solar heating of the ocean surface have contributed to an increase in upper ocean heat content in recent decades. For example, in the Canada Basin the near-surface temperature maximum increased by $0.13 \text{ }^\circ\text{C}$ per year between 2004–2007 (Jackson et al. 2010), and the heat content of the Beaufort Gyre halocline

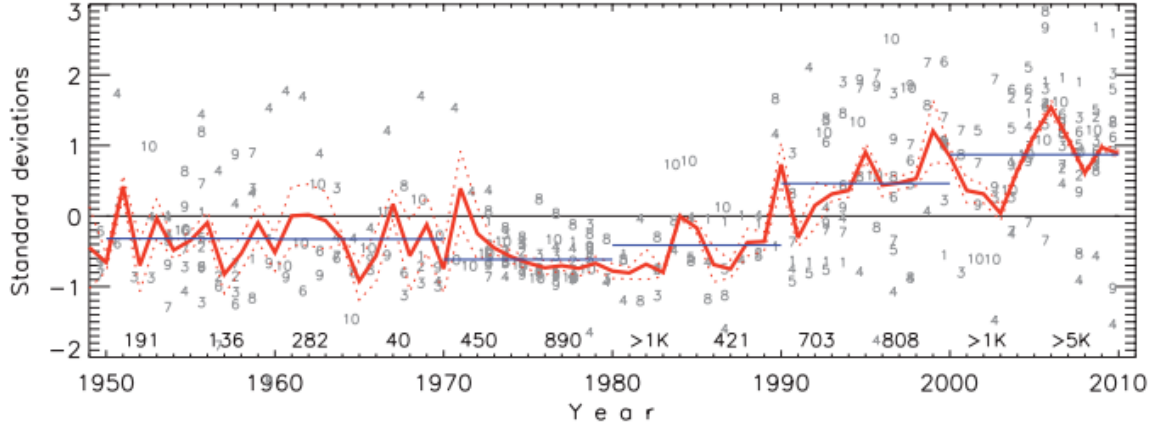


Figure 1.9: Figure from (Polyakov et al. 2012b) showing annual normalised maximum temperature of the AW layer (red), from a pan-Arctic observational dataset. Decadal-mean anomalies (blue) also given. A general warming of the AW layer temperature maximum can be seen from around 1980 onwards.

has doubled over the past three decades due to increased solar heating of source waters (Timmermans et al. 2018). In general the annual amount of solar energy absorbed by the upper Arctic Ocean increased from 200 MJ m^{-2} in 1979–1992 to 400 MJ m^{-2} in 1992–2005 (Perovich et al. 2007). The decrease in ocean surface albedo has contributed to this warming (Pistone et al. 2014; Woodgate et al. 2010). In the western Arctic, increases in the heat content of inflowing Pacific Water have also added to increases in upper ocean heat content, and have contributed to increased sea ice melt here (Perovich et al. 2015; Woodgate et al. 2010). Changes in the heat content of the Arctic shelf seas have been linked to the phase of the AO, with a positive phase resulting in higher heat content and vice versa (Steele et al. 2008).

The heat content of the AW, the focus of this thesis and most significant oceanic heat source in the Arctic, has also increased. According to marine sediments in the Fram Strait, the AW there is warmer than it has been at any point in the past 2000 years (Spielhagen et al. 2011). Observational evidence agrees with this assessment, showing a general AW warming trend during the 20th century and early 21st century (e.g. Polyakov et al. (2004) and Polyakov et al. (2012b)), as shown in Figure 1.9.

Figure 1.9 shows that these increasing trends emerge beyond natural temporal variations in AW temperature, which consist of both short term and long term oscillations. Short term inter-annual temperature oscillations have been observed in the AW boundary current, and can be traced back to the AW inflow at the Fram Strait and St. Anna Trough gateways

(Dmitrenko et al. 2008; Karcher et al. 2003; McLaughlin et al. 2009; Polyakov et al. 2004; Schauer et al. 2002b). Long term oscillations have periods of around 50–80 years (Polyakov et al. 2004), and appear in the observational record as the Early Twentieth Century Warming between the 1920s and 1940s (Bengtsson et al. 2004) and the cooler AW period that commenced in the 1970s (Ikeda 1990). AW temperature fluctuations on even longer timescales of around 1500 years have also been responsible for drastic changes in the Arctic climate (Jensen et al. 2018).

The increasing AW temperature trend in the Arctic is related to an increase in ocean heat transport through the Fram Strait and Barents Sea Opening. Increases in both temperature and volume transport of AW from the Nordic Seas into the Arctic have contributed to this enhanced ocean heat transport (Årthun et al. 2019; Barton et al. 2018; Muilwijk et al. 2018; Tsubouchi et al. 2021). Mean AW temperatures increased by 1–1.5 °C at the Fram Strait and Barents Sea Opening between 1980 and 2012 (Muilwijk et al. 2018), although quantifying changes in volume transport from observations is more difficult.

The key drivers of these increases in AW volume transport and inflow temperature have been touched upon in section 1.3.4.3 - namely, enhanced cyclonic wind stress over the Nordic Seas resulting in an increase in the strength of the Greenland Sea gyre (Chatterjee et al. 2018; Muilwijk et al. 2018). Increased air temperatures over the Nordic Seas have also been found to drive an increase in AW inflow temperatures into the Arctic (Furevik 2001), while reductions in the efficiency of AW cooling in the Barents Sea (due to changes in local winds and increases in air temperature) have further increased the heat transport through the St. Anna Trough (Skagseth et al. 2020). Furthermore, a reduction in sea ice import to the Barents Sea can cause a local increase in AW temperature, salinity, and hence density (Barton et al. 2018). Increases in AW volume transport have also been attributed to the reduction in sea ice export south through the Fram Strait and the resultant reduction in the halosteric height of the Greenland Sea (Wang et al. 2020).

1.4.3 Stratification

Recent reductions in stratification strength in the eastern Arctic have accompanied the increases in AW heat content described above. In the eastern Eurasian Basin, AW temperature increases have occurred alongside AW shoaling, a weakening halocline, and enhanced verti-

cal heat transfer from the AW to the surface layer. Vertical heat flux from the AW was in fact two to four times larger in 2014–2015 than in 2007–2008 (Polyakov et al. 2017). These changes have been ongoing since the 1970s, and have resulted in a reduction in winter sea ice formation and deepening winter convection (Lind et al. 2018; Polyakov et al. 2017; Polyakov et al. 2010). This process is thought to have preconditioned the ice for the record (at the time) summer Arctic sea ice retreat of 2007, by thinning it and making it more susceptible to wind and atmospheric thermodynamic forcing (Polyakov et al. 2010).

This process, known as “Atlantification”, reflects a north-eastward progression of AW influence in the Arctic, with conditions in the eastern Eurasian Basin seemingly transitioning to become more like those in the western Nansen Basin near the AW inflow, where AW exists at/near the surface (Polyakov et al. 2017). The increase in convective entrainment during winter due to the reduction in sea ice is a positive feedback loop driving further local increases in vertical heat flux from the AW layer. One model study suggests that this northward progression of AW influence on the upper water column will reach the Lomonosov Ridge by 2070, resulting in a 1.2 *m* reduction in winter sea ice thickness and an increase in ocean-to-ice heat flux in the Eurasian Basin from present values of around 0.5 Wm^{-2} to 5 Wm^{-2} (Årthun et al. 2019).

Atlantification of the Arctic also represents a northward progression of alpha oceans. Alpha oceans are stratified mainly by temperature, owing to their warmth (typically $> 10\text{ }^{\circ}C$) and the nonlinear increase of the thermal expansion coefficient of sea water with temperature (Carmack 2007). As mentioned, the Arctic is a beta ocean. Given that the front between alpha and beta oceans marks the most southerly latitude where sea ice can form, Atlantification has important consequences for sea ice extent. Furthermore, the difference in the nature of convection in alpha and beta oceans is important for heat flux to the Arctic Ocean surface – the warm, saline AW would be less dense than the cool, fresh polar water in an alpha ocean, and hence further increases in AW ventilation and vertical heat flux would be expected.

1.4.4 Forced trends and natural variability in the AW layer

Natural (or internal) variability in the climate system can act to strengthen or mitigate any forced trends, such as those from climate change, and arises due to inherent random fluctuations in large scale atmospheric or oceanic circulations, and changes in ocean dynamics. An

example might be a cool summer in spite of an overall increase in long-term air temperatures.

As mentioned, natural variability is present in AW temperature records, with oscillations in AW temperature spanning several years to millennia. Establishing whether any trends in AW properties or circulation are part of this natural variability or lie outside of it as an externally forced trend is key in determining whether the changing climate is contributing to these trends. Observations have already shown evidence of AW temperature increases outside of the range of historic internal variability (Polyakov et al. 2012b) (see Figure 1.9).

Natural variability appears to have played a more important role in recent reductions in sea ice cover, being responsible for approximately half of the September sea ice extent decline to date, with this proportion varying across the Arctic (England et al. 2019; Kay et al. 2011). Internal variability in AW inflow via the Barents Sea (itself driven by variability in the AMOC and atmospheric circulation) has played an important role in this, and is expected to continue to do so throughout much of this century (Årthun et al. 2019). Using an ensemble of runs of the same model, each perturbed by a small change in initial conditions, is a useful way of assessing natural climate variability, as demonstrated by many of the studies cited in this subsection.

1.4.5 Consequences of a changing Arctic

Many of the local and wider scale consequences of these Arctic changes have been described above. Sea ice loss, in terms of extent, area and volume in both winter and summer, is a key change local to the Arctic (Kwok 2018). As mentioned earlier, the resultant decrease in ocean surface albedo has already resulted in a notable increase in Earth’s radiative heating (Pistone et al. 2014) - a more wide scale impact. Sea ice loss and the resulting increase in freshwater export can cause a freshening of the North Atlantic, with the potential to weaken North Atlantic overturning (Jahn et al. 2013). Deeper Arctic outflows, namely those of AW, could also reduce North Atlantic deep water density, as AW density anomalies within the Arctic are able to be maintained throughout their advection within, and out of, the Arctic (Karcher et al. 2011). These could reduce the density of the Denmark Strait overflow (Jeansson et al. 2008; Karcher et al. 2011), reduce boundary densities southeast of Greenland (Hirschi et al. 2007), and weaken surface-forced water mass transformation in the Irminger and Labrador Seas (Petit et al. 2021; Yang et al. 2016) - any of which could weaken the AMOC.

Returning to the Arctic itself, a potential introduction and spread of alpha stratification in the Arctic would not only increase vertical heat flux from the AW layer towards the surface, but would also result in changes in ocean biogeochemistry and productivity, as detailed in Carmack (2007) and Polyakov et al. (2020a). Polyakov et al. (2020a) found that the Atlantification of the eastern Arctic, increasing influence of Pacific Water on the western Arctic halocline, and reduction in Arctic sea ice, have resulted in primary productivity increases over recent years in large parts of the Arctic. The increased stratification of the Amerasian Basin limits nutrient flux to the surface here, which does not favour an increase in productivity. The opposite is true in the Eurasian Basin. However, eddies and Pacific Water pathways also play a key role in supplying nutrients to the upper halocline of the Amerasian Arctic, and can drive local increases in nutrients (Mathis et al. 2007; Polyakov et al. 2020a). Specific examples of biological changes include increases in zooplankton biomass in the Chukchi Sea driven by increases in Pacific Water inflow (Polyakov et al. 2020a); a shift to a dominance of boreal species of zooplankton, pelagic fish and marine mammals over Arctic species in the northern Barents Sea (Eriksen et al. 2017; Kovacs et al. 2011); and an eastward shift in the range of Atlantic cod along the outer shelf of the Eurasian Basin (Hollowed et al. 2013).

1.5 Arctic Ocean observational challenges and model performance

Given the perennial and seasonal sea ice covering much of the Arctic, both remote sensing and the collection of ship-based measurements of the Arctic Ocean are challenging. As a consequence, the Arctic has historically been less well sampled than other oceans. There are very few Argo floats in the region, owing to the sea ice, but conductivity-temperature-depth profiles taken by equipment mounted within sea ice (known as ice-tethered profilers) have resulted in a significant growth in year-round observations (Krishfield et al. 2008; Toole et al. 2011). There are a handful of moorings at continental shelves in both the eastern and western Arctic which also provide year-round measurements, along with more summer-biased ship-based measurements. Overall, the general lack of observational data – particularly preceding the 21st century – together with its temporal and seasonal inhomogeneity present a challenge for pan-Arctic scale observational analysis of the ocean.

More spatially and temporally homogeneous data can of course be obtained from models.

However, as is often the case, biases are present in ocean general circulation models (OGCMs) of the Arctic. Generally, OGCMs produce an AW layer that is too thick and too deep (Ilicak et al. 2016). The coupled climate models used in the Coupled Model Intercomparison Project 5 (CMIP5) have the same issue, along with biases and low inter-annual variability in AW temperature, underestimated heat transport into the Barents Sea, overestimated sea ice extent, and a large inter-model spread in hydrography (Li et al. 2017; Shu et al. 2019). There is little improvement in the Arctic hydrography of the models in CMIP6, with similar problems persisting (Khosravi et al. 2022). Gaining a better observational picture of the Arctic Ocean is an important step towards improving and bench-marking these models, and is a key motivating factor behind further-enhancing observational coverage of the Arctic Ocean.

Despite the biases described above, the CMIP5 multi-model mean is able to reproduce climatological patterns of the temperature and depth of the AW core (Shu et al. 2019). One model involved in CMIP5 is the Community Earth System Model version 1 (CESM1). We use data from an ensemble of runs of this model - the CESM Large Ensemble (CESM-LE) in our analysis. The CESM1 (and CESM-LE, specifically) perform generally well in the Arctic, with CESM1 reproducing observed Arctic warming well (Jahn 2018), and CESM-LE capturing inter-annual variability in sea ice (Barnhart et al. 2015; England et al. 2019) and producing a realistic sea ice cover response to AW heat inflow (Årthun et al. 2019). Therefore, although biases should be kept in mind when interpreting model representations of the Arctic, models are a very useful tool for studying the Arctic Ocean.

1.6 Thesis Aims

This thesis will focus on building a picture of the current state and likely future changes of the AW layer within the Arctic Ocean, given the emerging importance of its role in sea ice loss and recent reports of increases in its temperature detailed above. This is crucial, not only due to the importance of AW within the Arctic, but also because of its impact downstream after exiting the Arctic, and the wider impacts of Arctic sea ice loss.

This will help to fill a key gap in current research. Few studies have taken a pan-Arctic view of the AW layer and assessed how the layer itself varies spatially and temporally based on observations, or have used model data to investigate the future evolution of pan-Arctic

AW and the wider Arctic Ocean water column under a changing climate.

The main aims of this thesis are therefore:

- To build a detailed pan-Arctic observational picture of the AW layer. This will be used to compare and contrast temporal changes in AW in the eastern and western Arctic. It will also be used to assess the merit of metrics commonly used to describe the AW layer.
- To investigate recent and projected future changes in the AW layer during the 21st century in a coupled climate model under both historical and RCP 8.5 forcing (where RCP 8.5 stands for Representative Concentration Pathway 8.5, and represents a climate subject to continuous increases in greenhouse gas emissions that would result in a radiative forcing increase of 8.5 Wm^{-2} between pre-industrial times and 2100).
- To determine whether any trends in AW layer properties are the result of natural variability or are driven by external anthropogenic forcing, using a large ensemble of model runs.
- To assess the impact of any AW changes on the upper water column and Arctic Ocean stratification.
- Overall, to assess the fate of AW heat within the Arctic, and how this fate might change in the future.

1.7 Thesis Structure

The structure of this thesis is as follows. Chapter 2 will present a pan-Arctic observational picture of AW within the Arctic using 40 years of observations spanning 1970s to 2018. This will not only provide a strong basis for the remaining analysis, but will give key insight into how the AW heat, salinity, depth and density vary in time and space across the whole basin. The impact of AW heat on the upper water column and accuracy of common key metrics used to assess the AW layer will also be investigated. This chapter is published in the Journal of Geophysical Research as the manuscript “Spatial and Temporal Variability of Atlantic Water in the Arctic From 40 Years of Observations” on which I took the lead. Chapter 3 moves on to model data. Here, the Community Earth System Model Large Ensemble (CESM-LE) will

be used to identify trends in AW layer properties within the Arctic and investigate the causes of any changes in AW heat content. The ensemble of model runs will be utilised to determine whether any trends in AW properties are part of the natural variability of the climate system or are driven by anthropogenically forced trends (the CESM-LE uses a RCP 8.5 forcing for much of the 21st century). The impact of any AW changes on the halocline, surface layer, and stratification of the Arctic Ocean will be investigated using the same model ensemble in Chapter 4. This chapter will answer important questions on the potential impact of AW changes on both sea ice loss and other key Arctic hydrography. This is particularly important given that the true role of AW heat in a changing Arctic heavily depends on the evolution of the strong stratification provided by the halocline. Chapter 5 then pulls the findings from all three results chapters together, providing a combined observational and model perspective on the state of the AW layer across the Arctic, and summarising how the fate of AW heat in the Arctic may evolve in a changing climate.

Chapter 2

Spatial and temporal variability of Atlantic Water in the Arctic from forty years of observations

2.1 Introduction

Atlantic Water (AW) is the most significant source of oceanic heat in the Arctic (Carmack et al. 2015). As detailed in Chapter 1, AW warming and shoaling has been linked to the drastic sea ice loss observed in the Arctic in recent decades (Mulwijk et al. 2018; Polyakov et al. 2017; Polyakov et al. 2012b; Polyakov et al. 2010; Spielhagen et al. 2011). Obtaining a detailed picture of the AW layer throughout the Arctic and how it has evolved over recent decades in different regions, and impacted the water column, is therefore important in expanding our understanding of how AW heat influences sea ice across the Arctic region. This will also help us understand more about the processes that influence AW heat within the Arctic Ocean itself.

Understanding how AW heat is likely to change in the future is a key part of predicting what will happen to the Arctic in the years to come. Models, of course, will play a key part in this. However, as described in Chapter 1, there is large discrepancy and bias amongst coupled climate model representations of AW in the Arctic, with the AW layer generally being too deep and thick (Ilıcak et al. 2016). The AW temperature biases are primarily due to inaccurate representation of sea ice coverage and surface cooling in the Barents Sea,

formation of cold and dense water in the Barents Sea, and AW inflow temperatures through the Fram Strait (Ilıcak et al. 2016; Shu et al. 2019). It is therefore particularly important to have an observational description of AW to help evaluate these models.

Oceanic observations in the Arctic are sparse and often seasonally biased, and many observational studies of the Arctic focus on specific regions or transects (e.g. Anderson et al. (1994), Beszczynska-Moller et al. (2012), Li et al. (2020), Lind et al. (2018), McLaughlin et al. (2009), and Polyakov et al. (2020b)). However, the number of Arctic Ocean observations has increased in recent years. This chapter aims to synthesise data from various sources across the Arctic from the 1970s to 2018 to give a pan-Arctic, up-to-date description of the AW layer and its impact on the rest of the water column. Diagnostics derived from these observations, such as the temperature, salinity and depth at the AW temperature maximum (the AW core) and AW heat content are used to characterise the spatial and temporal variability of the AW and are described in section 2. The spatial variability of the AW properties is investigated in section 3, with temporal variability in both the eastern and western Arctic described in section 4. Observed changes in the AW and the heat distribution within the water column at moorings and at repeated CTD transects are discussed in section 5. Section 6 explores regional correlations between AW core metrics and vertically integrated AW layer properties to investigate both how representative the AW core temperature is of AW heat content, and regional differences in mixing. Conclusions are given in section 7.

2.2 Data and Methods

Conductivity-temperature-depth (CTD) observations from across the Arctic are combined from four different sources: the ice-tethered profiler (ITP) program ((Krishfield et al. 2008; Toole et al. 2011), <http://www.whoi.edu/itp>) and the Beaufort Gyre Exploration Project (BGEP, <https://www.whoi.edu/beaufortgyre>) - both based at the Woods Hole Oceanographic Institution - the Nansen and Amundsen Basins Observational System (NABOS, <https://uaf-iarc.org/nabos>), and data from the NOAA World Ocean Database (WOD, <https://www.ncei.noaa.gov/products/world-ocean-database>, (Boyer 2018)). The WOD collates oceanic observational data from a wide range of sources. The WOD data used here are those from CTD profiles, drifting buoys, and ocean stations. Any ITP, BGEP or NABOS data were removed from the WOD dataset before use to avoid duplication. The BGEP and NABOS

datasets include data from both moorings and ship surveys.

Throughout this chapter, salinity is given in Practical Salinity Units, and potential temperature (when not available directly from the observational data product), heat content and potential density are computed using the Thermodynamic Equation of Seawater 2010 (TEOS-10) (IOC et al. 2010).

All data used in this chapter are processed versions of the raw data gathered in the field. Details of these procedures can be found in the sources referenced above, but all involved calibration, sensor-correction and the removal or flagging of obviously erroneous data. In addition to this initial processing, further routines were applied to the data and profiles were smoothed for much of our analysis - details of which are given below. Profiles with more than 10 % of data masked or flagged as suspicious were omitted from the analysis and, unless otherwise stated, monthly mean data from moorings were used so as not to bias any regional analysis to the mooring locations due to the relative high sampling frequency here compared to other locations. This resulted in about 59000 profiles for analysis.

Here we define the Atlantic Water layer as that portion of the water column that lies below the halocline and has potential temperature above 0 °C. The top and bottom of the layer are the 0 °C crossing points, either side of the potential temperature maximum, and the AW layer thickness is the distance between them. As density is driven by salinity in the Arctic, potential temperature is effectively a passive tracer. The potential temperature maximum, referred to here as the AW core, is commonly used to follow the circulation and transformations of AW (Karcher et al. 2003), and is the focus of much of this manuscript.

The AW core in a given profile is defined as the point at which the maximum potential temperature is reached in the portion of the profile with salinity greater than 34.7 (in order to avoid surface temperature maxima), following (Lique et al. 2012). Figure 2.1 shows the distribution in time and space of profiles where the AW core was identified, with mooring locations shown as black squares.

To ensure the AW core identified in each profile was not an artefact of limited sampling, profiles were required to start above 100 *m* and cover a depth range of at least 500 *m* before being used to identify the AW core (this also eliminated data from the surrounding shelf seas, allowing us to focus on AW within the Arctic basin only). This latter step resulted in about 44000 AW core data points. Before identifying the core, profiles were smoothed

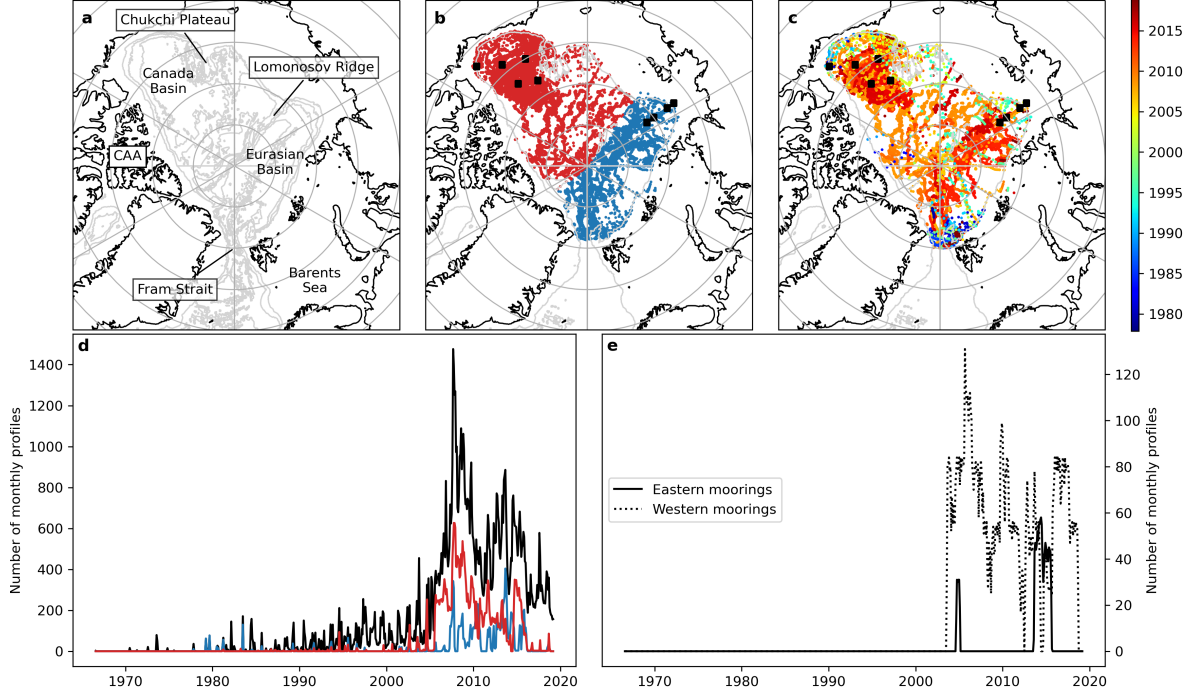


Figure 2.1: (a) Map of the Arctic with bathymetric contours for every 1000 m shown along with relevant geographic features. (b) Spatial distribution of all AW core data points coloured by region, with east in blue, west in red, and mooring locations marked with black squares. (c) Spatial distribution of all AW core data points coloured by the year in which the measurements were obtained. (d) Time series showing number of monthly profiles, with east in blue, west in red and all Arctic data in black (some of which lay outside of both the eastern and western regions, so are not shown on map or used in analysis). (e) Time series showing number of monthly mooring profiles in the eastern and western regions.

over a vertical distance of 80 m by taking the mean of the profile data within 40 m of each data point. This length scale was chosen as it was the best at preserving the general shape and magnitude of the temperature profile, while removing the spikes due to features such as thermohaline intrusions and eddies (although please note that this smoothing was not applied to the profiles in Figure 2.4). This is important as the main aim of this chapter is to get a general picture of patterns and long-term trends in AW core properties, and features such as intrusions can disproportionately affect the depth of the AW temperature maximum in basin interiors in such a way as to detract from this. Figure 2.2 shows the impact of different smoothing lengths on a temperature profile from the Canada Basin in April 2006. The jagged original profile is shown in black, with various smoothing lengths used for the blue and orange profiles. The larger the smoothing length, the less accurate the magnitude of the AW core temperature, with a shorter smoothing length in this case leading to a less accurate AW core depth. An 80 m smoothing length was the best compromise. Any profiles

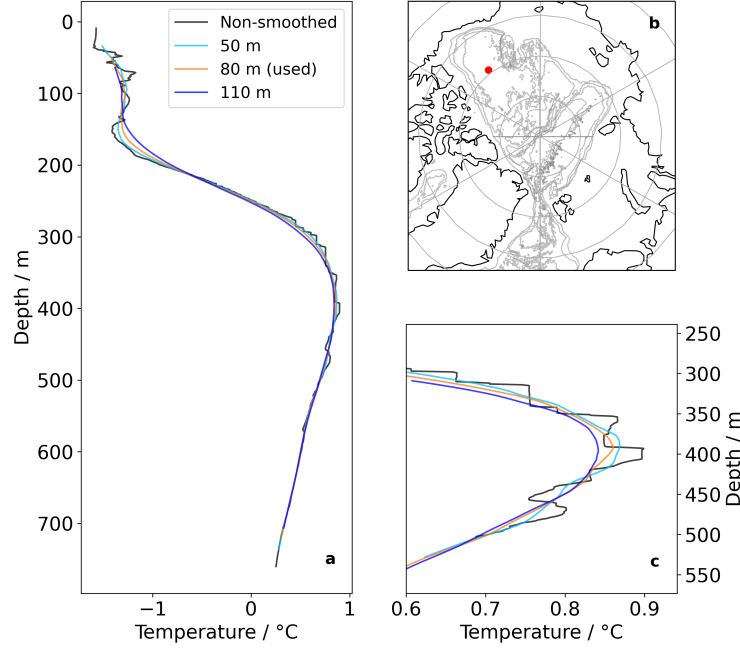


Figure 2.2: (a) A profile from April 2006 in the Canada Basin in its original form (black), smoothed over 50 *m* (light blue), smoothed over 80 *m* (orange), and smoothed over 110 *m* (dark blue). (b) Profile location is shown with a red dot. (c) More detailed profiles around the AW core depth.

used in the analysis should be assumed to be smoothed unless stated otherwise.

The depth coverage varies between data sources - ITP profiles cover the upper 800 *m* of the water column, whereas many data from CTD stations and moorings extend down to around 2000 *m*. This variation in depth range does not affect the analysis in this chapter given that the AW core can be identified in both cases, and any profiles that do not sample the whole AW layer are omitted from the AW heat content analysis in sections 5 and 6. Although ITPs and moorings provide year-round measurements, there remains a spring/summer bias in data from other sources. However, this is unlikely to impact results due to the negligible seasonality of AW when compared to its overall variability in space and time (Lique et al. 2012).

Heat content, HC, was computed for various portions of the vertical temperature profiles according to

$$HC = \int_a^b \rho_\theta(z) c_p T(z) dz \quad (2.1)$$

where ρ_θ (in kgm^{-3}) is potential density, c_p (in $Jkg^{-1} °C^{-1}$) is specific heat, T (in $°C$) is

potential temperature, z (in m) is depth, and a and b are the depth bounds defining the layer in question. Approximately 1500 profiles sampled the entire AW layer and allowed for the computation of total AW heat content. To account for differing profile lengths above the AW layer (due to variation in upper depths of profiles and the depth of the AW layer itself), heat content density is used to evaluate the heat stored in this upper layer in a similar way to (Polyakov et al. 2011), where the heat content derived for this upper portion of the water column is divided by the depth range over which it is computed. This quantity is proportional to the average temperature over that depth range.

2.3 Results and Discussion

2.3.1 The Atlantic Water core across the Arctic

Investigating how the hydrographic properties of the AW core change across the Arctic can give a good picture of the behaviour of the AW layer in general. Figure 2.3 shows maps of the potential temperature, salinity, depth and potential density anomaly of the AW core from all observations, giving an idea of the general spatial distribution of properties at the depth of the AW temperature maximum.

Figure 2.3b highlights the temperature difference between AW core properties in the Eurasian Basin and western Arctic. The core loses heat as it is advected around the basin - its temperature in the Canada Basin is approximately 1.5 °C lower than where it is first subducted under the fresh surface layer at the southern boundary of the Eurasian Basin. The most significant heat loss is seen here, where, much like in the Nordic Seas (Lind et al. 2018), the AW loses heat to the atmosphere and through mixing with the cooler surface layer. The higher AW core temperature along the Lomonosov Ridge relative to the western Arctic boundary suggests that the AW that recirculates back along the ridge is warmer than that which continues towards the western Arctic.

The salinity of the core (Figure 2.3c) decreases on its journey around the Arctic, particularly in the Eurasian Basin where it mixes with fresher surface waters upon subduction. Turbulent mixing may play an important role in AW freshening in parts of the western Arctic - the difference in AW core salinity (and temperature) between the boundary and interior of the Canada Basin is indicative of this. Whereas the AW in the interior of the Canada

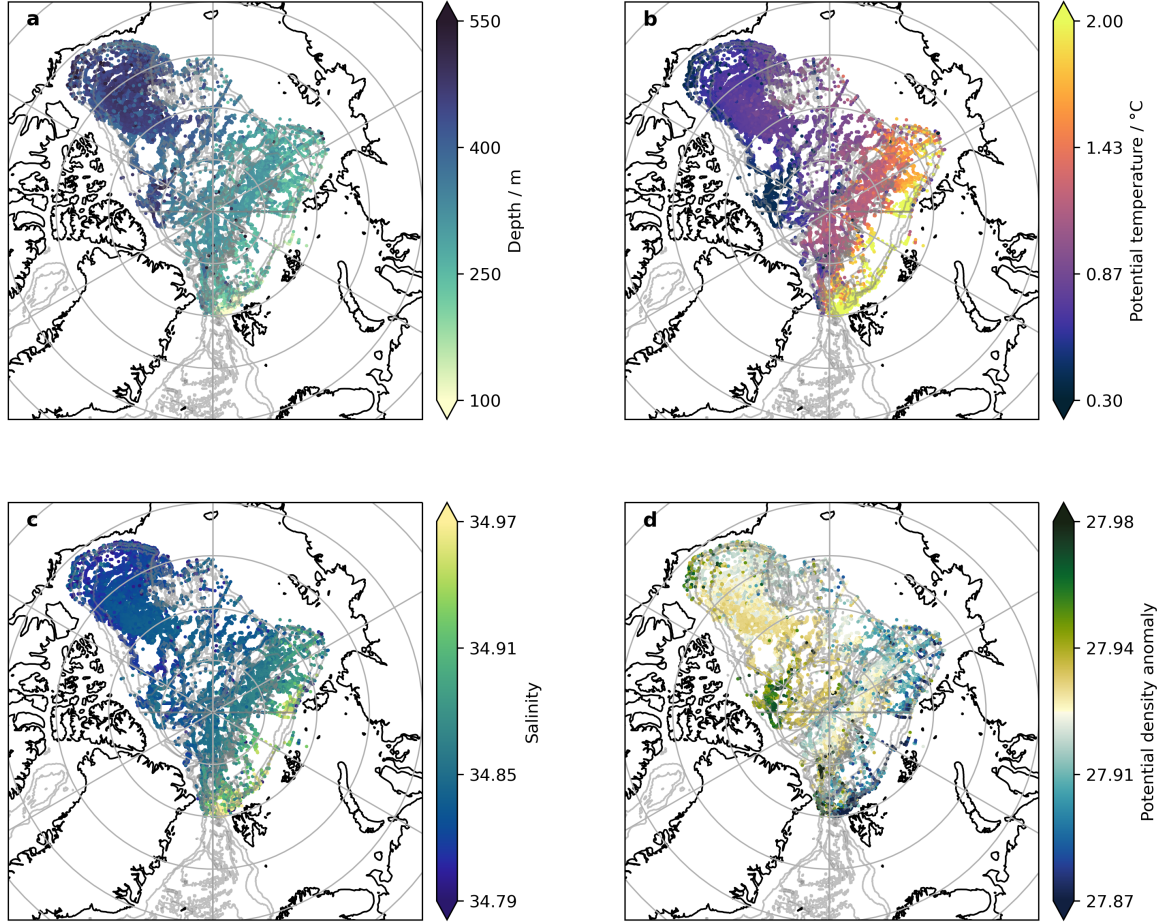


Figure 2.3: Maps showing (a) depth (b) potential temperature (c) salinity and (d) potential density anomaly of all AW core data points used in this chapter.

Basin has travelled around the north of the Chukchi Plateau, the AW at the boundary has travelled over the Chukchi Plateau's complex bathymetry (Li et al. 2020; McLaughlin et al. 2009). The relatively low temperature and salinity of this boundary AW can be explained by enhanced mixing experienced over this rough topography upstream. Another factor that likely contributes to the differences in AW at the interior and boundary of the Canada Basin is the presence of a cold, fresh Pacific Water boundary current at the boundary which likely increases vertical heat and salt fluxes from the AW layer due to its impact on temperature and salinity gradients here.

Despite the AW core freshening along the AW advection pathway, the density of the core appears to increase from its relatively low value along the southern Eurasian Basin boundary to higher values in the western Arctic and Eurasian Basin interior (Figure 2.3d). This is because AW core cooling upon advection has a greater effect on core density than the core freshening does due to the large magnitude of the cooling relative to the freshening. There are

particularly dense, cold regions along the western shelves just north of the Canadian Arctic Archipelago (CAA) and Greenland. Heat loss from the top of the AW layer to the fresher, cooler water above through vertical mixing upon AW advection could explain this AW core cooling and freshening. This would deepen the AW core without the AW layer as a whole getting denser, as heat lost preferentially from the upper AW would result in the core being found deeper within the AW layer itself (and on denser isopycnals due to its cooling). This could explain the cooling, freshening and increase in density of the AW core seen along the AW advection pathway, with the coldest and densest regions towards the end of its journey (Aksenov et al. 2011).

Another process which might be contributing to the cooling and increase in density of the AW core during its advection is interaction between AW and cold, dense cascading flows from the shelves (formed from brine rejection during sea ice formation). These flows have been modelled throughout the Arctic (Luneva et al. 2020), and have been observed to interact with and modify AW in the Eurasian Basin (Ivanov et al. 2007). Observations of this interaction in the western Arctic are limited due to the sparsity of observations in these shelf regions, but dense water cascades have been observed to interact with waters as deep as the lower halocline here (Ivanov et al. 2004; Luneva et al. 2020; Melling et al. 1995), suggesting that interaction with upper AW is feasible.

Due to limitations in our dataset, we are unable to quantify the relative importance of these two potential mechanisms through which the properties of the AW core are modified. Although observational evidence of interaction between shelf flows and AW is limited in the western Arctic, it is likely that both heat loss from the top of the AW layer and interaction with shelf flows contribute to the AW core becoming cooler and denser along the AW advection pathway.

The AW core depth exhibits a bimodal structure, as shown in Figure 2.3a, being much deeper in the Canada Basin (approx. 500 *m*) than the Eurasian Basin (approx. 300 *m*) due to the Ekman pumping associated with the winds which drive the Beaufort Gyre. The effect of the Beaufort Gyre on the AW in the Canada Basin can also be seen in the (un-smoothed) vertical temperature profiles in Figure 2.4, where the cool waters of the gyre push down the AW layer to a much greater depth than that at which it resides in the eastern Arctic. However, the important role that the halocline plays in isolating the AW from the surface

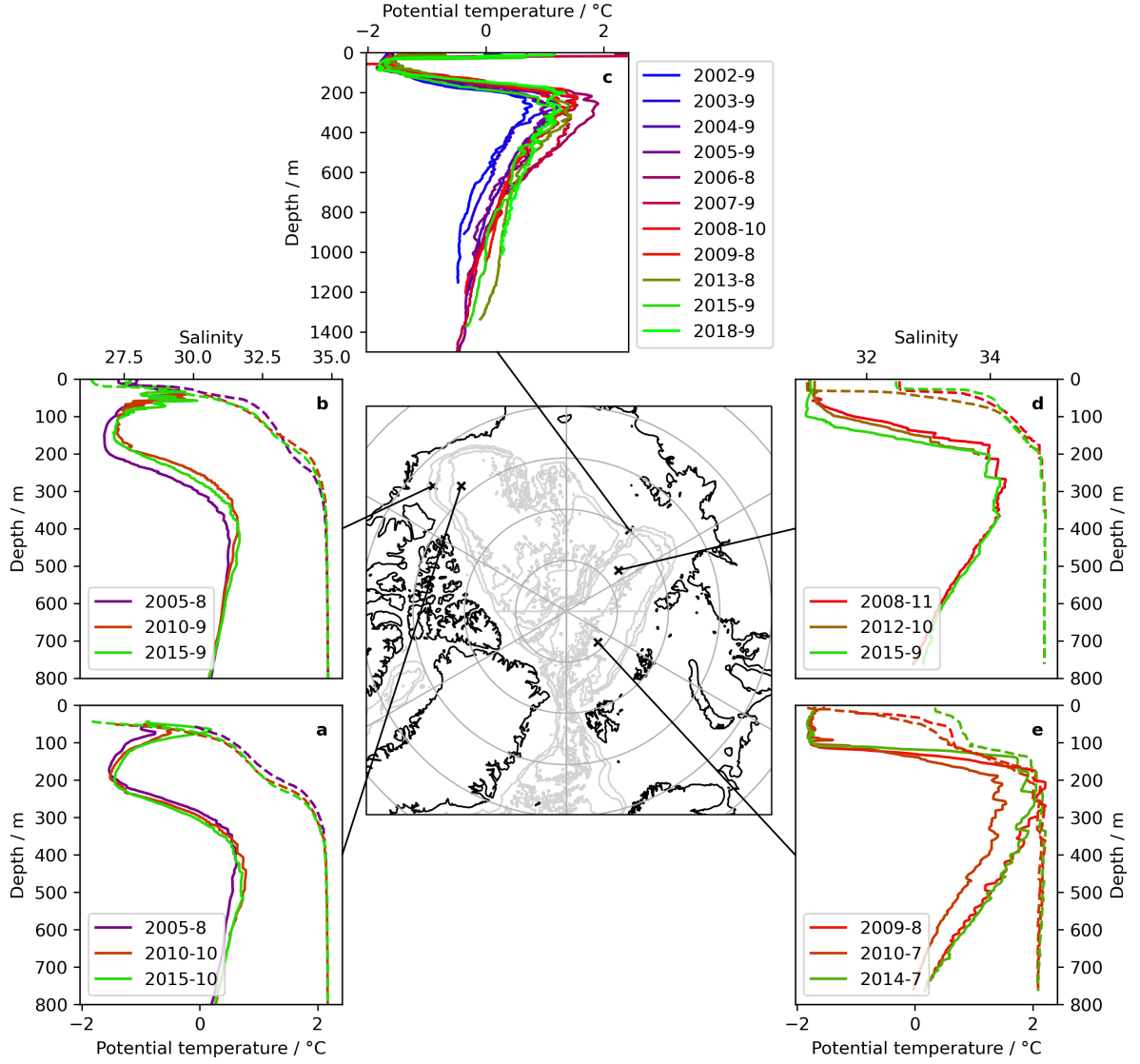


Figure 2.4: Un-smoothed potential temperature (solid) and salinity (dashed) profiles (a) in Canada Basin interior, (b) on Canada Basin shelf, (c) at the Siberian end of Lomonosov Ridge, (d) in the eastern Eurasian Basin, and (e) in the western Eurasian Basin. Profiles are coloured by the year in which they were measured, with year and month given in the legend. Note the change in y-axis scale in (c).

can be observed across the whole Arctic (see Figure 2.4).

Zig-zags and staircase features in these un-smoothed profiles also indicate the presence of thermohaline intrusions and diffusive convection, respectively (Bebieva et al. 2017). Thermohaline intrusions form in the presence of temperature and salinity gradients along isopycnals (Ruddick 1992), and are an important mechanism for AW transport from the boundary to the interior of both the Canada and Eurasian Basins (Kuzmina et al. 2011; McLaughlin et al. 2009). They are often found near the AW core depth (unlike staircases which tend to be found above the AW core depth where diffusive convection is supported (Bebieva et al. 2019)).

These intrusion signatures are seen in Canadian Basin data in the 2000s. Intrusions are also seen in the Eurasian Basin throughout the time period covered in this chapter, providing strong evidence for their long-term presence in this region.

2.3.2 Regional differences in Atlantic Water properties and their temporal variability

Although the maps in Figure 2.3 give an idea of the general spatial variability of AW, they do not indicate how the AW has changed over the time period studied. This section will explore the temporal variability observed in the AW layer in different regions of the Arctic.

Figure 2.4 shows profiles from the same locations measured in different years. The uppermost plot in Figure 2.4 uses profiles from moorings at the Siberian end of the Lomonosov Ridge, sampling the AW boundary current. The depth range sampled by the moorings captures both AW branches - the Fram Strait Branch Water (FSBW) and the Barents Sea Branch Water (BSBW), centered at around 200-500 *m* and 750-1000 *m*, respectively. The temperature of these two branches here appears to vary independently in time. The BSBW shows a general warming trend throughout the period sampled (although this conclusion is tentative as we cannot rule out an influence of transient features such as eddies). This warming could hint at a more systematic change in BSBW temperature, which could be explained by surface air temperature increases over the Barents Sea (Skagseth et al. 2020) or reductions in sea ice import to the region (Lind et al. 2018). The FSBW temperature is more variable from year to year, reflecting the variability of AW inflow temperature through the Fram Strait (Ivanov et al. 2012), although local transient features may also play a role in these profile differences. The heat loss experienced by the AW in the Barents Sea may act as a buffer for BSBW against high-frequency variation in upstream AW temperature.

Building on the regional differences in AW shown in Figure 2.3, Figures 2.5–2.8 highlight how the properties of the AW core in the eastern (blue) and western (red) Arctic evolve with time. Canada Basin mooring data is shown in black. Maps and annual normalised histograms show how the potential temperature, salinity, potential density anomaly and depth of the AW core change. The period covered by each map is indicated by grey lines enveloping the corresponding annual histograms - these periods were chosen to account for the varying amount of data available during each period. The reader is referred to Figure 2.1

for time series of the amount of data available from each region.

The year-to-year spatial variation in data distribution in the eastern Arctic makes inferring any trends from the histograms for the eastern Arctic difficult, and no significant trend can be found. After applying our processing scheme for core detection, only three years of eastern Arctic mooring data provides information on AW core properties (see Figure 2.1), so no meaningful trends can be identified from this fixed-point data either. However, the more consistent spatial distribution of data in the western Arctic and long time-series from mooring data allow trends to be inferred for the Canada Basin - these mooring data are described in more detail in the next section of this chapter. The differences in the range of temperature and salinity data between the east and west highlights the transformation undergone by the AW as it travels around the basin, reinforcing what was shown in Figure 2.3, with AW core salinity and temperature decreasing due to mixing, and AW core density increasing.

