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Weakening of cold halocline layer exposes sea ice to oceanic heat in the eastern Arctic Ocean

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Abstract:

A 15-year duration record of mooring observations from the eastern (>70ºE) Eurasian Basin (EB) of the Arctic Ocean is used to show and quantify the recently increased oceanic heat flux from intermediate-depth (~150-900 m) warm Atlantic Water (AW) to the surface mixed layer (SML) and sea ice. The upward release of AW heat is regulated by the stability of the overlying halocline, which we show has weakened substantially in recent years. Shoaling of the AW has also contributed, with observations in winter 2017-2018 showing AW at only 80 m depth, just below the wintertime surface mixed layer (SML), the shallowest in our mooring records. The weakening of the halocline for several months at this time implies that AW heat was linked to winter convection associated with brine rejection during sea ice formation. This resulted in a substantial increase of upward oceanic heat flux during the winter season, from an average of 3-4 W/m^2 in 2007-2008 to >10 W/m^2 in 2016-2018. This seasonal AW heat loss in the eastern EB is equivalent to a more than a two-fold reduction of winter ice growth. These changes imply a positive feedback as reduced sea ice cover permits increased mixing, augmenting the summer-dominated ice-albedo feedback.

Keywords: Arctic Ocean, climate change, stratification, sea ice, ocean mixing

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1. Introduction

In recent decades there has been a dramatic decline in seasonal sea ice extent in the Arctic Ocean, with a more recent year-around decline in sea ice extent, area and volume (Kwok 2018; Stroeve and Notz 2018). This change has shifted the local radiative balance resulting in a positive ice-albedo feedback mechanism as increasing lead fraction and surface melt pond areas in decaying Arctic sea ice facilitate enhanced upper-ocean solar heating and more rapid melting of ice floes (e.g., Perovich et al. 2008; Toole et al. 2010). Moreover, it was hypothesized that the declining sea ice has larger scale hemispheric impacts on the North Atlantic Oscillation and, in consequence, mid-latitude weather patterns (e.g., Francis et al. 2017; Garcia-Serrano et al. 2015; Kolstad and Screen 2019).

Heat associated with oceanic currents originating from lower latitudes provides an important, and year-round, source of heat to the Arctic Ocean (e.g., Carmack et al. 2015). The dominant external source of oceanic heat is the warm (temperature >0^{o}C) and salty water of Atlantic origin (Atlantic Water, AW) which is distributed throughout the deep basins at intermediate depths (~150-900 m, **Fig. 1**) and holds sufficient heat to melt the Arctic sea ice 3-4 times over (Carmack et al. 2015). Across much of the eastern (>70^{o}E) Eurasian Basin (EB) this heat is isolated from the surface, and hence the sea ice, by large vertical density gradients associated with the Arctic halocline (60-150 m, **Fig. 1**). The presence of the halocline impedes the transport of AW heat upward towards the surface across much of the Arctic Ocean (e.g., Fer 2009). The exception to this is the western (<70^{o}E) Nansen Basin where substantial turbulent mixing linked to the tides (Fer et al. 2010; Padman and Dillion 1991; Rippeth et al. 2015; Renner et al. 2019) and wind events (e.g. Provost et al., 2017; Graham et al., 2019) supports heat fluxes in excess of 50 W m^{-2}

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Inflowing AW is warming (Barton et al. 2018) driving a regime shift in sea ice cover over the past decade in the Barents Sea (Onarheim et al. 2018). There is also a growing body of evidence that the characteristics of the Arctic halocline are changing; for example, the halocline has weakened in the eastern EB since the 1970s (Steele and Boyd 1998; Polyakov et al. 2010). These changes have accelerated over the past decade (Polyakov et al. 2020a) with continuous time series from moored instruments capturing the significant weakening of the cold halocline layer (the upper part of the halocline with temperatures near freezing and negligible vertical temperature gradient) and shoaling of the AW in 2013-2015 (Polyakov et al. 2017).

The combination of weaker stratification and shoaling of the AW in the EB, coupled with the loss of sea ice, has allowed progressively deeper winter ventilation in the eastern EB in recent years (Polyakov et al. 2017). This process further enhances the annually averaged upward AW heat fluxes. The shift in sea ice state and upper ocean stratification to conditions previously unique to the western Nansen Basin has been termed 'atlantification' (Polyakov et al. 2017) and represents a transition toward a new Arctic climate state, in which the geographical influence of the AW heat on sea ice volume is spreading eastwards.