The mooring data in Figure 2.5 reveal a gradual cooling in the Canada Basin after 2002, presumably after the arrival of the AW core warm temperature anomaly which entered the Canada Basin in the early 2000s (after having entered the Arctic through the Fram Strait in 1990) (Li et al. 2020; McLaughlin et al. 2009). The maps in Figure 2.5 show the spread of this anomaly from the northern edge of the Chukchi Plateau into the interior of the Canada Basin in 2000–2004, with a more homogeneous AW core temperature field in 2005–2009.

Figures 2.6, 2.7 show an increase in AW core salinity and density in the interior of the Canada Basin throughout the mooring time period (2003–2018). This could be a downstream impact of Eurasian Basin Atlantification. Local increases in AW salinity in the Eurasian Basin and surrounding seas due to reduced Eurasian Basin sea ice have been documented (e.g. (Barton et al. 2018; Lind et al. 2018)), but the impact of these changes on AW downstream in the western Arctic is not yet fully understood.

Alternatively, an increase in the AW core salinity and density within the Canada Basin could be due to the AW core moving deeper as heat is transported away from it. Unlike at the boundary, where turbulent mixing results in similar changes in AW core temperature and salinity (Li et al. 2020; McLaughlin et al. 2009), in the interior diffusion governs the vertical transport of temperature and salinity. The diffusive convection regime in the upper AW layer (Bebieva et al. 2019) results in temperature being diffused more effectively than salt, causing the core to cool but not freshen. The warmest point is then located at a deeper, more saline

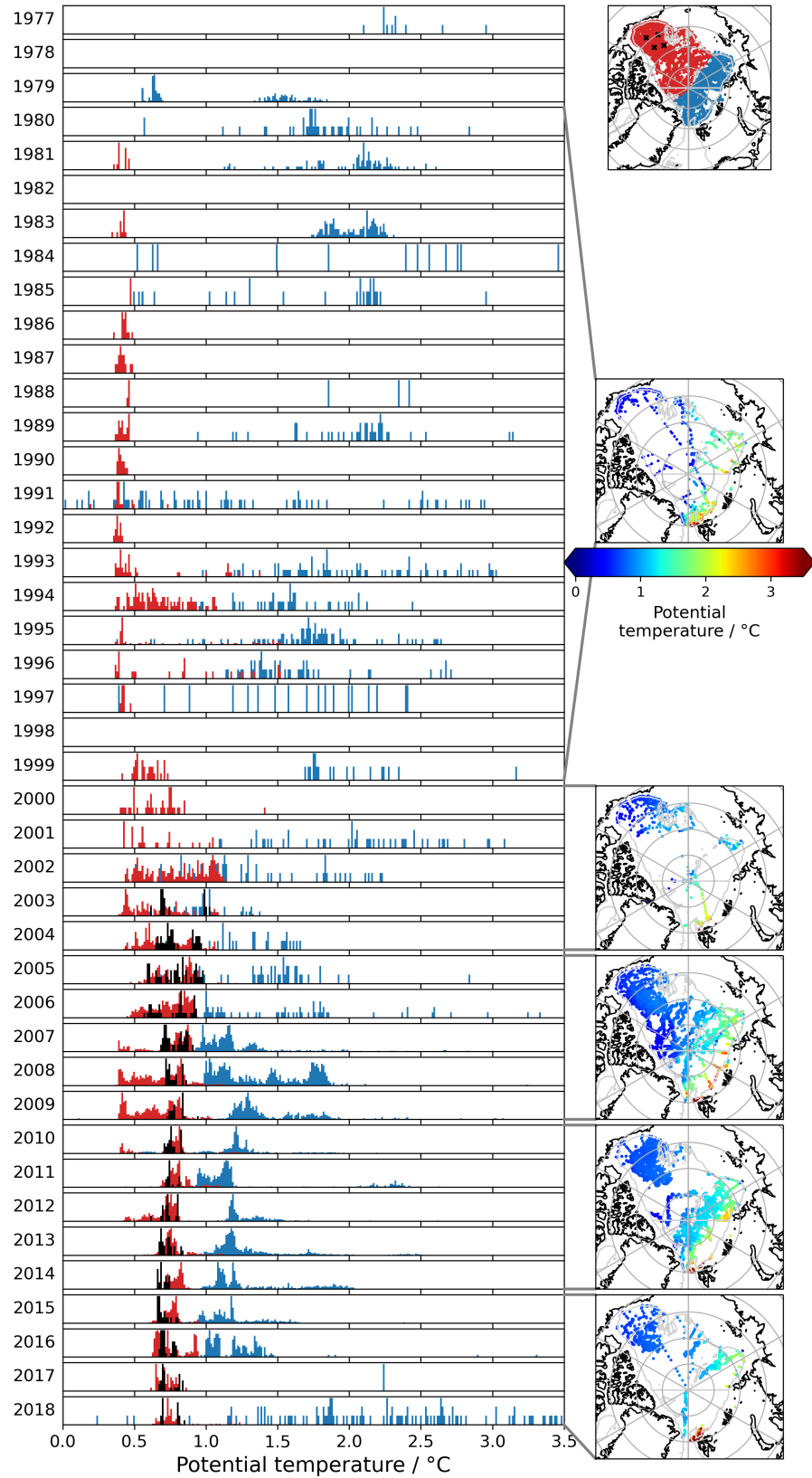


Figure 2.5: Annual normalised histograms of AW core potential temperature. Histogram data is coloured by region, with blue for eastern data, red for western data, and black for western mooring data - as shown in the map at the top right. The remaining maps on the right show the spatial distribution of the data in the histograms contained within each pair of grey lines.

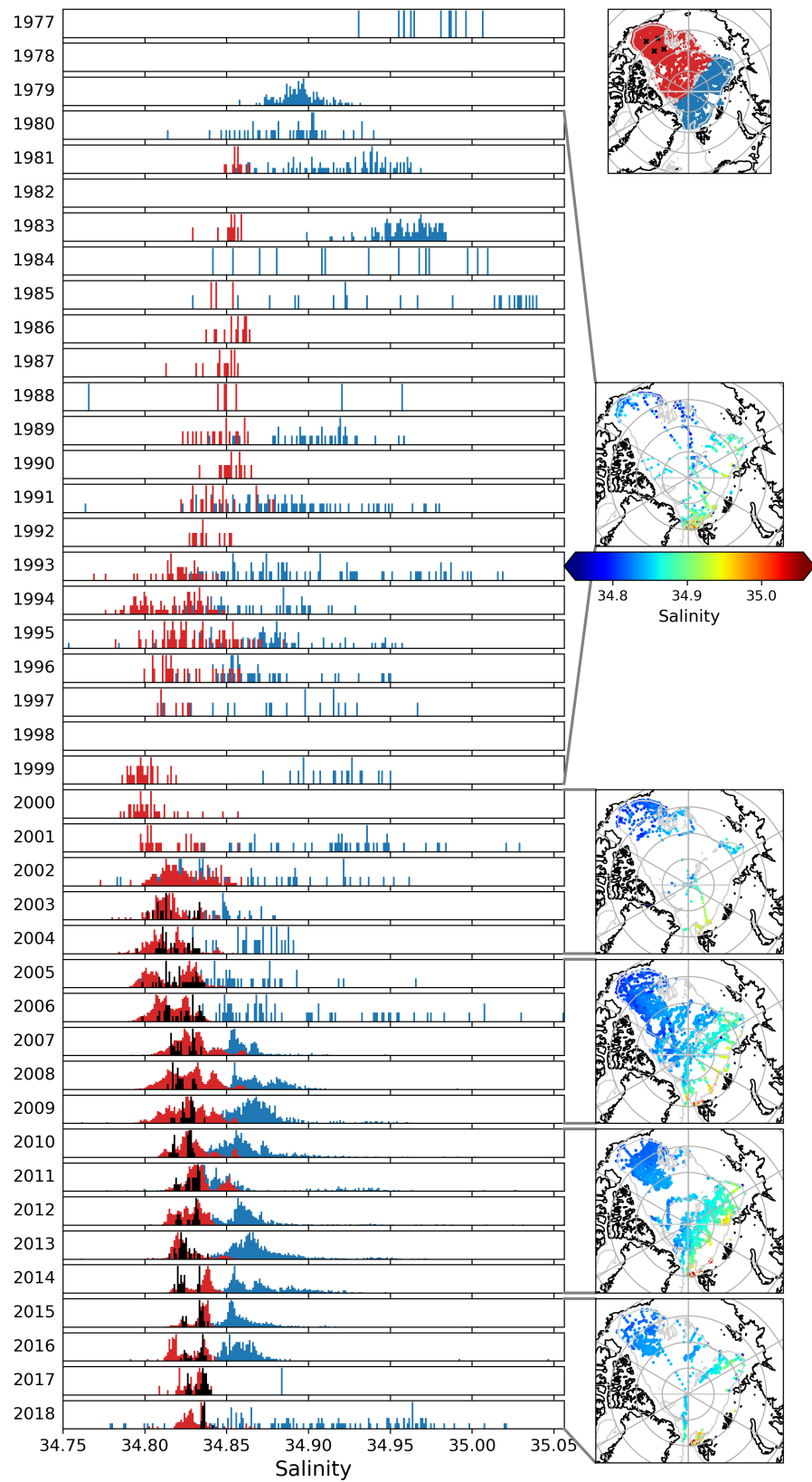


Figure 2.6: As Figure 2.5 for AW core salinity

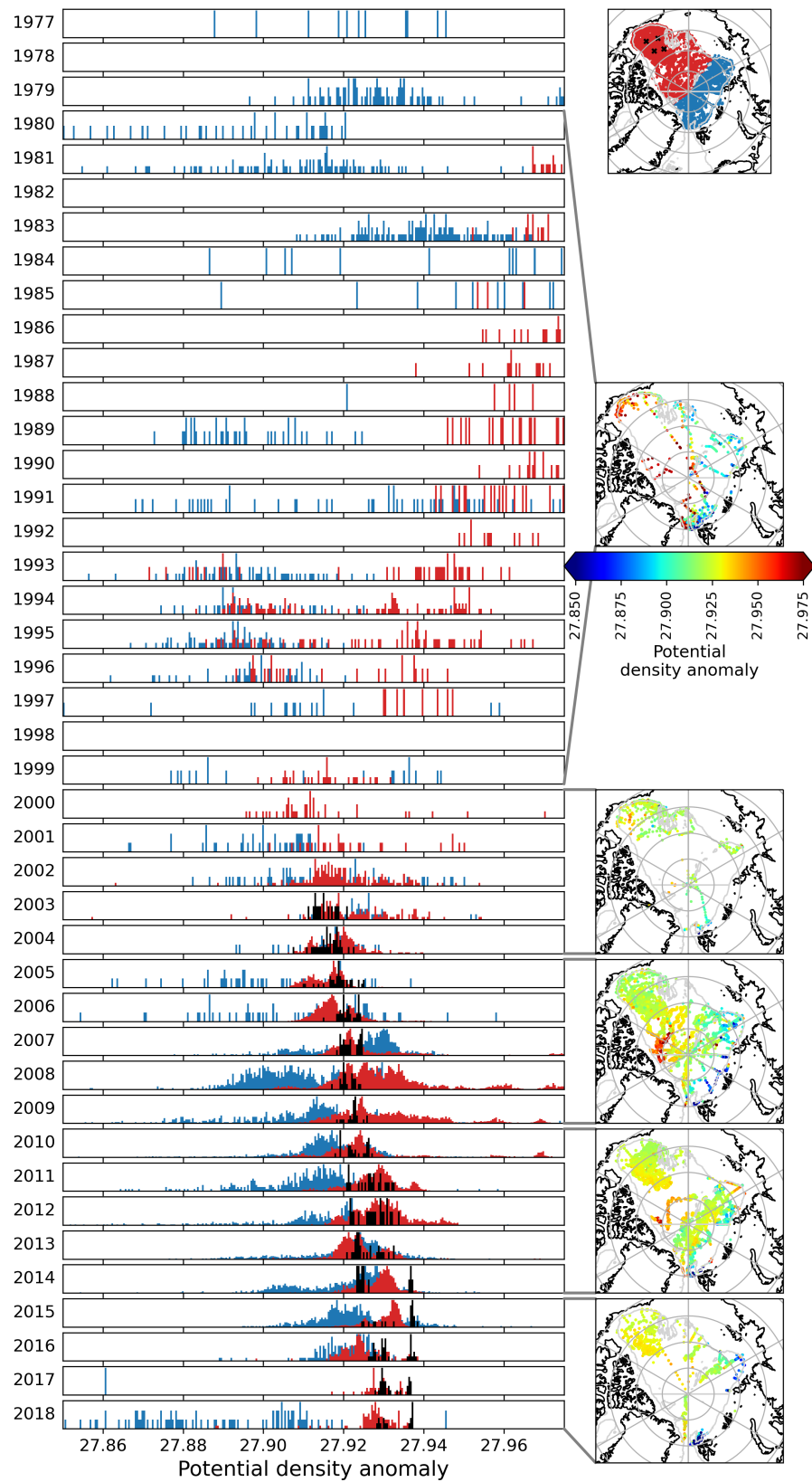


Figure 2.7: As Figure 2.5 for AW core potential density anomaly

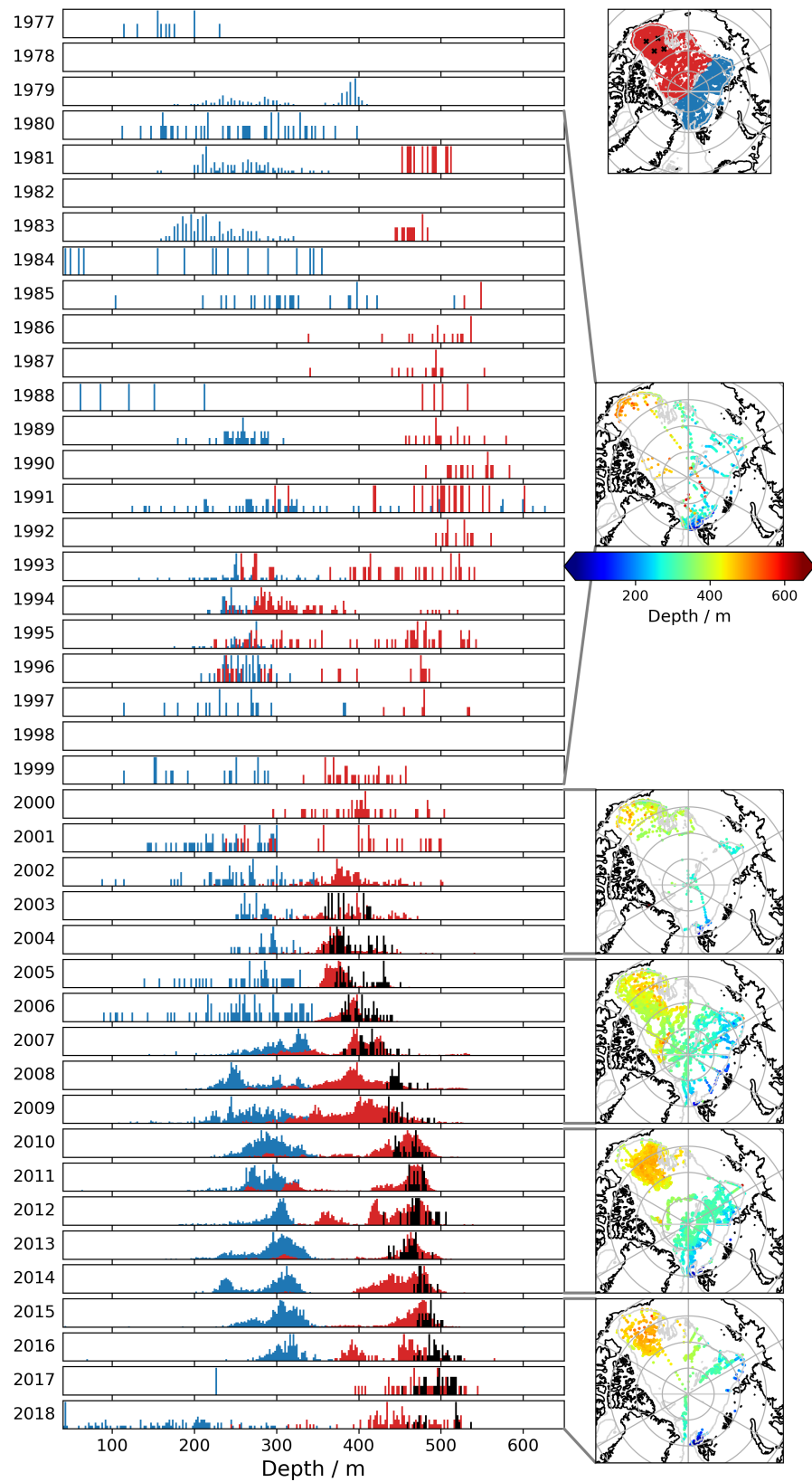


Figure 2.8: As Figure 2.5 for AW core depth

part of the AW.

The most prominent trend in these histograms is the increasing depth of the AW core in the Canada Basin mooring data (Figure 2.8). Zhong et al. (2014) showed that the AW deepening caused by the spin up of the Beaufort Gyre dominates over the influence of AW core density on depth if the gyre intensifies sufficiently, with AW position in relation to the gyre centre becoming more important than its density from 2007 onwards. This means that when the gyre is sufficiently intense, AW suppressed by the gyre can reside deeper than other AW that is denser (e.g. that at the Canada Basin boundary which the gyre does not suppress). Figure 2.8 shows that the deepening of the AW core coincides with the isopycnal deepening related to the Beaufort Gyre spin-up reported by Zhong et al. (2014), Proshutinsky et al. (2019) and others. The effect of the AW core potential density increase on this core deepening will be investigated in the following section of this chapter, where the whole water column will be considered.

2.3.3 AW variability at transects and moorings

Broad regional trends in AW core properties have been discussed above. To investigate these further, and put AW core property changes within the context of the wider water column, the temporal variability of data at moorings and across regularly repeated CTD transects is investigated below. This reveals more about the implications of AW changes for water column stratification and heat distribution. Trends from individual Canada Basin moorings (black data in Figures 2.5–2.8) are also discussed in more detail.

2.3.3.1 Eurasian Basin

The potential temperature and salinity along a NABOS CTD transect repeated from 2002–2018 across the AW boundary current in the eastern Eurasian Basin is shown in Figure 2.9. The year of each transect is given in the plot, and the AW core depth is identified with a black dot. The vertical black lines near the surface of the transects show the location of the CTD profiles. Data between these profiles have been interpolated using a Delaunay triangulation grid. The AW layer warms in general throughout the time period (when comparing the start and end years). However, this warming is pulse-like rather than continuous, with one warm pulse peaking in 2007–08 (likely the same warm pulse of AW that entered through Fram

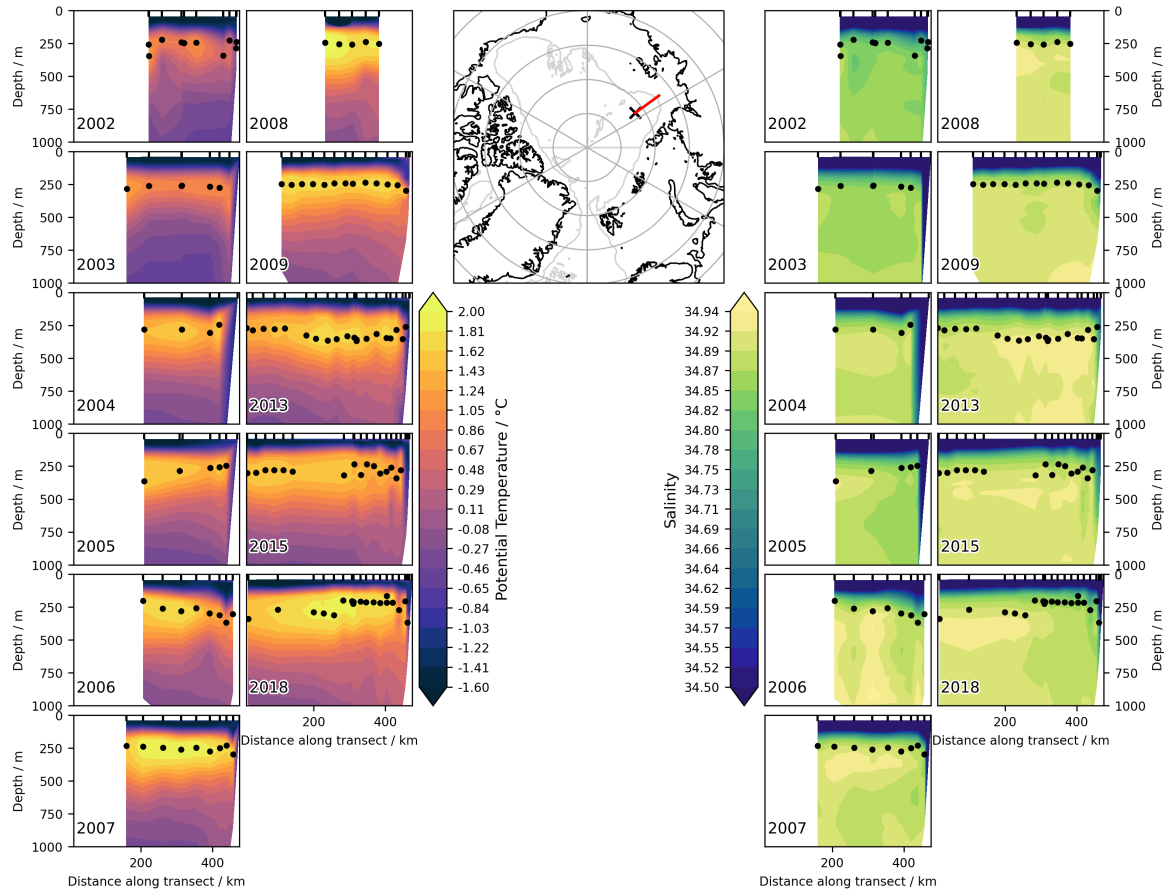


Figure 2.9: Potential temperature and salinity along a repeated CTD transect in the eastern Eurasian Basin. The year in which each transect was taken is given at the bottom left of each plot. In all years, transects were measured in August, September or October. The origin of the x-axis of the transect is marked with a black cross on the map, so that the x-axis origin is at the most offshore station. CTD profile locations are marked on the transect plots with vertical black lines at the surface. The black dots on the transect plots denote the location of the AW core.

Strait around the year 2000 (Polyakov et al. 2011; Polyakov et al. 2005)) and a second in the 2018 section. The AW core in 2018 is $1\text{ }^{\circ}\text{C}$ warmer than that in 2002. The salinity of the AW also shows an increasing trend throughout the time period covered in Figure 2.9, although regions and years of high salinity are not coincident with regions or years of high temperature.

Figure 2.10 allows for a more quantitative assessment of the changes in the water column at this location. The first three panels of this figure show the evolution of AW core properties across the transect. This figure shows more clearly that the core freshens onshore in most years, suggesting that AW is mixing with fresher waters from the shelf or that the AW that reaches the shelf is that which is fresher. As above, the core temperature (and AW

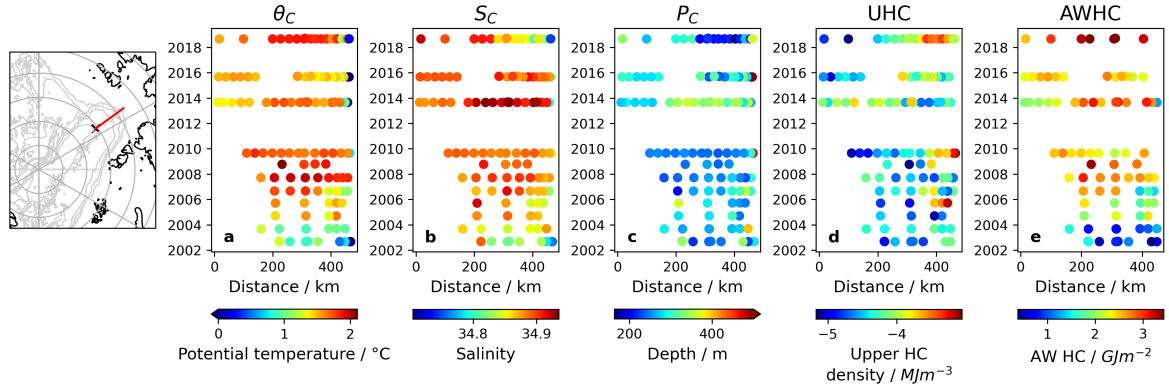


Figure 2.10: Water column properties across a repeated CTD transect in the eastern Eurasian Basin. Year markers denote the start of that year. Transect location is shown on the map, with a black cross denoting the x-axis origin of the transect plots. The x-axis origin is therefore at the most offshore station. Transect plots show (a) AW core potential temperature, (b) AW core salinity, (c) AW core depth, (d) heat content density of the water column above the AW layer, denoted as upper heat content (UHC), and (e) total heat content of the AW layer.

layer heat content shown in panel five) exhibits warm pulses which are superimposed upon a general warming trend across the period. Between 200–300 km along the transect, AW core temperature and salinity have linear trends of $4.13 \times 10^{-2} \text{ }^{\circ}\text{C}$ per year (resulting in an increase of $0.7 \text{ }^{\circ}\text{C}$ between 2002–2018) and 3.63×10^{-3} psu per year, respectively. Nearer the boundary, greater than 350 km along the transect, these trends are reduced to $1.52 \times 10^{-2} \text{ }^{\circ}\text{C}$ per year and 7.64×10^{-4} psu per year. This reduction in magnitude of AW core property trends away from the interior is likely due to enhanced mixing near the boundary. Notably the heat content of the AW layer increases to three times its 2002 value in 2018. The salinity and depth of these warm pulses differ, however - the AW core during the warm pulse in 2018 is fresher and shallower than the one in 2008. A weakened halocline may have allowed the warm AW to shoal higher in the water column and mix with the fresher surface layer, as reported by (Polyakov et al. 2020b). The 2013 transect, although slightly cooler than those from 2008 and 2018, has a comparatively salty, deep AW core. This non-coincidence of AW core salinity and temperature changes suggests that even enhanced mixing due to a weaker halocline does not mask the signal of these warm AW pulses.

The fourth panel of Figure 2.10 shows the “heat content density” (the heat content of a portion of the water column divided by the height over which it is computed) of the sampled water column above the AW layer, denoted as upper heat content (UHC) in the figure. This quantity is proportional to the average temperature of the surface layer and halocline. UHC increases when the AW core salinity is low and AW core depth is shallow - perhaps

an indication that a shallower AW layer transfers more heat to the halocline and surface layer. Despite the similarities in AW core temperature of the 2009 and late-2013 transects, the fresher/shallower core in 2009 coincides with a larger surface layer heat content density. This emphasises the strong link between halocline strength and the depth and salinity of AW. This is also highlighted in recent work by Polyakov et al. (2018), where halocline stability, quantified using density anomalies throughout the layer, is identified as a key climate change indicator in the region. The implication that AW shoaling is more influential than AW temperature change on surface layer heat content is not surprising given the low levels of vertical mixing throughout the Arctic (Fer 2009). This is also reflected in the dissimilarity between variations in AW layer heat content and UHC seen in Figure 2.10.

2.3.3.2 Canada Basin

Figure 2.11 shows the same analysis applied to two repeated CTD transects in the Canada Basin. This allows for detection and comparison of any signals advected downstream from the Eurasian Basin transect in Figure 2.10. The length of the transects also enables comparisons between AW found on the boundary and within the interior of the Canada Basin.

As in the Eurasian Basin, Figure 2.11 shows evidence of the pulse-like nature of AW core temperature evolution, with warm AW core values in the interior in the mid-2000s indicative of the warm anomaly that arrived in the Canada Basin in the early 2000s (McLaughlin et al. 2009). As seen in Figure 2.3, AW at the Canada Basin boundary (>1000 km along section A, and the furthest few data points of section B) is cooler and fresher than that in the interior due to the enhanced mixing it experiences upstream over the rough bathymetry of the Chukchi Plateau (Li et al. 2020; McLaughlin et al. 2009). This enhanced mixing and cooling is the likely reason why the warm temperature anomaly is not seen at the boundary in Figure 2.11 - the temperature signal is much weaker there than it is in the interior.

On both transects, at the boundary, AW core temperature and salinity vary similarly, presumably because both of these properties are governed by the same mixing processes upstream. This is not the case in the interior, however, where from 2012 onwards both transects see an increase in AW core salinity which is not reflected in the temperature. This increase in salinity (and density) and decrease in temperature in the interior was also seen in the Figures 2.5–2.7. T-S plots of un-smoothed profiles from the ends of each annual

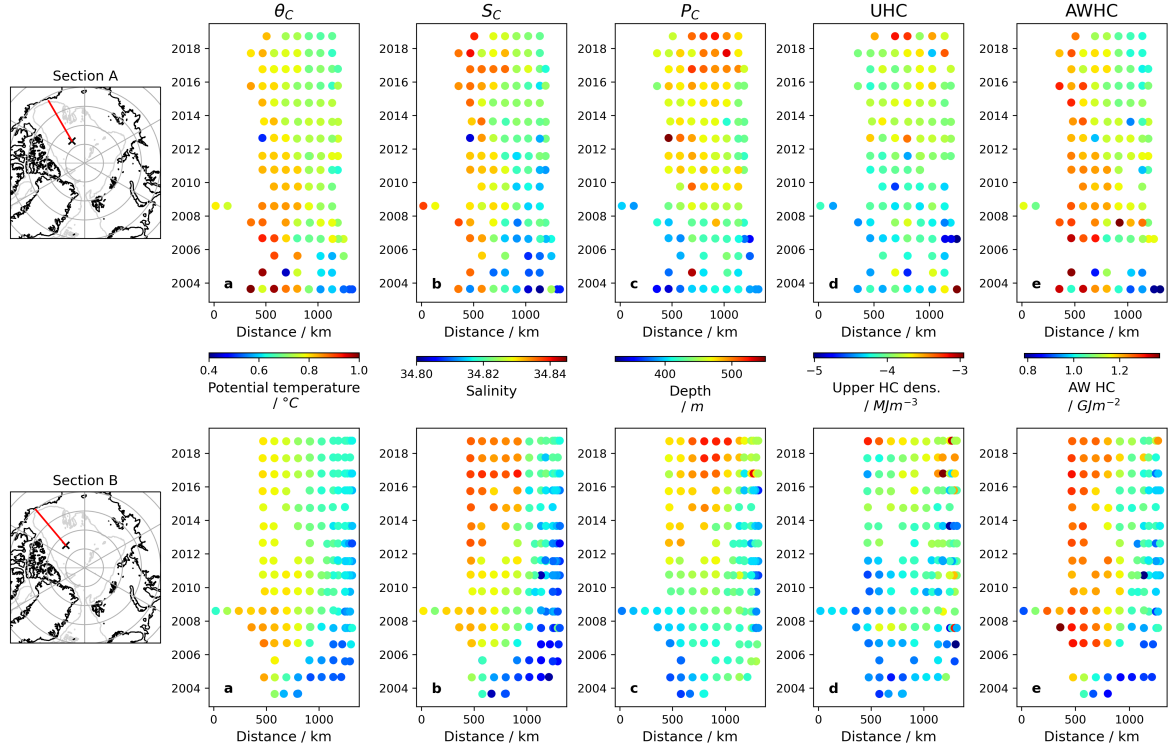


Figure 2.11: Water column properties across two repeated CTD transects in the Canada Basin. Year markers denote the start of that of year. Transect location is shown on the maps, with black crosses denoting the x-axis origin of each transect. The x-axis origin is therefore at the most offshore station. Remaining panels show (a) AW core potential temperature, (b) AW core salinity, (c) AW core depth, (d) heat content density of the water column above the AW layer, denoted as upper heat content (UHC), and (e) total heat content of the AW layer.

transect (not shown here) reveal that the increase in interior salinity is accompanied by a disappearance of thermohaline intrusions i.e. the commencement of AW core cooling after the warm anomaly has arrived. This confirms the mechanism proposed earlier - of heat diffusing away from the AW core resulting in the core residing at a more saline part of the AW layer - as the most likely explanation for these trends. The idea that this AW core salinity increase is related to AW core heat loss, and is not related to changes in the AW layer as a whole, is also supported by a lack of trend in AW layer mean salinity in the Canada Basin interior over this period (not shown).

The slight warming in 2016 of the AW core temperature in the interior of section B of Figure 2.11 could be evidence of the AW warm anomaly observed upstream of the Chukchi Borderlands in 2010 (after having entered the Arctic Ocean through the Fram Strait around 2000, (Li et al. 2020)). It can also be seen at the Eurasian Basin transect in Figure 2.10 around 2008, and until now has not been conclusively observed in the Canada Basin interior. This

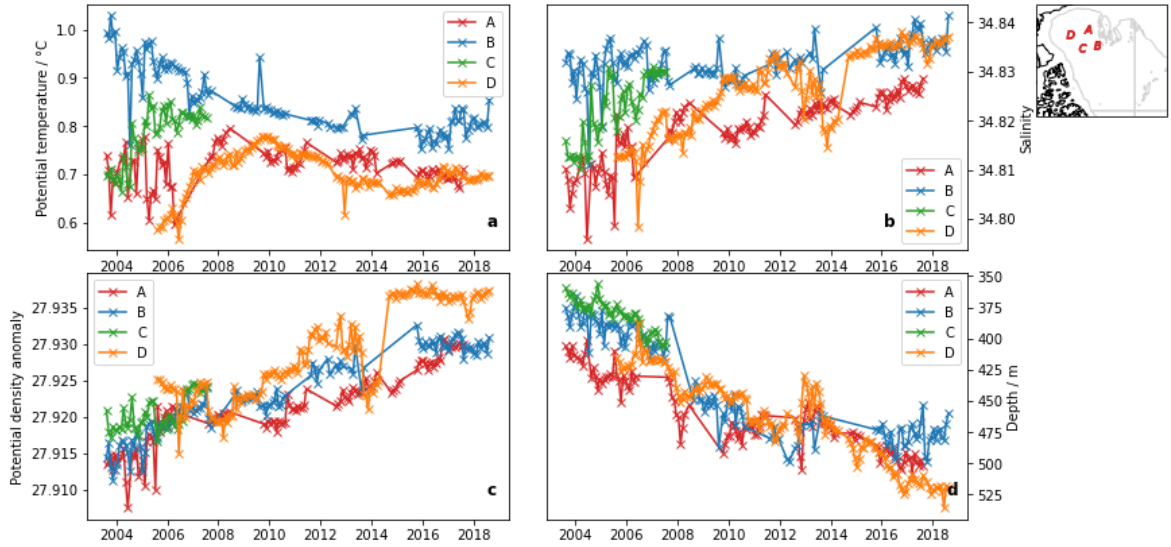


Figure 2.12: Time series of monthly mean AW core (a) potential temperature, (b) salinity, (c) potential density anomaly, and (d) depth from the four Canada Basin moorings. Mooring locations (A-D) are shown on the map at the top right.

gives AW advection timescales from the eastern Eurasian Basin to the north of the Chukchi Borderlands and around the north of the Borderlands into the Canada Basin interior of order 8 years, in agreement with other observational studies (Li et al. 2020; Polyakov et al. 2011). The amplitude of the warming is low compared to that of the previous warm temperature anomaly, despite the second anomaly being 0.24 °C warmer than the first in the Eurasian Basin (Polyakov et al. 2010). This could be due to enhanced heat loss experienced by the AW during its advection, associated with increased ventilation, enhanced mixing with cooler, fresher water above, and/or interaction with shelf flows in the eastern Arctic (Ivanov et al. 2007). Further observational data would be needed to confirm the presence of this second AW warm anomaly in the Canada Basin interior, however.

As discussed in Figures 2.5–2.8, the AW core in the Canada Basin cooled, freshened and deepened from the early 2000s onward. Trends from individual Canada Basin moorings (data in black in Figures 2.5–2.8) are more clearly shown in the time series in Figure 2.12 and are given explicitly along with their R-squared values in Table 2.1.

As seen in the maps in Figure 2.5, the warm anomaly that entered the Canada Basin in the early 2000s spread into the interior from the Chukchi Plateau before cooling (McLaughlin et al. 2009). This is reflected in the consistent, significant cooling trend at mooring B (Figure 2.12 and Table 2.1), and the pre-2010 warming at moorings C and D (Figure 2.12). There is an increase in salinity at all moorings, with significant trends at moorings A, C and D. As

Table 2.1: Trends in monthly mean AW core potential temperature, salinity, potential density anomaly, and depth from each of the four Canada Basin moorings. R-squared values for the fit of each trend are given in parentheses. Note that mooring C data only cover five years (see Figure 2.12).

Mooring	θ trends / °C	Salinity trends / psu year ⁻¹	σ_θ trends / kg m ⁻³ year ⁻¹	Depth trends / m year ⁻¹
A	$2.81 \times 10^{-4}(0.00)$	$2.22 \times 10^{-3}(0.76)$	$1.79 \times 10^{-3}(0.88)$	9.204(0.79)
B	$-1.81 \times 10^{-2}(0.70)$	$6.90 \times 10^{-4}(0.26)$	$1.70 \times 10^{-3}(0.89)$	12.24(0.82)
C	$3.82 \times 10^{-2}(0.61)$	$4.74 \times 10^{-3}(0.70)$	$1.33 \times 10^{-3}(0.43)$	9.80(0.79)
D	$2.17 \times 10^{-4}(0.00)$	$1.96 \times 10^{-3}(0.73)$	$1.52 \times 10^{-3}(0.75)$	9.01(0.88)

discussed above, this seems to be related to the cooling of the core after the spread of the warm anomaly. The resultant increase in AW core density can be clearly seen in Figure 2.12, with significant trends at moorings A, B and D (note that C has a shorter time series). The trends in depth in both Figures 2.8 and 2.12 are also significant (Table 2.1), and could be attributed to the increase in AW core salinity. However, comparing the patterns of change in AW core depth and salinity in Figure 2.11 suggests that the salinity increase is not the most important driver of this depth increase. As discussed, the spin-up of the Beaufort Gyre and associated enhanced downwelling influences AW depth (Lique et al. 2015a; Lique et al. 2015b; Zhong et al. 2014), so the increase in core depth is likely explained by gyre intensification.

The Hovmöller plots in Figure 2.13 reveal more about the mechanisms behind the AW core deepening in the Canada Basin, and also help explain the increase in upper layer heat content density. This figure shows Hovmöller plots of potential temperature profiles from three of the BGEP moorings in the Canada Basin (mooring C having been omitted due to the shorter time series available there), with white lines marking isopycnal depths. The core depth varies in concert with isopycnal depth at all moorings, again emphasising that changes elsewhere in the water column, rather than those of the AW properties themselves, are likely driving AW depth changes here. The increase in AW core salinity/density does not appear to play an important role, as seen in Figure 2.11. With the exception of mooring D (which the gyre moved away from during the start of the period shown (Regan et al. 2019)), the isopycnals and hence AW deepen over the time period covered. The temporal pattern of this deepening is in agreement with previous papers (Proshutinsky et al. 2019; Zhang et al. 2016) and can be attributed to an overall increase in Ekman pumping (and hence freshwater content) in the central Canada Basin between 2003-2018 due to a combination of changes in

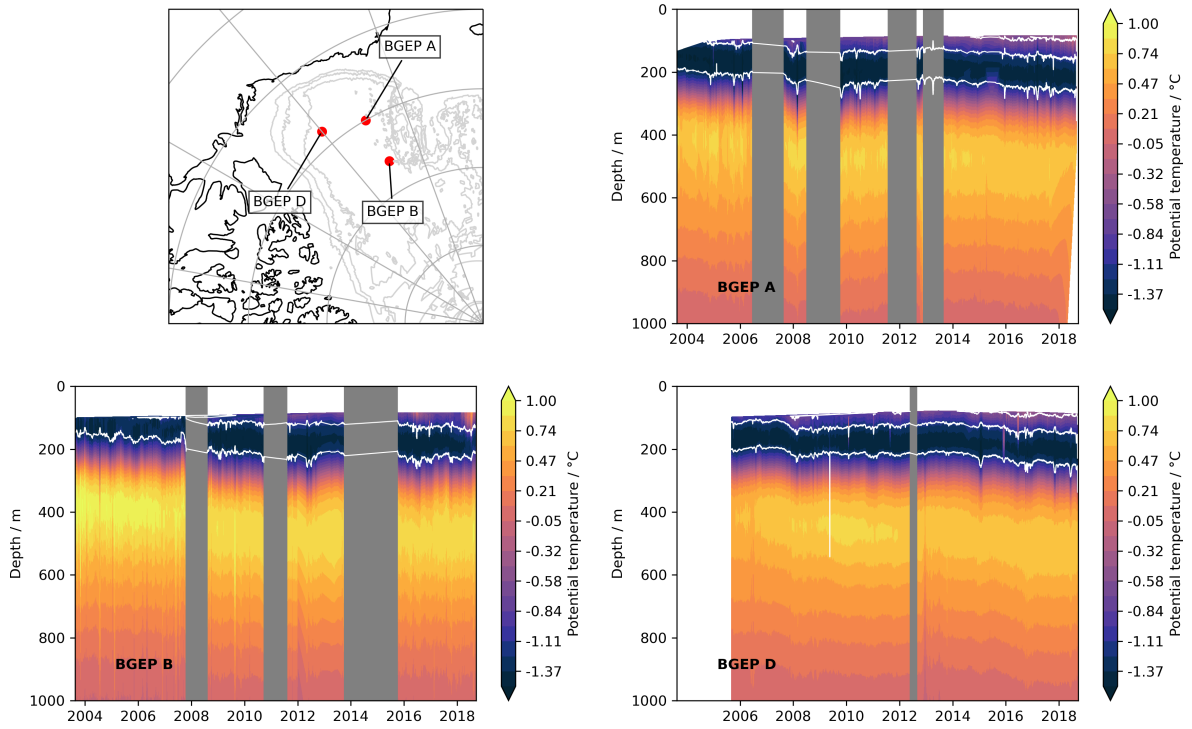


Figure 2.13: Hovmöller plots of potential temperature from three moorings in the Canada Basin (BGEP moorings A, B and D - locations shown on map). White lines denote the isopycnals - the deepest at 27 kg m^{-3} , contour interval 1 kg m^{-3} . Grey regions cover time periods with insufficient data.

sea ice conditions and the strength of the Beaufort Gyre (Proshutinsky et al. 2019).

The increase in upper layer heat content density seen in Figure 2.11 can also be attributed to Beaufort Gyre intensification. Pacific Water is subducted into the halocline of the interior Canada Basin via isopycnals which outcrop in the Chukchi Sea (Timmermans et al. 2014). The increase in Ekman transport has therefore resulted in thicker Pacific Water layers within the Canada Basin halocline (Timmermans et al. 2014), and the warming of Pacific Summer Water (which can be seen as the warm shallow water mass that appears in later years of Figure 2.13) between 2003–2014 has subsequently resulted in an increase in halocline heat content (Timmermans et al. 2014; Timmermans et al. 2018). Changes in upper layer heat content density are therefore independent of AW heat content, as seen in Figure 2.11. This suggests that increases in sea ice bottom-melt reported in the Beaufort Sea (Perovich et al. 2015) are likely due to Pacific Summer Water (along with other local features such as the near-surface temperature maxima (Jackson et al. 2012; Timmermans 2015)), not AW warming. The deepening of the AW, and its increased isolation from the surface in the Canada Basin, is in stark contrast to the concurrent Atlantification seen in the Eurasian Basin.

Although spatial and temporal patterns of AW core potential temperature and AW layer heat content in Figure 2.11 are similar, they do not match as closely as might be expected. We explore this further in the following section.

2.3.4 Relationships between profile metrics

Much of the analysis in this chapter has involved the use of AW core properties to infer AW layer properties within the Arctic. It is therefore important to investigate how representative AW core properties are of the AW layer in general. Here we compute correlations between AW core and integrated AW layer metrics, which also shed some light on how the AW layer loses heat in each region.

The maps in Figure 2.14 show total AW layer heat content during different time periods, chosen to give a roughly even data distribution between panels. As most observational profiles do not sample deep enough to cover the entire AW layer, there are substantially less AW heat content data (1500 data points) than AW core data. This, of itself, emphasizes the usefulness of AW core data in assessing the pathways and evolution of the AW layer.

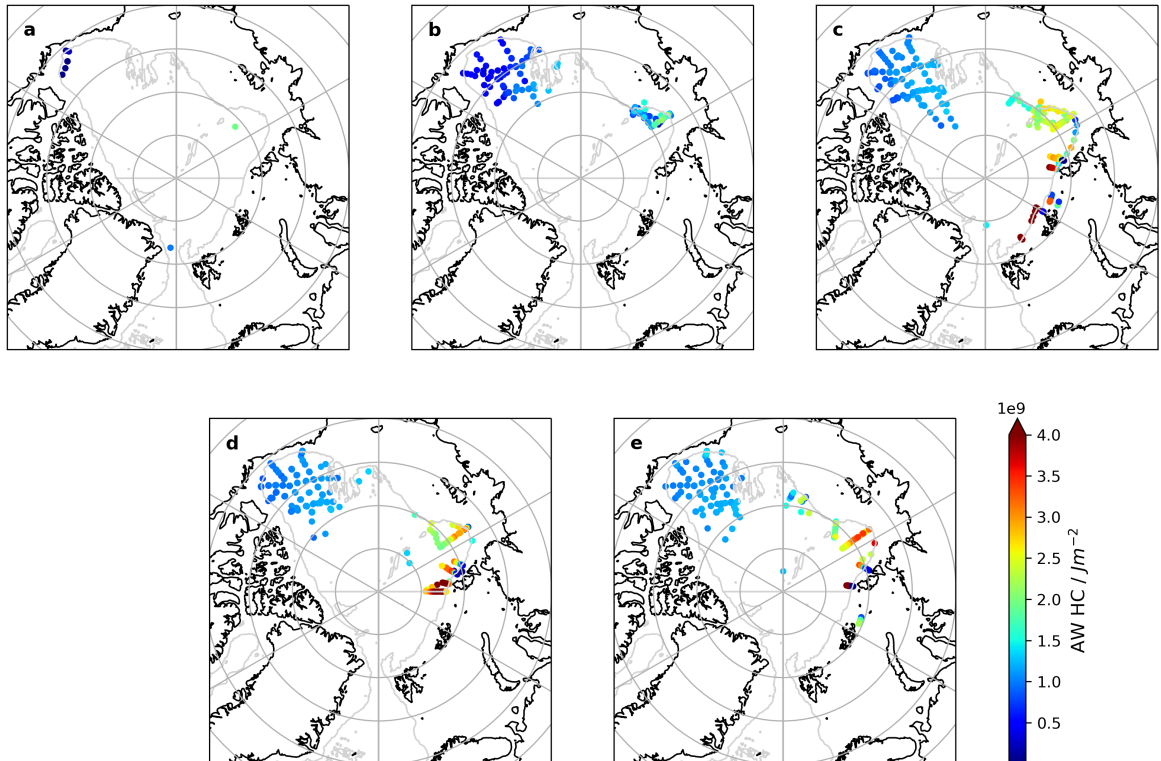


Figure 2.14: Maps of total AW layer heat content from all profiles which sampled the entire AW layer (defined as the layer between the two 0°C crossing points either side of the AW core depth) for (a) 1980-1999, (b) 2000-2004, (c) 2005-2009, (d) 2010-2014, and (e) 2015-2018.

In the Canada Basin, total AW layer heat content increased in the mid-2000s (Figure 2.14, panels b and c) after the arrival of the AW warm anomaly (Li et al. 2020; McLaughlin et al. 2009) and has since remained at that higher level of approximately $1.5 \times 10^9 \text{ Jm}^{-2}$, with no long term trend observed. The eastern Eurasian Basin (EEB) (i.e. the portion of the Eurasian Basin east of 90°E which has good data coverage for much of Figure 2.14) saw an increase in AW heat content throughout the period studied, in-line with the reported Atlantification of the region (Lind et al. 2018; Polyakov et al. 2017; Polyakov et al. 2010). Figure 2.14 shows a stark difference between AW heat content in the eastern Eurasian and Canada Basins, implying that the AW that bifurcates and recirculates towards the Fram Strait along the Lomonosov Ridge is warmer than the AW in much of the western Arctic. The heat content maps in Figure 2.14 are very similar to the AW core potential temperature maps in Figure 2.3, further suggesting that the AW core temperature captures the general pan-Arctic spatial pattern of AW heat content variability well. However, the cooling of the AW core in the Canada Basin (Figures 2.5 and 2.12) does not seem representative of the steady western AW heat content values in Figure 2.14, providing more evidence that the decrease in temperature (and increase in salinity) of the AW core do not reflect temporal changes in the AW layer itself.

To be more quantitative in assessing how well the AW core temperature represents AW heat content, correlations between these two metrics were computed. Figure 2.15 shows scatter plots between pairs of variables computed from profiles in both the eastern (restricted to EEB only due to constraints on spatial data coverage - see Figure 2.14) and western Arctic. These regions are defined in Figure 2.1. R-squared values for each of the plots are given, along with regression lines. The relationship between total AW heat content and the potential temperature of the AW core is shown in Figure 2.15a. There is a strong correlation between these two variables in both the EEB and western Arctic, highlighting the general effectiveness of the AW core temperature as an easily measurable metric for assessing changes in AW heat content. As noted above, however, caution must be taken when using AW core properties to infer temporal trends in the AW layer, as the cooling (and resultant increase in salinity) of the AW core does not necessarily correspond with a cooling (or salinification) of the AW layer more generally.

Figure 2.15b shows AW layer thickness against AW layer heat content, with a moderate

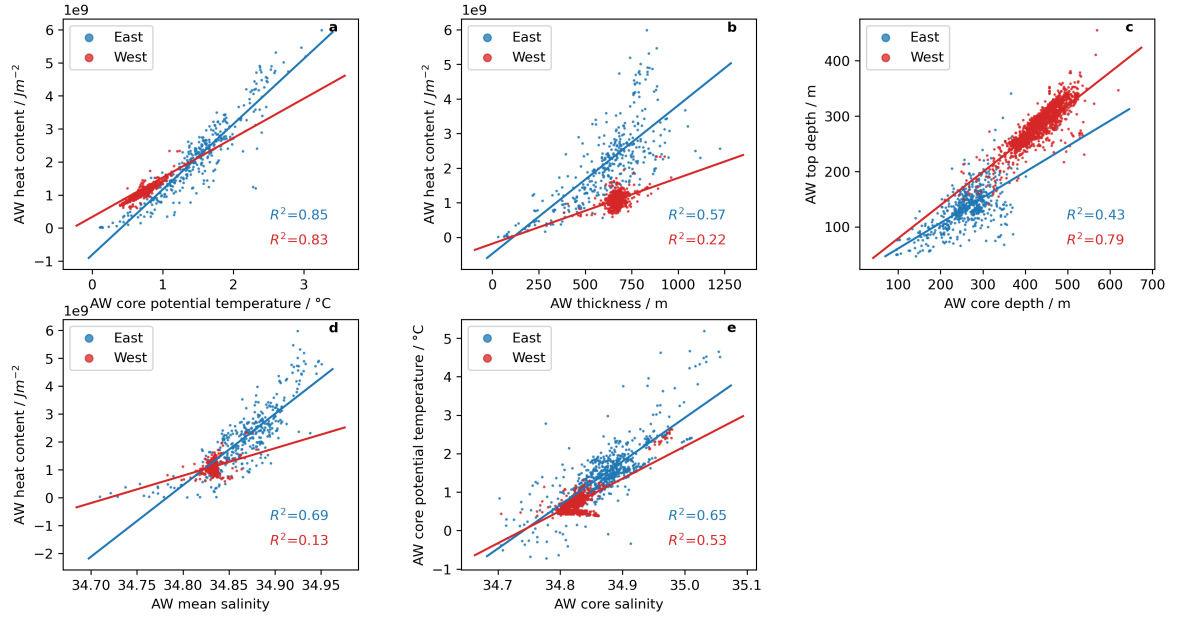


Figure 2.15: Scatter plots between (a) AW core potential temperature and total AW heat content, (b) AW layer thickness and AW heat content, (c) AW core depth and AW top depth, (d) AW mean salinity and AW heat content, and (e) AW core salinity and AW core potential temperature. Blue data is from the eastern Arctic, with red data from the western Arctic (regions defined in Figure 2.1). R-squared values and regression lines are shown for each scatter plot.

correlation in the EEB and a weak correlation in the west. This is an important reminder that although the potential temperature of the AW core will give a good idea of how AW heat content may vary (see Figure 2.15a), AW heat located away from the AW core also affects AW heat content. This is particularly true in the EEB where BSBW heat in the deep AW (which does not vary in concert with FSBW/the AW core, as seen in Figure 2.4) affects AW heat content.

Figure 2.15c shows the relationship between AW core depth and the depth of the upper boundary of the AW layer (i.e. the 0°C crossing point above the AW core depth). AW temperatures in both the eastern and western Arctic appear to be affected by heat loss from the top of the AW layer, causing the temperature maximum to deepen, although not necessarily to the same extent as the AW top (which is defined by a fixed temperature). This lowers the correlation between these two variables. However, there is a much stronger correlation in the west than in the EEB, as AW core depth and AW top depth in the Canada Basin are both greatly influenced by the Beaufort Gyre which affects the two metrics in the same way. The comparatively low correlation between AW core depth and AW top depth in the EEB highlights the care that should be taken when using AW core depth to assess AW

layer shoaling here.

Comparing the mean salinity of the AW layer with the AW layer heat content can give an idea of the role that mixing with fresher waters plays in AW heat loss. Figure 2.15d shows that, while there is a relatively high correlation between these variables in the EEB, the correlation is negligible in the west. This implies that although mixing with fresher waters is important for AW heat loss in the Eurasian Basin - as would be expected given that the AW subducts beneath the cooler, fresh polar waters here, losing a lot of heat - it is not as important in the western Arctic.

In Figure 2.15e, AW core salinity is compared to AW core potential temperature. In the EEB this reflects what is seen in the integrated AW layer (Figure 2.15d). In the west however, a stronger (moderate) correlation exists between the AW core data than that between the AW layer data. This could be related to the thermohaline intrusions which spread the warm AW core anomaly into the Canada Basin during the study period (McLaughlin et al. 2009). However it also highlights once again the freshening of the AW core, relative to the AW layer, as it is advected around the Arctic.

2.4 AW core properties and their relationship to AW temperature profile shape

As mentioned throughout this chapter, changes in AW core properties do not necessarily represent changes in the AW layer as a whole. A useful way to think about this is to consider the shape of the AW layer temperature profile. A temperature profile with a larger gradient in the upper AW layer will result in an AW core that resides higher-up within the AW layer (e.g. the eastern Arctic temperature profiles in Figure 2.4) than a similar profile with a less steep upper temperature gradient (e.g. the western Arctic temperature profiles in Figure 2.4). This can result in changes in AW core depth differing from those of the AW top, as seen in Section 2.3.4.

A steeper temperature gradient will likely increase the rate of vertical diffusion of heat away from the upper AW. This cumulative diffusion of heat, combined with mixing, as AW is advected towards the western Arctic is likely what causes these regional profile differences in Figure 2.4. The Figure 2.4 profiles also show that the temperature (and salinity) gradient is greater at the top of the AW than towards the bottom. This is a signature of differing

mixing rates at the top and bottom of the AW. Considering AW temperature and salinity profile shape is therefore not only a useful way of putting AW core changes into context, but also provides insight into the physical processes occurring in the AW layer.

2.5 Conclusion

The analysis in this chapter has used all available hydrographic profiles from across the Arctic from the 1970s to 2018 to build a picture of AW in the Arctic Ocean entirely from observations, and to investigate its spatial and temporal variability. Much of the analysis has focused on the properties of the AW core (the depth at which the maximum potential temperature occurs). This was found to be a generally effective and easily detectable metric to assess the heat content of the AW layer. However, the depth of the AW core does not always reflect the depth of the top of the AW layer, particularly in the eastern Arctic, and care must be taken when using temporal trends in AW core properties to assess trends in the AW as a whole - a cooling or increase in salinity of the core does not necessarily translate to a cooling or increase in salinity of the entire AW layer.

In general, as the AW is advected around the Arctic the potential temperature and salinity of its core decrease. Despite freshening, the AW core density increases along its advection pathway. This is partially due to the preferential loss of heat and salt from the top of the AW layer to the fresher, cooler water above through vertical mixing along the AW advection pathway. This likely deepens the core without the AW layer as a whole getting denser - upper AW cools such that the AW core (temperature maximum) is found on deeper (denser) isopycnals. Interaction with dense shelf flows formed by brine rejection during sea ice formation may also play an important role in the cooling and freshening of the AW core during its advection around the basin.

The evolution of AW has differed between the eastern and western basins of the Arctic. In the Eurasian Basin, AW core temperature and AW heat content increased from 2002–2018, with the former increasing by approximately $0.7\text{ }^{\circ}\text{C}$ during this period. Warm pulses were superimposed upon this trend. Instances of high upper ocean heat content in the east were found to be associated with shallower, fresher AW. In contrast to this, and similar reports in the literature of eastern Arctic Atlantification, the western Arctic saw AW core temperatures decrease from a previous warm peak by approximately $0.1\text{ }^{\circ}\text{C}$ between 2008–2018 (although

AW layer heat content increased), and also saw AW heat become more isolated from the surface. This increased isolation was due to Beaufort Gyre intensification which deepened the halocline. These findings suggest the emergence of two different regimes - with AW affecting sea ice in the east, and Pacific Water influencing sea ice in the west. This implies that the future evolution of the Eurasian Basin will strongly depend on AW, whereas Pacific Water and the Beaufort Gyre will be the biggest drivers of change in the Canada Basin. This contrasting regional evolution is in agreement with other recent studies, which describe halocline weakening, AW shoaling, and increased sub-Arctic influence in the Eurasian Basin, contrasting with a freshening and deepening of the surface layer in the Amerasian Basin driven by local atmospheric conditions (Polyakov et al. 2020a; Polyakov et al. 2020b).

Despite the limitation of sparse, temporally inhomogeneous oceanographic measurements in the Arctic, pan-Arctic observational analysis can give useful insights into the overall temporal and spatial patterns of heat distribution in the Arctic Ocean. Given the challenges of realistically representing the AW layer in forced ocean-sea ice and coupled climate models, and the stark regional differences emerging in the Arctic Ocean, the use of pan-Arctic observations for model validation and benchmarking will be essential. Only by combining insight from observations and models will we be able to accurately determine what the future Arctic will look like under a changing climate, which is important both for the region itself as well as for the wider climate system.

This thesis will now put these observational results into a longer term context through the use of model data spanning 1920–2100. Natural oscillations are a known feature of temporal variations in AW temperature (Polyakov et al. 2004). Therefore, determining whether the trends found in this chapter represent forced change or just natural variability will be an important objective of the model data analysis. The temporal and spatial homogeneity in coverage afforded by the model data will also allow us to investigate potential local and remote drivers of change, and will also provide crucial insight into how AW heat may evolve in the future.

Chapter 3

The evolution of the AW heat in the 21st century

3.1 Introduction

This chapter will focus on the evolution of AW layer properties over the next century and how these relate to AW inflow into the Arctic. Observational analysis in the previous chapter showed AW warming in recent decades within the Arctic Ocean. Results also suggested that, despite an increase in the influence of AW heat on the upper ocean in the eastern Arctic, AW heat in the western Arctic became more isolated from the upper ocean - and crucially, sea ice - over the study period (1980–2018). Three key questions arise from these results, which this chapter will aim to shed light upon.

Firstly, what will the future hold for the state of the AW layer in the Arctic? Will the trends seen in Chapter 2 continue, with AW shoaling and warming in the Eurasian Basin (Polyakov et al. 2017) and deepening and slightly cooling in the Canada Basin?

Secondly, what is driving these changes? Clearly, drivers can be split into local and remote processes. Therefore, can we link any evolution found in AW properties to those at the inflow, or is there some unexplained portion that must be influenced by processes local to the Arctic? Årthun et al. (2019) suggested that increasing AW heat transport through the Barents Sea Opening will result in an advective progression of AW influence on the upper ocean all the way to the Lomonosov Ridge by 2070. Furthermore, winds over the Nordic Seas influence AW advection into and within the Arctic. Enhanced cyclonic wind stress in this

region, and the resultant increase in Greenland Gyre strength, have been found to increase both AW volume transport into the Arctic and AW inflow temperature (Chatterjee et al. 2018; Muilwijk et al. 2018). Muilwijk et al. (2019) found that Nordic Seas wind stress can in fact influence AW advection as far as the Canada Basin. However, Richards et al. (2022) (and Chapter 2 of this thesis) suggested that local processes result in notable changes to AW layer properties along its advection pathway within the Arctic. Both mixing at the top of the AW layer, and interaction between AW and dense cascading shelf flows, were proposed as the two likely local processes at play. Cascading shelf flows have been modelled throughout the Arctic (Luneva et al. 2020) and have been observed to interact with AW in the Eurasian Basin (Ivanov et al. 2007). Despite sparse observations in the western Arctic, cascading shelf flows have been observed to interact with lower halocline waters here (Ivanov et al. 2004; Luneva et al. 2020; Melling et al. 1995), implying that interaction with AW is possible - particularly if the AW layer were to become less dense. Determining the relative role of upstream conditions (specifically, AW heat transport into the Arctic and atmospheric circulation) will therefore be an important focus of this chapter.

Thirdly, what role does climate change play in AW property changes in the Arctic? Are AW trends forced, or are they part of the natural variability in the climate system? Natural variability, which arises in the climate system due to random fluctuations in atmospheric or oceanic circulation, or due to changes in ocean dynamics, can strengthen or reduce the short-term magnitude of forced trends. Temperature records of AW in the Arctic oscillate with a period of around 50–80 years (Bengtsson et al. 2004; Ikeda 1990; Polyakov et al. 2004), with longer oscillatory periods of around 1500 years also apparent (Jensen et al. 2018) - both are signatures of natural variability. Observations show evidence of recent AW warming outside of the range of these natural oscillations, however (Polyakov et al. 2012b). Assessing the relative role of natural variability in AW property changes compared to any externally driven forced trends, especially during the 21st century, is an important part of predicting the full impact of anthropogenic climate change in the Arctic - particularly given the significance of AW heat in the Arctic Ocean (Carmack et al. 2015).

Model outputs from the Community Earth System Model Large Ensemble (CESM-LE) will be used in this chapter, marking a shift from the observational analysis of the previous chapter toward more of a model-data focus for the remaining thesis. This model will be

presented in more detail in section 3.2, but one of the main reasons we chose to work with the CESM-LE was because of how well-suited large ensembles are to studying this natural variability.

3.2 Data and methods

3.2.1 Model description

The CESM-LE (Kay et al. 2015) consists of 40 runs of CESM version 1 (CESM1), a coupled climate model with the Community Atmosphere Model version 5 (CAM5), Los Alamos Sea Ice Model (CICE4), Community Land Model version 4 (CLM4), and Parallel Ocean Program version 2 (POP2) - all run at 1° horizontal resolution (Hurrell et al. 2013). POP2 has 60 vertical levels, varying in thickness from 10 *m* near the surface to 250 *m* at depth. The model runs from January 1920 to December 2100 (although we only include data up to December 2099 in our analysis in order to work with periods of consistent length). The model is forced using historical data from 1920 to December 2005, with Representative Concentration Pathway 8.5 (RCP8.5) forcing used from 2006 onwards. RCP8.5 represents a climate system subject to continued increases in greenhouse gas emissions that would result in an 8.5 Wm^{-2} increase in radiative forcing by 2100 compared to pre-industrial times. Each member of the 40-member ensemble differs from other members due to perturbations in their initial conditions - specifically round-off level differences in air temperature fields. Therefore differences between members reflect the natural/internal variability of the climate system due to atmospheric fluctuations. Kay et al. (2015) give a more detailed overview of the CESM-LE.

3.2.2 Model assessment

Although climate models - including CESM-LE - have been found to generally underestimate internal variability in heat transport into the Arctic via the Barents Sea (Li et al. 2017), the CESM-LE has been found to give a more realistic representation of the internal variability of sea ice extent and thickness than other models involved in the Coupled Model Intercomparison Project 5 (CMIP5) (of which CESM1 was a part) (England et al. 2019). In fact, several studies have found that CESM-LE performs generally well in the Arctic. It realistically captures inter-annual variability in sea ice concentration and sea ice decline in all months

(Barnhart et al. 2015; England et al. 2019), simulates a realistic sea ice cover response to AW heat inflow into the Barents Sea (Årthun et al. 2019), and also produces a mean sea level pressure pattern in general agreement with observations (DeRepentigny et al. 2016). CESM1 itself simulates the present Arctic sea ice state realistically (although exhibits a positive bias in mean September sea ice extent) (Jahn et al. 2016), and can replicate the observed cloud response to sea ice variability (Morrison et al. 2019).

Despite the generally good performance described above, CESM1 does have some biases that should be borne in mind throughout this (and the next) chapter. The model has a cold surface air temperature bias over the Arctic (Park et al. 2014), and recent trends in Barents Sea winter sea ice extent are larger than those modelled by any of the CESM-LE members (Onarheim et al. 2017). Some biases also exist in the AW layer of CESM1. When assessing the performance of models used in CMIP5, Shu et al. (2019) found that all CMIP5 models underestimate the inter-annual variability in AW core temperature (i.e. the AW temperature maximum) compared to observations, and model the AW layer as being too deep and thick. Furthermore, the ocean component of the CESM1 was found to have a warm bias in AW (of approximately 0.5 - 1 °C at 400 m depth) throughout the Arctic basin due to insufficient cold water entering the Arctic basin through the St. Anna Trough, itself the result of insufficient cold, dense water formation in the Barents and Kara Seas and excessively warm AW inflow in the Fram Strait (İlıcak et al. 2016). This warm bias was also seen in the coupled CESM simulations themselves (Danabasoglu et al. 2012). AW was also found to have a salty bias in the Nordic Seas and western Eurasian Basin, and a fresh bias in the western Arctic, with biases up to +0.1 *psu* in the Eurasian Basin and -0.23 *psu* in the western Arctic (İlıcak et al. 2016). Other known CESM1 issues relevant to this thesis include the absence of a summer Pacific Water signature in the halocline of the western Arctic (Lavoie et al. 2022), and unrealistically deep vertical mixing resulting in biased surface freshwater stratification (Rosenblum et al. 2021) (although these will be more relevant in Chapter 4).

The model overestimation of AW depth and thickness can be seen in temperature profiles at locations in both the eastern and western Arctic when comparing ensemble mean CESM-LE profiles with smoothed observational profiles from Chapter 2 (see Figure 3.1), and is unsurprising given the coarse vertical resolution of the model - Shu et al. (2019) proposed that spurious diapycnal mixing in coarse-resolution CMIP5 models results in a too-deep AW

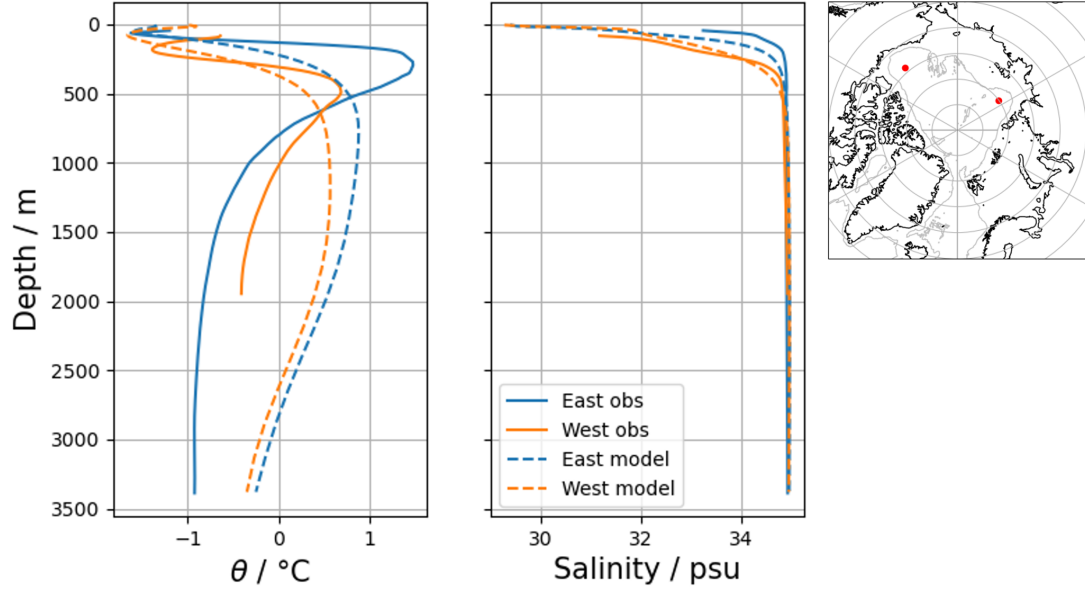


Figure 3.1: A comparison of temperature and salinity profiles from the Eurasian Basin in the eastern Arctic (blue) and from the Canada Basin in the western Arctic (orange) - in the ensemble mean CESM-LE data (dashed) and observations (solid). Observational profiles (taken from a NABOS mooring in the east and a BGEP mooring in the west) have been smoothed over a vertical distance of 80 *m*. All profiles are for September 2015. Profile locations are shown with red dots.

layer bottom. Furthermore, a model with coarser vertical resolution would be more likely to smear out temperature gradients. We find that the depth of the AW temperature maximum (i.e. the AW core, used in much of the observational analysis in Chapter 2) is consistently estimated as too deep in the model compared to observations. It is for this reason that integrated AW layer properties are used throughout this chapter, rather than the properties of the AW temperature maximum.

3.2.3 Statistics

In this chapter, we define the AW layer in a similar way to the previous chapter. The AW layer is classified as the layer warmer than 0 °C and below the temperature minimum of the upper 500 *m* of the water column. The exception to this rule is if the temperature minimum of the upper 500 *m* is at 500 *m*, in which case we assume the AW layer is at the surface, as is the case in the Nordic Seas for example. This latter criteria ensures that no warm water above the cold halocline (e.g. that warmed by solar heating) is incorrectly classified as AW. The 500 *m* criteria gave reliable classification, and was chosen as the AW layer was consistently thicker than 500 *m* in the model data. Using the halocline depth as classified

by e.g. Bourgain et al. (2011) was considered, but significant changes in temperature and salinity over space and time in the CESM-LE data meant that no consistent criteria could be found that accurately identified the halocline base. Salinity criteria were not used in the AW layer definition, due to both the changes in hydrography over time and the simplicity afforded by the use of one variable in the definition. The AW layer in this chapter can therefore be thought of as AW heat. Locations where the ocean depth is shallower than 100 *m* were discounted from the analysis.

Temperature transport (defined later in the text), rather than heat transport, is used to quantify heat transported through the gateways from the Atlantic to the Arctic. This is because we compute this transport over a section with non-zero volume transport, not a closed volume. The resultant values are therefore dependent on an arbitrarily chosen temperature scale, and hence should not be interpreted as a heat transport (Schauer et al. 2009). Temperature transport is an acceptable metric to use in this work given that we are interested in temporal changes in AW heat entering the Arctic, which are fully captured by temperature transport. The locations of these gateway sections (the Fram Strait, Barents Sea Opening (BSO), and St. Anna Trough) are shown in orange in Figure 3.2. Also shown in Figure 3.2 are the western and eastern regions used in regional analysis later in this chapter. Similarly to chapter 2, these broad regions are separated by the Lomonosov Ridge.

We quantify internal/natural variability as the standard deviation of the trends in detrended ensemble member data - detrended meaning that the ensemble mean (which we take to represent any externally-driven/forced trends in the simulation) has been subtracted from each ensemble member.

Ensemble trend correlation is used in this chapter to quantify the link between 30-year trends in oceanic heat transported into the Arctic and 30-year trends in AW layer temperature elsewhere in the Arctic. AW layer temperature trends at grid points across the Arctic are taken over all 30-year time periods starting in January of each year between 1920 and 2070. The inflow trends used in this analysis are taken over: 1920–1949, 1950–1979, 1980–2009, and 2010–2039. These intervals ended here as periods later than this did not result enough regions of meaningful correlation due to the short time difference between the inflow trend period and the local trend period. The 30-year length was chosen as it aligns with many periods used in other analysis in this chapter and, more importantly, typical variations in

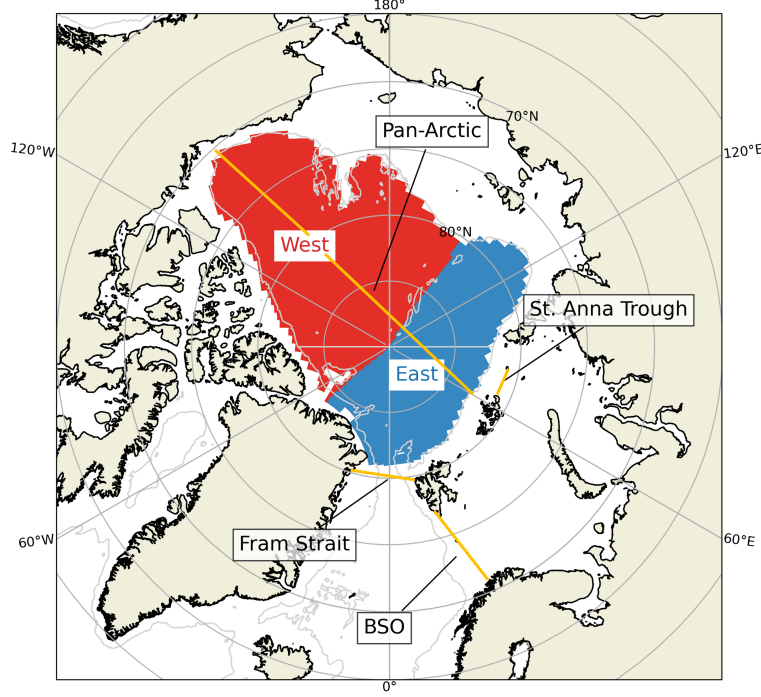


Figure 3.2: Map showing the key sections and regions used in this chapter. The grid cells used in the eastern (blue) and western (red) regions are coloured, along with the Fram Strait, Barents Sea Opening (BSO), St. Anna Trough and pan-Arctic sections. The average latitude of the Fram Strait section is 80 °N.

inflow temperature are of the order of a few decades. The Pearson correlation coefficient at a given location for a given combination of inflow time period and local time period is calculated by first computing the 30-year trends in both inflow temperature transport and local AW layer temperature for each ensemble member, and then computing the correlation between these two variables over the ensemble members. A similar technique is used by Wettstein et al. (2014) to investigate the role of atmospheric forcing in driving Arctic sea ice loss. At each grid point, the combination of inflow period and local period with the highest correlation is noted.

The dominant timescales of variability in inflow temperature and AW layer temperature elsewhere in the Arctic are also calculated. These are computed for each ensemble member, first by detrending the data by subtracting from it a fitted 5th order polynomial - this order was chosen as it tended to capture the general trend without over-fitting and removing the higher frequencies. A Fourier transform was then performed on the detrended data and taking e.g. the highest peak of this Fourier spectrum as the dominant time period of variability. The mode (i.e. most common), mean and standard deviation of the two most dominant time scales of variability amongst the ensemble members can give insight into how inherent these

timescales are to the AW temperature variability.

Several parts of analysis compare data over the periods 1920–2000 and 2020–2100. This is so that an 80-year period in the historically forced part of the model can be compared with an RCP8.5 forced period. These periods are the same length so that the Fourier spectra obtained can be fairly compared. In many other parts of analysis, we often group data into the following periods: 1920–2009, 2010–2039, 2040–2069, 2070–2099; rather than equal-length 30-year periods. This is because we find that minimal changes occur during the 20th century.

3.3 Results and Discussion

3.3.1 Hydrographic properties of the AW layer within the Arctic in the CESM-LE and their natural variability

3.3.1.1 The evolution of AW layer properties

To set the scene and give a descriptive introduction to this (and the next) chapter, a spatial overview of the temporal changes in the AW layer is given in Figures 3.3, 3.4, and 3.5, which show ensemble mean maps of AW layer properties over various time periods.

Ensemble mean maps of depth-mean AW potential temperature over different periods are shown in Figure 3.3. As expected, AW is at its warmest near its inflow in the Nordic Seas and loses much of its heat in the Barents Sea and Fram Strait before entering the Arctic Basin. The AW layer temperature increases during the 21st century - initially in the Nordic Seas - before temperature increases are seen across the Arctic by the end of the century. AW warming within the Arctic therefore appears to be related to a warming inflow.

Similar ensemble mean maps of AW layer mean salinity are given in Figure 3.4. The AW layer freshens across the Arctic in the 21st century. That this freshening appears most pronounced on the shelves is the result of the fresher parts of the AW moving onto the shelves, and also potentially interacting with shelf flows. Although the evolution of water column structure will be investigated in more detail in Chapter 4, it is worth mentioning here that this freshening of the AW layer is partly due to AW heat shoaling in to the halocline - this will be shown and discussed later.

Similar ensemble mean maps of the depth at the top of the AW layer - i.e. at the 0 °C crossing point above the vertical maximum in potential temperature - show the AW heat

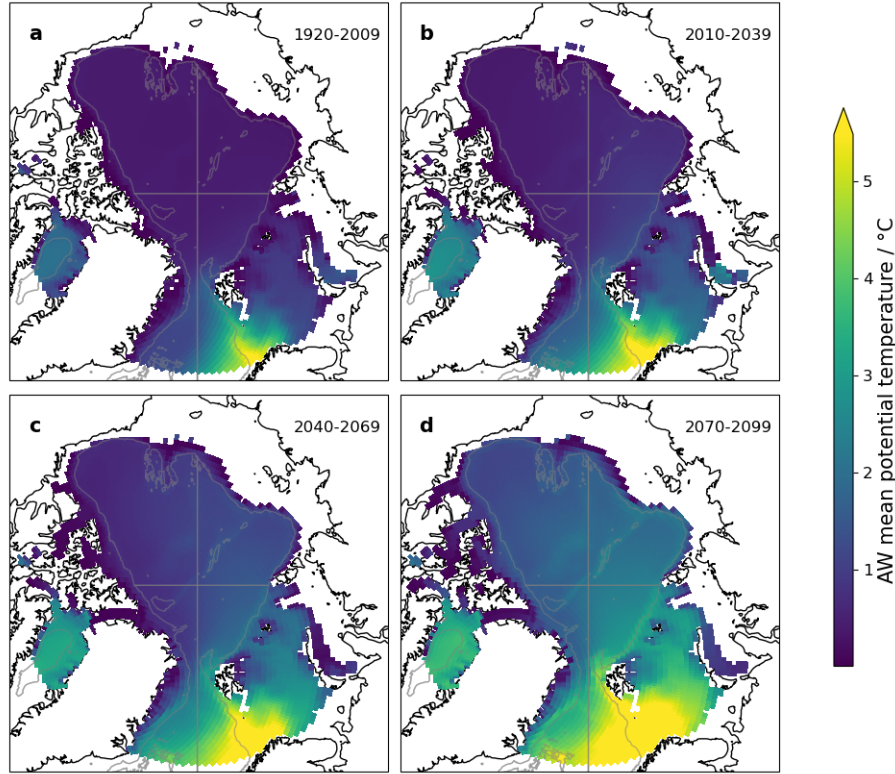


Figure 3.3: Ensemble mean maps of depth-mean potential temperature of the AW layer, averaged over the periods denoted at the top right of each panel. The first period is longer than other periods due to the relatively small changes in properties during this period (see Figure 3.8). The AW layer is defined as the layer between the two 0°C crossing points either side of the vertical maximum in potential temperature.

shoaling, first in the east and then the west (Figure 3.5). The eastern shoaling of AW in recent decades has been widely reported in observational studies (e.g. Lind et al. (2018), Polyakov et al. (2017), and Polyakov et al. (2010)). In the western Arctic, however, the AW heat sits beneath the thick halocline of the Beaufort Gyre, sitting deeper in the interior Canada Basin than in the Eurasian Basin due to Ekman pumping as the gyre intensifies and freshwater content in the region increases (Zhong et al. 2014). Observational studies have suggested that vertical heat flux from the AW layer in the Canada Basin is low - just a tenth as large as local heat flux from the surface layer to sea ice (e.g. (Timmermans et al. 2008)), highlighting the relative insignificance of AW heat to the surface layer here. The western shoaling of AW heat in Figure 3.5 during the 21st century is therefore a notable result. Figure 3.6 shows that this shoaling is not due to a thinning of the halocline, but a vertical extension of AW heat into the halocline layer in both the east and the west. This implies that AW heat may become an important factor in western - not just eastern - Arctic for driving the variations

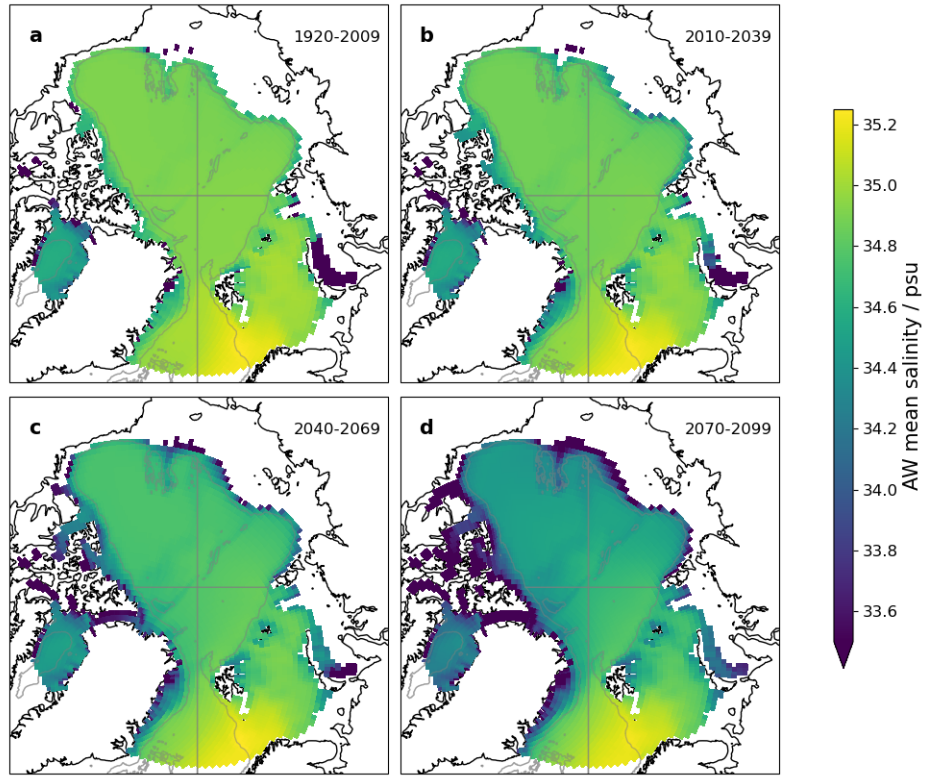


Figure 3.4: As in Figure 3.3 but for AW layer mean salinity.

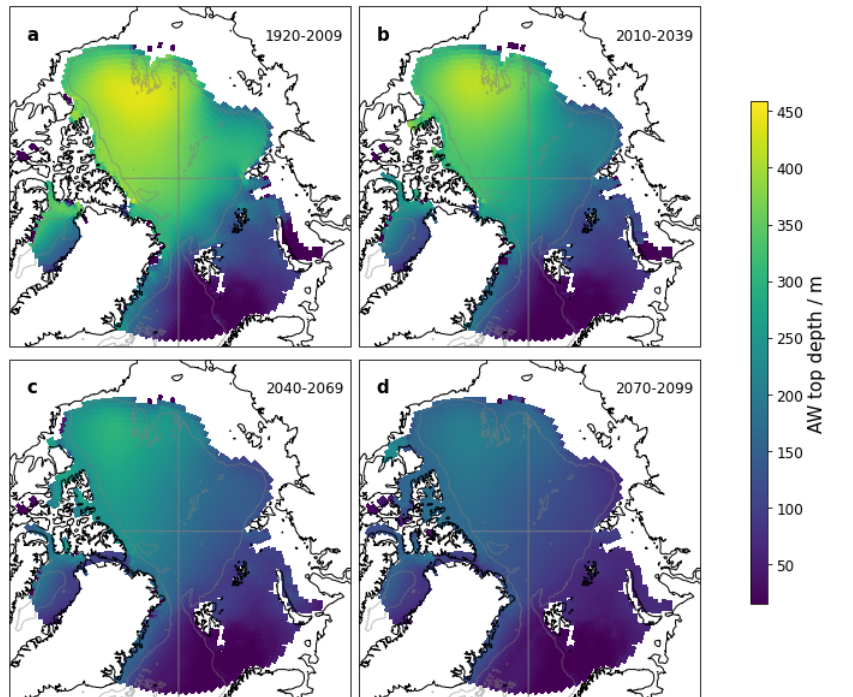


Figure 3.5: As in Figure 3.3 but for AW top depth (corresponding to the 0° isotherm above the AW temperature maximum).

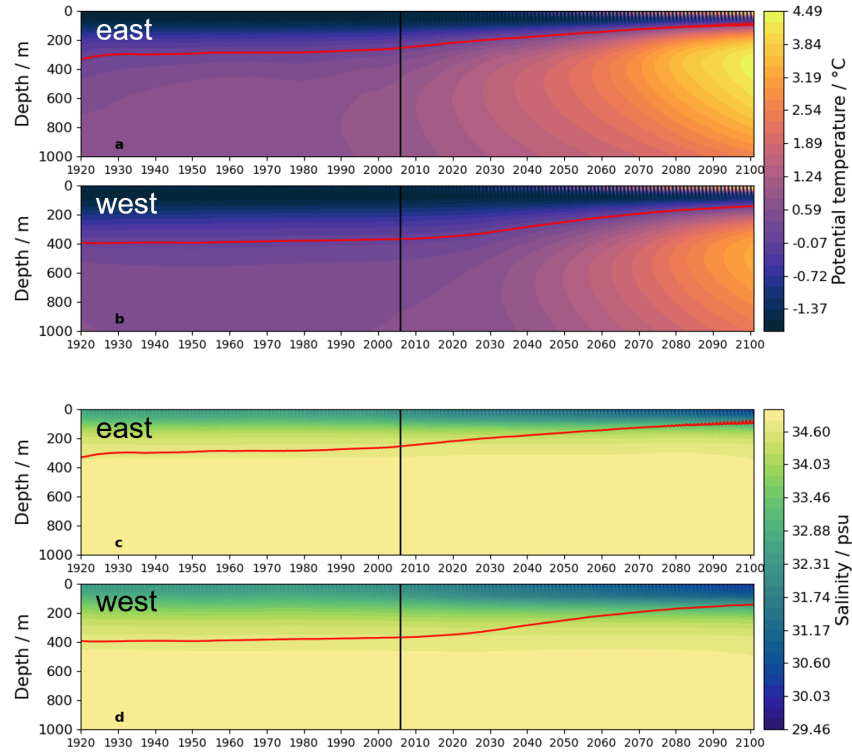


Figure 3.6: Hovmöller plots of ensemble mean (a)-(b) potential temperature and (c)-(d) salinity in the upper 1000 *m* of the water column, averaged over eastern ((a) and (c)) and western ((b) and (d)) regions of the Arctic Ocean as defined in Figure 3.2. The vertical black line denotes where model forcing changes from historical to RCP 8.5. The red contour line denotes the 0 °C isotherm (top of the AW layer).

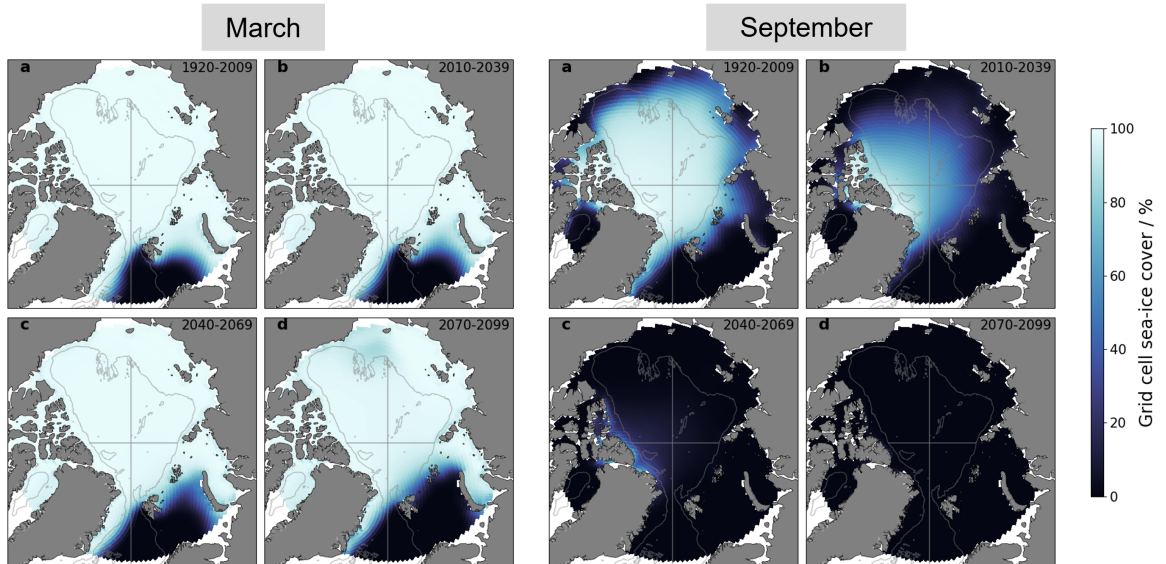


Figure 3.7: Ensemble mean maps of sea ice cover in March and September for periods: (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, (d) 2070–2099.

and trends of sea ice extent and thickness later this century. Changes in halocline structure and stability will be investigated in Chapter 4.