Since the increased oceanic heat fluxes associated with atlantification drive sea ice melt, and reduced sea ice increases oceanic heat fluxes through increased convective entrainment in winter, this process represents a positive ice/ocean-heat feedback mechanism. This mechanism is analogous and complementary to the ice-albedo feedback, in which atmospheric warming leads to a reduction of ice and snow coverage and decreasing albedo, resulting in further snow and sea ice retreat (Manabe and Stouffer 1980).

The strength of the ice/ocean-heat feedback is determined by the vertical flux of AW heat across the halocline into the surface-forced seasonal convective layer. Polyakov et al. (2017)

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estimated seasonal changes of heat content Q in the eastern EB halocline (65-130 m) and an equivalent divergent heat flux (the difference of fluxes at two depth levels for which a 1D equation of heat balance for a unit-area water column is integrated) of δF_h ~12 W m^{-2} over this depth range for winter 2013-2014, and ~8 W m^{-2} for winter 2014-2015. They argued that these inferred values of δF_h exceeded previous regional estimates (e.g., Lenn et al. 2009; Polyakov et al. 2013) by a factor of 2-4 and potentially account for an additional loss of up to 18-40 cm of sea ice over this period of time associated with the increase in upward AW heat transport. In consequence the impact of the oceanic heat flux on sea ice formation in 2013-2015 was comparable to that of the atmospheric thermodynamic forcing (Polyakov et al. 2017).

The aim of this paper is to quantify the changes in the upper ocean heat content, and the consequent release of heat from the AW up into the halocline and to the surface mixed layer in the key eastern Eurasian Basin of the Arctic Ocean. We improve on the Polyakov et al. (2017) study by including new data collected over the period 2015-2018 to quantify changes in the upper ocean heat content, and the consequent release of heat from the AW up into the halocline and to the surface mixed layer in the EB. We then compare these regional estimates with earlier estimates.

2. Data

Our analyses utilize observations of ocean temperature, salinity, and currents from moorings deployed in the eastern EB (**Fig. 2, Table 1**). Observations at the $M1_4$ mooring site began in August 2002, with several co-located moorings deployed and recovered annually prior to 2009, and longer duration of deployments since 2013 (**Table 1**).

Moorings deployed in summer 2013 and recovered in summer 2015 provided two-year long records for most instruments except for the $M1_5$ upper ocean Acoustic Doppler Current Profiler (ADCP), which worked for 10 months only. Mooring M3 located off Severnaya Zemlya, was

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deployed at water depth of 1350 m. Six moorings ($M1_1$-$M1_6$) formed a ~350-km cross-slope section spanning from the 250 m to 3400 m isobaths along 126°E. Topographically steered boundary current flows along slope across this section (Pnyushkov et al. 2015, 2018). Averaged over 2013-2015, the maximum current speed of ~11 cm/s was found at the shallowest mooring $M1_1$ (on the 250 m isobath), with only ~0.5 cm/s in the deep basin at moorings $M1_5$ and $M1_6$. The AW core defined by the maximum water temperature is typically located at the $M1_5$ mooring site at a depth of ~250 m.

Deployment of moorings in 2015-2018 repeated the mooring distribution used for 2013-2015 except that the $M1_6$ mooring was not re-deployed (**Table 1**). Almost all mooring instruments provided full three-year long records; the $M1_3$ McLane Moored Profiler (MMP) stopped recording after two years. In addition, a short-term mooring, $M1_{4\text{-short}}$, was deployed for 18 days only (September 2–20, 2018) close to the $M1_4$ climatologic mooring site (**Table 1**). The short-term mooring was designed to provide current and CTD data with the most rapid possible sampling rate in the upper 200 m.

Mooring Conductivity-Temperature-Depth (CTD) data: The MMP-based moorings at the $M1_4$ mooring location in 2002-2009 collected temperature, salinity and current velocity profiles once per day. Four 2013-2015 moorings ($M1_2$, $M1_3$, $M1_5$, and $M1_6$) and two 2015-2018 moorings ($M1_3$ and $M1_5$) provided vertical MMP profiles with two-day sampling interval with a ~0.25 m spacing. The MMPs on most moorings sampled the 50–700 m depth range; however, the 2015-2018 $M1_5$ mooring missed its target depth and the MMP record only reached to ~170 m below the surface. The MMP on the $M1_{4\text{-short}}$ mooring sampled about every 18 minutes and obtained 1369 profiles. The MMP temperature and conductivity calibrated measurement accuracies are ±0.002˚C and ±0.002 mS/cm.