It is worth mentioning that, given the presence of warm Pacific Water in the halocline of the western Arctic (e.g. Timmermans et al. 2014; Timmermans et al. 2017), the apparent AW shoaling in the western Arctic could be explained by an increase in temperature of this Pacific Water in the model. However, the shoaling is not due to Pacific Water in this case - local temperature-salinity profiles (not shown) reveal that Pacific Water is not present at these depths. An absence of a Pacific Water signature beneath the mixed layer in CESM1 in the Canada Basin has also been reported in other studies (e.g. Lavoie et al. (2022), who found that this was due to inaccuracies in the simulation of Pacific Water pathways in the western Arctic).

Also worth mentioning is the relationship between this spatial pattern of AW shoaling and the diminishing September sea ice extent (see Figure 3.7 for maps of September and March ensemble mean sea ice coverage). The shoaling of AW heat in Figure 3.5 does not align with the retreat of the September sea ice edge. This implies that local increases in sea ice formation and melt and resultant local changes in water column stability are not responsible for the shoaling of AW heat seen in Figures 3.6 and 3.5, and vice versa.

3.3.1.2 Forced trends and natural variability of the AW layer in the 20th and 21st centuries

To get an idea of how the ensemble mean trends in AW properties shown above compare to natural variability, Figure 3.8 shows regionally-averaged monthly time series of AW properties from every ensemble member (blue and red for east and west, respectively) and the ensemble mean (black). For definitions of each region, see Figure 3.2.

To get a more detailed spatial picture of the relative dominance of natural variability and forced trends, Figures 3.9–3.12 show the ensemble mean trend, the standard deviation of the trends in the detrended ensemble member data (representing natural variability) and the ratio between these two quantities. A ratio greater than one implies that forced changes are responsible for the ensemble mean trends, with a ratio of less than one implying that natural variability is more important than any forced trend at that location. The figures have been split into two time periods - 1920–2000 and 2020–2100 - to represent the periods of little and

notable trend seen in the time series of Figure 3.8.

3.3.1.2.1 AW temperature and heat content

Figure 3.8a shows mean potential temperature across the AW layer. Before 2000, the slight increase in AW temperature associated with the Early Twentieth Century Warming between 1920s–1940s (Bengtsson et al. 2004) can be seen. The subsequent cooler AW during the 1970s (Ikeda 1990) can also be seen in the eastern time series of Figure 3.8a. These natural oscillations in AW temperature are negligible compared to the trend in temperature that commences in both regional averages later in the time series, which lies well outside of the envelope of natural variability represented by the extent of the coloured lines. This anthropogenically forced trend results in 2 °C and 1.6 °C increases in AW layer mean potential temperature in the eastern and western Arctic, respectively, during the 21st century.

Although the AW layer warming seen in Figures 3.3 and 3.8 was most noticeable in the 21st century, Figures 3.9a and d show that there is an AW warming trend across much of the 20th century Arctic as well, particularly at the inflow - as seen earlier in Figure 3.3. (Note that the low heat content values in shallow regions such as the Barents Sea and over ridges are related to the shallow water depth here). Advective spread of natural oscillations in AW inflow temperature are the key to the spatial differences in AW temperature trends. The remnant of the cool AW of the 1970s (Ikeda 1990) can be seen in the western Arctic in the region furthest from the inflow during this period. By the year 2000, the AW in the eastern Arctic has just begun warming (see Figure 3.8a), and so the eastern and much of the western Arctic sees an overall warming trend (Figures 3.9a and d). This suggests that if the trends were quantified over several centuries prior to the 20th century, multi-decadal natural variability would dominate across the whole Arctic prior to the year 2000.

Figure 3.9i shows that the AW warming is a forced trend everywhere during the 20th century period apart from at the inflow and in the Barents Sea. Note that the blue strip in the western Arctic in Figure 3.9i is just an artefact due to the change of sign of the ensemble mean trend. AW heat build-up is also anthropogenically forced in all basins. Natural variability dominates near the AW inflow and in the Barents Sea where AW is at the surface and in contact with the atmosphere, as any inter-member differences in atmospheric temperature or the presence of sea ice and polar water will have a large impact on AW temperature here.

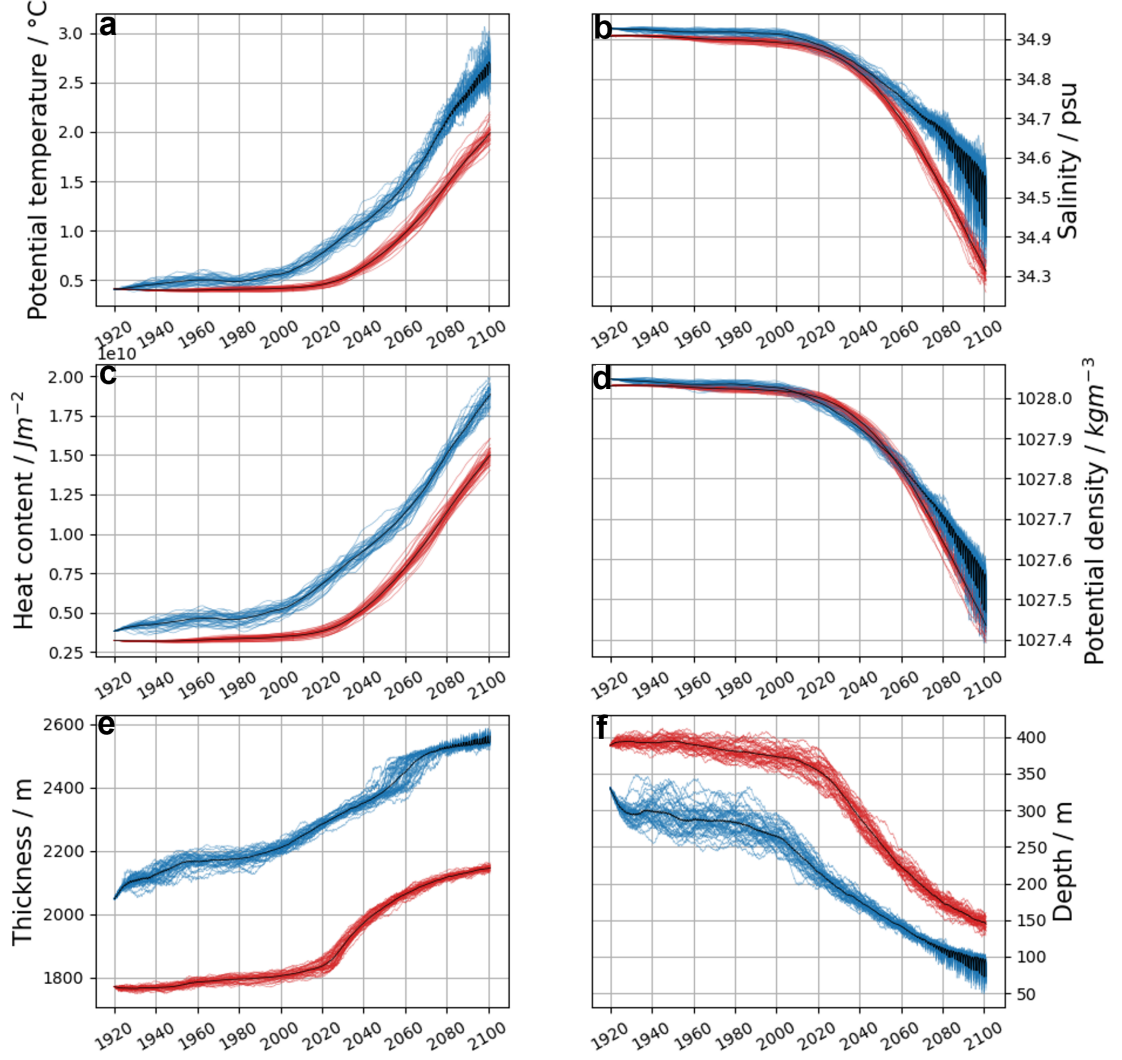


Figure 3.8: Time series of hydrographic properties averaged across the eastern and western Arctic (see Figure 3.2 for regional definitions). Coloured lines represent data from individual ensemble members, with the eastern average for each ensemble member in blue and western average in red. The black lines are the ensemble means for these regions. Quantities plotted are (a) AW layer mean potential temperature; (b) AW layer mean salinity; (c) AW layer heat content; (d) AW layer mean potential density; (e) AW layer thickness; and (f) depth of the top of the AW layer. The AW layer is defined as the layer between the two 0°C crossing points surrounding the vertical temperature maximum.

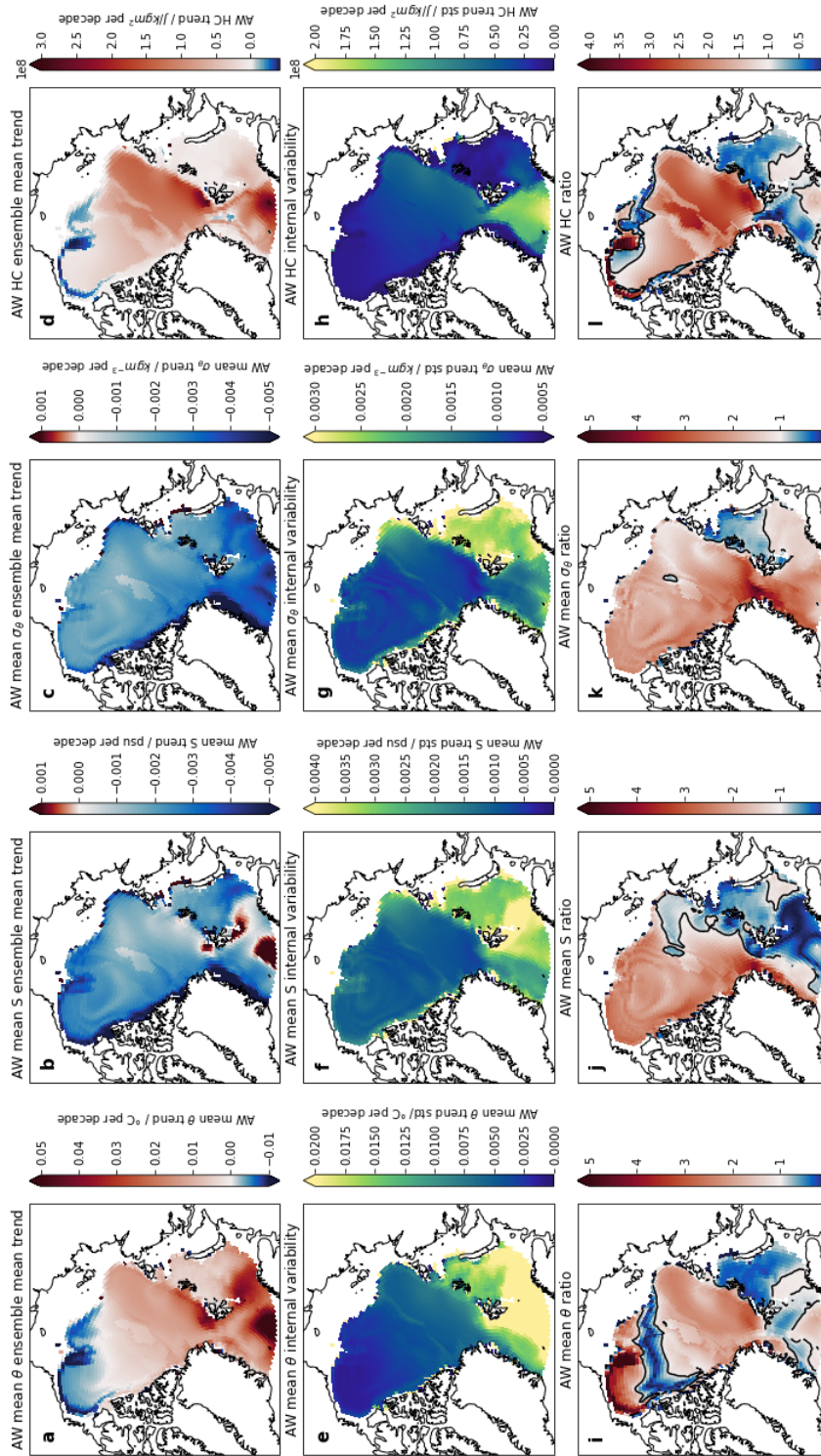


Figure 3.9: Ensemble mean (i.e. forced) trend over 1920–2000 (top row), internal variability over 1920–2000 (middle row), and ratio between forced trend and natural variability during 1920–2000 (bottom row) (obtained by dividing the top row by the middle row) in (a,e,i) AW layer mean potential temperature, (b,f,j) AW layer mean salinity, and (c,g,k) AW layer mean potential density, and (d,h,l) AW layer heat content.

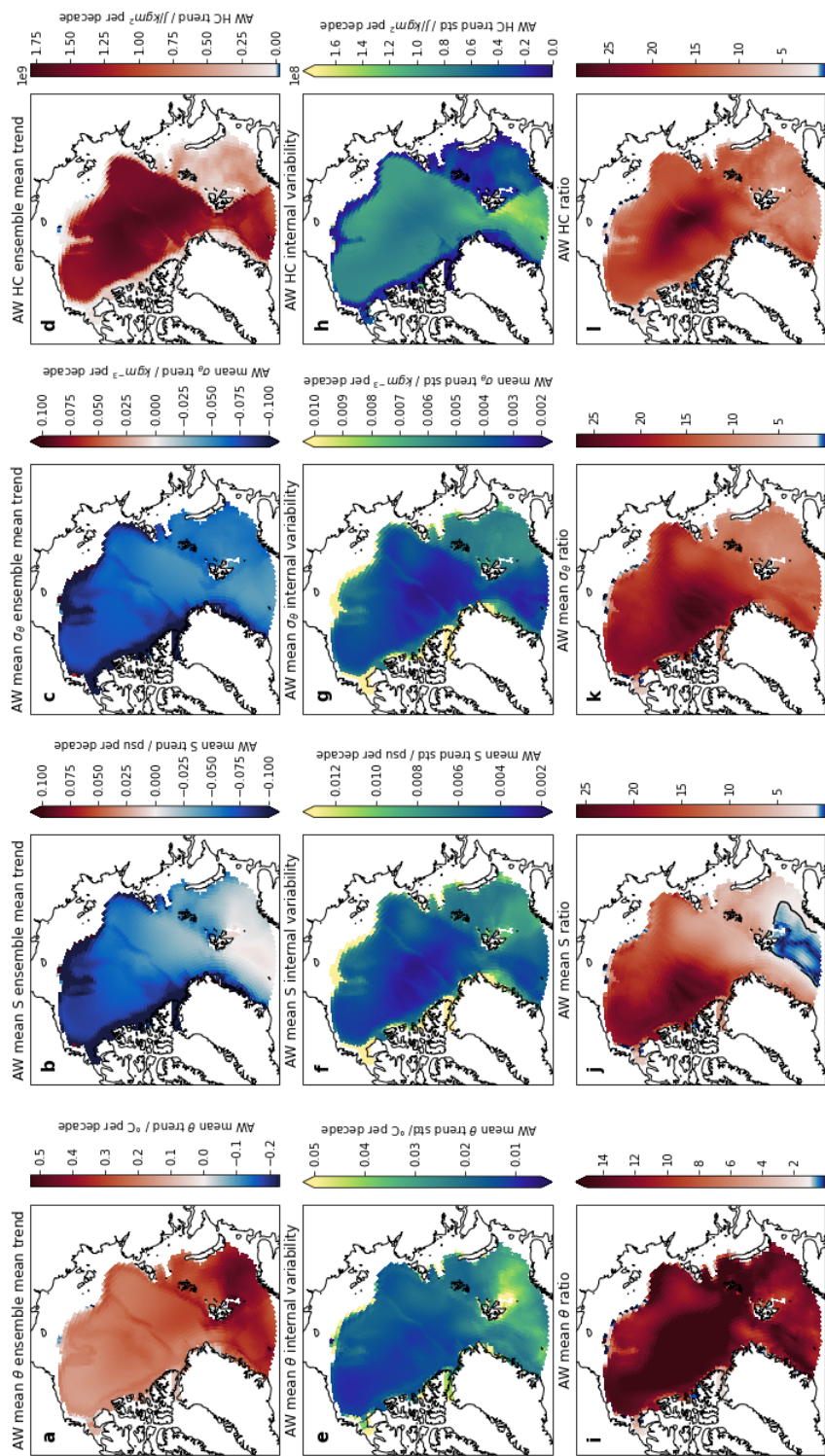


Figure 3.10: As in Figure 3.9 but for the period 2020–2100.

Notably, this low forced trend to natural variability ratio is not advected on to the rest of the Arctic, reflecting the significant amount of heat in the lower AW that is not affected by heat loss to polar water or the atmosphere.

Figures 3.10a and d show AW layer mean temperature and heat content trends during the 21st century period. Compared to 1920–2000 the trends are more homogeneous across the whole Arctic, with an increase in temperature and AW heat content across the whole Arctic Ocean and inflow region. These warming trends are roughly 10 times greater than they were in the previous century, of the order of 0.1 °C per decade. The Barents Sea warms at a much greater rate than in 1920–2000, with Barents Sea AW warming in 2020–2100 being larger than AW warming seen in the rest of the Arctic. Anthropogenically forced trends govern AW layer mean temperature and AW heat content everywhere in the Arctic during this period.

3.3.1.2.2 AW salinity and density

Figure 3.9 also shows the AW freshening and becoming less dense, as seen in Figure 3.8. The Nordic Seas and inflow regions are exceptions to this rule - AW becomes more saline here (salinity and temperature of AW inflow from the Nordic Seas have been found to fluctuate in phase (Mulwijk et al. 2018) so this is not surprising), although this change is dominated by natural variability. The freshening in the rest of the Arctic in spite of this increase in salinity highlights the important role of local Arctic processes in AW freshening. This is particularly true near the shelves where AW freshens the most. Away from the shelves, freshening trends are similar throughout the Arctic, indicating that AW freshening just after inflow (just upstream of red regions in Figure 3.9b), where it first meets and subducts under fresh polar water, is the main cause of AW freshening within the Arctic Ocean.

The freshening is driven by an anthropogenically forced trend everywhere apart from at the inflow region, Barents Sea, and eastern Eurasian Basin (see Figure 3.9j). At the inflow region and Barents Sea, natural variability is high once again due to the important role the atmosphere and polar water play in governing the salinity (and density) of AW here. Again, this regional difference in forced-to-natural variability ratio reflects the significant part of the lower AW which is not affected by this surface freshening and is advected on to other parts of the Arctic. The important role of natural variability in governing AW salinity in the eastern Eurasian Basin must either be related to variability in halocline thickness and

the resultant impact on AW depth and salinity, or to the variability of BSBW salinity that's advected downstream. There is no seasonal cycle in ensemble mean AW layer salinity here at this time (see Figure 3.8b), so it is not related to inter-member seasonal AW salinity differences. Notably, where freshening is highest at the shelves, natural variability is also high relative to other regions. This suggests that more variable local controlling factors such as winds/halocline depth and shelf flows here play an important role in governing local AW freshening and potential heat transfer to the surface.

Note that, in contrast to the natural variability-driven AW salinity in the eastern Eurasian Basin, there is a forced trend in density driven by the forced AW warming here. The different drivers of water column density and stratification will be investigated in more detail in the next chapter.

In the 21st century, a forced freshening trend (i.e. a trend outside of the range of natural variability) emerges in both regions of the time series in Figure 3.8b. There is also a lag between the freshening in both regions, with the eastern regional average freshening slightly before the west. Figure 3.10b shows decreases in AW layer mean salinity and density across the whole Arctic Ocean and inflow region. The freshening trends (and density trends - which have a similar pattern to those of salinity within the Arctic in the 21st century) are forced and are about twice as large as they were in 1920–2100 - 0.1 *psu* per decade compared to 0.05 *psu* per decade. The salinity of AW in the Nordic and Barents Seas is an exception to these rules. AW salinity increases slightly here, although natural variability here is greater than this trend. AW freshening within the Eurasian Basin and Barents Sea is therefore still important for AW freshening trends across the Arctic basin in the 21st century. Again, the greatest freshening trends are on the shelves, suggesting that the increasing interaction between AW and shelf flows continues into the 21st century, and/or that as AW gets fresher more AW heat is able to shoal up onto the shelves. Either way, this implies that more heat is shoaling onto the shelf regions which could have important consequences for heat reaching the surface here. Ensemble variability of shelf freshening (and density) is high compared to variability in the rest of the Arctic, suggesting that external factors such as winds and halocline depth are important, and ultimately might play a key role in AW surface heat loss on the shelves.

3.3.1.2.3 AW depth and thickness

Figures 3.8e and f show ensemble member and ensemble mean AW layer thickness and AW top depth averaged across the eastern and western Arctic regions. Again, the AW layer is defined using temperature only, so these are essentially measures of the change in depth and thickness of AW heat. The AW top shoals in both regions across the whole time series, but particularly after the early 21st century. Over the course of the 21st century, the top of the AW layer shoals by over 200 *m* in the eastern Arctic and approximately 175 *m* in the western Arctic. Figure 3.6 suggests that this shoaling represents AW heat shoaling in to the halocline in regional averages, as opposed to halocline thinning. The onset of shoaling therefore coincides with the onset of AW freshening (see Figure 3.8b), with eastern and western Arctic seeing 0.4 *psu* and 0.58 *psu* reductions in AW layer mean salinity over the 21st century, respectively. There is also a resultant decrease in AW layer density in both the eastern and western Arctic (Figure 3.8d). Combined with the concurrent AW warming trend, these trends are signals of “Atlantification”. It is interesting to note that up to the present day, this process has only been observed in the eastern Arctic (e.g. Polyakov et al. 2017; Polyakov et al. 2018). This could be because, according to these model data, we are currently in the period that lies between these changes occurring in the east and in the west. This lag can be explained by the time taken for enough warm AW to advect from the east to the west for a signal to appear in the regionally-averaged western data. Although it is important to note that local factors affecting water column stratification (e.g. the impact of the Beaufort Gyre on AW depth in the western Arctic) also play an important role in western AW depth (Zhong et al. 2014).

Natural variability of the eastern AW top depth time series (Figure 3.8f) is relatively large during the 20th century compared to any forced (ensemble mean) trend. This is not related to the presence of a seasonal cycle in the ensemble mean. Atmospheric pressure patterns are known to affect the advection of shelf waters into the eastern Arctic and hence affect the thickness of the halocline here (Steele et al. 1998) and therefore the depth of the top of the AW layer. Figure 3.8f therefore implies that prior to the year 2000, natural variations in atmospheric pressure over the eastern Arctic, along with other processes affecting AW depth such as subduction, play an important role in AW depth. Similarly, in the western Arctic, the wind-driven Beaufort Gyre impacts AW depth (Zhong et al. 2014), although natural

variability in AW depth in the west is not as great as that in the east, and does not notably change during the study period. A forced shoaling trend then emerges around 2000 and 2020 in the east and west, respectively.

The AW layer thickens both due to this shoaling and due to a spread of AW heat downwards from the bottom of the AW layer/deepest 0 °C point. This is evident from the increase in AW layer thickness which is greater than the amount of AW shoaling - shoaling in both the east and west across the study period is around 250 *m*, whereas AW layer thickening is around 500 *m* and 350 *m* in the east and west, respectively. The shoaling of the AW top and AW layer thickening coincide roughly with the onset of AW warming and freshening in both regions, as mentioned. The decrease in rate of thickening in the eastern data around 2080 is interesting given that the depth of the region (plus AW top depth) is much greater than this, so it is not indicative of AW reaching the sea floor. It could suggest that there is a limit to the amount of heat that can spread into the deepest basin water here as the AW layer becomes fresher and the stratification in the lower water column strengthens. Given the steady increase in AW layer heat content seen in Figure 3.8c, a limit on the vertical spread of AW heat would likely further contribute to the increases in AW layer temperature. This effect appears to occur in the western Arctic as well, albeit less abruptly.

The coarse vertical resolution of CESM-LE is important to bear in mind when interpreting these results however, given that this likely leads to inaccuracies in AW top depth and AW layer thickness, as mentioned earlier.

Investigating the natural variability and forced trends in AW depth and thickness in more detail using Figure 3.11 again suggests that local winds played an important role in AW properties in certain regions during the 20th century. In part of the Canada Basin, AW thins and deepens between 1920–2000 (this is masked in the regional average time series). This indicates a thickening of the halocline layer above, which itself is highly dependent on the wind-driven flow of the Beaufort Gyre. It is not surprising, then, that these AW deepening and thinning trends are dominated by natural variability. In the eastern Eurasian Basin, natural variability also dominates AW depth and thickness. As mentioned earlier, local winds can play an important role in the state of the halocline (and hence AW depth) in this region (Schauer et al. 2002a; Steele et al. 1998), and hence likely play a key role in this high natural variability of AW depth and thickness.

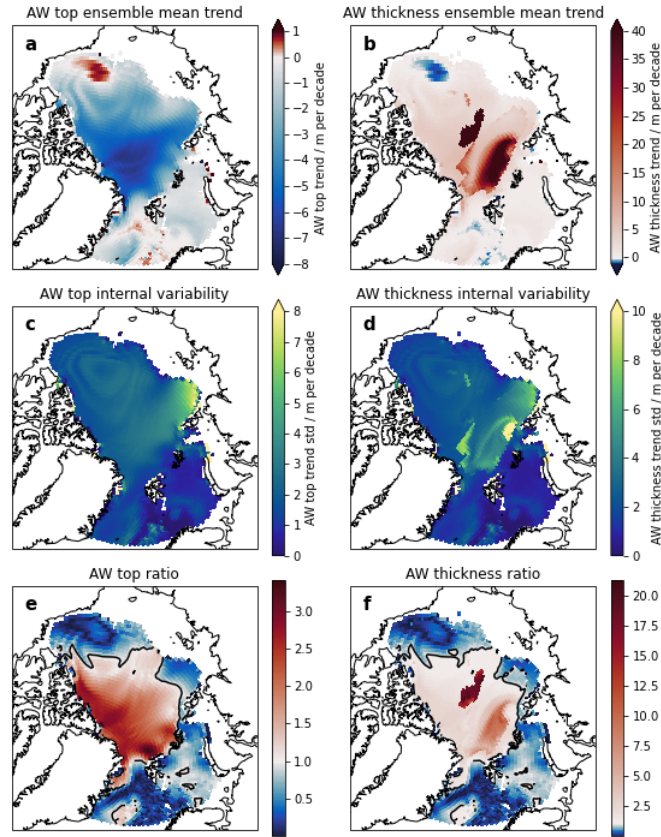


Figure 3.11: As in Figure 3.9 but for (a,c,e) the depth of the top of the AW layer, and (b,d,f) AW layer thickness. Shoaling and thinning trends are represented as negative (blue).

The apparent role of local winds (Schauer et al. 2002a; Steele et al. 1998) in the natural variability of AW depth, salinity and density in the 20th century in the eastern Eurasian and Canada Basins suggests that, combined with forced warming trends, winds played an important role in governing heat transfer to the surface in these regions.

Elsewhere in the 20th century period, apart from in the Nordic and Barents Seas where natural variability dominates as AW is at the surface, Figure 3.11 shows anthropogenically forced trends in AW thickness and depth. Where there is a forced trend, AW shoals and thickens. This is particularly pronounced in the western Eurasian Basin and Makarov Basin as heat spreads downwards here. Despite some variation between members on the spatial extent of the deepening in each basin, this is a strong forced signal across the ensemble. This suggests that the spread of warm AW into these deep regions is an important way in which the Arctic stores AW heat.

Trends and natural variability in AW top depth and AW thickness over 2020–2100 are shown in Figure 3.12. The top of the AW layer shoals due to a forced trend in both the

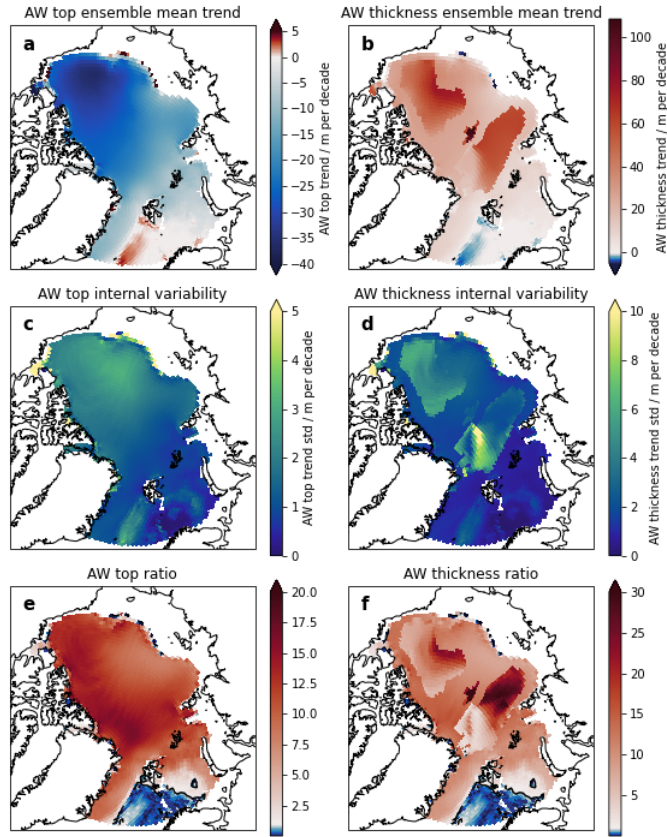


Figure 3.12: As in Figure 3.11 but for the period 2020–2100.

eastern and western Arctic and Barents Sea, with particularly strong shoaling in the west of up to 40 *m* per decade. As discussed above, this is AW heat shoaling into the halocline (see also Figure 3.6). At the inflow regions, the top depth and thickness of the AW layer are both dominated by natural variability, which again is not advected on to the rest of the Arctic. This is partly due to the fact that AW is already at the surface in these inflow regions, and so we would not expect to see a shoaling trend here, unlike in the rest of the Arctic where the AW layer is able to become more shallow.

Figure 3.12 shows AW layer thickening throughout the Arctic during 2020–2100, particularly in the Canada Basin and eastern Eurasian Basin (and again in the Makarov Basin). Comparing the AW shoaling trend with the AW thickness trend in Figure 3.12b suggests that the bottom of the AW layer is deepening in these regions more than it is elsewhere. This is in contrast to what was seen in Figure 3.11b where AW layer thickening (and implied AW bottom deepening) during 1920–2000 was greatest in the western Eurasian Basin (and Makarov Basin). Combined with the 2020–2100 warming trends in Figure 3.10a and d, which show greatest AW temperature increases along the shallower ridges and shelves but

consistent build-up of AW heat throughout the Arctic, this provides further evidence that the Arctic “stores” heat in deep basins (up to around 3000 *m* depth by the end of the century - not shown) but is unable to do so at the shallower boundary current pathways resulting in temperature increases there. Despite high natural variability at the edges of these deepening regions, where again there is variation amongst the ensemble members on the exact lateral extent of the deepening - anthropogenically forced trends drive this increase in bottom depth.

3.3.1.2.4 Seasonal cycle emergence

A notable feature of the eastern time series of AW layer mean temperature, salinity, density and AW depth of Figures 3.8a, b, d and f is the emergence of a seasonal cycle in the ensemble mean during the middle of the 21st century. This is another indication of AW heat shoaling into the surface mixed layer in the eastern Arctic. During freezing season, brine rejection and the resultant deepening of the surface mixed layer allows the AW heat to shoal into the mixed layer, losing heat and becoming fresher and less dense. The AW top (i.e. 0 °C isotherm) then deepens during the winter and the subsequent melt season when convection becomes less vigorous. The spread in the time series of the ensemble members (i.e. natural variability) increases when the seasonal cycle is present, as the AW layer is affected by the perturbed atmosphere of each ensemble member. Notably, despite the increase in natural variability of the AW layer due to this enhanced seasonal cycle, the forced trends are still dominant.

There is no seasonal cycle in the western time series of any property in Figure 3.8. This is not surprising given that, although AW heat shoals into the halocline here, the surface mixed layer in the western Arctic is isolated from the AW by a much thicker halocline layer. When comparing the relative magnitudes of the eastern and western forced trends, it is therefore best to compare the western values with the freezing-season eastern values, i.e. the warmer, saltier, denser eastern values, so as not to incorporate the surface mixed layer into the comparison when it is not involved in the western data. Taking this into consideration, we see that, although the magnitudes of the ensemble mean temperature trends in both regions are comparable, the freshening trend seen in the west is greater than that in the east. This is in agreement with the freshening of the BSBW, as discussed in the previous section, and the freshening of AW upon advection within the Arctic. For most of the time period covered, the ensemble mean density of AW in the east is greater than that in the west due to this

freshening upon advection. However, between 2010 and 2055 the density is lower in the east due to the earlier onset of warming and freshening here. Geostrophically, this would favour an enhanced recirculation of AW along the Lomonosov Ridge during the 2010–2055 period. However, we will find later that this is not what happens.

As mentioned earlier when discussing Figure 3.4, the spatial pattern of ensemble mean shoaling does not align with that of ensemble mean sea ice retreat (Figure 3.7). However, England et al. (2019) found that natural variability can play an important role in sea ice extent in parts of the Arctic. It is therefore worth noting that in Figure 3.10, the ratio of forced trend to natural variability in AW temperature, salinity and density is smallest (although still greater than 1) in the sector to the north of the Siberian shelves due to relatively large natural variability here. This aligns with the large variability of summer sea ice extent in this region in the CESM-LE (England et al. 2019). However, as seen in Figure 3.8, forced AW freshening and warming trends dominate over these seasonal oscillations. The impact of the upper ocean on AW will be investigated in more depth in Chapter 4.

3.3.2 Gateway transports and AW advection

So far in this chapter we have seen how the hydrographic properties of the AW layer are dominated by anthropogenically forced trends throughout the Arctic during the 21st century. The AW layer warms, freshens and shoals in the ensemble mean. The increase in AW heat in the Arctic is related to an increase in inflow temperature, as seen above. Hence, in this section, ensemble mean changes in AW heat being advected through the gateways between the Arctic and Atlantic will be investigated to get a better picture of what is happening at these gateways. The impact of inflow transports on the rest of the Arctic will be investigated in more detail in the final section of this chapter.

3.3.2.1 Hydrography and velocity at the gateways

Figure 3.13 shows ensemble mean hydrographic properties and velocity along a section across the Fram Strait (see Figure 3.2 for location) averaged over 1920–2009 (there was little change during this period) and 30 year periods from 2010 to 2100. The Fram Strait is an important gateway between the Atlantic and Arctic Oceans, being the deepest channel connecting the Nordic Seas and the Arctic Ocean. AW enters the Arctic on the eastern side of the Fram

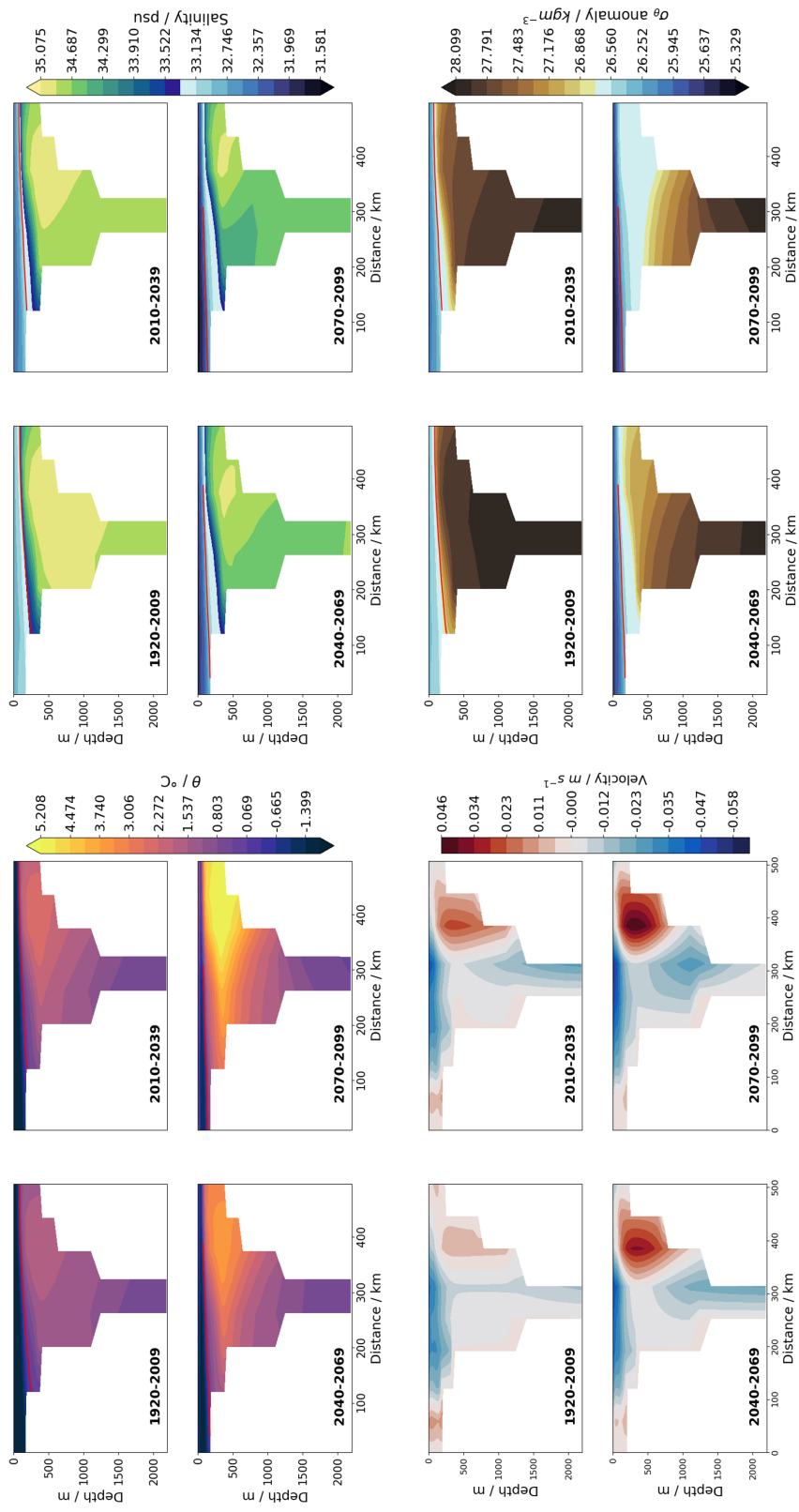


Figure 3.13: Ensemble mean hydrographic properties across the Fram Strait (see Figure 3.2, looking north with an average section latitude of 80 °N) averaged over: (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, and (d) 2070–2099. Red contour lines mark the location of the 0 °C isotherm. The upper left quadrant shows potential temperature, the upper right shows salinity, and the lower right quadrant shows potential density. The lower left quadrant shows velocity, with velocity into the page in red and out of the page in blue.

Strait as the West Spitsbergen Current (WSC) - this can be seen in red on the right of the velocity sections in Figure 3.13. The CESM-LE ensemble mean gives the volume transport of this current through the Fram Strait as 0.9 *Sv* averaged over 1980–2009. Southward outflow near the surface in the central and western Fram Strait is the East Greenland Current (EGC) (which transports fresh polar water and modified BSBW southward out of the Arctic (Schauer et al. 2004)), which has a volume transport of 1.7 *Sv* in the 1980–2009 ensemble mean. The southward flow at greater depths just to the west of the inflowing AW is FSBW returning south towards the Nordic Seas - either after recirculation within the Fram Strait or after advection around the Eurasian Basin - with southward flow to the west of this being the returning BSBW mentioned above (Schauer et al. 2009).

Comparing the 1980–2009 model WSC and EGC volume transports to observational values suggests that the model underestimates the volume transport of these two currents. The WSC has been observed to have a volume transport of around 7 *Sv* (Beszczynska-Moller et al. 2012), and the EGC carries around 9 *Sv* (de Steur et al. 2014; Schauer et al. 2009). This underestimation in volume transports is not surprising given that it is a common feature of coarse-resolution ocean-sea ice components of climate models (Ilıcak et al. 2016). Furthermore, the Fram Strait section used here is around 80 °N, whereas the observations of the EGC transport referenced above are from 78 °50' N, and it is known that this current weakens with latitude here as some recirculates (de Steur et al. 2014). Overall, Figure 3.13 shows that the CESM-LE gives a reasonable representation of the observed key currents through the Fram Strait (Schauer et al. 2004) given its coarse resolution. However, the underestimation of inflowing AW volume transport through the Fram Strait should be borne in mind throughout this chapter, given that AW is the focus of our analysis.

Figure 3.13 shows an increase in the velocity of the AW entering the Fram Strait during the 21st century, along with an increase in the outflow at depth. This intensifying deep outflow is unlikely to be water recirculating within the Fram Strait due to the difference in temperature and depth of this outflow compared to the inflow, and is hence AW which has recirculated around the broader Arctic. The EGC also appears to intensify during this period - combined with the surface freshening, this implies an increase in freshwater output into the Nordic Seas from the Arctic. The temperature sections in Figure 3.13 show an increase in temperature of both the eastern and western intermediate-depth waters in the Fram Strait

and at the surface during the 21st century. This warming is particularly prominent in the inflowing AW, which warms by more than 3.5 °C across the period studied. The freshening seen across the Fram Strait during this period is not surprising given expected increases in seasonal sea ice melt and the resultant increase in freshwater within, and outputted from, the Arctic (Jahn et al. 2020; Stroeve et al. 2007). Despite the freshening and density decrease seen at depth in Figure 3.13, the concurrent freshening and density decrease of the surface layer is more significant, resulting in an increase in local water column stability (not shown). Comparing the density sections in Figure 3.13 with the temperature and salinity sections hints at an interesting change in the nature of water column stratification at the Fram Strait. The structure of the early density sections at intermediate depths is very similar to the structure of the early salinity sections here, indicating that the water column density is predominantly determined by salinity. However, the later intermediate-depth density sections resemble the temperature field. This apparent switch in the nature of stratification from a beta to an alpha ocean is notable. If it occurs on a wider scale throughout the region then it could have important consequences for convection. The changing roles of temperature and salinity in determining the density field are investigated in Chapter 3, when the main focus of the analysis will shift to stratification changes.

Figure 3.14 shows the same sections across the BSO (location given in Figure 3.2), the gateway through which the AW enters the shallow Barents Sea. The temperature, salinity and density sections show a warming and decrease in density across the section throughout the 21st century. The salinity decreases across most of the section, with the exception being the surface which becomes more saline - as expected given predicted future sea ice loss in the region (e.g. Figure 3.7, Onarheim et al. (2017) and Smedsrud et al. (2013)). The velocity field across the BSO is complex but consistent across time, with the Norwegian Coastal Current at the far south (right) of the section bringing in the warmest AW. There is some recirculation near Bear Island (Skagseth 2008), seen here at the shallower bathymetry towards the northern end of the section. The two persistent neighbouring positive and negative velocities that extend throughout the water column in all time periods are barotropic flows circulating in a trough.

The other key entrance to the Arctic Ocean basin for AW is the St. Anna Trough (see Figure 3.2). At least 60 % of the AW that enters the Barents Sea through the BSO cools

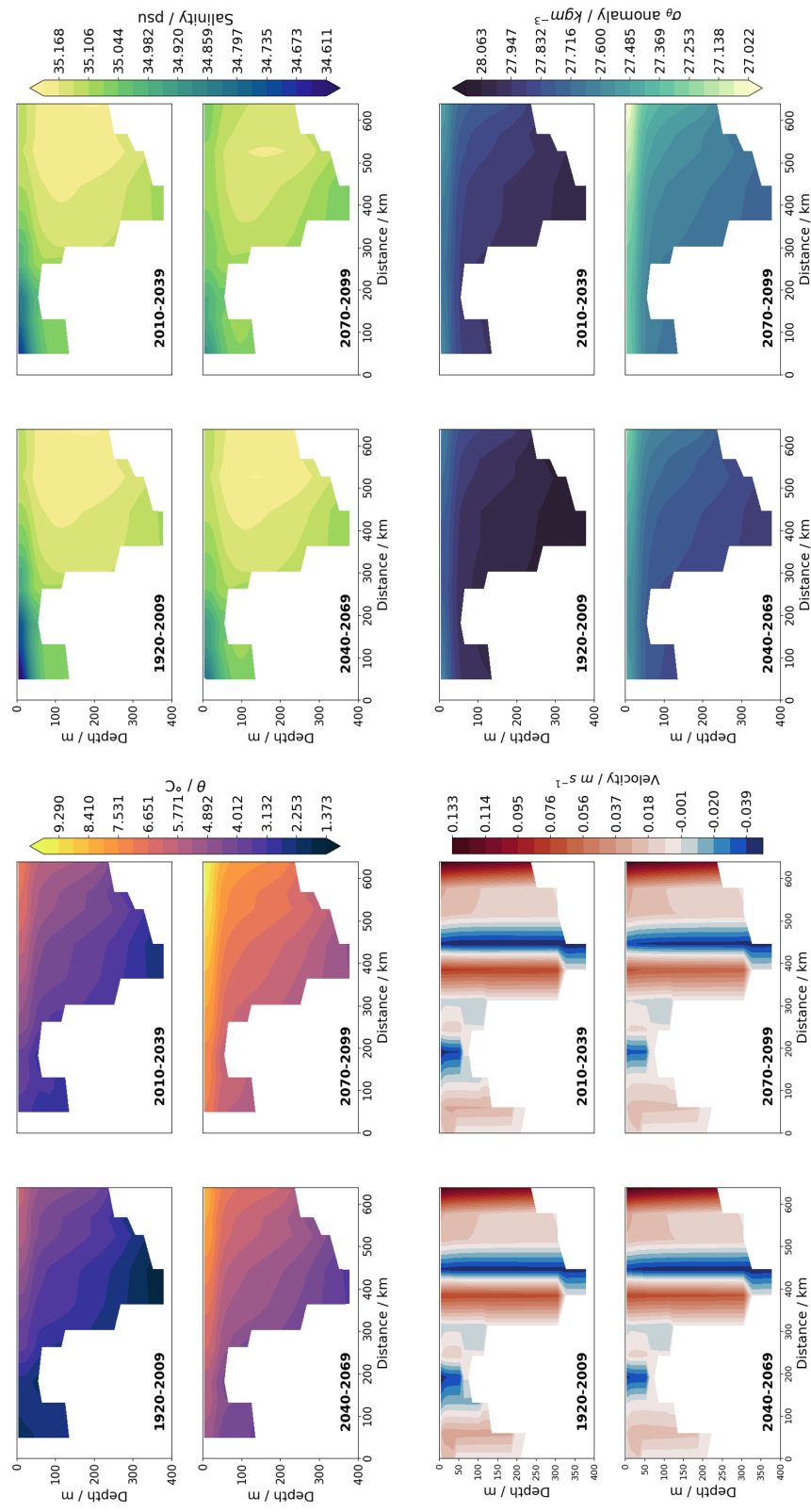


Figure 3.14: As in Figure 3.13 but for the Barents Sea Opening (see Figure 3.2 for location), looking east.

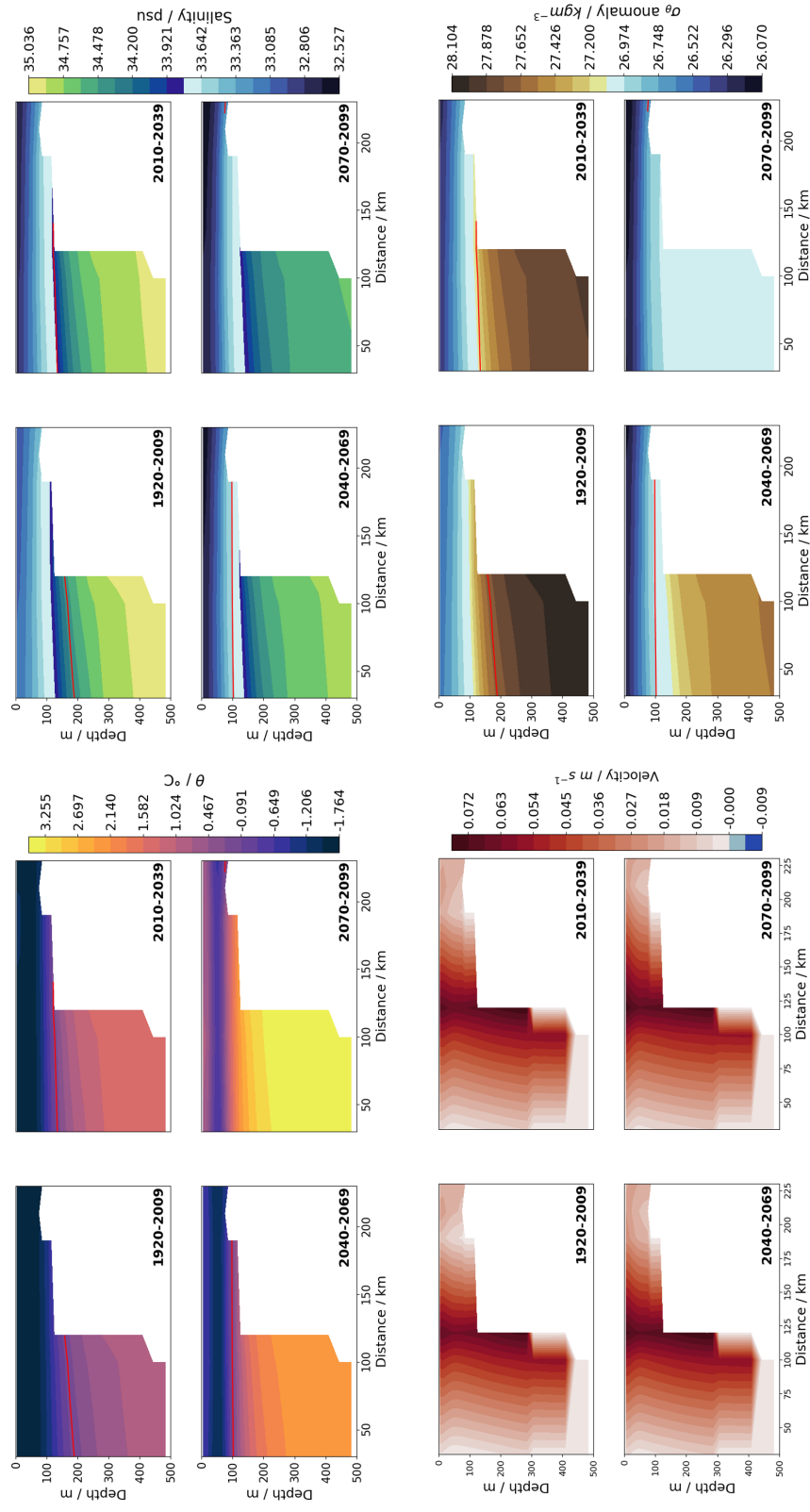


Figure 3.15: As in Figure 3.13 but for the St. Anna Trough (see Figure 3.2 for location), looking north.

and gets transformed into dense water that descends into the Arctic Ocean (Smedsrud et al. 2013) - most of which enters via the St. Anna Trough (Dmitrenko et al. 2015).

Currents in St. Anna Trough flow northward into the Arctic basin. Temperature, salinity, density and velocity sections of ensemble mean data across the St. Anna Trough averaged over the same time periods as earlier are given in Figure 3.15.

The velocity field does not change much in each time period. However, warming and freshening trends are seen across the section in the 21st century, with the 2070–2099 panel being particularly warm and fresh. Notably, the density of water flowing through the St. Anna Trough decreases - this could have important consequences downstream as this water is advected through the Arctic Ocean as BSBW. The warming of water at depth in the St. Anna Trough is in line with Skagseth et al. (2020), who predict that increases in atmospheric temperatures and reduction in sea ice in the Barents Sea will result in a warmer BSBW. This reduction in the efficiency of AW cooling in the Barents Sea has already been observed (Skagseth et al. 2020).

It has also been suggested that the BSBW will not only become warmer, but also more saline in the future as less melt water is entrained into it (Aagaard et al. 2001; Dmitrenko et al. 2015). This is not what is seen in Figure 3.15, where both the surface layer and deeper water freshen. Freshening of the inflowing AW at depth was also seen upstream at the BSO in Figure 3.14, and, notably the depth of the halocline in the St. Anna Trough section does not appear to change (Figure 3.15). Both of these suggest that the freshening seen at depth in the St. Anna Trough is at least partly related to processes upstream of the Barents Sea. This implied freshening of AW in the Nordic Seas is surprising given that temperature and salinity anomalies flowing in the Nordic Seas towards the Arctic have been found to fluctuate in phase (Mulwijk et al. 2018), and suggests that a freshening of the Nordic Seas due to the increase in freshwater outflow into this region (Figure 3.13), and/or freshening coastal currents in the region, may be driving this change. The freshening seen at the BSO in Figure 3.14 is less than that seen at the St. Anna Trough (Figure 3.15), however. Therefore both upstream freshening in the Nordic Seas and processes within the Barents Sea play roles in this freshening. The Barents Sea becomes ice-free year-round in the ensemble mean by 2070–2099 (see Figure 3.7), and between 2061 and 2088 amongst ensemble members (Onarheim et al. 2017), so a reduction in local brine rejection each freezing season seems a likely contributor

to the local component of the freshening.

3.3.2.2 Gateway transports

The main focus of this work is on AW heat within the Arctic and how this might change in the future. Therefore the key question that arises from the above hydrographic sections is: what do these changes in the temperature and intensity of gateway currents tell us about future temperature transport into the Arctic from the Atlantic?

Figure 3.16 shows time series of the ensemble mean volume and temperature transports across the Fram Strait. Temperature transport (TT) is calculated in the same way as heat transport, and is computed relative to $0\text{ }^{\circ}\text{C}$ as shown in Equation 3.1

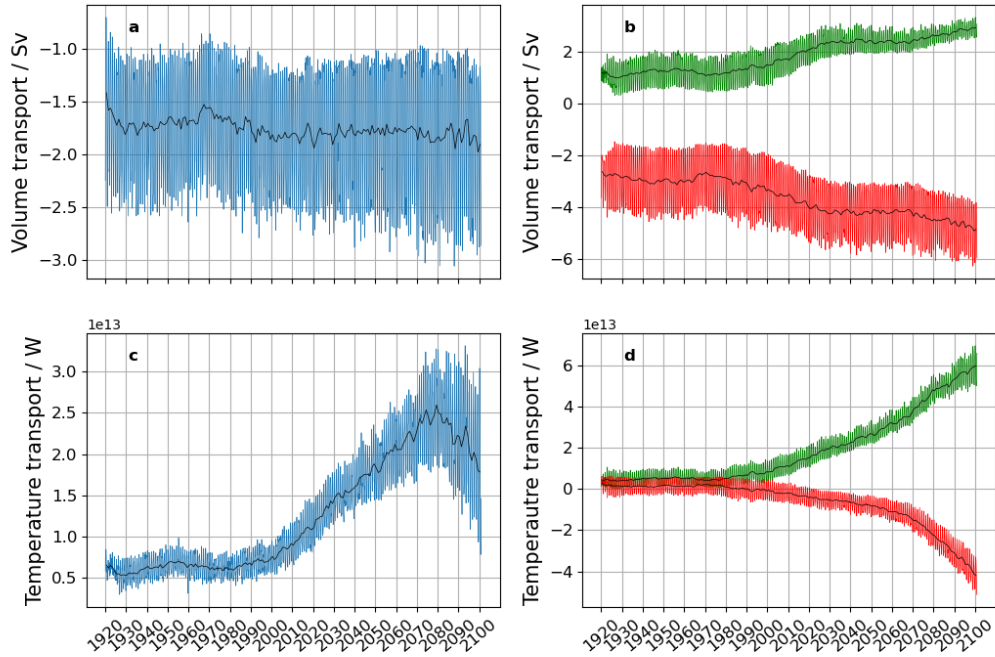


Figure 3.16: Time series of ensemble mean volume and temperature transport through the Fram Strait (location given in Figure 3.2). Quantities plotted are (a) net volume transport, (b) volume transport into (green) and out of (red) the Arctic (c) temperature transport and (d) temperature transport into (green) and out of (red) the Arctic. Black lines show annual mean. A negative temperature transport out of the Arctic can be interpreted as heat being removed from the Arctic. Likewise a positive temperature transport out of the Arctic represents cold water advected out of the Arctic. Temperature transports were computed relative to $0\text{ }^{\circ}\text{C}$.

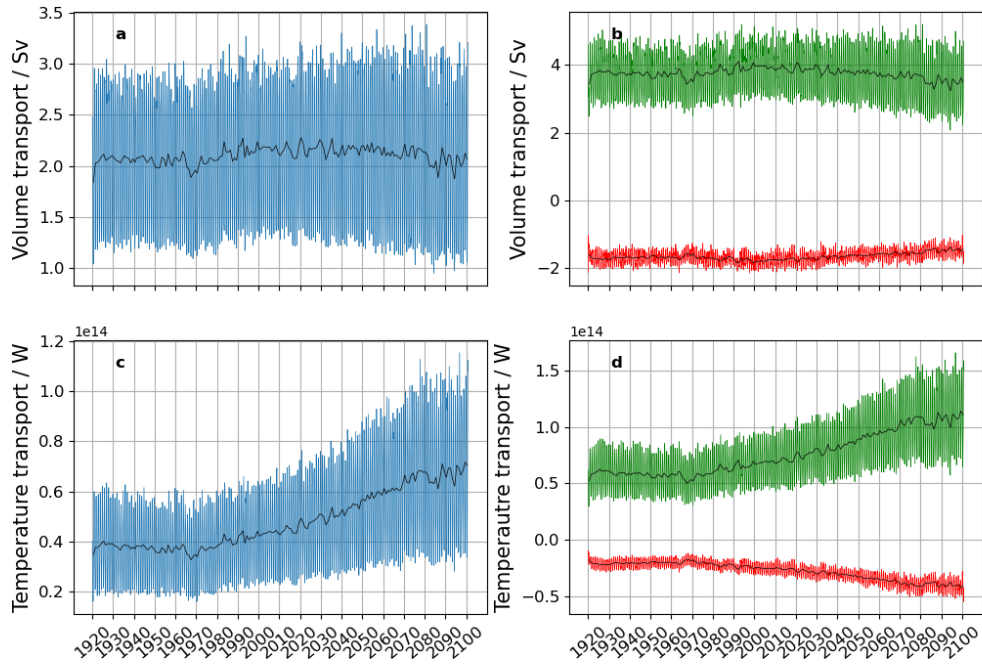


Figure 3.17: As in Figure 3.16 but for the Barents Sea Opening.

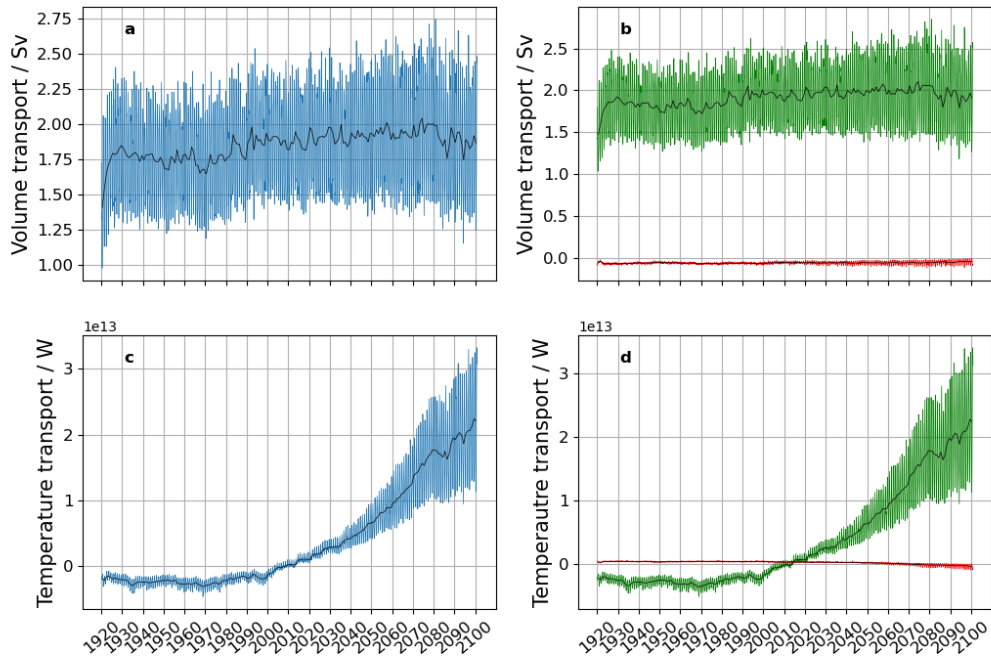


Figure 3.18: As in Figure 3.16 but for the St. Anna Trough.

$$TT = \int \int \sigma_{\theta} c_p \theta v_p dz dx \quad (3.1)$$

where σ_{θ} is potential density, c_p is the specific heat capacity of sea water in the model, θ is potential temperature, v_p is velocity perpendicular to the section, x is distance along the section, and z is depth.

Figure 3.16a shows a southward net volume transport (at all depths, not just the surface EGC) through the Fram Strait of around $1.8 Sv$, with significant monthly variability, in agreement with observations (Schauer et al. 2009). In fact, the magnitudes of net volume and temperature transports computed at both the Fram Strait and BSO are in general agreement with observations (as will be shown). Eddies, although parameterised and not explicitly represented in CESM-LE (POP2 parameterises submesoscale mixed layer eddies following Fox-Kemper et al. (2008), uses the parameterisation described in Danabasoglu et al. (2008) for near-surface boundary eddy flux, and uses Gent-McWilliams parameterisation for horizontal tracer diffusion used elsewhere in the ocean), play an important role in recirculation of AW within the Fram Strait (Hattermann et al. 2016), so this agreement is notable. These realistic volume and temperature transports therefore suggest that the eddy parameterisation is accurate. It should be noted that the agreement between observations and volume/temperature transports through these straits is in contrast to the model underestimations of individual current transports described earlier. Figure 3.16b shows the breakdown of net volume transport into northward transport through the Fram Strait into the Arctic (green) and southward transport out of the Arctic (red). The increasing trends in these directional volume transports from the late 20th century are symmetrical across the x-axis, suggesting that the same process(es) are responsible for the increase in magnitude of both the northward and southward volume transport. Anti-correlation between inflow and outflow here is in fact often seen in models (e.g. Jahn et al. (2012)).

Note that seasonal variations in volume transport out of the Fram Strait in Figure 3.16 are larger than those of the inflow. This may be the result of seasonal differences in how much water is recirculated within the strait, which is dependent on the wind-driven Greenland Sea gyre (de Steur et al. 2014).

The net temperature transport into the Arctic through the Fram Strait increases steadily between 2000–2070 (see Figure 3.16c) as would be expected given the increase in temperature

of the inflowing AW seen in Figure 3.13. However, after 2070 temperature transport begins to decrease. Figure 3.16d suggests that this plateau and subsequent decrease in temperature transport into the Arctic via the Fram Strait is related to an increase in the temperature transport out of the Arctic. After 2070, the magnitude of temperature transport out of the Arctic increases at a greater rate than the temperature transport into the Arctic, resulting in the net decrease in temperature transport through the Fram Strait. The reasons for this increase in outflowing temperature transport will be discussed later.

Figure 3.17 shows the same time series for transports through the BSO (with green lines representing transport eastward over the section towards the Arctic/in to the Barents Sea, and vice versa for red plots). The BSO volume transport shown in panels a and b of Figure 3.17 has no notable trend, in contrast to the volume transports seen above at the Fram Strait, and its magnitude over the 20th century is again in agreement with observational estimates of a 2 Sv net inflow (Ingvaldsen et al. 2002; Schauer et al. 2002b). The increase in net temperature transport across the BSO into the Barents Sea seen in panel Figure 3.17c is therefore a result of the temperature increase at the section seen in Figure 3.14. This increase in net temperature transport begins around 1970, although is still smaller than month-to-month variability.

Time series of volume and temperature transports through the St. Anna Trough are shown in Figure 3.18. There is a small positive trend in volume transport here. Temperature transport through the St. Anna Trough increases from around 1970 by a similar magnitude to the increase in temperature transport at the BSO and Fram Strait. In contrast to the BSO, the monthly variability in ensemble mean temperature transport at the St. Anna Trough increases markedly during the latter half of the 21st century. As mentioned, individual CESM-LE ensemble members show the Barents Sea becoming ice free throughout the year for the first time between 2061 and 2088 (Onarheim et al. 2017). Given that much of the increase in monthly variability in St. Anna Trough temperature transport occurs around this period, changes in sea ice conditions - and hence the magnitude of winter AW heat loss - are likely to be the primary cause of this increase in the seasonal cycle of temperatures. Despite this apparent increase in winter AW (BSBW) heat loss, AW heat flowing into the Arctic through the St. Anna Trough increases during the 21st century.

The atmosphere likely plays a key role in the gateway transport time series of Figures

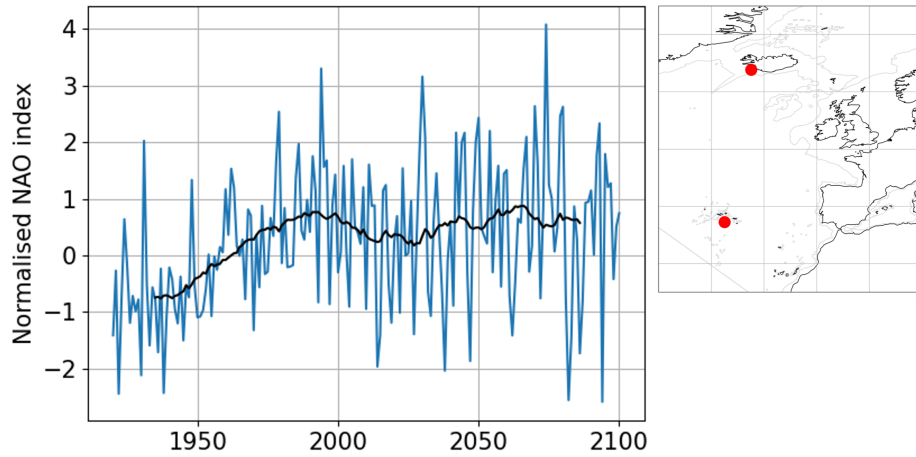


Figure 3.19: Time series of normalised NAO index (blue) and its running 30-year mean (black), computed using ensemble mean data from a point-wise sea level pressure difference between Iceland and the Azores. Red dots on the map show the locations used.

3.16–3.18. AW inflow to the Nordic Seas and Arctic Ocean is primarily governed by wind stress over the Nordic Seas and hence the strength of the Greenland Sea Gyre (e.g. Chatterjee et al. (2018), Muilwijk et al. (2019), Muilwijk et al. (2018), and Nøst et al. (2003)), along with the halosteric height of the Greenland Sea (itself governed by sea ice export) (Wang et al. 2020). In the CESM-LE specifically, inflow to the Nordic Seas from the south increases up until at least 2080, with strengthened flow within the Nordic Seas consistent with a decrease in local sea level pressure and a cyclonic circulation anomaly (Årthun et al. 2019). The increase in volume transport both into and out of the Arctic via the Fram Strait seen in both the velocity sections of Figure 3.13 and the time series in Figure 3.16a and 3.16b is consistent with an increase in the strength of the Greenland Sea Gyre. This would result in an equal increase in both northward and southward volume transport, which is indeed what is seen in these figures. At least part of this inflow increase can also be attributed to an increase in baroclinic flow as the density gradients across the strait strengthen. Given that AW inflow is related to Nordic Seas wind stress, AW volume transport through the Arctic gateways is linked to the NAO (Muilwijk et al. 2018), which is known to have a positive trend in CMIP5 models, including CESM1 (using RCP 4.5 scenario) during the 21st century (Gillett et al. 2013). Trends in gateway volume transports seen in Figures 3.16–3.18 are therefore likely related to the NAO.

Figure 3.19 shows the normalised NAO index along with its 30-year running mean, com-

puted from ensemble mean data. In order to compute the index, sea level pressure for each month at a location just south west of Iceland and another in the Azores was first normalised. This was done by subtracting the monthly means and dividing by the monthly standard deviations (both computed over the historically-forced 1920-2005 period) for the corresponding month of each year. The monthly normalised sea level pressure at the Icelandic location was then subtracted from the Azores location. An annual value was obtained by averaging the December–March values and assigning this to the year in which the January falls. The resulting NAO index therefore reflects the state of the NAO relative to its mean state over 1920–2005.

The NAO generally increases over the time period studied (Figure 3.19), in line with what was expected based on the general increase in AW volume transport through the Fram Strait. However, while the AW inflow increase mainly occurred between 1970–2030 and from 2070 onwards, the NAO increased between 1920–2000, after which it remains above the historical average. This temporal disagreement isn’t surprising given the method used in the NAO quantification. Using a “station-based” (i.e. fixed location) method doesn’t always fully capture the variations of the NAO, as the centers of the pressure systems involved in the oscillation can shift laterally. Sea level pressure fields will be examined in more detail later, which will reveal more about the link between atmospheric circulation and AW advection than can be captured through a station-based NAO index.

The lack of trend in BSO volume transports in the CESM-LE, despite the increase in flow into and within the Nordic Seas discussed above, is attributed by Årthun et al. 2019 to local changes in sea level pressure at the BSO itself in the model.

Unlike volume transport, the increase in southward heat transport through the Fram Strait seen around 2070 in Figure 3.16 cannot be described by changes in the Greenland Sea Gyre/Nordic Seas wind stress, given its dissimilarity with the inflowing temperature transport. This increase is also seen when only water deeper than 200 *m* is considered in the computation, so it is not related to the increase in freshwater outflow seen at the surface of the section. It is instead the result of an increase in outflowing temperature transport at depth. Based on Figure 3.13, the increase in southward heat transport through the Fram Strait therefore appears to be related to the outflow at depth in the central Fram Strait - i.e. AW returning to the Fram Strait after being advected within the Arctic Ocean. Therefore,

the increase in temperature of the outflowing AW from 2070 onwards is greater than the increase in inflowing AW temperature. Based on the circulation changes seen above, and the AW layer temperature maps in Figure 3.3, both the FSBW or BSBW could contribute to this (given that the BSBW can take several decades to travel around the Arctic (Wefing et al. 2020)).

The increase in heat transported out of the Fram Strait could be caused by the warming of the inflowing BSBW at the St. Anna Trough, as seen above, both through the BSBW retaining some of its higher temperature throughout its advection around the Arctic, and through the interaction between BSBW and FSBW near the St. Anna Trough (Schauer et al. 2002a). It could also be a result of reduced heat loss from the AW within the Arctic. This is surprising given that the increase in ventilation of the AW layer due to Atlantification is expected to result in a cooler, fresher AW (Pérez-Hernández et al. 2019), but it may be that towards the end of this century surface air temperatures increase enough to cause reductions in how much heat is lost from the AW to the atmosphere even when it is directly ventilated - a northward extension of the reduction in AW cooling efficiency already observed in the Barents Sea (Skagseth et al. 2020). However, maps of sea ice concentration (see Figure 3.7) reveal that the Eurasian Basin is still fully covered by sea ice at the end of winter in the ensemble mean during 2070–2099. Given that the increase in outflowing temperature transport through the Fram Strait is seen across the range of seasons in the time series (Figure 3.16), increased seasonal AW ventilation within the Arctic is unlikely to be the cause. Therefore, a combination of warming BSBW inflow and increases in the temperature of other water masses interacting with AW within the Arctic (e.g. the halocline and laterally-advected shelf flows) are the most likely causes of the change in heat transport through the Fram Strait from 2070. Halocline changes will be investigated in more detail in Chapter 4, which focuses on the upper portion of the water column.

Regardless of its causes, the reversal of the increasing trend in net heat transport through the Fram Strait shows that there is a limit to the amount of oceanic heat that can be “absorbed” by the Arctic Ocean (either through a limit on temperature or through a limit on AW layer thickness related to heat build-up in the deep basins, as seen earlier). It does not appear to be the case that increasing trends in the heat content of inflowing FSBW and BSBW will necessarily result in a compensatory increase in heat loss from the AW to other

water masses within the Arctic. This suggests that the Nordic Seas may experience increased inflow of heat southward from the Arctic in the future, not just from the Atlantic.

3.3.2.3 AW advection within the Arctic

We now turn our attention away from gateway transports and to the AW velocity field within the Arctic. The velocity field at 500 *m* depth averaged over several periods is given in Figure 3.20, with the difference between the 1920–1949 field and the 2070–2099 field (latter minus former) given in Figure 3.21. This depth was chosen as it is included within the AW layer at all times and locations in the deep basins, and hence gives an idea of upper AW advection throughout the Arctic.

Figure 3.20 reveals that, during the 20th century simulation, the upper AW circulation in the Arctic exists in two distinct regimes - with a cyclonic circulation in the Eurasian Basin, and an anticyclonic circulation in the Amerasian Basin. AW appears to penetrate into the Amerasian Basin through gaps in the Lomonosov Ridge. However, in the 21st century this circulation drastically changes. A pan-Arctic boundary current is established, with AW being advected over the Lomonosov Ridge near the Laptev Sea, travelling around the Canada Basin and back to the Fram Strait. The strong cyclonic recirculation in the Eurasian Basin seen in previous periods diminishes to just the western Eurasian Basin, with FSBW advecting over the central Lomonosov Ridge into the western Arctic rather than recirculating throughout the Eurasian Basin interior and along the ridge.

It was seen above (Figures 3.16–3.19) that trends in AW advection through the Fram Strait may be linked to an increase in the NAO index. But does this impact AW advection elsewhere in the Arctic? In a study by (Mulwijk et al. 2019), the NAO was found to be positively correlated with AW flow even as far as the Canada Basin in some models, with these models showing an increase in AW volume transport in western basins when there was anomalously cyclonic wind stress over the Greenland Sea - although CESM was not included in this study. Observations have shown that AW intrudes into the Makarov and Canada Basins during the positive phase of the AO (the NAO and AO are directly linked (Deser 2000)), with flow along the Lomonosov Ridge weakened and most AW crossing the Lomonosov Ridge into the Canada Basin rather than being advected along it (Karcher et al. 2012). This is much like what is seen in the CESM-LE AW velocity field in later periods

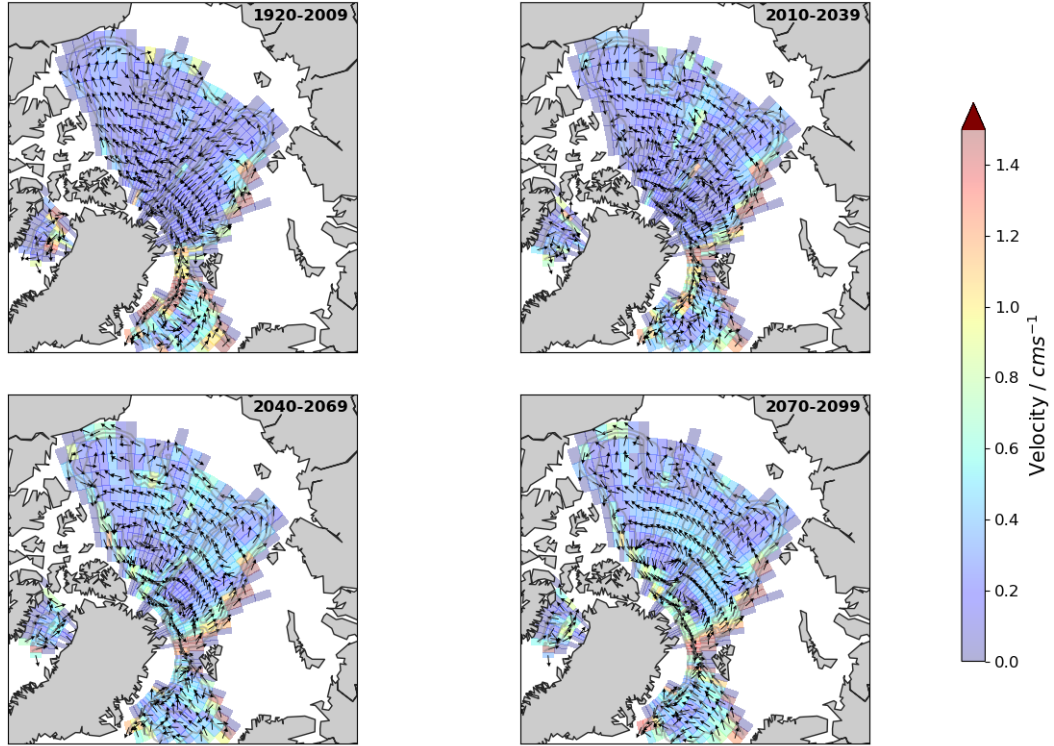


Figure 3.20: Ensemble mean velocity at 500 m depth, averaged over the periods denoted at the top right of each panel. Vectors give the direction of the velocity, with colours denoting speed. To aid clarity, the grid has been sampled at every third point.

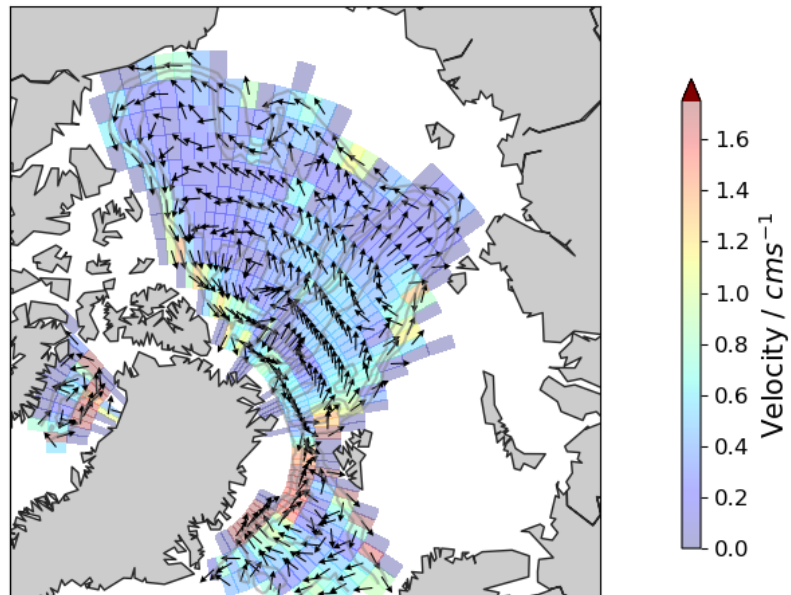


Figure 3.21: Change in ensemble mean velocity at 500 m depth between 1920–1949 and 2070–2099. Vectors give the direction of the velocity, with colours denoting speed. To aid clarity, the grid has been sampled at every third point.

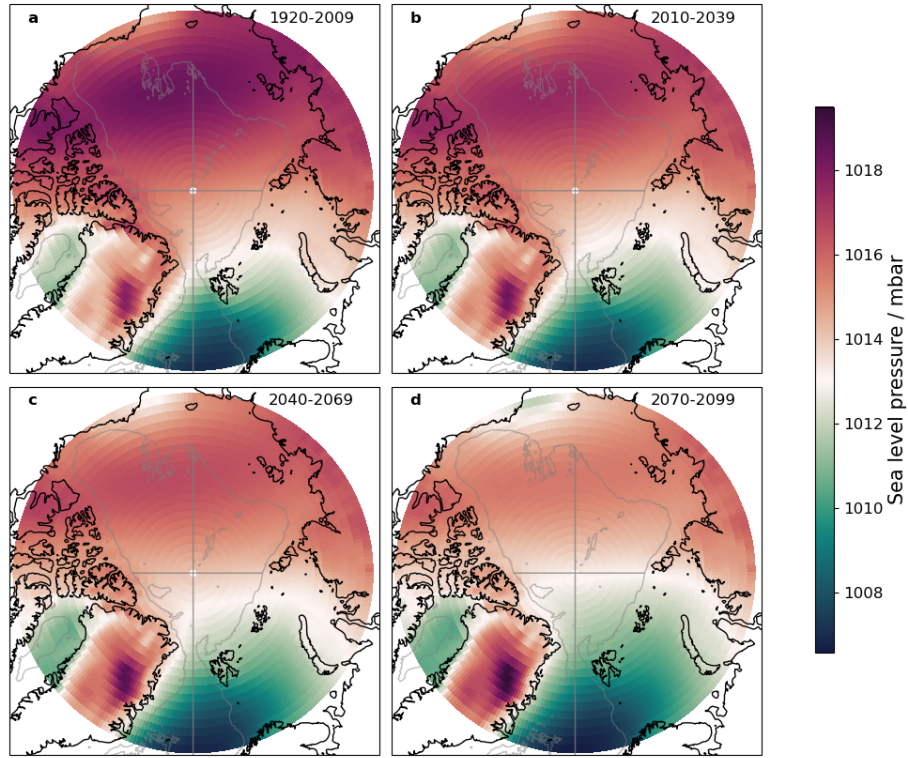


Figure 3.22: Maps of ensemble mean sea level pressure averaged over the periods denoted at the top right of each panel. Green colours show low (below average) pressure with red areas showing high pressure.

of Figure 3.20. Karcher et al. (2012) found that these changes come about in part due to a weakened Beaufort High. As mentioned earlier, the station-based NAO index used in Figure 3.19 does not always give the full picture of atmospheric circulation changes. Therefore to get an idea of whether the atmosphere may play a role in the shift in AW velocity field seen in Figures 3.20 and 3.21, the ensemble mean sea level pressure fields averaged over the same periods are shown in Figure 3.22. The typical pattern of low pressure over the Nordic Seas and high pressure centered over the western Arctic can be seen in all time periods. However, during the 21st century, the high pressure weakens and the low pressure region advances northward and eastward, encroaching into the Eurasian Basin and enhancing over the Barents Sea. This shift coincides with the shift in AW circulation (Figure 3.20) and aligns with what would be expected from a general shift towards a more positive NAO/AO. That this shift was not captured by the NAO index used in Figure 3.19 is likely related to the apparent eastward shift of the low pressure region in these maps (although our maps do not show data far south enough to confirm). Nevertheless, these sea-level pressure fields, combined with the apparent impact of the NAO on pan-Arctic AW circulation (Muilwijk

et al. 2019), suggest that the trend towards a more expansive low over the Nordic seas and a weaker Beaufort High in this and other CMIP5 models during the 21st century (Gillett et al. 2013) plays an important role in the changes in AW inflow and circulation seen in Figures 3.16–3.18, 3.20 and 3.21.

Another argument supporting wind-driven changes in AW advection is given by the velocity field at 5m (not shown). This is more consistent across time than the 500 *m* velocities, but shows the lateral extent of the Beaufort Gyre reducing during the 21st century - a sign that the reduction in intensity of the Beaufort High impacts the Beaufort Gyre, which would explain the increase in the cyclonicity of AW flow in the western Arctic in later periods, given that it is thought that the Beaufort Gyre flow influences the flow of upper AW beneath it (Karcher et al. 2012; Lique et al. 2015b). A reduced time-averaged Beaufort High is perhaps not surprising in a warming Arctic given that Moore et al. (2018) linked warm surface air temperatures in the Arctic and reduced sea ice extent to an intensified low over the Barents Sea and enhanced propagation of cyclonic weather systems into the western Arctic, reducing the average sea level pressure here.

Overall, ensemble mean warming of AW in the Arctic - seen earlier in the chapter to be related to increasing inflow temperatures at the Atlantic-Arctic gateways - is in line with an increase in temperature transport through these gateways during the 21st century. This temperature transport increase is driven by increasing temperature of AW upstream of the gateways, and an increase in the volume transport of this warmer inflowing water due to changes in atmospheric circulation that resemble a positive trend in the NAO/AO. This large-scale atmospheric circulation pattern also causes a change in the circulation of AW within the Arctic, and would perhaps explain the shift in AW layer thickening from the western to eastern Eurasian Basin seen in Figures 3.11 and 3.12. However, comparing changes at the inflow with those seen at various locations in the Arctic can be done more broadly by utilising the ensemble members themselves. This will be the focus of the next section.

3.3.3 The link between AW variability in the Arctic and gateway transports

This section will focus in more detail on the impact of AW inflow on the properties of the AW elsewhere in the Arctic. We have seen so far that the warming of AW across the Arctic

during the 21st century in the CESM-LE ensemble mean can be attributed to a warming inflow from the Nordic Seas. However, questions remain over the spatial extent of the influence of the inflow from each gateway, and how strong this influence is throughout the Arctic. For instance, although the western Arctic, far along the AW advection pathway, warms due to warming AW inflow (see above), does it warm to the same extent as the inflow? Are temporal variations in AW inflow temperature reflected in temporal variability in AW temperature at other locations in the Arctic? The individual ensemble members can be used to gain insight into these questions.

3.3.3.1 Natural variability in AW transports through gateways

Figures 3.23 and 3.24 show ensemble member volume and temperature transports through the Fram Strait and St. Anna Trough - the two main gateways through which AW enters the Arctic Ocean basin. Both net transport and directional transports into and out of the Arctic are plotted, along with the ensemble mean (which has been shown in Figures 3.16 and 3.18). The key takeaway from these figures is that, apart from St. Anna Trough volume transport,

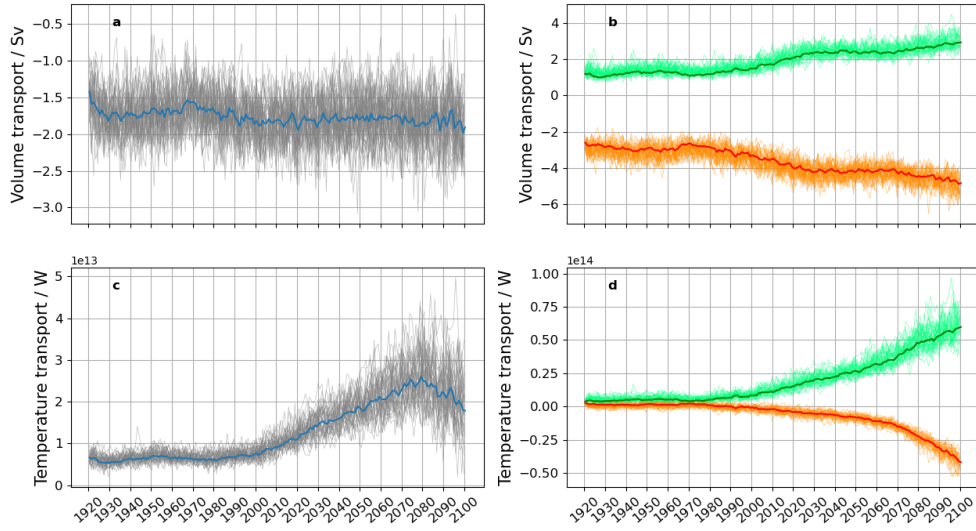


Figure 3.23: Time series of (a) ensemble member (grey) and ensemble mean (blue) net volume transport across the Fram Strait; (b) ensemble member volume transport in to (green) and out of (orange) the Fram Strait, with corresponding ensemble mean transports in dark green and red, respectively; (c) ensemble member (grey) and ensemble mean (blue) net temperature transport across the Fram Strait; (d) ensemble member temperature transport in to (green) and out of (orange) the Fram Strait, with corresponding ensemble mean transports in dark green and red, respectively.