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The moorings $M1_1$, $M1_2$, $M1_4$ and M3 with no MMP profilers deployed in 2013–2018, as well as mooring M1g, deployed about 12 km from mooring $M1_4$ in 2008 –2010 (**Fig. 2, Table 1**) were equipped with Seabird SBE-37 CTD instruments and provided records of conductivity, temperature and pressure with sampling interval of one hour or shorter, with measurement accuracies for temperature and conductivity of ±0.002˚C and ±0.003 mS/cm, respectively.

Mooring current data: Most moorings used in this analysis included 300 kHz Acoustic Doppler Current Profilers (ADCP) targeting the upper 50-60 m of the water column (**Table 1**). Moorings with no MMP were also equipped by long-range ADCP 75 kHz covering deeper layers (**Table 1**). ADCPs provided current velocities, averaged over 2-m (prior to 2013) or 4-m (after 2013) vertical cells, with 1-h time resolution. The manufacturer's estimates for ADCP accuracies are 0.5% of measured speed and 2° for current direction.

Moorings equipped with MMPs provided current velocity profiles with above mentioned profiling intervals and 0.25 m vertical resolution. The MMPs were equipped with a Falmouth Scientific Inc. (FSI) micro-CTD sensor in 2002-2004 and a Sea-Bird Electronics (SBE) 41CP CTD sensor starting from 2004, with temperature and conductivity measurement accuracies of about ±0.002˚C and ±0.0003 S/m, respectively. Prior to 2013, the MMPs carried the FSI Acoustic Current Meters (ACM); after 2013, the ACMs were substituted with the FSI ACM-PLUS-MP (http://www.falmouth.com/product-information.html). The velocity precision of the FSI ACM (ACM-PLUS-MP) carried on the MMP are reported to be ±2% (1%) of reading and ±0.5 cm/s for velocity resolution. Compass accuracy is ±2˚. All MMP sensors were calibrated before their deployment and immediately after their recovery using McLane facilities.

Ship-borne CTD data: Mooring observations were complemented by repeated hydrographic profiles collected using a Seabird SBE911plus CTD system in 2013, 2015 and 2018 at $M1_4$

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mooring site (**Fig. 2**). The effective vertical resolution, considering the different sensor characteristics, is about 25 cm. Individual temperature and conductivity measurements are accurate to ±0.002˚C and ±0.0003 S/m.

3. Methods

Defining a proxy for Richardson (Ri) number: *Ri* is a measure of the stability of the water column, *ie*. when $Ri < 0.25$ the vertical shear in the flow is sufficient to generate instability and turbulent mixing. As such *Ri* estimates provide a useful indicator for the likelihood of shear instability/ mixing. The correct scale for the estimation of *Ri* is the Ozmidov scale (which in this case we estimate to be O(0.1m)). However, the vertical resolution of the *Ri* estimate is limited by the positions of instruments on the moorings which have a vertical resolution of 20m. Whilst the 20 m *Ri* estimates are likely to smooth out the fine structure of individual instabilities, we argue that the smaller the large-scale *Ri* value is, the greater the likelihood of shear instability (and so turbulence and mixing). As such the 20m *Ri* provides a useful proxy for the likelihood of shear instability. Moreover, trends in the 20m (proxy) *Ri* estimate will expose trends in the likelihood of shear instability, the key interpretation here. This approach is supported by direct comparisons of dissipation and low resolution *Ri* estimates (e.g. Mead Silvester et al., 2014).

The mooring-based estimates of *Ri* (**Fig. 5**) are based on MMP measurements of stratification and velocity. Stratification over the 100-140 m layer is quantified using buoyancy frequency (N), $N^2 = -(g/\rho_o)\partial\rho/\partial z$, where ρ is the potential density of seawater, ρ_o is the reference density (1030 kg m^{-3}), and g is the acceleration due to gravity. The limited depth range of 100-140m was chosen due to insufficient data coverage in early years (see **Table 1**). The *Ri* proxy was estimated as $Ri = N^2/|\mathbf{U}_z|^2$, where $|\mathbf{U}_z|$ is the magnitude of the vertical shear of the horizontal

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currents; $|\mathbf{U}_z|$ and N were calculated averaging gradients over 20 m vertical scale for all points within the 100-140 m depth range.