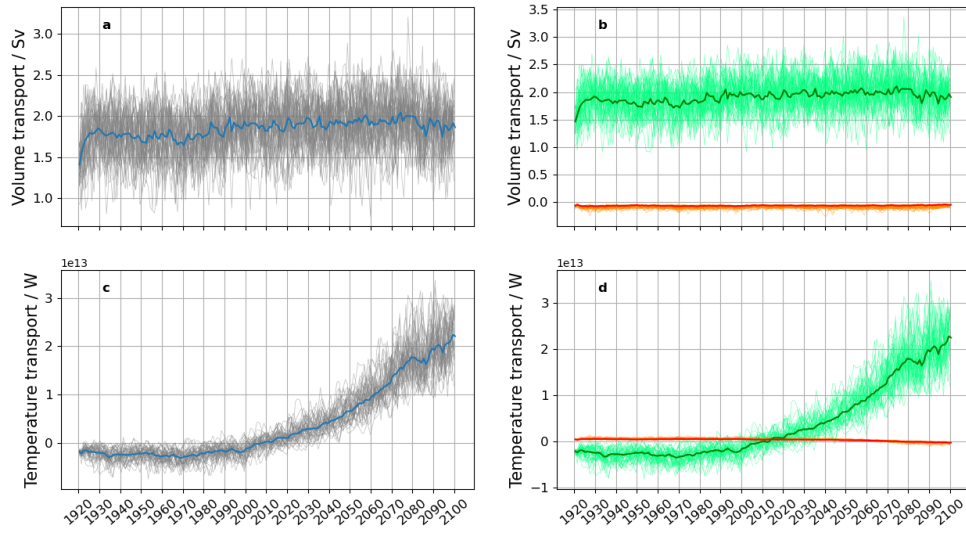


Figure 3.24: As for Figure 3.23 but for the St. Anna Trough.

natural variability in the directional volume and temperature transports in to the Arctic through these gateways is small compared to the anthropogenically forced trend shown by the ensemble mean in the 21st century, as expected based on the results earlier in this chapter. Net volume transports through both gateways have high variability between ensemble members, however this is not surprising given the role that wind can play in inter-annual variability of AW inflow (Chatterjee et al. 2018). This high volume transport variability is likely the main cause of the high natural variability in Fram Strait net temperature transport towards the end of the 21st century, given the high temperature of both the inflow and outflow at this time (Figure 3.13). These individual ensemble member time series will be used as the basis for the remaining analysis in this section.

3.3.3.2 30-year temperature trends

Figure 3.25 shows the maximum (out of all possible integer-year time lags) 30-year ensemble trend correlations between Fram Strait inflow and AW layer-mean temperature. The corresponding time lags at which the maximum correlation occurs are also plotted. The latest inflow period used was 2010–2039 so as to allow reasonable time for AW advection within the Arctic before the end of our data set.

Figures 3.25a and e show that temperature trends in Fram Strait advection in 1920–1949 were moderately correlated with those in the rest of the Arctic with correlation coefficients

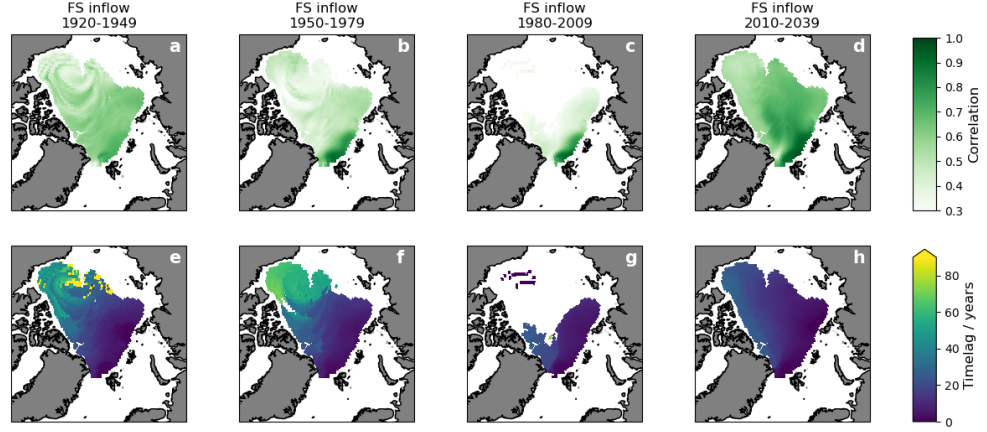


Figure 3.25: (a)-(d) maximum ensemble trend correlations between 30-year trends in Fram Strait AW inflow temperature and AW layer mean temperature at each grid point. (e)-(h) time lag corresponding to the maximum correlation. If the trend at a given location is not significant in at least half of the ensemble members, the region is stippled (however this criteria was not met in this figure). Correlations less than 0.3 are deemed weak and hence are not plotted. Correlation values are Pearson’s correlation coefficient.

between 0.3 and 0.7, and highest correlation near the inflow, as expected. Figure 3.25e implies that FSBW is recirculated within the Eurasian Basin within 20 years, whereas it takes much longer (several decades) for advection into the western Arctic. It is not surprising, therefore, that correlation here is smaller than in the Eurasian Basin after heat loss during advection. Figures 3.25b and f show that the correlation between inflowing Fram Strait AW temperature trends and those within the Arctic weakens for the period 1950–1979 across most of the Arctic, with the correlation just downstream of the inflow increasing to over 0.8. Advection time scales between the Fram Strait and the Canada Basin reduce compared to 1920–1949 to around 60–70 years. Figures 3.25c and g show that this correlation reduces even further for AW that enters the Fram Strait over 1980–2009, with many locations having a maximum correlation coefficient of less than 0.3 (again with the exception of the region downstream from the inflow where the correlation remains high).

There’s a notable change in this weakening trend for the 2010–2039 inflow however. Correlations between inflowing Fram Strait temperature trend and AW temperature trends elsewhere in the Arctic are moderate–strong everywhere, with highest correlations generally in the eastern Arctic. Implied advection timescales in the western Arctic have reduced markedly compared to the 1920–1949 period, being less than 30 years everywhere.

The change in advection timescales in Figure 3.25 can largely be explained by the shift

in circulation seen in Figure 3.20. The long advection time between the Fram Strait and western Arctic in Figures 3.25e and f represents the time taken for AW to enter the Eurasian Basin and reside there until the AW becomes advected more strongly into the western Arctic at the turn of the century (Figure 3.20). The short advection time scales between the inflow and the western Arctic in 2010–2039 show that the pan-Arctic AW circulation regime of the 21st century results in a much quicker transfer of AW to the Canada Basin. This is reflected in the higher correlations between inflowing and local temperature trends during this period. It appears that, similarly to (Chatterjee et al. 2018) who found that faster through-flow of AW in the Nordic Seas resulted in warmer AW entering the Arctic, a faster advection of AW from the Fram Strait towards the western Arctic results in more similar temperature trends between these two locations. The apparent role of sea level pressure fields in AW circulation changes discussed earlier (see Figure 3.22) therefore also appears to accelerate warming in the western Arctic.

The reduction in correlation coefficient between temperature trends in Fram Strait inflow and temperature trends elsewhere during the 20th century is likely related to the Atlantification of the western Eurasian Basin/region just downstream from the inflow, where correlation increases during this time. As the AW becomes more ventilated here, changes in temperature in this region may be damped by the atmosphere, resulting in AW downstream of this region not reflecting the inflow temperature change. The spread of the high correlation region from the inflow to elsewhere in the Arctic in Figure 3.25 matches the spatial pattern and timing of Atlantification described in literature (e.g. Polyakov et al. 2017) where shoaling and warming of AW starts in the late 20th century in the western Eurasian Basin before spreading to the eastern Eurasian Basin. The spread of the high correlation to the western Arctic in our model data is likely due to the large scale changes in AW advection described earlier.

Figure 3.26 shows the same quantities computed for AW inflow through the St. Anna Trough (i.e. BSBW). Figures 3.26a-d show that BSBW influence on AW temperature trends within the Arctic is initially widespread, but - similar to Fram Strait inflow - its spatial influence diminishes to nearer its inflow during the 20th century. Unlike the Fram Strait inflow, however, this influence shrinks further for inflow during 2010–2039. The warming seen in the western Eurasian Basin and western Arctic in the 21st century throughout this chapter (e.g Figure 3.3) therefore seems to be influenced more by FSBW than BSBW, despite

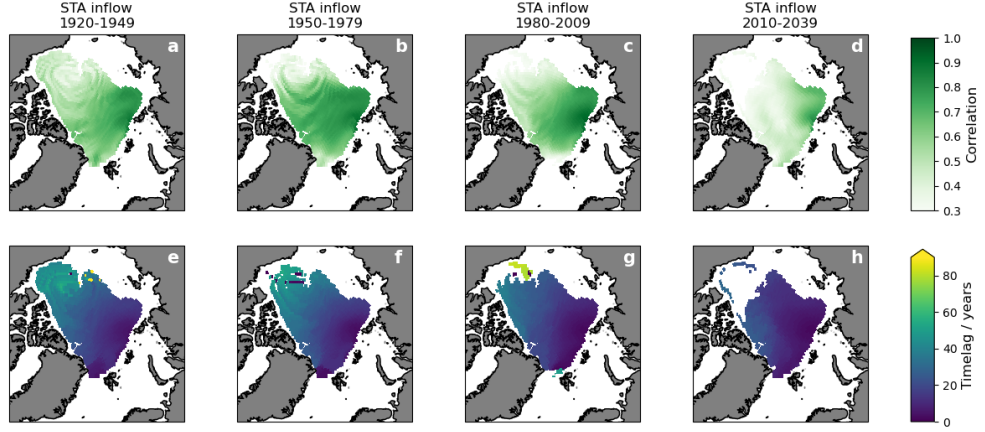


Figure 3.26: As for Figure 3.25 but for inflow through the St. Anna Trough.

increasing temperature transport to the Arctic via the St. Anna Trough during 2010–2039 (see Figures 3.18 and 3.24).

3.3.3.3 Dominant timescales of variability

We now turn our attention to whether temporal variations in AW inflow temperature are reflected in temporal variability in AW temperature at other locations in the Arctic.

Figure 3.27 shows the first and second-most dominant timescales of variability in AW layer mean temperature across the Arctic. The ensemble average, standard deviation and mode are plotted to give an idea of the distribution across the ensemble. Focusing on the ensemble averages in the upper left quadrant of the figure, we see that the timescales of variability are of the order of decades. The dominant and second-most dominant timescales of variability increase in the western Arctic from 1920–2000 to 2020–2100, while they decrease in the eastern Arctic. Along the western boundary, before the pan-Arctic boundary current flows in this region, the dominant timescale of variability is very short, indicating that highly variable local influences such as shelf flows influence the temperature of AW here. The shift to a longer timescale here during 2020–2100 - similar to that elsewhere in the Arctic - suggests that advected AW carries these decadal timescales of variability within the Arctic.

The small values of the ensemble standard deviation shown in the upper right quadrant of Figure 3.27 shows good agreement between ensemble members on these timescales of variability.

The low number of ensemble members with spatially consistent modal values of the

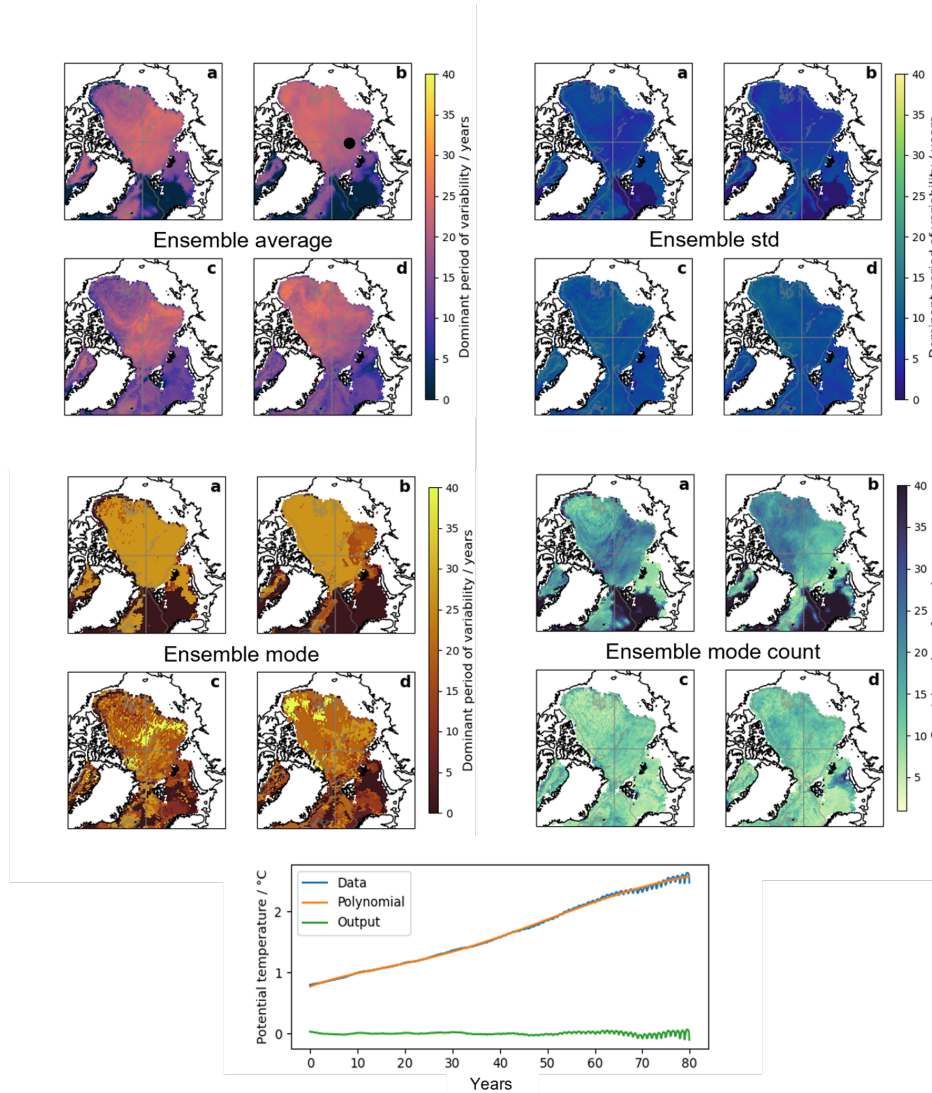


Figure 3.27: The attributes plotted in each labelled subplot are as follows: (a) dominant timescale of variability over 1920–2000, (b) dominant timescale of variability over 2020–2100, (c) second-most dominant timescale of variability over 1920–2000, and (d) second-most dominant timescale of variability over 2020–2100. The statistical parameters plotted in each quadrant are as follows. Upper left quadrant: ensemble average of the dominant/second-most timescale of variability in the mean potential temperature of the AW layer. Upper right quadrant: ensemble standard deviation in the dominant timescale of variability in the mean potential temperature of the AW layer. Lower left quadrant: ensemble modal value of the dominant timescale of variability in the mean potential temperature of the AW layer. Lower right quadrant: how many ensemble members had these modal values as their dominant timescale of variability. The time series in the bottom panel gives context to what is shown in the rest of this figure. The original 2020–2100 time series of AW layer mean temperature at the black point shown on subplot (b) of the upper left subplot is plotted (blue). A polynomial has been fitted to this data (orange) and then subtracted from the data to give a detrended output (green). A Fourier transform is performed on this output to quantify dominant periods of variability in the signal.

second-most dominant timescale of variability (Figure 3.27c and d) suggest that only the modal values of the dominant timescales of variability (Figure 3.27a and b) can be meaningfully interpreted. These modal values are, in general, larger than the ensemble average, indicating that the ensemble distribution has a negative skew, with some members having particularly short dominant timescales of variability. Most of the Arctic has a modal value of approximately 30 years. The western boundary has a shorter modal timescale, with good agreement between ensemble members. Regions with low mode count are the western Arctic in 1920–2000 and the eastern Eurasian Basin in 2020–2100. The latter also shows a lower modal timescale of variability, again indication that local influences, not variability advected with the AW, could be playing a role in this (similarly to the western boundary).

Upstream of the inflow regions, in the Nordic and Barents Seas, dominant timescales of variability are short, with good agreement (low standard deviation and high mode count) between ensemble members. This is not surprising as the AW is in direct contact with the atmosphere here so will experience a large seasonal cycle. However, we need to look at the variation of the inflow itself to evaluate how much of a role AW inflow plays in the pan-Arctic patterns of temporal variability in AW temperature seen in Figure 3.27. Spectral plots of AW inflow temperature variability through the Fram Strait and St. Anna Trough over 1920–2000 and 2020–2100 are therefore given in Figure 3.28.

The Fram Strait subplots (first two panels) of Figure 3.28 show that the dominant mode of variability in the Fram Strait inflow temperature is seasonal (i.e. with a period of one year) as expected given the proximity of AW to the surface here. This is true of both the 20th and 21st century periods. The 21st century shows a slight shift towards longer timescales of variability if this seasonal variation is neglected, with ensemble average and modal peaks being 13.6 years and 16 years in the 21st century, compared with 12.5 and 13.3 years in the 20th century period. This is also noticeable in the general shape of the spectra, and is in line with the shift seen in the western Arctic in Figure 3.27. However, although some ensemble members in Figure 3.28 do show spectral peaks of a few decades (which was seen to be typical of AW in the Arctic in Figure 3.27) the timescales of ensemble average and ensemble mode inflow variability are shorter than that seen in the Arctic Ocean. Local processes within the Arctic therefore seem to act as a low-pass filter for AW temperature fluctuations, absorbing some of the higher frequency variability in inflow temperature (along with seasonal variability,

as shown by Lique et al. 2012).

The St. Anna Trough subplots (final two panels) of Figure 3.28 show a similar strong seasonal peak - although notably this does not appear in the ensemble mode or mean (apart from the mode of the 2020–2100 period). Unlike at the Fram Strait, timescales of variability here become shorter in the 21st century, with the ensemble-average dominant (second-most dominant) timescale shifting from 12.3 (13.8) to 2.9 (7.7) years, and modal timescale from 6.7 (40.0) to 1 (10) years. Similarly to the Fram Strait, however, these variability timescales are generally shorter than those seen in the Arctic Ocean itself.

This final point is the key takeaway from Figures 3.27 and 3.28. The shorter temporal variability in inflow temperature compared to the temporal variability in AW temperature elsewhere in the Arctic demonstrates that local processes - for example, heat loss to the halocline when there is short-term fluctuation AW temperature - must play a key role in the latter. These act to absorb the shorter term variability in inflow, allowing the multi-decadal variability seen in the inflow temperature to emerge as the dominant mode of variability across much of the Arctic.

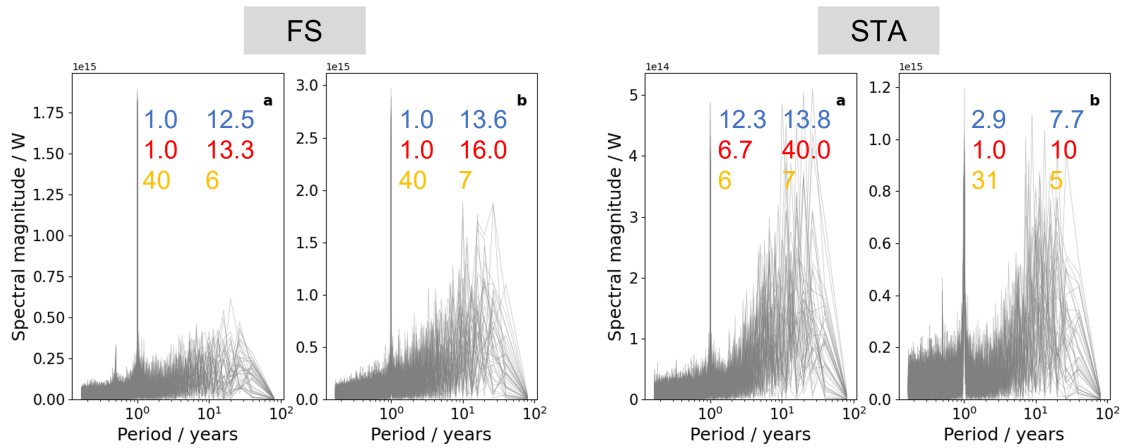


Figure 3.28: Spectral plots of AW inflow temperature variability through the Fram Strait (FS) and St. Anna Trough (STA) over (a) 1920–2000 and (b) 2020–2100 for every ensemble member. The first column of numbers in each plot gives: ensemble average dominant time scales of variability (i.e. highest peak of the spectrum) (blue), the ensemble mode of this i.e. the most common dominant timescale amongst the ensemble members (red) and the number of ensemble members that have this modal value as their spectral peak (orange). The second column of each plot gives the same quantities for the second-most dominant timescale of variability (i.e. the second-highest spectral peak) that is more than one year different from the first. This is due to the broad peaks in some spectra and because we are not concerned with sub-annual differences in variability timescales.

3.4 Conclusion

This chapter has focussed on AW layer properties in the CESM-LE simulations. Section 3.3.1 showed AW warming (linked to a warming inflow), along with the freshening and shoaling of AW heat. This occurred not only in the east but also in the western Arctic during the 21st century in the ensemble mean. In the 20th century, local changes in AW layer properties were often related to the advection of natural oscillations in AW properties around the Arctic. The local water column structure, e.g. depth of the halocline, played an important role in local AW layer depth and properties, particularly in the Canada Basin and eastern Eurasian Basin, where winds are known to affect halocline depth. However, anthropogenically forced trends became the dominant driver in AW layer changes of the 21st century. These results suggest that future forced changes in the climate system will drive a shoaling, warming, and freshening of the AW heat in the Arctic - not only in the east as has been observed already (e.g. Polyakov et al. (2017)), but in the western Arctic as well. It is important to bear in mind the caveat that Pacific Water is not present in the CESM-LE Canada Basin, however (Lavoie et al. 2022). This affects the properties and stratification of the Canada Basin and so is likely to influence the shoaling of AW heat in the model in this region. Revisiting this result in other coupled climate models with more accurate simulations of Pacific Water advection pathways would therefore be a useful next step in investigating the future evolution of AW heat in the western Arctic.

The role of AW inflow from the Atlantic in driving these ensemble mean (i.e. forced) changes was investigated in section 3.3.2. Results suggest that a 21st century shift to an atmospheric circulation resembling a more positive phase of the NAO/AO, with a weakened Beaufort High and a low over the Nordic Seas extending northeastward into the Eurasian Basin, is the likely driver of an increase in volume transport through the Fram Strait and a shift in the circulation of AW within the Arctic, with a pan-Arctic boundary current emerging. This enhanced volume transport, combined with an increase in temperature of inflowing water at all gateways, resulted in an increase in AW heat input into the Arctic.

In section 3.3.3, focusing on the influence of trends and variability in AW inflow temperatures on AW temperature in the rest of the Arctic revealed the importance of local processes (i.e. those downstream of the inflow) governing AW temperature within the Arctic. The circulation shift described above results in FSBW having an increased influence on AW

temperatures in the western Arctic. However, correlation between FSBW inflow temperature trends and those in the western Arctic are still only moderate at best. The influence of warming BSBW on AW warming within the Arctic diminishes to the eastern Eurasian Basin in the early 21st century. Finally, turning to the temporal variability of the detrended inflow, local processes again appear to be important. This time they act as buffers against high-frequency variability in AW inflow temperature from appearing in AW temperature variability elsewhere in the Arctic.

This sets the scene for the next chapter, where the focus will shift away from the AW layer and how changes in the AW inflow can influence those downstream, and towards changes in the upper water column, where local properties of the surface layer and halocline can play a key role in AW heat loss, salinity and depth. Investigating changes in stratification will be key to establishing whether the forced trends in AW properties seen in this chapter are likely to impact sea ice and the temperature of the surface layer, beyond what has already been observed in the Eurasian Basin (e.g. Polyakov et al. 2017). Results in this chapter have suggested that surface layer freshening is greater than AW layer freshening, and hence that water column stability will not be compromised. However, we have also seen suggestions that there may be a shift at intermediate depths from an alpha-ocean to a beta-ocean - i.e. that temperature may begin to play a more important role in stratification than salinity. If this is the case, then the forced changes in AW seen above could prove important not only for what outflows into the North Atlantic, but for the local Arctic itself.

Chapter 4

The impact of AW evolution on the stability of the upper water column

4.1 Introduction

The previous chapters have focussed on spatial and temporal changes in the AW layer and AW heat. Model data predict AW warming and freshening throughout the Arctic, predominantly over the 21st century (Chapter 3). A warming, freshening AW will have consequences for the climate system in terms of modified outflows into the North Atlantic and the resultant potential impact on meridional overturning (e.g. Karcher et al. (2011)). It will also, as already observed in the eastern Eurasian Basin (Polyakov et al. 2017), result in a warmer, more saline surface and hence reduced stratification. However, in much of the Arctic, the surface mixed layer may not be directly affected by this increase in AW layer heat content, particularly if the seasonal sea ice cycle and freshwater content continue to increase as observed in the western Arctic (Krishfield et al. 2014; Proshutinsky et al. 2019). The key question that this chapter aims to address, therefore, is: what do the long term past and future AW changes seen in previous chapters mean for the stratification of the upper water column in the Arctic?

Along with surface and AW properties, the relative roles of temperature and salinity in governing density are also key in determining water column stratification changes in the Arctic. As it is now, the Arctic is a beta ocean, meaning that it is stratified by salinity, as opposed to warmer subtropical oceans which are stratified mainly by temperature (alpha oceans) (Carmack 2007). This is the result of the non-linear increase of the thermal expansion

coefficient with temperature - as the water column warms, temperature plays an increasingly important role in density, and vice versa. These stratification types have important consequences for the physical and biogeochemical properties of oceans (Carmack 2007). Perhaps most important from the perspective of this thesis is the fact that sea ice can only form in beta oceans, given that the salinity stratification prevents deep convection (Bulgakov 1962). Furthermore, convection in an alpha ocean differs from a beta ocean in the sense that the temperature of water is more important than salinity in determining whether it gets ventilated. As the Arctic warms, this could mean an increase in AW ventilation, particularly in winter when surface temperatures are low.

Alpha-beta ocean stratification is important in the context of this thesis, given that in the previous chapter we saw signs that cross-sections of density began to resemble the temperature field more than the salinity field during the latter part of the 21st century at the Fram Strait in CESM-LE. In this chapter we will use the same model ensemble to determine whether the AW inflow regions may indeed shift to alpha stratification in the future, how and when it might happen, and whether this shift is present elsewhere in the Arctic given the northward progression of Atlantification into the Eurasian Basin.

The strongest stratification in the Arctic Ocean occurs across the halocline. As discussed previously in this thesis, the halocline acts as a key barrier between the AW heat and the cool surface and sea ice. Only a cold halocline can fully isolate the surface layer from AW heat, given that the upper thermocline/lower halocline is just the top of the AW layer. A seasonal cold halocline can emerge each summer due to the differences in density between the summer surface mixed layer and that from the previous winter (Steele et al. 1998), although this, by definition, is a transient feature and so is not a permanent barrier between the AW and the surface. However, a “true” cold halocline layer can be formed through local winter convection (Rudels et al. 1996) and advection of saline lower halocline waters (Aagaard et al. 1981) (both providing sources of lower halocline waters) combined with interleaving of freshwater further downstream along the halocline advection pathway. This gives a year-round cold halocline layer in regions such as the Makarov Basin, as proposed by Steele et al. (1998). In the Canada Basin, Pacific Water is the dominant water mass in the halocline (Timmermans et al. 2014; Timmermans et al. 2020). This Pacific Water is ventilated in the nearby Chukchi Sea before becoming entrained into the halocline beneath the Beaufort Gyre

(Timmermans et al. 2017). Before entrainment it is modified just north of the Chukchi Sea by mixing with halocline waters of Barents Sea origin (Woodgate et al. 2005). This demonstrates the widespread extent of Atlantic-derived halocline waters - both Fram Strait and Barents Sea origin halocline waters are advected throughout the Arctic Ocean (Rudels et al. 2004).

In order to answer the key question of this chapter and investigate what upper ocean stratification changes are predicted by the CESM-LE in the Arctic Ocean, this chapter will make use of both ensemble member and ensemble mean data to look into long term changes in stability, density, temperature and salinity profiles at various locations in the upper 500 *m* of the water column. The relative contribution of temperature and salinity to long term changes in the density structure of the water column will also be investigated. Finally, we will make use of a bulk metric to quantify overall upper water column stability, and use this to compare the roles of forced trends and natural variability in governing any stability changes. Throughout the analysis, both March and September data will be analysed independently. This is due to the difference in stratification between the freezing and melting seasons, and will allow us to identify any long term changes in the seasonal cycle.

4.2 Data and methods

The data used in this chapter are the same as those used in Chapter 3, i.e. from the CESM Large Ensemble, with data between 1920–2005 corresponding to historical forcing, and those spanning 2006–2099 using RCP 8.5 forcing. We will investigate changes in the ensemble mean and the spread across ensemble members. In this chapter, as the focus is on the impact of AW changes on the upper water column (where stability changes could influence the flux of AW heat to the surface and sea ice), we focus on the upper 500 *m* of the water column.

This chapter will utilise two different methods for quantifying the stability of the water column. The first of these is analysing vertical profiles of the square of the Brunt-Väisälä frequency (N^2). The Brunt-Väisälä frequency is defined as:

$$N = \sqrt{-\frac{g}{\rho} \frac{\partial \rho}{\partial z}} \quad (4.1)$$

where g is gravitational acceleration, ρ is potential density, and z is depth. N^2 is a measure of the local stability of the density stratification, and represents the angular frequency of

oscillation of a vertically displaced water parcel. A positive N^2 indicates stable stratification, with a higher value indicative of stronger local stability.

Although profiles of N^2 give an idea of the local vertical stability of the water column, quantifying the overall water column stability from these profiles can be difficult. In a recent paper, Polyakov et al. (2018) proposed that available potential energy (APE) is a good measure of bulk halocline stability in the Arctic, and so we also calculate this. In our case, we define APE as:

$$APE = \int_{z_2}^{z_1} g(\rho - \rho_{ref})z dz \quad (4.2)$$

where z_2 is the surface, z_1 is 483 m, g is gravitational acceleration, ρ is potential density, ρ_{ref} is the potential density at 500 m, and z is depth. The choice of reference depth is related to the model grid depths. As we only focus on the upper 500 m of the water column in this chapter, we take the reference density as the depth point closest to 500 m, which in this case is 483 m.

APE, as it is defined above, will be negative in a stable water column, with a more negative value implying greater stability in line with the standard physical definition of the potential energy of a system. Similarly a positive APE would be indicative of an unstable water column.

We chose to use a fixed base depth of 483 m in our computation of APE, rather than using a set of criteria to identify the cold halocline base for the lower integral limit, as in Polyakov et al. (2018). This was due to the broad temporal and spatial scope of our data, over which the properties of the halocline change significantly. However, there is certainly scope to use the curvature of the temperature profile in future work to more accurately quantify the stability of the halocline and surface layer if using a less varied selection of profiles. Using a fixed-depth base was sufficient for this analysis, given that 483 m is within the AW layer (when defined as the layer between the two 0 °C crossing points either side of the AW temperature maximum) at all times and locations, as seen in Chapter 3; and that the density gradient within the AW layer itself is small compared to that in the halocline (as shown later in e.g. Figure 4.3).

Assessing the relative contributions of temperature and salinity to density throughout the upper water column is another key part of this chapter. In order to do this, we make the

assumption of a linear equation of state for seawater:

$$\rho = \rho_s + \rho_s(-\alpha(\theta - \theta_s) + \beta(S - S_s)) \quad (4.3)$$

where α and β are the thermal expansion and haline contraction coefficients, respectively, both computed using the Thermodynamic Equation of Seawater 2010 (TEOS-10) (IOC et al. 2010); ρ is potential density, ρ_s is the potential density at the surface, θ is potential temperature, θ_s is potential temperature at the surface, S is salinity, and S_s is salinity at the surface. Please note that “surface” here refers to the uppermost oceanic depth level of the model data, which in the case of the CESM-LE is 5 *m*.

We define the temperature and salinity contributions to vertical variations in density using the temperature and salinity terms in equation 4.3, i.e. we take the temperature contribution as $\rho_s(\alpha(\theta - \theta_s))$ and the salinity contribution as $\rho_s(\beta(S - S_s))$.

4.3 Results and discussion

4.3.1 Water column stability and its drivers

To get an overall idea of the properties and stability of the Arctic Ocean upper water column, and how these change over time, Figure 4.1 shows ensemble mean data for March (end of freezing season) averaged over set periods along the pan-Arctic section shown in Figure 3.2. Similar sections for September are very similar so are not shown here.

Many of the AW changes discussed in the previous chapter are visible in Figure 4.1. The AW warms, freshens and shoals during the 21st century periods. AW heat shoals into fresher, less dense water, and becomes less dense at greater depths as a result of this warming and freshening.

The surface also becomes less dense during the 21st century in both March and September. This appears to be related to surface freshening. Notably, the depth of the isohalines in Figure 4.1 does not change significantly over time. Figure 4.2 shows that the magnitude of the stratification maximum of the water column increases over time in both March and September, as this surface freshening is greater than that of the AW. The AW (the top of which is denoted in these figures with a red line) remains below this stratification peak in all time periods, despite shoaling. September surface warming occurs during time periods b-d

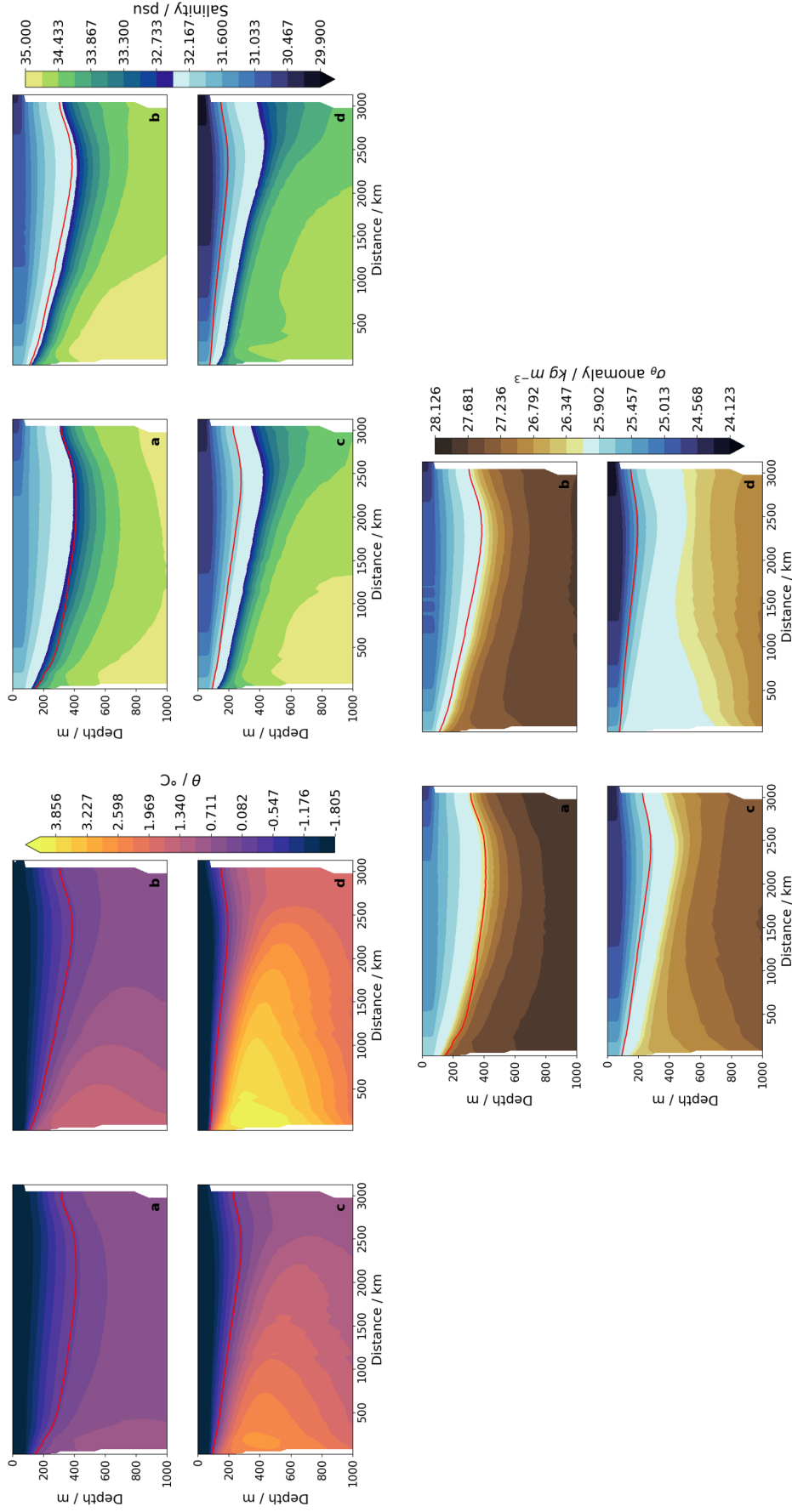


Figure 4.1: March ensemble mean pan-Arctic sections of potential temperature (upper-left), salinity (upper-right), and potential density (bottom). Each subplot shows the ensemble mean data averaged over the following periods: (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, (d) 2070–2099. The red contour denotes the AW top i.e. the $0\text{ }^{\circ}\text{C}$ isotherm. The section runs from east to west so that the viewer is looking towards Greenland/Canada. See Figure 3.2 for section location.

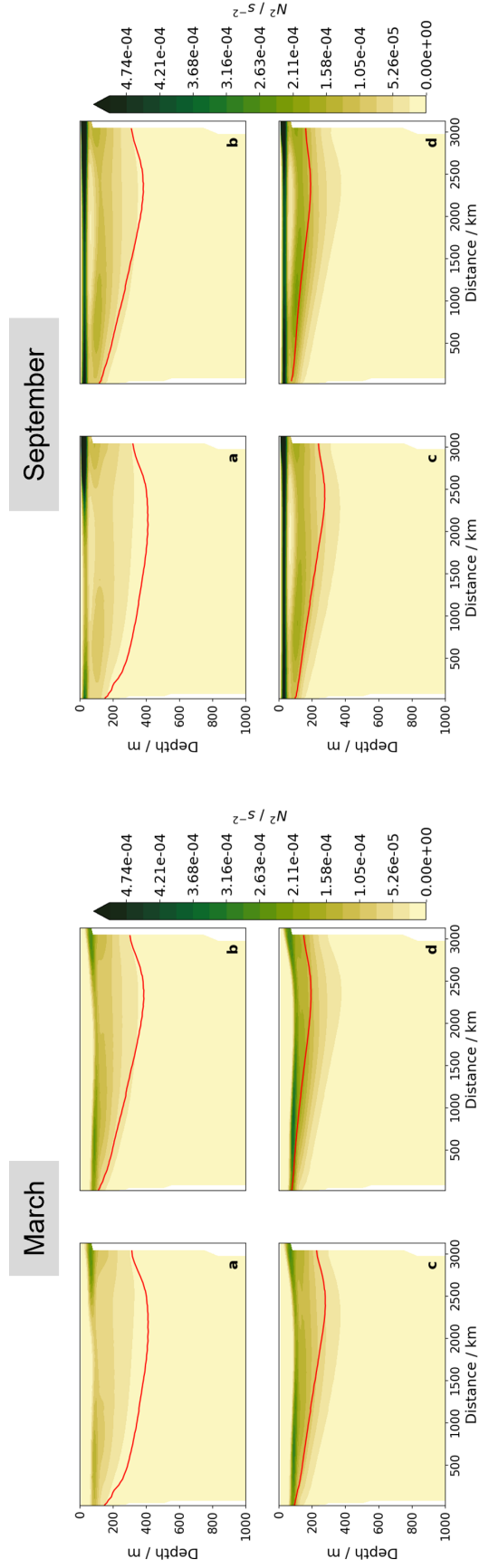


Figure 4.2: March and September ensemble mean pan-Arctic sections of the square of the Brunt-Väisälä frequency (N^2) over the following periods: (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, (d) 2070–2099. The red contour denotes the AW top i.e. the 0 °C isotherm. The section runs from east to west so that the viewer is looking towards Greenland/Canada. See Figure 3.2 for section location.

(2010-2099), but is not related to AW warming, as the warming AW and warming surface remain separated by a cold halocline (which forms a local maximum in stratification). It is instead likely the result of increased solar heating due to reductions in summer sea ice coverage (see Figure 3.7).

Notably, the density structure of the water column in Figure 4.1 starts to resemble the temperature field in the periods c-d (2040–2099), similarly to what was found across the Fram Strait in Chapter 3 (see Figure 3.13). The only notable differences between the March (Figure 4.1) and September ensemble means is the warmer surface layer in September in later periods, as mentioned above, and the second stratification peak beneath the surface mixed layer in September in Figure 4.2. However, data from both months suggest that throughout the year, AW heat has little impact on the surface ocean and that water column stability increases over time - even as it warms and freshens during the 21st century.

Figures 4.3–4.8 give a more detailed picture of the properties shown above, including the variation within the ensemble, and give insight into what drives the density peaks seen in Figure 4.1.

Figure 4.3 shows the March and September N^2 (i.e. local stability) and density profiles from a location away from the AW inflow in the eastern Eurasian Basin. This location is representative of the properties of the interior of the basin, and is a region currently undergoing significant hydrographic change (Polyakov et al. 2017). The magnitude of the largest stratification peak in both March and September increases outside of its range of natural variability (represented by the differences between ensemble member profiles) when comparing the 1920–2009 and 2070–2099 periods, with a slight decrease in September between 2040–2069 and 2070–2099. September profiles show a second local stratification maxima below this largest peak. This too increases over time, although remains within the range of natural variability demonstrated by its varying magnitude amongst ensemble members.

The density profiles of Figure 4.3 (expressed here and in similar figures using the potential density anomaly) can reveal more about what drives the changes in the local water column stability represented by N^2 , given that N^2 peaks are located at the maxima in vertical density gradient. The whole water column at this eastern Eurasian Basin location becomes less dense over time, with this density decrease laying outside of the range of variability. However, the most significant changes in density occur higher up in the water column i.e. in the surface

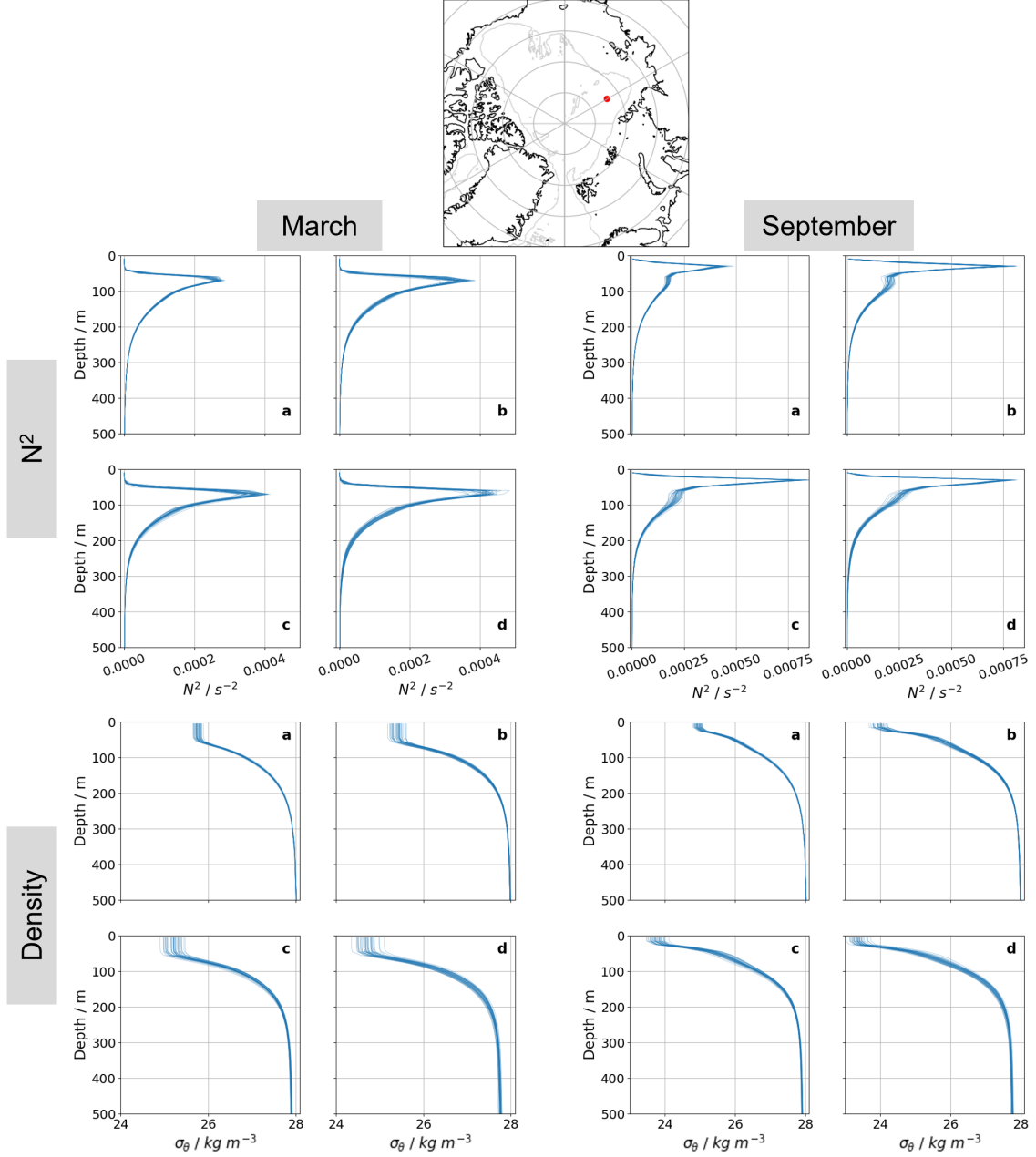


Figure 4.3: Eastern Eurasian Basin profiles of Brunt-Väisälä frequency squared (N^2) (top row) and potential density (bottom row) from each ensemble member averaged over the following periods: (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, (d) 2070–2099. Only data from March (left column) or September (right column) are included. Profile location is denoted with a red dot on the map at the top of the figure.

mixed layer, where density decreases by over 1 kg m^{-3} in March and over 1.5 kg m^{-3} in September over the time periods. Hence the reduction in density of the surface in relation to the AW (which sees a density decrease of approximately 0.3 kg m^{-3}) is the driver of the increase in the N^2 maximum. As mentioned, this is the case in both March and September - the surface freshening trend is not limited to one season.

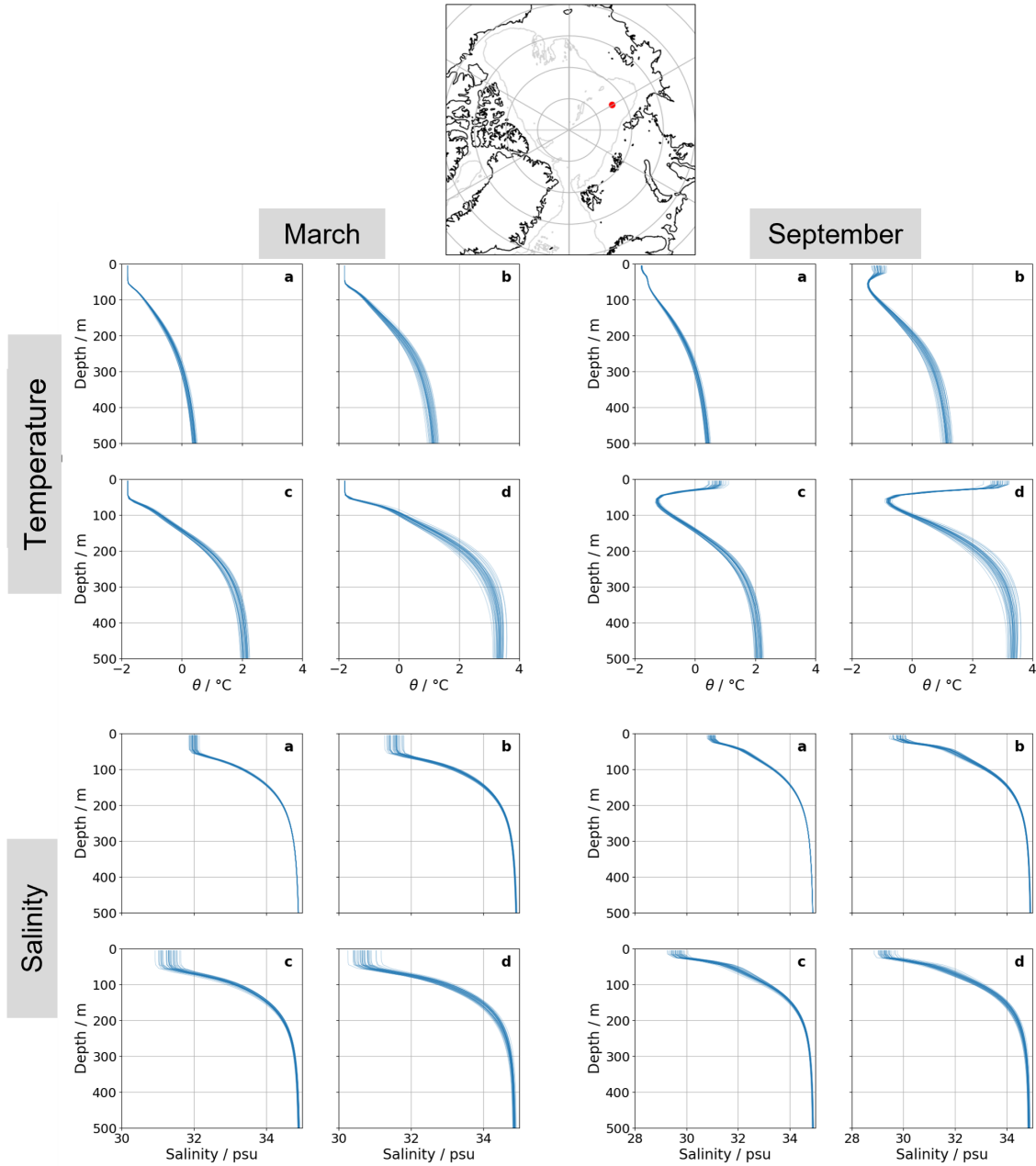


Figure 4.4: As for Figure 4.3 but for temperature (top row) and salinity (bottom row).

Figure 4.4 shows the temperature and salinity profiles from all ensemble members at the same location as above in the eastern Eurasian Basin. These profiles will add context to the density profiles, and hence local stratification profiles, of Figure 4.3.

Focusing first on the temperature profiles of Figure 4.4, we see that the March surface mixed layer does not warm over time, despite significant AW warming ($3\text{ }^{\circ}\text{C}$ over all time periods at 500 m) beyond the envelope of ensemble variability. This results in a stronger gradient across the thermocline between the surface mixed layer and AW temperature max-

imum. September sees a similar degree AW warming; however, this is less than the surface layer warming (which is $> 4^{\circ}$) during this month. This results in a September temperature minimum between the surface mixed layer and AW, which itself warms, although by around 1°C , so less than the surface and AW warming. This temperature minimum is the cold halocline, although given that this disappears by March it is a seasonal cold halocline. This, as mentioned above, does not provide a year-round barrier between AW heat (the top of the thermocline) and the surface mixed layer. The seasonal differences in temperature profile evolution are due to the presence of sea ice. In March, sea ice is present (see Figure 3.7) and so the surface is at freezing temperature. Sea ice is not present in Figure 3.7 during the later September periods however, allowing the atmosphere to warm the surface mixed layer.

The salinity profiles of Figure 4.4 show surface freshening beyond the range of ensemble variability in both months. Surface waters freshen by just over 1 psu between the 1920–2009 average and the 2070–2099 average in March and September, whilst AW shows negligible freshening at 500 m.

Tying these temperature, salinity, density and N^2 profiles from Figures 4.3 and 4.4 together, we see that the N^2 peak at this location lies just below the surface mixed layer in both March and September. The second largest N^2 peak in September lies just beneath the temperature minimum (i.e. just beneath the seasonal cold halocline). The increase in the magnitude of the largest N^2 peak in both March and September is the result of the reduction in surface layer density being larger than the reduction in density of the water column beneath it. This itself is due to enhanced surface freshening in March and a combination of surface freshening and warming in September relative to the halocline below. The relative importance of temperature and salinity in governing density (and hence local stability) will be investigated later in this chapter. This will also allow us to look more closely at the reduction in the largest September N^2 peak magnitude in the 2070–2099 period of Figure 4.3 - the cause of which (beyond halocline and AW freshening relative to the surface) is difficult to decipher from these plots.

Figures 4.5 and 4.6 show similar profiles for a location in the western Eurasian Basin - a region under the direct influence of AW inflow - and generally paint similar picture to that of the eastern Eurasian Basin. The main N^2 peak lies just beneath the surface mixed layer in both months, although, unlike in the eastern Eurasian Basin, this peak shoals over time in

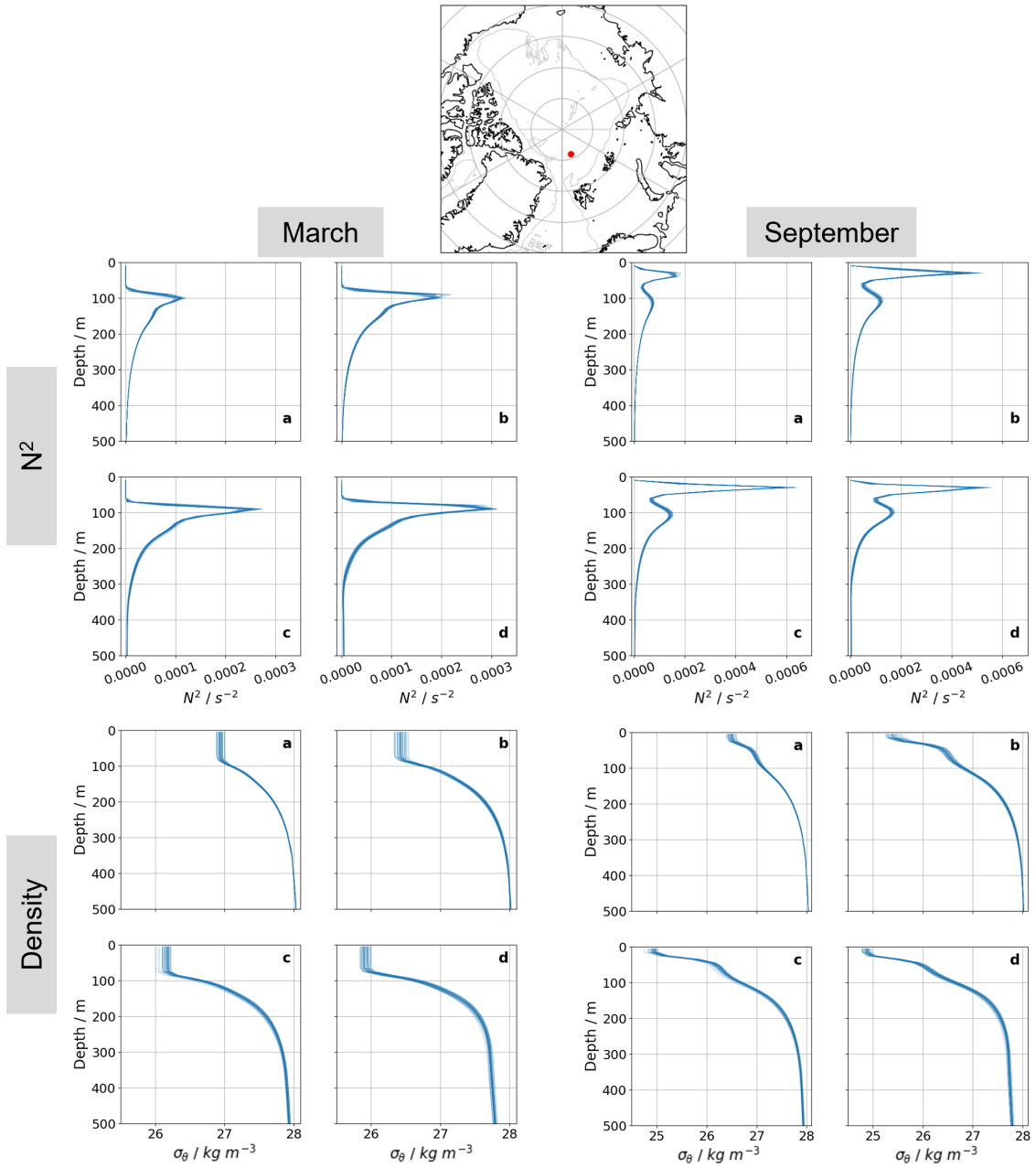


Figure 4.5: As in Figure 4.3 but for a location in the western Eurasian Basin.

March as the surface mixed layer thins. The salinity profiles in Figure 4.6 suggest that this is related to a reduction in freshwater at the surface, perhaps due to sea ice loss in the region. Based on the AW shoaling and constraint beneath the water column stability maximum seen in Figure 4.1, this suggests that a reduction in Arctic freshwater content could be enabling shoaling.

The overall cause of these local stability increases is the reduction in surface density in relation to the density of the halocline and AW below it. Again, the extent to which warming

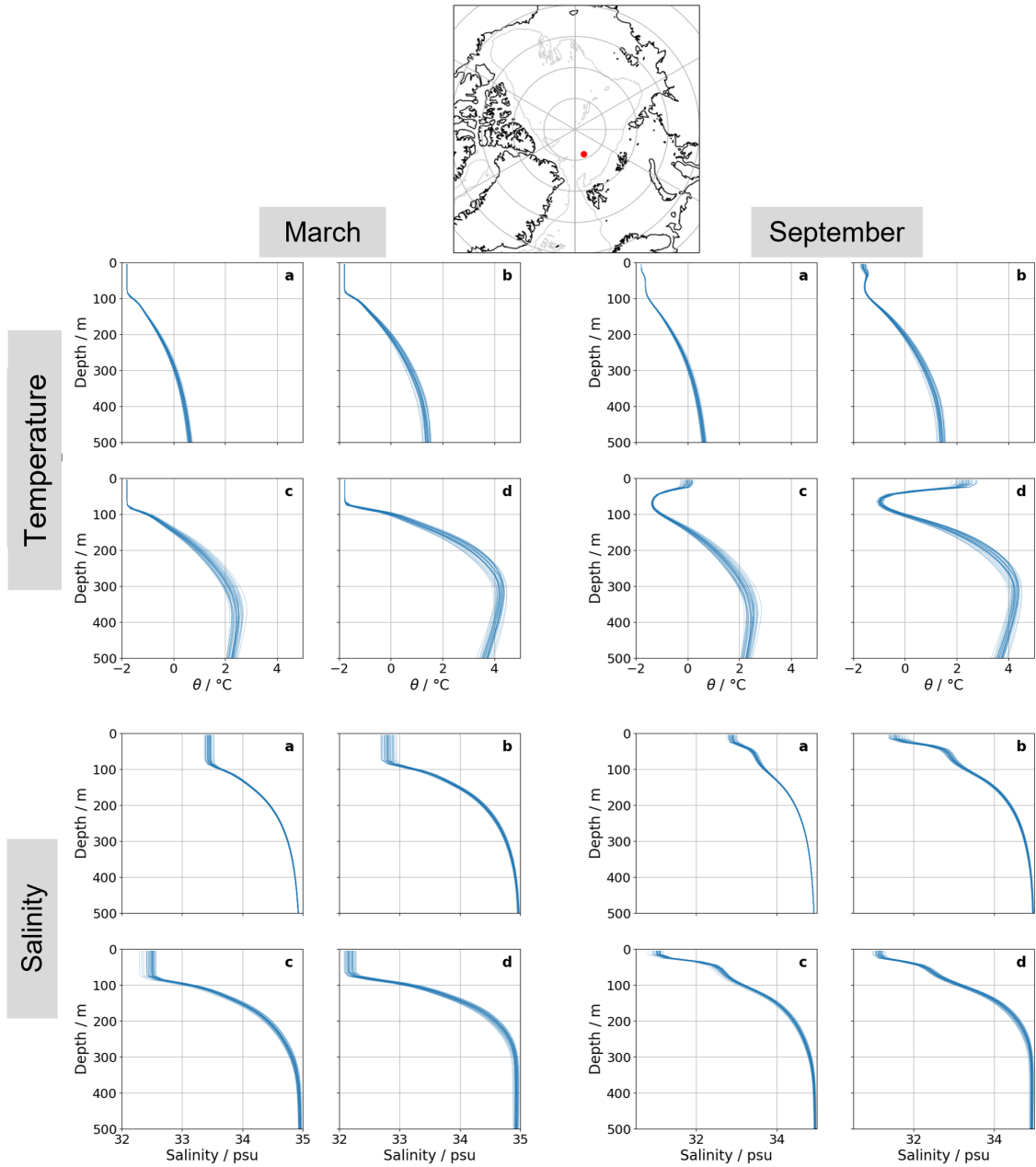


Figure 4.6: As in Figure 4.4 but for a location in the western Eurasian Basin.

and freshening govern these changes will be investigated later.

As in the eastern Eurasian Basin, the magnitude of this main stratification peak reduces at the end of the 21st century. It can be seen from the density, temperature and salinity profiles of Figures 4.5 and 4.6 that this is due to the density of the (again, seasonal) cold halocline reducing compared to the surface during the latter time period. This is also the likely cause of the reduction in the main stratification peak in the eastern Eurasian Basin, although it is more easily seen here in the western Eurasian Basin. The cause of the reduction

in cold halocline density compared to the surface appears to be related to a freshening of this layer in relation to the surface, which itself is likely caused by the fresher winter surface mixed layer (which forms the seasonal cold halocline during melt season, e.g. Steele et al. (1998)). So the surface freshening, which initially drives an increase in local stability beneath the surface mixed layer, also contributes to a later reduction in this local stratification maximum in summer as the halocline begins to freshen.

A second local stratification peak is present in September at the bottom of the cold halocline - the stability here increases with time due to the cold halocline freshening. So these profiles could imply that the overall stability between the AW and surface during summer still increases, as can also be implied by the increase in density difference between the surface and, say, 200 *m*.

Figures 4.7 and 4.8 show the same profiles for a location in the Canada Basin. Compared to the Eurasian Basin, these profiles show stronger surface warming in September (around 6 °C) but weaker AW warming (around 1 °C in both months). This larger surface warming could be explained by the longer retention time of surface waters in the Beaufort Gyre region which gives more time for absorption of solar heat. Surface salinity shows a smaller decrease than in the Eurasian Basin. The March peak in N^2 increases during the 21st century, but this increase lies within the range of natural variability represented by the range of ensemble member profiles. This peak is in the halocline, and is again the result of a reduction in surface mixed layer density relative to that of the halocline and AW. A deepening of the surface mixed layer causes this maximum in water column stratification to deepen over time, in contrast to the western Eurasian Basin profiles. A second stratification peak emerges in March just below this main peak during the 21st century.

The September N^2 peak increases up to the 2010–2039 period before decreasing so that most ensemble members have an N^2 peak in 2070–2099 that is smaller in magnitude than the in 1920–2009. Similarly to what was seen in the western Eurasian Basin (Figures 4.5 and 4.6) this is due to the cold halocline freshening, reducing the density gradient between itself and the surface, and also strengthening the density gradient between itself and the AW layer resulting in a slight increase in the magnitude of the second-highest N^2 peak. This freshening of the halocline layer is likely due to a freshening of the halocline rather than a strengthening of the Beaufort Gyre, given the relative stability in isohaline depths in the region (Figure

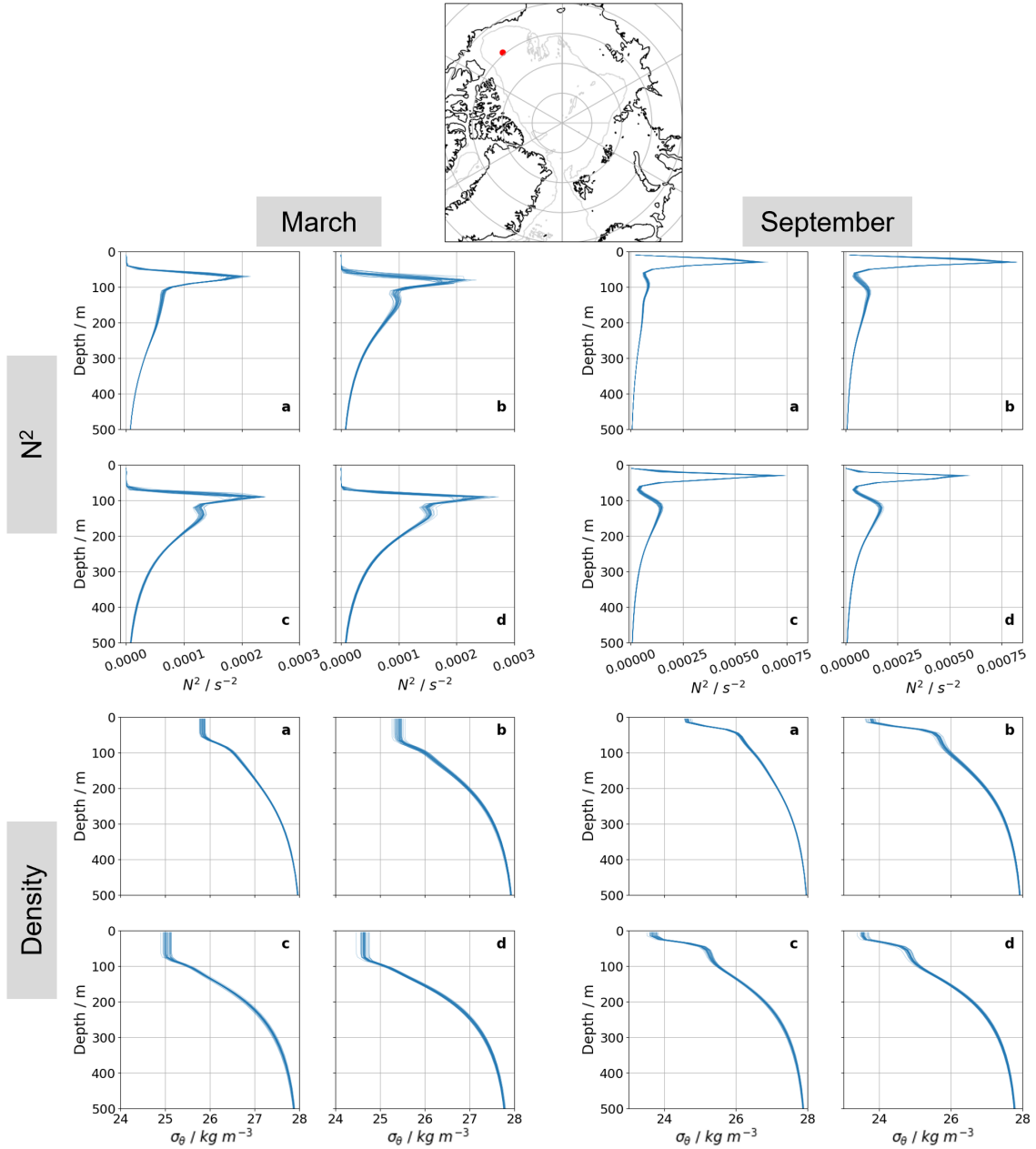


Figure 4.7: As in Figure 4.3 but for a location in the Canada Basin.

3.6).

The September freshening can also be seen in the ensemble mean TS diagrams in Figure 4.9. The cold halocline base, which is often defined using the “elbow” of a TS diagram (e.g. Kikuchi et al. (2004)), freshens in this figure. As mentioned earlier in this thesis, there is an absence of Pacific Water signature in the western Arctic halocline in CESM-LE up until at least the 2010s (Lavoie et al. 2022). Figures 4.9c and d suggest that this lack of Pacific Water continues throughout the remaining 21st century in CESM-LE, with no distinct Pacific Water signature in these subplots in March nor September. This highlights an important caveat for

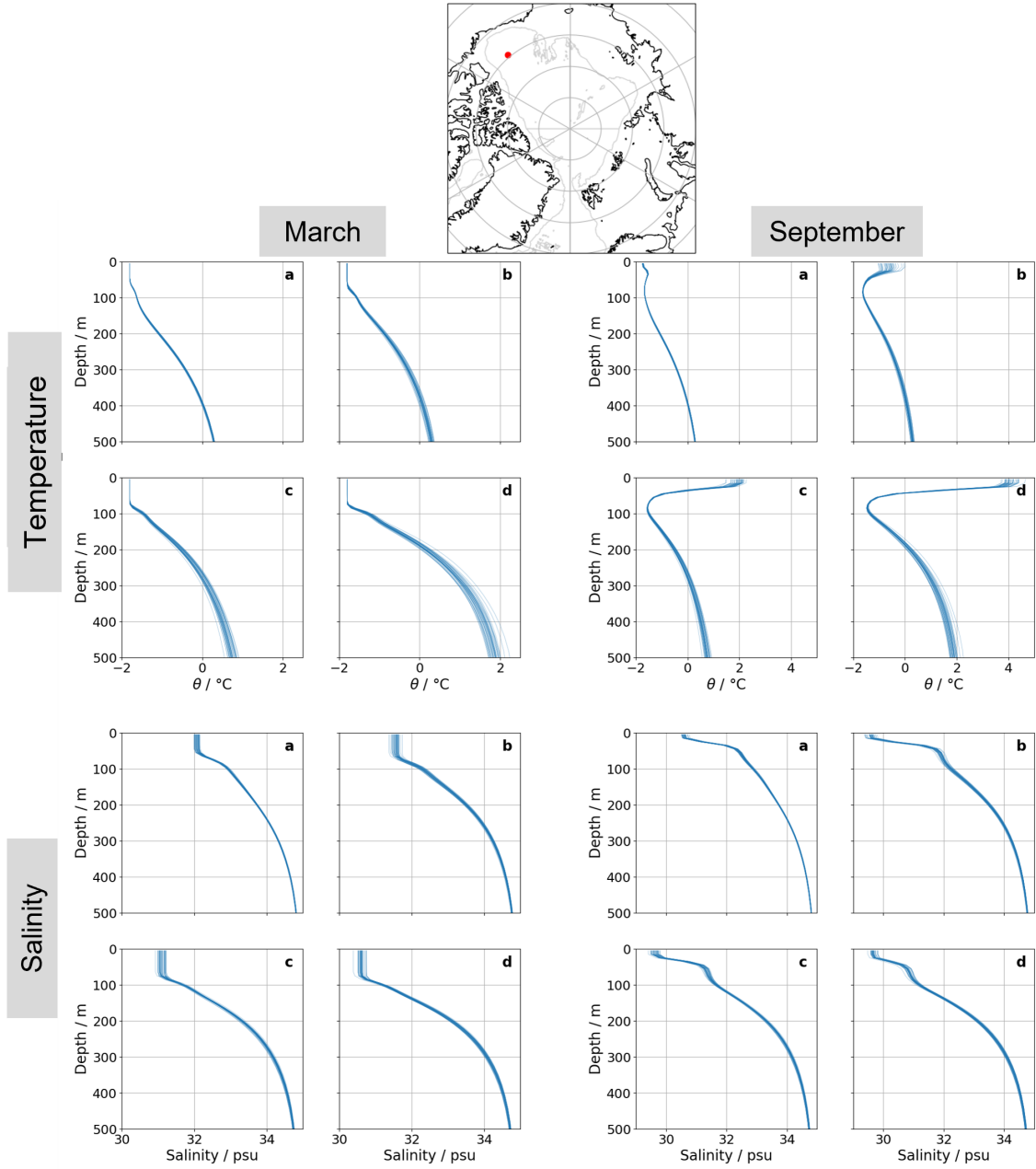


Figure 4.8: As in Figure 4.4 but for a location in the Canada Basin.

our results. The absence of Pacific Water signature in the model profiles could result in a cooler and potentially thinner Canada Basin halocline than exists in real ocean. This may result in a bias in later bulk water column stability metrics in the Canada Basin.

Looking at the key properties and stability of ensemble member profiles at various locations in the Arctic has revealed that, in general, stratification between the surface and AW increases over time. This owes to an increase in the density gradients in the upper water column as the reductions in surface layer density are greater than reductions in density else-

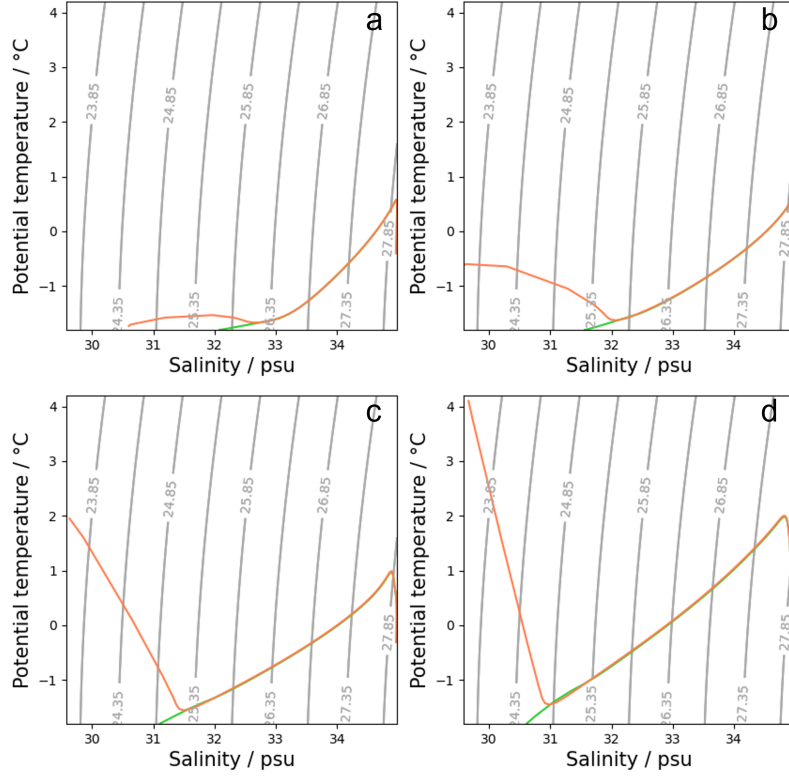


Figure 4.9: TS diagrams of ensemble mean data from March (green) and September (orange) at the same Canada Basin location as Figures 4.7 and 4.8, averaged over (a) 1920–2009, (b) 2010–2039, (c) 2040–2069, (d) 2070–2099.

where. This appears to be mainly related to the relatively high degree of surface and/or halocline freshening compared to freshening elsewhere in the water column.

The above stratification profiles (Figures 4.3, 4.5 and 4.7) have shown how a decrease in one local stratification maximum can directly lead to a change in a second local stratification maximum elsewhere in the halocline. This poses an interesting question when trying to assess changes in stability between the surface and AW. A decrease in one stratification peak between these two water masses may not necessarily result in an overall reduction in stratification between them if another peak emerges concurrently. This sets the scene for the final section of this chapter which will use a bulk measure for water column stability.

First, however, the relative roles played by temperature and salinity in water column stratification will be investigated. Whether changes in temperature or salinity are the main cause of the changes in density profiles seen above can not be fully determined from temper-

ature and salinity profiles alone, although based on the shape of those seen above it appears - unsurprisingly - that reductions in salinity play a key role. The next section aims to determine whether this is indeed the case in the future Arctic given the apparent degree of warming, and to what degree the influence of these two properties on density may change in the future.

4.3.2 The role of temperature and salinity in governing stratification

To investigate the roles that temperature and salinity play in the density changes seen above, ensemble mean profiles averaged over the same periods and locations as above will be plotted along with the relative contribution of temperature and salinity to this density profile, as explained in the Data and Methods section of this chapter. These contributions can be thought of as the density profile without the contribution of the other property. Alternatively, they can be thought of more literally as the contribution of each property to water column density (relative to the surface). The roles played by temperature and salinity in water column stratification will be shown more clearly by investigating the contributions of these two properties to how density changes with depth relative to the surface.

The Arctic Ocean is a beta ocean due to its low temperatures, and salinity has been/is the dominant factor governing ocean stratification at past and present times. Given that the Arctic Ocean is warming, this analysis will investigate if/when any part of Arctic may shift toward an alpha regime, with stratification mainly governed by temperature. This shift would be important not just for the Arctic, where an ocean stratified by temperature would result in surface AW, different convection regimes, and no sea ice formation (sea ice can only form in beta oceans (Bulgakov 1962)), but would also impact the global thermohaline circulation and biogeochemical fluxes (Carmack 2007).

Figure 4.10 shows how density profiles (along with the temperature and salinity contributions) in the eastern Eurasian Basin change relative to time period-a (1920–1949). We plot the density anomaly relative to the surface in order to assess water column stratification and its contributors. Given that the linear equation of state (we use Equation 4.3) is only an approximation of density, and that the density plotted is the potential density from the model data rather than from the equation of state in Equation 4.3, the contributions of temperature and salinity to density changes with depth may not always be exact. However, it will suffice

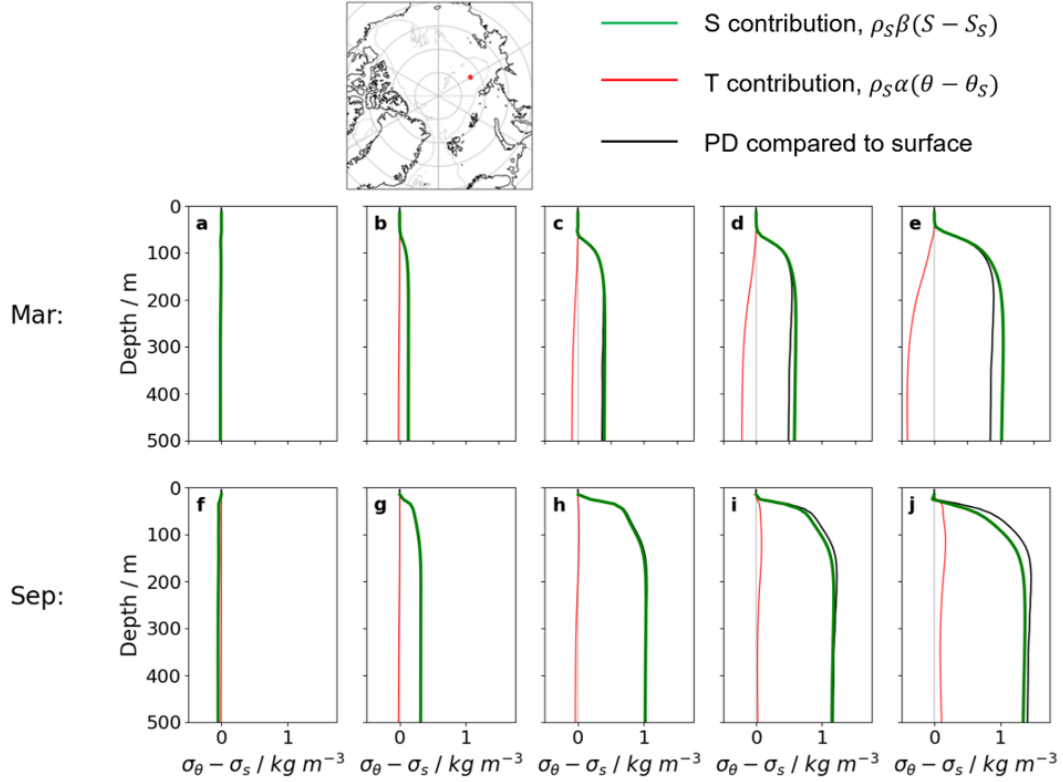


Figure 4.10: Ensemble mean density difference from the surface density, and the temperature (red) and salinity (green) contributions to this density difference - all relative to the 1920–1949 average. Contribution lines in bold denote the main contributor to the density at that depth in the profile. Ensemble mean data has been taken from either March or September, and has been averaged over the following 30-year periods: (a) and (f) 1950–1979, (b) and (g) 1980–2009, (c) and (h) 2010–2039, (d) and (i) 2040–2069, (e) and (j) 2070–2099. Profiles are from a location in the eastern Eurasian Basin, denoted with a red dot on the map at the top of the figure.

for this analysis as the aim is to get an idea of the relative importance of each property.

Figure 4.10 shows that salinity is the dominant factor governing stratification in the eastern Eurasian Basin throughout the year over the whole time period covered by the model simulations. This can be seen not only through the boldness of the salinity contribution line, but also through the close resemblance between the density and salinity-contribution profile shapes. The salinity profile drives an increase in water column stratification. The contribution of temperature to density becomes more significant over time. In fact, in March it is a key driver of the reduction in density of the lower water column, as seen by how much less dense the 2070–2099 March density profile is compared to the salinity contribution alone (Figure 4.10e).

The September profiles of Figure 4.10 show that there is a seasonal difference in the

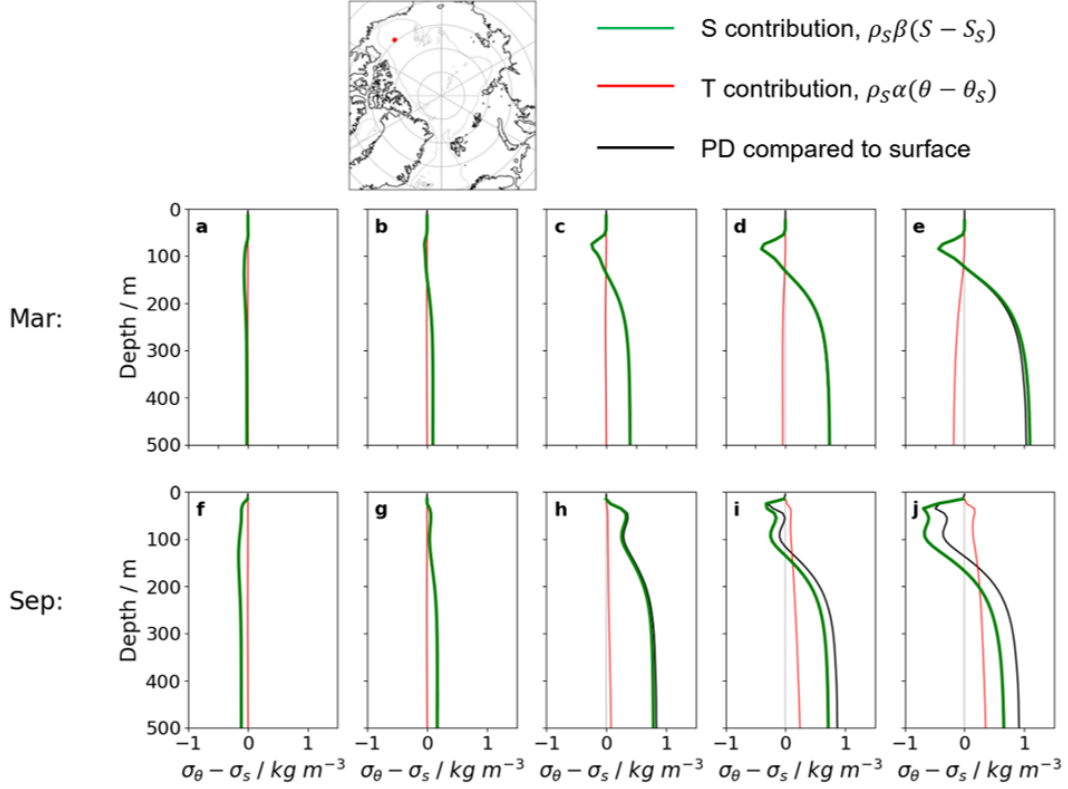


Figure 4.11: As for Figure 4.10 but for a location in the Canada Basin.

evolution of the temperature contribution. In contrast to what was just described in March (with warming AW contributing to a decrease in water column stability, i.e. a negative density gradient, below the halocline), September sees the surface warming relative to AW, resulting in the temperature changes enhancing water column stability. However, the dominance of salinity at this location at all times means that the freshening of the surface (seen earlier in Figure 4.4) relative to the AW is what drives an increase in density gradient and hence water column stability between the surface and AW, in what remains a beta-stratified water column.

Very similar results to above are obtained at the western Eurasian Basin location (not shown).

Figure 4.11 shows the same analysis for the Canada Basin location. Again we see the importance of salinity in governing density. At all times and depths, salinity is the most important contributor to density. The salinity contribution to stratification becomes less stable (i.e. negative) over time in the top 100 m as the surface mixed layer, and hence halocline, deepens (March) and halocline freshens relative to the surface (September) - as seen earlier in Figure 4.8. However, the density difference between the surface and AW

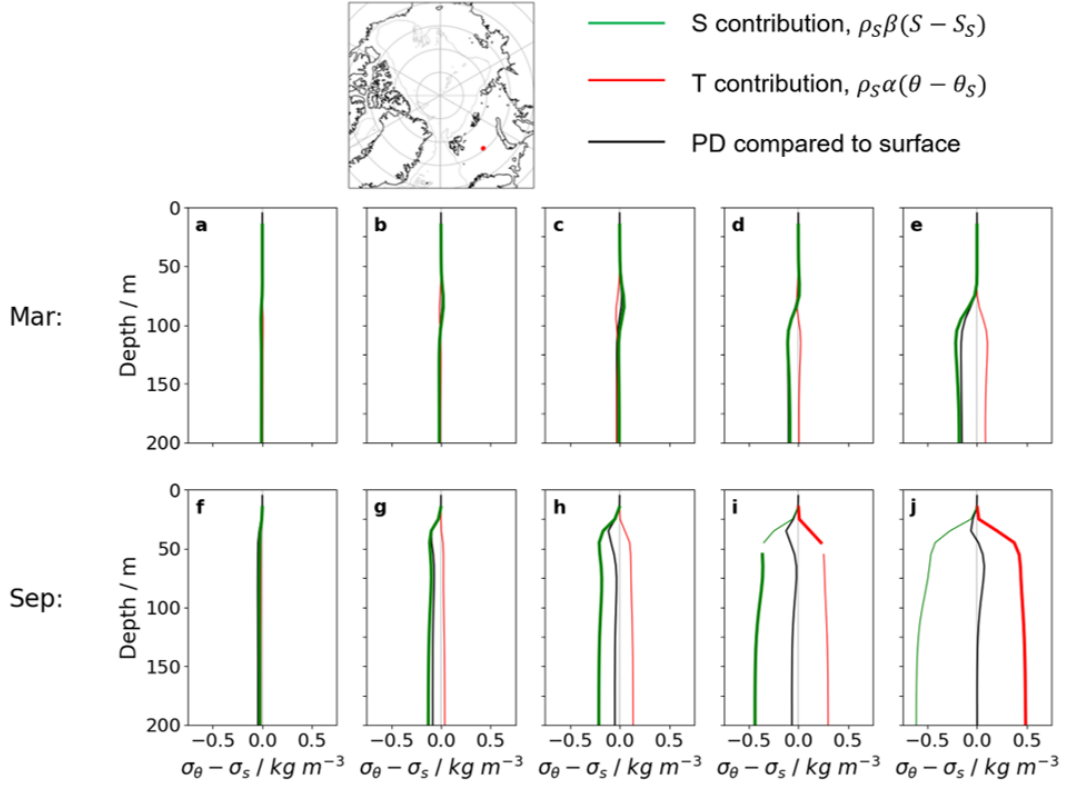


Figure 4.12: As for Figure 4.10 but for a location in the Barents Sea.

increases with time, as seen at other locations, and this is predominantly due to a freshening of the surface compared to the AW. Unlike in the Eurasian Basin, where the enhanced role of temperature in stratification is most apparent in March, in the Canada Basin this temperature contribution change is most notable in September. This is due to the relatively large degree of surface warming (seen earlier in Figure 4.8) compared to AW warming. Again, we attribute this to the longer retention time of surface waters in the Beaufort Gyre region. Compared to the surface, the AW (and halocline) warms less, and so the temporal anomalies of temperature contribution to density (i.e. red profiles in Figure 4.11) have a positive gradient with depth. The two key takeaways from the Canada Basin profiles are therefore that the stratification between the surface and AW appears to strengthen over time, but that there may be a decrease in stratification strength between the surface and the summer halocline as the latter water mass freshens. This could have important consequences for the role of the halocline waters in local sea ice melt.

Finally, the same analysis is computed for a new location in the Barents Sea (Figure 4.12). We have seen above that although temperature begins to play more of an important role as the water column warms, density and stratification are still dominated by salinity both in the

eastern and western Arctic. Here in the Barents Sea, however, there is a shift during 2040–2099 to a water column where density is predominantly influenced by temperature during September. In Figure 4.12i and j this is shown by part, or all, of the temperature contribution being bold where it is larger in magnitude than the salinity contribution. The ensemble mean temperature and salinity profiles in Figure 4.13 show that this shift first occurs in the warmest part of the water column in the 2040–2069 average, before the temperature dominates the stratification of the whole water column during the 2070–2099 average.

Despite this shift in stratification regime, compensating changes in temperature and salinity in the Barents Sea September mean that the density profiles here do not change dramatically over time. In fact there is little overall change in the density gradient in this month, as seen in the lower right quadrant of this figure. This finding is similar to that of Barton et al. (2020), who used observations to infer that the impact of recent increases in Barents Sea heat content on steric height have been offset by salinity changes. Our results imply that this may continue throughout this century.