Defining timing and depth of seasonal upper ocean ventilation and divergent heat flux δF_h *:* For this analysis, temperature observations carried out by $M1_2$, $M1_3$, $M1_4$ and M3 moorings in 2013–2018 were used. SBE-37 data from non-MMP moorings $M1_2$ (2015–2018), $M1_3$, $M1_4$ and M3 were complemented by MMP profiles from $M1_2$ (2013 –2015) mooring. SBE-37 observations were linearly interpolated to match the MMP vertical resolution. We are interested in the analysis of seasonal ventilation of the halocline. Accordingly, temperature observations were filtered using wavelet transformations to keep seasonal variations only (and thus different temporal sampling by MMP and SBE-37 did not affect our results). A standard package of wavelet programs was used based on the DOG Mother function. Estimates of heat content (Q, J/m^3, with freezing point taken as a reference temperature at a given salinity) for the halocline (65-140 m) are shown in **Fig. 6**. To assure that the use of SBE-37 point measurements with relatively coarse vertical resolution and continuous MMP profiles for estimates of Q did not affect our results we calculated Q using MMP temperature record from $M1_2$ mooring (2015–2018) twice, first time with original MMP resolution and another one with sub-sampled coarser resolution matching SBE-37 depth levels (**Table 1**). Results of Q integrated over the halocline depth range and averaged in time over the entire record length differed by 8%.

The aim is to define the timing and amplitude of upward heat flux associated with winter ventilation. To this end, we identified timing and amplitude of the maximum Q (as accumulated over the warm phase of the seasonal cycle) and the minimum of Q (associated with winter ventilation) using Q vertically integrated over 65-140 m. The depth of the ventilation is defined

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as the deepest point where a distinct minimum of Q was found. The maximum of vertically integrated Q was then re-calculated using the depth of ventilation.

Following Polyakov et al. (2017), we limited the boundary of the winter ventilation layer to 140 m. For some years, the boundary of the layer was deeper (as shown in **Fig. 6** by the black horizontal segments located at the very bottom of the panels with Q). Therefore, our choice of the ventilation layer is conservative and estimates of divergent heat fluxes δF_h derived from change of heat content ΔQ during each winter season represent the lower bound, consistent with the objectives of the study. For the upper boundary of the layer for which Q is estimated, we selected the depth 65 m, chosen because this best determines the halocline layer in which heat from the AW is stored and released (Polyakov et al. 2013, 2017). We evaluated the sensitivity of our estimates to the choice of the boundary of the ventilation layer by calculating δF_h for 65-140 m and 65-150 m layers. The 10 m increase in layer thickness increases δF_h by less than 8%.

Following Polyakov et al. (2013), we estimated δF_h (W/m^2) between two depth levels as the change, in time, of vertically integrated Q. This approach is based on the assumption that all change in heat content is due to vertical exchange (so 1D). Note that these values are flux *differences* between two depth levels, and total heat fluxes may be larger than these values due to additional non-divergent heat transports; thus, our inferred estimates of divergent heat fluxes represent *lower* bounds for the total heat flux (for details, see Polyakov et al. 2013).

4. Results

a. AW warming and weakening of halocline stratification in the eastern Eurasian Basin

Time series of the AW temperature show significant interannual variability (**Fig. 3a**). The AW in the eastern EB began warming in the early 2010s, with the AW temperature in 2018 being, on

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average, 0.5–0.7°C higher than in 2011 (**Fig. 3a**). This recent warming is particularly noticeable at shallower depths, with the increase in temperature at 150 m exceeding 1.5°C between 2011 and 2018. This warming over the depth range 150-750 m between September 2013 – May 2014 and September 2016 – May 2017 is partially associated with shoaling of the upper halocline boundary (**Fig. 3c**) and a substantial increase in AW layer thickness (**Fig. 4**).

Cross-correlation analysis of time series of AW temperature measured at 250m from 1997-2018 in Fram Strait, the entry point of AW into the Arctic, and from 2002-2018 in the eastern EB (red time series in **Fig. 3a**) shows the strongest correlation, $R = 0.67$, for a lag of 682 days (Fram Strait series leads, **Fig. 3b**). The fit between the two time series is better over the last 7-8 years than it is over the earlier period. The ~2 year lag suggests that warm pulses of AW that entered the Arctic Ocean through Fram Strait, are traveling towards the eastern EB at a speed 2-2.5 time faster than that estimated for a warm AW pulse which entered the eastern EB in 2004 (Polyakov et al. 2005). This implies that the rate of advection has increased over time. However, noisy data due to gaps in the EB record preclude meaningful statistical analysis using just the early part of the time series. Assuming that the lagged correlation between the two time series will persist in the near future, the latest part of the Fram Strait series (not shown) implies that the AW temperature in the eastern EB reached its peak in late 2018 (these data are not yet available) and will slowly decrease over the next 1-2 years.

Temperature and salinity profiles in the eastern EB from CTD during 2013-2018 and MMP during 2003-2018 recorded a decline of stratification (N^2) over the 110-140m depth range of the halocline (**Figs. 1c, 5a,b**) which may be a result of both the shoaling of AW and weakening of halocline stratification. Polyakov et al. (2018) used available potential energy defined for the variable-depth halocline to show overall weakening stratification in the EB since the 1980s, with

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accelerated tendencies in the 2010s compared with the 2000s. However, the substantial weakening of halocline stability from 2013 to 2015 (Polyakov et al. 2017) which continued in 2015–2018, and which was also partially associated with shoaling of the AW (**Fig. 4**) found at 80 m depth, as inferred from the most recent observations in winter 2017-2018 (**Figs. 3c**). This represents the shallowest depth the AW has been observed in the 15 years of mooring deployments. As these estimates used a linear interpolation of CTD time series made at 38m and 107m at mooring $M1_4$, we are not able to definitively conclude that the cold halocline layer was present (albeit very thin) during the winter of 2017–2018. However, the record suggests the extreme thinning (or even absence) of the Arctic cold halocline layer for several months at this time (**Figs. 3c, 4**) implying that AW heat was exposed to winter convection associated with sea ice formation and brine rejection.

b. Increased oceanic heat fluxes and ice loss in the eastern Eurasian Basin

The weakening stratification, shoaling of the AW layer and increase of current shear in recent years (e.g., Polyakov et al. 2020b) have altered the seasonal cycle of upward AW heat transport (**Fig. 6**). Estimated change in heat content (Q) from the halocline (65-140 m) during winter, averaged at four moorings, is equivalent to mean divergent heat fluxes (Section 3) of δF_h = 12.0±5.5, 3.5±2.2, 3.0±1.9, 12.9±1.7 and 20.6±6.8 W/m^2 for five winters from 2013-2014 through 2017-2018 (**Figs. 6, 7**). For three of these winters (2013-2014, 2016-2017, and 2017-2018), δF_h greatly exceeded (3- to 5-fold) the previous estimates derived from summer 2007-2008 microstructure observations over the Laptev Sea slope (Lenn et al. 2009; Polyakov et al. 2019) and winter 2009-2010 ITP-37 observations in the central Amundsen Basin (Polyakov et al. 2013). For the winters of 2014-2015 and 2015-2016, estimates of δF_h were comparable to upward heat fluxes of about 3-4 W m^{-2} from 2007-2008. We attribute the decrease of δF_h in

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2015-2016 (cf. Polyakov et al. 2017) to an anomalous freshening event in the upper ocean. This freshening is evident in data collected at mooring $M1_3$ (**Fig. 8)** which shows that strong upper (<75 m) ocean stratification (evidenced by high N^2 values) in 2016 precluded seasonal ventilation beyond the SML. Stronger stratification in winter 2015 (compared with winters of 2014 and 2017, **Fig. 8d**) limited seasonal ventilation to the upper ~115m, thus not extending deeply enough to reach the main pool of AW heat (**Fig. 8b**). In consequence the heat flux is limited. The strongest heat flux is inferred for winter 2017-2018 and is associated with the weakest stratification (**Fig. 5**), providing further evidence for the key role of stratification in mediating upper ocean ventilation.

The new estimates of seasonal ventilation of heat evaluated from the δF_h for the winter seasons of 2016-2017 and 2017-2018 are equivalent to 78±4 and 93±29 cm reductions in ice growth, respectively, for the eastern EB (**Fig. 7)**, given that one year of a heat flux of 1 W/m^2 in isolation is equivalent to about 10 cm of sea ice loss. This represents a two-fold increase in the sea ice loss rate compared to that estimated for 2013-14 (54 cm) and 2014-2015 (40 cm) (Polyakov et al. 2017), and so partially explains intensified eastern EB sea ice loss in more recent years (Onarheim et al. 2018; Stroeve and Notz 2018).