However, unlike elsewhere in the Arctic, the stability of the water column in March reduces in the Barents Sea (based on the negative parts of the density profile seen in the March profiles of Figure 4.12). Here in March, salinity still drives density changes, and this

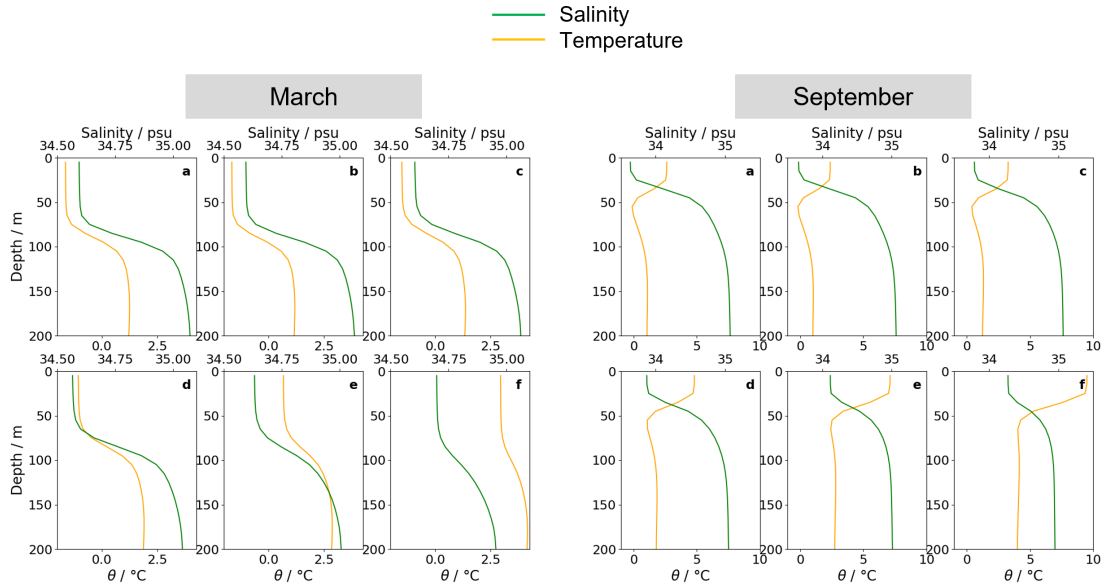


Figure 4.13: March and September ensemble mean profiles of potential temperature and salinity for the same location in the Barents Sea as in Figure 4.12. Data have been averaged over the following 30-year periods: (a) 1920–1949, (b) 1950–1979, (c) 1980–2009, (d) 2010–2039, (e) 2040–2069, (f) 2070–2099.

reduction in winter water column stability is due to the freshening of the deepest Barents Sea waters and salinification of the surface compared to 1920–1949, as seen in Figure 4.13. This can be explained by the reduction in winter Barents Sea sea ice coverage (e.g. (Onarheim et al. 2017) and Figure 3.7), with less brine rejection from sea ice formation to salinify the deeper Barents Sea, and less sea ice present to melt upon interaction with the AW. This reduction in March water column stability - along with the warming of the surface layer shown in Figure 4.13 - implies that there is an increase in the heat transferred to the Barents Sea surface (and potentially the atmosphere, although this is affected by the efficiency of Barents Sea heat loss (Skagseth et al. 2020)) during the 21st century in winter.

The same is not the case in the summer. The high surface temperatures seen in Figure 4.13 are most likely related to increased local summer atmospheric temperatures given their magnitude compared to the temperature of the AW at depth. The shift to a temperature-dominated summer stratification (seen earlier in Figure 4.12) would mean that this solar heating inhibits the transfer of AW heat towards the surface, despite the increase in surface salinity, as the deeper AW is cooler than these warm surface waters. This has limited implications for local atmospheric temperatures, given that they are likely warmer than AW anyway, but could result in warmer Barents Sea Branch AW downstream in the Arctic than there would have been had the stratification remained salinity-governed with the surface becoming less fresh.

Analysis in this section has shown that, although temperature does play more of a role in density as the Arctic Ocean warms, salinity is still the key driver of vertical density gradients. The main driver of the increase in density gradient between the surface and AW is the freshening of the surface relative to the AW layer over time. However, the same can not be said of the Barents Sea, where warming results in temperature becoming the dominant factor in governing density stratification during summer towards the end of the 21st century - i.e. the ocean stratification changes from beta to alpha. Whether this shift will continue north as temperatures continue to increase in the Arctic in the future (see Chapter 3) remains an open question, but these results suggest that it is indeed a possibility - particularly if the Arctic becomes ice-free during summer.

4.3.3 Quantifying water column stability and its trends

We now move away from investigating water column structure and towards using a bulk metric to quantify the stability of the water column between the AW and the surface. As seen earlier in this chapter when looking at Brunt-Väisälä frequency squared (N^2) profiles, there are often two stratification peaks in the water column - one between the surface mixed layer and the cold halocline, and one between the cold halocline and the AW. A decrease in the N^2 value of one peak often results in an increase of the other as the relative densities of these water masses change. Furthermore, changes in the vertical distance between the AW and surface mixed layer can affect the overall stability between the two water masses. Quantifying the overall stability between the AW and the surface is therefore not straightforward.

As mentioned, available potential energy (APE) was found by (Polyakov et al. 2018) to be a good measure of bulk stability across the halocline. In the following analysis, APE (see Equation 4.2) will be used to quantify the stability of the upper water column and any change in water column stability. Although this will not provide as detailed a look at the stability of different sections of the upper water column as N^2 profiles, APE can provide a definitive measure to address the key question of this chapter: will the increase in AW heat content impact the Arctic Ocean surface by changing the stability of the upper water column? Details of the APE calculation can be found in the Data and Methods section of this chapter. A more negative APE represents higher stability.

Figure 4.14 shows time series of regionally-averaged APE anomaly relative to 1920 (regions are the eastern and western regions defined in Figure 3.2) for each ensemble member and the ensemble mean. This figure shows that the APE becomes more negative over time in both regions in March and September, implying that there is an increase in the stability of the upper 500 *m* of the water column. This appears to be a forced trend, laying outside of the range of natural variability represented by the envelope of ensemble member time series. The APE decreases commence during the 2010s. This is around the same time as the start of the forced trends in regionally averaged AW layer salinity and density seen in the Chapter 3 (Figure 3.8). Therefore, given that much of the change in density profiles seen earlier in this chapter was due to surface freshening, this implies that AW freshening and surface freshening are, unsurprisingly, linked - be that through increased AW freshening upon subduction beneath the Arctic surface layer, or through increased AW freshening in the

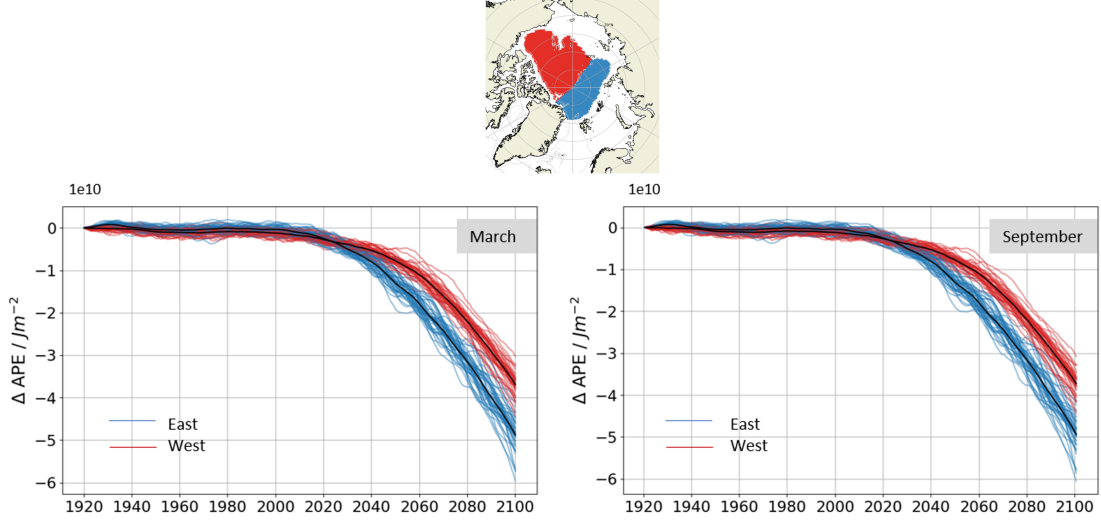


Figure 4.14: Time series of March and September ensemble mean available potential energy anomaly, relative to 1920, averaged over the eastern (blue) and western (red) Arctic (regions defined in map). Coloured lines are time series from each ensemble member, with black lines denoting the ensemble mean.

Nordic Seas.

The similarities between the March and September APE time series are striking (the differences between a typical ensemble member time series in March and September are about 2% of the average value of the time series, so are not noticeable). This highlights that the changes in the overall upper water column stability are not seasonal, despite stark differences in the temperature and salinity profiles between seasons (particularly at the surface).

An important take-away from Figure 4.14 is that the stability increases despite a reduction in magnitude of some of the N^2 peaks during latter half of the 21st century (although, as mentioned, the impact of this on overall water column stability was ambiguous given that N^2 represents local point-wise stability, and that other N^2 peaks increased in magnitude at the same time). This emphasises the importance of not basing overall stability on the shape of water column profiles alone without proper quantification. The APE takes away any ambiguity and provides a useful bulk measure of stability changes (at the expense of vertical detail in water column stability). The APE decreases, and stability increases, over time in the upper water column essentially because of the large degree of surface freshening.

This stability increase implies that increasing AW heat content has little direct impact on the local surface layer and sea ice across the Arctic. This is due to both the surface freshening and the proceeding halocline freshening seen in Figures 4.4, 4.6, and 4.8. Given that much of

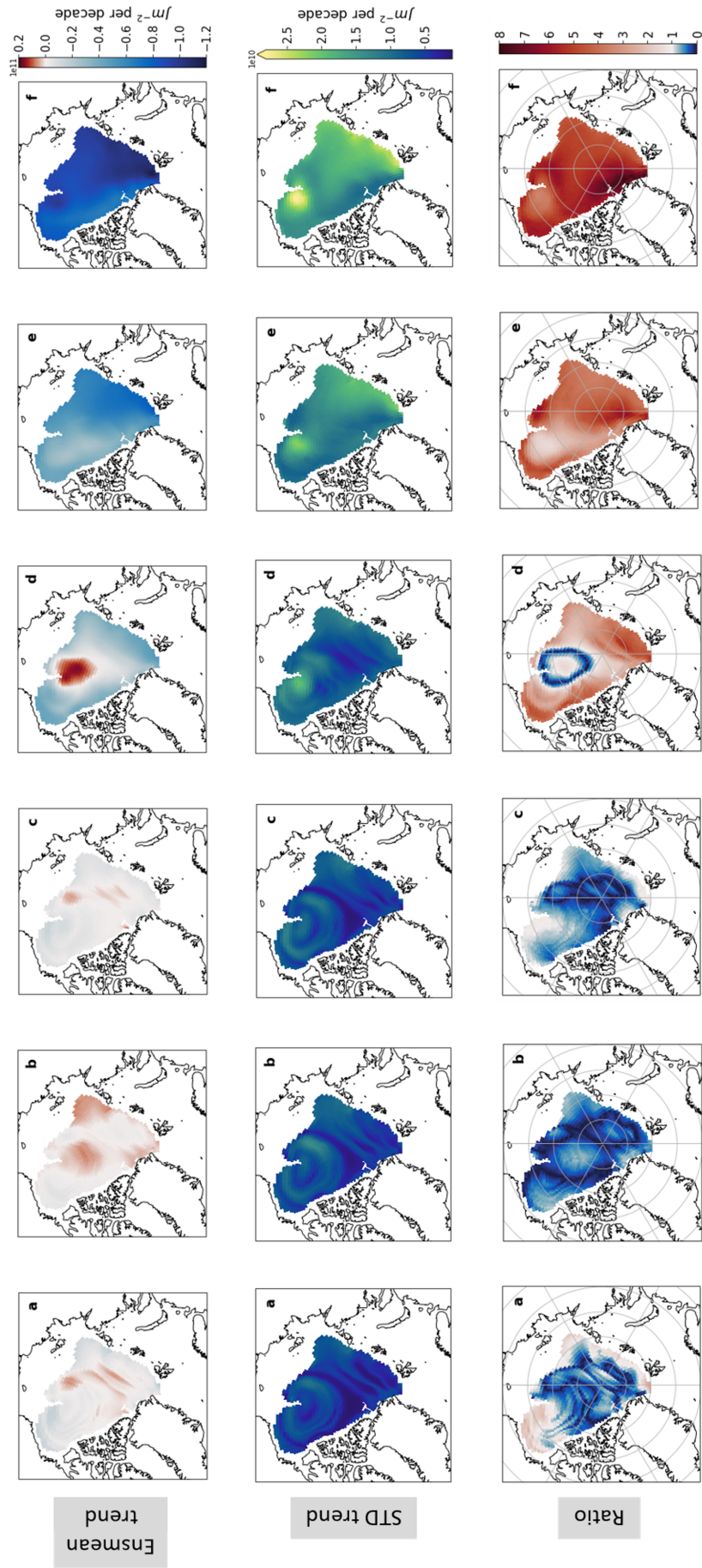


Figure 4.15: March ensemble mean trend, representing forced trend (top row); standard deviation of trends amongst detrended ensemble members, representing natural variability (middle row); and the forced trend divided by the natural variability (bottom row) of APE. Data have been averaged over the following 30-year periods: (a) 1920–1949, (b) 1950–1979, (c) 1980–2009, (d) 2010–2039, (e) 2040–2069, (f) 2070–2099.

the summer sea ice loss will have already occurred by the end of the 21st century according to this model (see Onarheim et al. (2017) and Figure 3.7), the salinity of the surface layer in winter will become more important in maintaining this strong stratification as AW warms and freshens (as seen in Chapter 3). This is because not only does the winter surface layer salinity affect winter stability, but it also impacts the density of the summer cold halocline which, as seen above, is important in maintaining the stability between the AW and the surface. Of course, it is likely that there is an indirect impact of AW heat content increase on sea ice given that a warmer AW layer would transfer more heat to the atmosphere at AW inflow regions where it is at the surface, which may then affect surface air temperatures and resultant sea ice melt. However, this is out of the scope of this chapter.

To investigate these APE/stability changes in more spatial detail, and to confirm quantitatively whether these are part of an anthropogenically forced trend or natural variability, Figure 4.15 shows the ensemble mean trend, standard deviation of the trends of the detrended ensemble members, and the ratio between these two quantities. This ratio represents the relative size of any stability trend compared to the natural variability of the stability. A ratio larger than one implies the influence of a forced trend, with a ratio less than one indicating that natural variability is the dominant factor governing stability change. Data from March is shown in this figure - September results are very similar given the similarity in APE anomaly.

Although Figure 4.15 shows an increase in APE and hence decrease in stability across much of the Arctic during the 20th century, dominated by natural variability, there is a strong anthropogenically forced stability increase (APE decrease) across the Arctic during the 21st century. A notable exception is the patch of stability decrease in the Makarov Basin during 2010–2040. This could be related to the decrease in Beaufort High winds (see Figure 3.22) during this period, reducing the accumulation of freshwater here, before this is overridden by the decreases in surface and halocline salinity seen above due to sea ice melt.

It is important to note that the lack of Pacific Water in the western Arctic (Lavoie et al. 2022) and surface stratification biases (resulting in a more saline surface layer) due to unrealistic vertical mixing (Rosenblum et al. 2021) in CESM-LE may influence these APE values and trends, mainly by potentially impacting halocline thickness and temperature, and surface salinity. We do not think that this would impact our conclusions, although repeating similar analysis with other models would be a very useful avenue of further work.

Overall, we see that the increases in regional stability in Figure 4.14 are widespread throughout much of the Arctic, with all regions seeing a forced increase in the overall stability of the upper 500 *m* of the water column at some point during the 21st century.

4.4 Conclusion

This chapter aimed to answer the following question: what do the AW changes seen in previous chapters mean for water column stratification in the Arctic? The answer to this is that, despite the AW freshening and warming, the stability of the upper 500 *m* of the water column increases during the 21st century, mainly due to the freshening of the surface and halocline in relation to the AW layer. This is true in both the eastern and western Arctic, despite regional differences emerging in a similar metric computed from observations up to the present day (Polyakov et al. 2018).

Although the role of temperature in determining density increases over time in both the Eurasian and Canada Basins, salinity still governs the water column structure. This beta ocean stratification maintains the stratification between the freshening surface and the AW, which does not freshen to the same extent over the model run. However, the Barents Sea water column switches from a beta to an alpha ocean during September in the latter half of the 21st century. This coincides with when the Barents Sea first becomes ice free year-round (see Figure 3.7 and (Onarheim et al. 2017)), implying that this transition could occur in regions such as the Eurasian Basin in the future if summer sea ice loss and temperature increases in the upper water column during summer continue. This could have important implications for convection regimes within the Arctic, with enhanced heat loss to the atmosphere as cold surface water would be more dense than the warmer water below.

Chapter 5

Summary and conclusions

The Arctic is experiencing the effect of climate change on increasing air temperatures more than any other region on Earth (Overland et al. 2019; Rantanen et al. 2022; Serreze et al. 2011). One of the most noticeable signatures of this is the drastic sea ice loss seen in the Arctic over recent decades (Deser et al. 2008; Notz et al. 2016; Perovich et al. 2019). The importance of oceanic heat - particularly the heat contained in the AW layer - in this sea ice loss is an open question. AW is the most significant source of oceanic heat in the Arctic (Carmack et al. 2015), and is separated from the cold surface mixed layer and sea ice across most of the Arctic basin by a strong halocline. The AW both within the Arctic and inflowing into the Arctic is warming (Mulwijk et al. 2018; Polyakov et al. 2012a), and is currently warmer than it has been for at least 2000 years (Spielhagen et al. 2011). Observations suggest that this AW warming, along with associated AW shoaling, in the eastern Arctic over recent decades has been linked to increased vertical heat flux toward the surface in this region (Polyakov et al. 2017; Polyakov et al. 2010). A changing Arctic is important both locally and globally - albedo reduction due to sea ice loss (Pistone et al. 2014), and freshening outflows into the North Atlantic and their potential impact on overturning (Jahn et al. 2013) being two key impacts of sea ice loss. Studying the emerging role of AW heat on Arctic sea ice loss is therefore increasingly important. This thesis set out to do just that, specifically to investigate the fate of AW heat in the Arctic Ocean and how this might change in the future.

A key feature of this thesis is its use of both observational and model data to gain insights from both perspectives. By first building a pan-Arctic picture of AW in the Arctic over recent decades, we aimed to gain an understanding of how AW heat has varied temporally

and spatially over recent years. These results were then put into a longer-term perspective using data from a large ensemble of model runs. Here, we focused on how the AW layer may change in the future and to what extent any changes are driven by anthropogenically forced climate change. The impact of this AW layer evolution on the upper water column via stratification changes was then investigated.

This work has revealed some key insights into the regional differences in AW trends to date, how the distribution of AW heat may change in the future, and the extent of the potential impact of these changes on water column structure. This chapter will summarise key results, put them into a wider context, and discuss potential future work.

5.1 Thesis summary and discussion

In Chapter 2, 59000 observational profiles from across the Arctic Ocean taken between the 1970s and 2018 were compiled and analysed to give an up-to-date pan-Arctic picture of the AW layer in the Arctic, and also to investigate how the AW layer has varied temporally and spatially over this time period. The broad scope of the observations also allowed us to evaluate the performance of metrics commonly used to assess the AW layer.

Beyond providing and publishing an extensive picture of the spatial and temporal variability of AW in the Arctic in recent decades (with the AW core data available as an open access dataset here: <https://doi.org/10.5287/bodleian:wxv8GA7Mk>), which will be a useful resource for future AW studies, two key findings emerged from this work. One of these was the differing evolution of the AW layer in the eastern and western Arctic over recent decades (i.e. the regions on the Eurasian and Amerasian side of the Lomonosov Ridge, respectively, see Figure 3.2). In the eastern Arctic, 2002–2018 saw a $0.7\text{ }^{\circ}\text{C}$ increase in AW core (i.e. at the AW layer temperature maximum) temperature, and an accompanying increase in AW layer heat content. Instances of high upper ocean heat content in the east coincided with fresher, shallower AW - a reflection of the influence of AW on the upper ocean here. Variations of the salinity in the AW layer here do not seem to be linked to its temperature, suggesting that AW salinity is linked to how much AW heat is transferred to the surface. This was investigated further in Chapter 4. These results are in line with the Atlantification of the eastern Eurasian Basin, reported in several studies (e.g. Polyakov et al. (2017) and Polyakov et al. (2010)). Although western AW layer heat content also increased over a similar period

(2008–2018), AW core temperatures here decreased by $0.1\text{ }^{\circ}\text{C}$ over these years. Crucially, AW heat became more isolated from the surface layer in the Canada Basin, in contrast to what was seen in the east. This is due to isopycnals deepening in the Canada Basin as freshwater accumulated in the halocline and surface layers, forcing the AW layer to reside deeper here. We attribute this to a concurrent spin-up of the Beaufort Gyre reported by Proshutinsky et al. (2019) and Zhong et al. (2014) and others. Along with the cooling AW core in the western Arctic, the AW layer heat content in the west does not increase to the same degree as in the east. These results suggest that two different AW regimes have recently emerged in the Arctic Ocean. One in the east, where AW is having an increasing influence on the surface mixed layer, and another in the west where AW is becoming more isolated from the surface and warming Pacific Water in the halocline plays a more important role in sea ice melt (Perovich et al. 2015; Timmermans et al. 2014; Timmermans et al. 2018). The future evolution of the Beaufort Gyre, both in terms of its intensity and position, will also be a key factor governing how the hydrography of the western Arctic changes (Doddridge et al. 2019). These regional differences are in agreement with recent studies by Polyakov et al. (2020a), Polyakov et al. (2018), and Polyakov et al. (2020b), with our analysis going into more detail at specific locations in order to give a more comprehensive picture of the spatial and temporal variability of various properties of the AW layer throughout the Arctic.

The second key observational finding was on how AW is transformed during its advection around the Arctic. The AW core cools and freshens upon advection, but despite this freshening the AW core becomes more dense. We propose that there are two likely mechanisms contributing to this. The first is heat and salt being lost preferentially from the top of the AW layer as the AW mixes with cooler, fresher water above it along its advection pathway. This results in a deeper core existing on denser isopycnals as heat and salt are lost from the upper AW without the AW layer as a whole getting denser. A second mechanism likely to be contributing to this transformation is interaction between AW and cold, dense cascading shelf flows. These shelf flows are the result of brine rejection from sea ice formation on the shelves (Turner 2010), and have been observed to interact with and modify AW in the Eurasian Basin (Ivanov et al. 2007). In the western Arctic, despite limited observations, interaction between these shelf flows and the lower halocline has been observed (Ivanov et al. 2004; Luneva et al. 2020; Melling et al. 1995), suggesting that interaction with upper AW

here is indeed feasible. The transformation of the AW core during its advection highlights the importance of local processes in modifying the AW within the Arctic. The relative importance of local processes and AW inflow in setting AW properties within the Arctic will be discussed when summarising Chapter 3.

As implied above, the properties of the AW at its temperature maximum (or core) do not always reflect those of the AW layer as a whole. However, properties of the AW core are commonly used in observational studies to assess the state of the AW layer and to study its circulation and transformation (Karcher et al. 2003). Given that our observational dataset contained a significant number of profiles which sampled the whole AW layer, we were able to assess the effectiveness of the AW core measurements in capturing the state of the AW layer. We found that the potential temperature of the AW core captured the spatial pattern of AW heat content well, but did not capture the temporal changes of AW heat content in a given region. In terms of depth, although AW core depth reflected the depth of the top of the AW layer (defined as the 0°C isotherm above the AW temperature maximum) in the western Arctic, this was not the case in the eastern Eurasian Basin. Our results are therefore a reminder that care should be taken when working with AW core data. The AW core is overall a useful metric for assessing AW layer properties, and is far more easily measured than the entire AW layer itself - as demonstrated by our dataset of 59000 profiles resulting in 44000 successfully identified AW core data points but only around 1500 profiles sampling the full AW layer. However, caution should be used when utilising AW core properties to draw conclusions about temporal AW layer changes in particular.

Chapter 3 moved away from observational analysis and instead involved data from the Community Earth System Model Large Ensemble (CESM-LE), which consists of 40 runs of a coupled climate model with 1° horizontal resolution, each perturbed by round-off level differences in initial air temperature (Kay et al. 2015). Variation between ensemble members therefore represents natural climate variability due to atmospheric fluctuations, and allowed us to investigate the relative roles of natural/internal variability and forced/externally-driven trends. The model is run using historical data between 1920–2005 and RCP8.5 forcing from 2006–2100.

The longer term perspective provided by the model allows us to set the changes seen in earlier observational analysis into a broader context. A key result of Chapter 3 is that,

contrary to what was suggested by the observational analysis in Chapter 2, the model does not show two opposing regimes in the eastern and western Arctic in the 21st century, with increased (decreased) of AW influence on the surface in the east (west). In fact, from around 2020 onwards, forced shoaling, warming and freshening trends emerge in AW heat in the western Arctic in the model data, lagging a few decades behind similar trends in the eastern Arctic. Therefore, we propose that the opposing regimes that appeared to emerge in the observational results are not the beginning of long-term differences in AW behaviour in the eastern and western Arctic, but are instead a result of our observational time period coinciding with this lag. Shoaling in both the eastern and western regions occurs as a vertical heat flux into the halocline, and is not the result of halocline thinning or isopycnal shoaling.

It is important to note that the shoaling of AW heat in the western Arctic does not necessarily suggest that Atlantification will progress into the western Arctic, due to the presence of Pacific Water in the western Arctic halocline. The Pacific Water layer here has thickened and warmed in recent years, contributing to sea ice melt in the region (Perovich et al. 2015; Timmermans et al. 2014; Timmermans et al. 2018; Woodgate et al. 2010). As detailed by Polyakov et al. (2020a), Pacific Water has had an increasing influence on the upper layers of the western Arctic Ocean in a similar way to the observed Atlantification of the eastern Arctic. The future spreading of AW heat into the halocline shown in our model data results would likely have little impact on local sea ice/surface layer heat content compared to this warming Pacific Water. However, our results show that AW should not be discounted when considering how the heat content of the upper water column of the western Arctic may change this century.

As mentioned, we found the pan-Arctic AW warming, freshening and shoaling trends in the 21st century model data to be forced/anthropogenically-driven. This is in contrast to the 20th century, where we found that local changes in the AW layer often relate to advected natural oscillations in AW properties or temporal variation in the local water column structure (related to wind-driven variations in halocline thickness). Although some studies have investigated the role of natural variability in sea ice decline in the Arctic (e.g. (England et al. 2019; Kay et al. 2011)), and even the role that AW heat transport plays in this (Årthun et al. 2019), few have investigated the role of natural variability in pan-Arctic AW changes in the 20th and 21st centuries. This finding was therefore a key part of our work. The switch

between the 1920–2000 period, where natural variability dominated AW properties in much of the Arctic, to the 2020–2100 period (simulated under an RCP8.5 scenario), with forced trends driving AW changes across the Arctic basin, provides a clear marker of the impact of climate change on the evolution of AW heat in the Arctic - the effects of which we are yet to see in the western region.

Our results suggest that an important driver of these 21st century changes appears to be a shift to an atmospheric circulation resembling a more positive NAO/AO. Various studies have found links between atmospheric conditions over the Nordic Seas and the state of AW flowing into and within the Arctic basin. Enhanced cyclonic wind stress over the Nordic Seas, associated with a positive NAO/AO, has been found to drive increased volume transport in the Arctic, resulting in more AW water being advected over the Lomonosov Ridge into the western Arctic, and increased AW inflow temperature (Chatterjee et al. 2018; Karcher et al. 2012; Muilwijk et al. 2019; Muilwijk et al. 2018). Furthermore, thermodynamic forcing due to surface air temperatures over the Nordic Seas greatly impacts the temperature of inflowing AW (Muilwijk et al. 2018). Our results agree with this correlation between large scale atmospheric circulation resembling the NAO/AO and enhanced AW volume transport into the Arctic, along with more AW transport into the western Arctic. We build on the results of the studies referenced above and investigate what the CESM-LE reveals about future AW layer changes. We propose that a trend towards a more positive NAO/AO-like atmospheric state in the 21st century, with enhanced cyclonic wind stress over the Nordic Seas and much of the eastern Arctic, and a reduced Beaufort High, drives enhanced volume transport into the Arctic during this period. It also results in the emergence of a pan-Arctic AW boundary current in the model data (AW circulation in the 20th century model data instead involved recirculation along the Lomonosov Ridge and anticyclonic AW flow in the western Arctic). The anthropogenically driven (i.e. forced) AW warming seen throughout the Arctic in the 21st century can be attributed to these drivers, with an increase in inflow volume transport coinciding with higher inflowing AW temperatures during the 21st century, resulting in an increase in heat transported into the Arctic.

Local processes (i.e. mixing at the top of the AW layer and interactions with cascading shelf flows) also appear to play an important role in AW properties, particularly further along the AW advection pathway, as suggested by the observational analysis in Chapter 2. AW

freshening trends in the model data were strongest at the shelves, implying that increases in seasonal sea ice formation in shelf regions are related to this freshening. Furthermore, despite an increasing correlation between AW inflow temperature trends through the Fram Strait and AW layer temperature elsewhere in the Arctic, correlation between inflow temperature trends and those elsewhere remained moderate at best across much of the Arctic. This is particularly pertinent given the expected increase in Arctic mixing levels as sea ice cover reduces (e.g. Rainville et al. 2011) and changes in the seasonal sea ice cycle on the shelves (see e.g. Figure 3.7) which could impact cascading shelf flows. The CESM-LE data also suggests that local processes act as a low pass filter on AW temperature variability, buffering AW temperatures within the Arctic against the higher frequency variability of AW inflow temperatures. Overall, these insights into the influence of local processes on AW properties show that atmospheric forcing is not the only factor affecting the fate of AW heat in the Arctic, and should provide further impetus for more studies into the role of these processes in the spatial and temporal variability of AW heat.

Chapter 4 also made use of CESM-LE data, this time to investigate the impact of AW warming and freshening on the upper water column, with a focus on whether these changes result in weaker stratification of the upper 500 *m*. This is key because, although changes in the properties of AW in the Arctic will likely impact the North Atlantic via Arctic outflows (Karcher et al. 2011), if stratification remains strong then the local impact of AW heat on sea ice within the Arctic will be limited. Reductions in stratification have already been observed in the Eurasian Basin (e.g. Polyakov et al. 2017; Polyakov et al. 2018; Polyakov et al. 2010). Therefore, investigating whether this reduction in stratification was also found in the model data, and how stratification is predicted to evolve in the 21st century alongside forced trends in AW properties, is highly relevant. Chapter 4 also quantified the relative roles of temperature and salinity in governing density (and hence stratification) at several locations. The Arctic is presently a beta ocean, with salinity governing density and stratification, allowing fresher, cooler water to exist above the warmer, more saline AW and enabling sea ice to form. Alpha oceans, meanwhile, are warmer (typically $> 10\text{ }^{\circ}\text{C}$) and are stratified according to temperature, owing to the nonlinear increase in the thermal expansion coefficient with temperature (Carmack 2007). A warming AW progressing further into the Arctic therefore raises the question of whether the Arctic may be heading towards alpha stratification. This

could have significant consequences for sea ice (which can only form in beta oceans (Bulgakov 1962)) and AW ventilation (given that warm water is more likely to be ventilated by convection in an alpha ocean).

The key outcome of the analysis of Chapter 4 was that stability in the upper 500 *m* of the water column increases during the 21st century, as summarised by Figure 4.14, primarily because the freshening in the surface mixed layer and halocline is greater than AW freshening. This stability increase is forced/anthropogenically-driven and is present in both the eastern and western regions from the 2020s onwards in both March and September. Polyakov et al. (2018) quantified a similar metric (available potential energy (APE) between the surface and the halocline base, as opposed to our APE which was computed between the surface and 500 *m* depth) using observations spanning 1981–2017, and found stability decreasing in the eastern Arctic while increasing in the western Arctic. Therefore, our results suggest that although current observational studies suggest that stability may be decreasing between the AW and the surface in some regions, this may not hold true later this century as the surface mixed layer and halocline continue to freshen.

Another key finding from Chapter 4 was that within the Arctic basin the ocean remains a beta ocean, despite warming that results in temperature playing an increasingly important role in density and stratification - a further indication that AW heat is destined to remain relatively isolated from sea ice across much of the Arctic. However, in the Barents Sea in September, the water column becomes alpha-stratified during the latter half of the 21st century, coinciding with the time period in which the Barents Sea becomes ice-free year round in the CESM-LE ensemble members. This implies that a similar transition could occur in the Eurasian Basin if summer sea ice loss and surface mixed layer warming continue after the 21st century (or potentially earlier given the CESM-LE caveats described below), resulting in a northward transition of the alpha-beta front (hence sea ice extent) and, as alluded-to above, a potential increase in AW ventilation. Furthermore, although we find stability increasing, AW heat still shoals into the halocline across the Arctic during the 21st century. Therefore perhaps a future focus on vertical heat flux alongside changes in water column stability is the best approach in investigating the fate of AW heat. Avenues of future work will be discussed in more detail later.

Taking a broader view, this thesis has used the complementary perspectives of observa-

tional data and model data to provide some key insights into the AW layer of the Arctic Ocean. Each data source has its own advantages - observational data providing a relatively (accounting for measurement errors) realistic representation of the real ocean, and model data providing temporally and spatially homogeneous fields of data, enabling a range of analysis techniques not feasible with observations, and allowing us to look into the future. However, each has drawbacks and caveats which are important to bear in mind when interpreting the results of this thesis. Unlike observations, CESM-LE - as with all models - systematically misrepresents some aspects of the hydrography of the Arctic. As seen in Figure 3.1, and reported by Shu et al. (2019), the AW layer in CESM-LE is too deep and thick. Furthermore, the model has a cold Arctic surface air temperature bias (Park et al. 2014) and underestimates sea ice loss in the Barents Sea (Onarheim et al. 2017), which suggests that simulated Atlantification may also be delayed. The model data also has a warm bias in the AW layer due to insufficient cold water inflowing through the St. Anna Trough - this is a common bias for a model of this type (Danabasoglu et al. 2012; Ilıcak et al. 2016). CESM-LE AW heat content values are also likely too large due to the simulated thickness of the AW layer. The poor temporal and spatial coverage of observations (particularly pre-2000 in the Arctic) provides a different set of caveats. Obtaining meaningful results from the pan-Arctic observational dataset for Chapter 2 was a challenge, especially as we chose not to interpolate between data points in order to introduce as little error as possible into our results. However, we were able to draw some key conclusions about how AW has evolved across the Arctic and have created a useful reference for further work - both for benchmarking model representation of AW against observational datasets, and for further observational analysis of AW properties.

Five key aims were listed in the introduction to this thesis. The first was to build a detailed pan-Arctic observational picture of the AW layer, which formed the cornerstone of Chapter 2, allowing us to describe spatial and temporal AW changes and evaluate the effectiveness of the AW core properties as a metric of AW variability. Chapter 3 tackled the second and third aims of this thesis: investigating recent and projected changes in AW layer properties in an ensemble of coupled climate model runs under historical and RCP8.5 forcing, and evaluating the roles of natural variability and anthropogenic forcing in these changes. We assessed the impact of the AW changes on the upper water column and stratification in Chapter 4, in line with the fourth thesis aim. Overall, all work in this thesis has been oriented around assessing

the current and future fate of AW heat in the Arctic, which was the fifth and overarching aim of this work.

5.2 Implications for the Arctic region and wider climate system

Many of the impacts of a changing Arctic Ocean on the general Arctic environment have been alluded to already. For example, the loss of sea ice in terms of extent and volume in both summer and winter due to the increasing direct impact of ocean heat on sea ice loss. Both Atlantic and Pacific Water warming have played a role in this, in the eastern and western Arctic, respectively (Perovich et al. 2015; Polyakov et al. 2020a). Our results suggest that the AW influence on the upper water column heat content (and hence sea ice) will increase this century, albeit not necessarily through a decrease in water column stability. A thinner, more mobile sea ice pack will also be unable to sustain as much stress which will impact the dynamics of the wind-driven Beaufort Gyre, potentially allowing more freshwater to be stored here (Doddridge et al. 2019; Meneghello et al. 2018). Freshwater storage will also be impacted by sea ice melt directly. Proshutinsky et al. (2019) found seasonal freshwater storage in the Beaufort Gyre to increase with increasing seasonal sea ice melt. Therefore, an increased influence of AW on sea ice loss will likely enhance the amount of freshwater stored within the Arctic.

Enhanced AW vertical heat flux to the Arctic Ocean surface will of course influence surface air temperatures in the region, which are already warming above the global average (Overland et al. 2019; Rantanen et al. 2022; Serreze et al. 2011), with the potential to further enhance “Arctic Amplification”. Results in this thesis have suggested that the Barents Sea might transition to an alpha ocean during summer later in the 21st century. This could potentially alter the local convection regime to one in which warm water (AW) is ventilated and cooled before sinking and being replaced by warmer AW - affecting the properties of AW flowing into the Arctic and enhancing heat loss from the AW layer to the atmosphere. Warming surface air temperatures will lead to more sea ice loss - this time as an indirect impact of Atlantification. However, a shift to an alpha stratification would not necessarily lead to this convective change, and more investigation is needed to evaluate the true impact of an alpha-stratified Barents Sea.

A change in the physical properties and distribution of the water masses within the Arctic Ocean will also result in changes in geochemical and biological components of the ocean. Biological productivity has increased over large parts of the Arctic in recent years due to the increased influence of Atlantic and Pacific Waters on the eastern and western Arctic, respectively, along with sea ice loss; with a shift to a dominance of boreal species over Arctic species in many shelf regions (Eriksen et al. 2017; Hollowed et al. 2013; Kovacs et al. 2011; Polyakov et al. 2020a). In the interior, Polyakov et al. (2020a) found that primary productivity increases occurred in regions that experienced an increase in vertical mixing (resulting in greater nutrient flux to the surface layer), i.e. much of Eurasian Basin and the parts of the Amerasian interior experiencing an increase in Pacific Water influence due to changes in Pacific Water pathways. Eddies also play a key role in supplying Pacific Water nutrients to the halocline in the Amerasian Basin (Mathis et al. 2007).

As mentioned, sea ice loss is expected to contribute to an increase in freshwater storage in the Arctic. However, freshwater storage in the Beaufort Gyre (the largest store of freshwater in the Arctic) also depends greatly on the intensity of the Beaufort High. Our work has shown that, according to CESM-LE, the Beaufort High is expected to reduce during this century. This would weaken the Beaufort Gyre and result in an export of freshwater into the North Atlantic (on top of increases in freshwater content that would be expected from sea ice melt being transported out of the Arctic via the transpolar drift). Furthermore, our results have implied that AW will continue to warm and freshen, and hence become less dense, in the 21st century. Both surface polar water and Atlantic origin water exiting the Arctic into the Nordic Seas will therefore be less dense. The less-dense Atlantic origin water can not only directly affect the density of outflows through the Denmark Strait (e.g. Jeansson et al. (2008) and Karcher et al. (2011)), but can also affect boundary densities in the region southeast of Greenland which play an important role in setting the strength of the Atlantic meridional overturning circulation (AMOC) (Hirschi et al. 2007). Further downstream, the potential increase in surface freshwater output from the Arctic via the East Greenland Current into the Irminger and Labrador Seas would also impact overturning strength via a weakening of the surface-forced water mass transformation that gives rise to the sinking limb of the AMOC (e.g. Petit et al. (2021) and Yang et al. (2016)). Given the important role played by the AMOC in global ocean heat and carbon uptake and redistribution, the impacts of a weakened

AMOC would be far reaching (Buckley et al. 2016; Jackson et al. 2018).

5.3 Further work

Several open questions remain about the fate of AW heat in the Arctic. Here, the main priority areas based on the findings of this thesis will be briefly described.

A key result of this thesis is that local processes - specifically vertical mixing between the top of the AW layer and the halocline above, and interaction between AW and cascading shelf flows - influence regional differences in AW temperature and salinity within the Arctic, and also play an important role in the dominant timescales of variability of local AW temperature. Determining which mechanism is more important and whether this varies spatially is an important next step. More widespread microstructure measurements across the Arctic Ocean will be key in aiding understanding of vertical mixing. Meanwhile, observations at continental shelves will be crucial in capturing any potential interaction between the AW and shelf flows (similar to what was achieved by Ivanov et al. (2004), Luneva et al. (2020), and Melling et al. (1995)), which is expected to become more likely as the AW layer becomes less dense. The relative contribution of each proposed mechanism to AW cooling and freshening could then be quantified, and the projected impact of these local processes on future AW changes could be better understood.

Improving our understanding of how these local processes could impact future AW properties, and their relative contribution to this, is also important. CESM-LE data has shown that AW freshening trends are strongest near the shelves, suggesting that cascading flows interact with the AW layer in a similar way to that proposed by Turner (2010). There would, however, be a limit to this freshening (and accompanying cooling) trend if the shelves became ice-free year round. If, instead, vertical mixing plays a greater role in AW freshening and cooling trends (with greater freshening near rougher topography), then the limit on AW freshening would be different. As mentioned previously, reductions in sea ice extent are expected to increase ocean mixing levels (Rainville et al. 2011). This could perhaps result in more widespread AW freshening and cooling, occurring in deep basins too, as sea ice cover reduces and the Arctic generally becomes more turbulent - even away from topographic features. Any increase in mixing levels would also of course increase vertical heat flux.

Vertical heat flux is also impacted by the difference between AW and halocline properties.

Given the important role of surface and halocline freshening in upper water column stability found in Chapter 4, predicting how these are likely to change in relation to AW salinity will be an important part of understanding how AW heat is likely to interact with the surface. Therefore, a key area of future work will involve modelling the future evolution of Arctic halocline salinity and of the winds that drive the advection of shelf waters key to halocline formation (Doddridge et al. 2019; Schauer et al. 2002b; Steele et al. 1998)

Returning to local processes, investigating how the interaction between AW and shelf flows and mixing at the AW top are likely to change in the future involves the use of models. Being a coarse-resolution model, the CESM-LE has obvious limitations in simulating the details of lateral and vertical mixing. A higher resolution model is therefore essential to investigate the role of mixing, eddies and thermohaline intrusions. Improving the resolution of coupled climate models in the Arctic is also a priority given that it is one of the key ways in which model performance in the region can be improved, as highlighted by Khosravi et al. (2022). Biases are present in the CESM-LE, as described earlier, so it is important to keep these in mind when interpreting our model results. The warm AW bias due to insufficient cold water entering the Arctic through the St. Anna Trough (Ilıcak et al. 2016) may mean that our results overestimate AW warming trends within the Arctic. CESM-LE also models the AW layer as being too deep, which may result in an underestimation of the impact of AW heat on the upper ocean - although Årthun et al. (2019) found that sea ice responded realistically to ocean heat transport through the Barents Sea in CESM-LE. Finally, CESM-LE lacks Pacific Water in the western Arctic halocline (Lavoie et al. 2022) and exhibits a surface stratification bias (Rosenblum et al. 2021), both of which likely affect our estimates of water column stability. Given these biases, further pan-Arctic AW studies using different coupled climate models would be useful to provide a different model perspective on projected AW/Arctic Ocean changes in the 21st century.

The use of higher-resolution models in the Arctic would also aid in another key area of future work that we propose: an investigation of the transfer of AW heat from the boundary to the interior basins. Eddies and intrusions are the primary ways through which AW heat is transferred from the boundary to the interior, and they are not explicitly modelled in coarse-resolution models such as the ocean component of CESM-LE. Analysing how heat transfer mediated by intrusions is likely to change as warmer AW is advected around the AW

boundary current, and the lateral temperature and salinity gradients between basin boundary and interior alter, would be very useful in enhancing our understanding of the spatial and temporal variability of AW heat across the Arctic. Similarly, it would be very insightful to explicitly model changes in eddy heat and salt fluxes in a warming, freshening Arctic - particularly at key locations such as the confluence of FSBW and BSBW just downstream of the St. Anna Trough, and over the rough topography of the Chukchi Plateau. Further investigating the transfer of AW heat from the Arctic boundaries to the basin interiors would also be something we would focus on using our observational dataset if we had more time. Using regional averaging to assess changes in boundary and interior heat content would provide insight into the magnitude and speed of heat transfer from the boundary to the interior, and how this may have increased as the boundary current has warmed in relation to the cooler, abyssal waters of the interior.

If more time were available, we would also investigate the causes of the increasing AW outflow temperature through the Fram Strait. In Chapter 3 we propose that this is likely related to the increased AW inflow temperature through the Barents Sea Opening, and/or the warming of other water masses within the Arctic that interact with AW. Investigating these two hypotheses using estimated advection timescales of recirculating BSBW, along with identifying and quantifying any reductions in AW heat loss within the Eurasian Basin, would be the first steps towards understanding the roles of each mechanism in the outflowing temperature increase. This would then allow us to better understand whether the increase in heat transport out of the Arctic into the Nordic Seas is likely to continue to increase beyond the modelled time period.

5.4 Final word

This thesis gives a detailed pan-Arctic observational picture of the spatial and temporal variability in AW heat within the Arctic, before using data from a coupled-climate model to put these changes into a longer-term perspective. Model results suggest that AW heat will have an increased influence on the upper water column across the whole Arctic in the 21st century. These trends lay outside of the range of natural variability, and can instead be attributed to a changing climate. Despite water column stability being maintained throughout the Arctic basin due to surface and halocline freshening, AW heat seems destined to play an increasingly

important role in a changing Arctic.

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