Time series from the shallower moorings ($M1_2$ and $M1_3$) show strong seasonal variations in the AW core temperature, which may be associated with seasonal displacement of the AW core relative to the slope (e.g., Baumann et al. 2018). However, the consistently low correlation between Q and the AW core temperature records, for all mooring sites (**Fig. 9**), implies that cross-slope shifts in AW temperature core are not a major driver of the seasonal variation in Q in the halocline. The correlation between Q and AW core temperature at the shallowest mooring ($M1_2$) where currents are strongest is also weak (R = 0.29) indicating that advection does not

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provide a significant contribution to the seasonal variability of *Q*. This evidence is consistent with the results of Polyakov et al. (2017) who argued that the in-phase seasonal maxima and minima of wavelet transforms of *Q* at all mooring sites suggests that the observed winter ventilation is driven by surface cooling and sea-ice formation—and not by lateral advection. They reasoned that spatially varying water transports across the slope, ranging from 13 cm/s (measured over the upper continental slope (250-700 m) by moorings $M1_1$ and $M1_2$) to 1-2 cm/s (measured at 2700 m and deeper, at mooring locations $M1_4$, $M1_5$ and $M1_6$) make the in-phase pattern of the seasonal signal at all moorings impossible to explain using the advective mechanism. Furthermore, mooring $M1_6$ which was farthest from the near-slope boundary current, in the ocean interior, yielded estimates for F_h which magnitudes and phases are consistent with estimates from the other moorings deployed on the eastern EB continental slope in 2013–2015 (Polyakov et al., 2017).

The one-dimensional approach adopted here can be further validated by considering the magnitude of the lateral temperature gradient necessary to explain the estimated heat flux, if advection were to dominate. In assuming an along slope current speed of 2 cm/s requires that the lateral temperature gradient dT/dx must be five times larger than that observed Fram Strait – central Laptev Sea slope temperature decrease of 1.8ºC [= 3.0 – 1.2] over ~2400km so dT/dx = 0.75×10^{-3} ºC/km to explain the estimated heat flux. Another potential contributor to the observed ventilation rates are lateral eddy fluxes. Ventilation of halocline by eddies is, however, difficult to quantify using available data. Nevertheless, considering that the typical time of eddy passing across the mooring site is about a week with the average frequency about one eddy per month (Pnyushkov et al., 2018b), it is unlikely that eddies can significantly contribute to changes

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of the heat content at seasonal time scales. These considerations imply the uncertainty in the 1D flux calculation from lateral advection and diffusion is small.

5. Discussion and Conclusions

Time series measurements from a 15-year mooring record in the eastern EB of the Arctic Ocean demonstrate that the previously identified weakening of stratification over the halocline, which isolates intermediate depth AW from the sea surface, over the period 2003 - 2015 (e.g., Polyakov et al. 2017, 2018), has continued at an increasing rate in more recent years (2015-2018). In consequence, oceanic heat fluxes for the winters of 2016-2018 are estimated to be greater than 10 Wm^{-2}. These fluxes are substantially larger than the previously reported winter estimates for the region for 2007-2008 of 3-4 Wm^{-2} (Lenn et al., 2009; Polyakov et al., 2019), and comparable to the estimates for the winters of 2013-2015 (Polyakov et al. 2017), implying a significant enhancement of the role of oceanic heat in this region in recent years.

Moreover, the increased vertical heat fluxes have been accompanied by increased upper-ocean current speeds $|\mathbf{U}|$ and the vertical shear in the horizontal velocities $|\mathbf{U}_z|$ over the period 2015–2018 (Polyakov et al. 2020b). Using mooring observations from 2003 to 2018, these authors showed that $|\mathbf{U}|$ and $|\mathbf{U}_z|$ in the upper 60 m of the water column increased by about 20% and 40%, respectively. In the lower halocline (110-140 m), $|\mathbf{U}|$ was generally larger after 2008, increasing on average from 2.5-3.5 cm/s in 2003–2008 to about 4-5 cm/s in 2009–2018 (**Fig. 5c,d**) although the change was not as strong in very recent years, 2016 and 2018 when compared to 2009–2015. There is also clear transition in $|Uz|$, with significantly larger shears evident post-2010, and in particular in the summer of 2018 (**Fig. 5c,d**). However, Pnyushkov et al. (2018) found no significant change in the mean along-slope water transport over the same period.

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The combination of reduced stratification and increased shear implies a decreases the gradient Richardson number, *Ri*, defined in section 3 (**Fig. 5e,f**), consistent with an increased turbulent heat flux, associated with vertical mixing by shear instabilities. Although the *Ri* estimates are based on 20m vertical resolution measurements, they show a clear trend towards reduced dynamic stability which may be interpreted as a tendency towards increased turbulent mixing in recent years, coincident with the increase in maximum halocline heat content (**Fig. 6**). This tendency is particularly strong in 2018 with amplified velocity shear in the relatively weakly stratified upper ocean (**Fig. 5**).

The increased shear and weakening of stratification as prerequisites for enhanced turbulent mixing are consistent with the recent transition in the upper ocean to conditions previously unique to the western Nansen Basin, a process called ‘atlantification’ (Polyakov et al. 2017). Our analyses confirm that, in part, the loss of stratification in the eastern EB halocline can be attributed to processes originating upstream. For example, the change in halocline salinity, the main contributor to water column stability in the eastern EB, is correlated with upper ocean salinity changes in the northern Barents Sea with a lag of approximately 2 years (**Fig. 10**) (Lind et al. 2018), revealing coherent interannual variability between the two regions. In the Barents Sea, these changes were found to be closely linked to declines in sea ice imports to the Barents Sea (Lind et al., 2018; Barton et al., 2018). The shift towards higher salinities in the eastern EB lag the changes in the northern Barents Sea by about 1 year (**Fig. 10**), implying an eastward lateral progression of the ‘atlantification’. Shelf-basin interactions may also be contributing to the observed warming (e.g. Timmermans et al., 2018).

Our observations point to the shift of this region of the eastern Arctic Ocean towards a new regime that is more typical of the continental slope regions of the western Nansen Basin where

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surface conditions are strongly influenced by oceanic heat imported from the Atlantic Ocean (**Fig. 11**). The flux of AW heat to the sea ice cover and the atmosphere has increased, during the winter season, from an average of 3-4 W/m^2 in 2007-2008 to >10 W/m^2 in 2016-2018, equivalent to more than a two-fold reduction of winter ice growth over the last decade.

The process described here represents a positive feedback, analogous to the ice-albedo feedback, since increased ocean heat flux to the sea surface reduces ice thickness and increases its mobility, increasing atmospheric momentum flux into the ocean and reducing the damping of surface-intensified baroclinic tides (Carr et al., 2019). We refer to this process as the "ice/ocean-heat" feedback. As with the ice-albedo feedback, the contribution of the ice/ocean-heat feedback to long-term sea ice trends depends on the seasonal variability of several factors that affect mixing rates including sea ice concentration and thickness, baroclinic tidal response to seasonally varying stratification, and wind stress impacts on sea ice and on AW shoaling. The transition in dominant mixing regime from double diffusion to shear-driven mixing also affects the relative magnitudes of buoyancy fluxes due to heat and salinity transports; the vertical diffusivities for heat and salt are the same in shear-driven turbulence, but are different for double diffusion (Kelley, 1984). Coincident vertical nutrient fluxes, which support oceanic primary productivity, food web structure and carbon export from the atmosphere to the seabed (Bluhm et al. 2015; Falk-Peterson et al. 2015), will also increase. Moreover, the nutricline has shoaled in recent years (relieving nutrient limitations, **Fig. 1d**) which coupled with declining sea ice cover (relieving light limitations), both influenced by atlantificiation, and so could lead to regional-scale enhancement of biological productivity in the central Arctic Ocean.

As ice thins – through atmospheric forcing, changing ocean heat fluxes, and feedbacks – upper-ocean stratification is responding and a new Arctic state is emerging which may not be

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easily reversed. For example, a large anomaly in AW heat input coupled with shoaling may lead, through the ice/ocean-heat feedback, to an expanding and more permanent Atlantic-dominated state wherein the hydrographic structure of the halocline no longer provides sufficient insulation between the intermediate depth AW and the sea ice, even when the heat flux associated with the AW is relaxed. This potential for a permanent transition of the eastern Arctic to a new state, emphases the pressing need for the incorporation of improved mixing schemes into Arctic climate models in order that they better simulate the evolving halocline stratification and its impact on sea ice state.

Appendix A1. Building Long-Term Time Series

Changes in the 110-140 m (halocline) layer at the $M1_4$ mooring site shown in **Fig. 5** were documented using MMP records for 2003-2007 and 2013-2018, SBE37 records from M1g mooring in 2008–2010, and ADCP records for 2008-2010. This layer is the key part of the lower halocline water (**Fig. 1a,b**) and has sufficient data coverage for the task. All original mooring data were processed to make them comparable. We filtered MMP vertical profiles with a running-mean filter to reduce resolution to 4 m, equivalent to the 2013-2018 ADCP observations. We subsampled ADCP and SBE37 data in time to match coarser MMP temporal resolution. The vertical shear is calculated consistently using gradients over 20 m vertical scale. Reconstruction of the record at the $M1_4$ mooring site in 2013-2018 using MMP data from nearby moorings is described below.

There were no MMP measurements within the 110-140m depth range at the $M1_4$ mooring in 2013-2015 and 2015-2018 (**Table 1**). Records for these years and depth range were reconstructed using weighted interpolated estimates from the neighboring $M1_3$ and $M1_5$ moorings.

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This approach is justified by the observed monotonic cross-slope change of current speed from $M1_3$, $M1_4$, and $M1_5$ mooring records for the depth ranges where overlapping data is available for the three moorings (**Fig. A1**). Estimates of buoyancy frequency *N* derived from temperature and salinity provided by these three moorings are statistically indistinguishable (**Fig. A1**).

Multiple regression is used to further validate the use of records from moorings $M1_3$ and $M1_5$ to reconstruct time series of temperature, salinity, and current speed at mooring $M1_4$ for 2013-2018. The model of multiple regression is

$$Y = \beta_o + \beta_1 X_1 + \beta_2 X_2 \,, \tag{1}$$

where $\beta_1 = \frac{r_{YX_1} - r_{YX_2} r_{X_1X_2}}{1 - r^2_{X_1X_2}} \frac{\sigma_Y}{\sigma_{X_1}}$, $\beta_2 = \frac{r_{YX_2} - r_{YX_1} r_{X_1X_2}}{1 - r^2_{X_1X_2}} \frac{\sigma_Y}{\sigma_{X_2}}$, and $\beta_o = \overline{Y} - \beta_1 \overline{X}_1 + \beta_2 \overline{X}_2$, overbar denotes means, σ denotes standard deviations, *r* is used to denote cross-correlation coefficients, and random error term is neglected. For independent parameters X_1 and X_2 time series from $M1_3$ and $M1_5$ moorings are used, time series from $M1_4$ is used as the dependent variable *Y*. We neglected the high-frequency part of the records by applying low-pass three-month running mean filtering to each time series used in the tests because in this study we mainly focus on longer-term (interannual) trends and changes. Evidence for the validity of this approach is provided in **Fig. A2**.

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Figure legends

Figure 1: Vertical profiles of (a) potential temperature θ, (b) salinity S, (c) the logarithm of squared Brunt-Väisälä frequency (N^2, s^{-2}, a measure of water column stability; 5-point smoothing is applied) and (d) nutrients at $M1_4$ mooring site made on August 27, 2013, September 20, 2015, and September 2, 2018. Circulation of the intermediate Atlantic Water (AW) in the Arctic Ocean is shown schematically in (e) by red arrows. In (a), the upper part of halocline is cold halocline layer (CHL) in which salinity increases with depth while temperature remains near the freezing point. The blue box indicates the area of the Arctic Ocean with mooring positions shown in **Fig. 2**. The Canada Basin (CB), Chukchi Sea (CS), East Siberian Sea (ESS), and Barents Sea (BS) are indicated.

Figure 2: Map showing the focus of the study together with the positions of moorings and location of CTD (Conductivity-Temperature-Depth) profiles made in summer 2013, 2015, and 2018 reported in this study. The Gakkel Ridge (GR) divides the Eurasian Basin (EB) into the Nansen Basin and the Amundsen Basin. The Lomonosov Ridge (LR), Novosibirskiye Islands (NI), Severnaya Zemlya (SZ), Franz Joseph Land (FJL), and Makarov Basin (MB) are indicated. Grey solid lines show depth in meters. The eastern EB region used for calculation of blue time series in **Fig. 10** is identified by green line.

Figure 3: Composite 2002 –2018 time series of (a) monthly mean potential water temperature (θ) and (c) daily depth of the lower halocline boundary (H_{base}) defined by 0°C isotherm at $M1_4$ mooring location (for location, see **Fig. 2**). (b) Comparison of de-seasoned monthly mean time series of normalized θ anomalies from 250m of M14 mooring of the eastern EB (EEB) and lagged by 678 days (as obtained from correlation analysis) F2-F3 moorings of Fram Strait; time series are normalized by their standard deviations.

Figure 4: Depth–time diagram of potential temperature θ (°C) from $M1_3$ mooring. Black lines show the depth of the halocline base and lower Atlantic Water boundary both defined by 0°C isotherms.

Figure 5: Estimates of (left) annual and (right) summer mean (a,b) squared buoyancy frequency N^2 (10^5 s^{-2}), (c,d) current magnitude $|\mathbf{U}|$ and squared vertical shear of horizontal currents $|\mathbf{U}_z|^2$, and (e,f) proxy of Richardson number Ri for the 110-140 m depth range for the $M1_4$ mooring location. Statistical significance of means is shown at the 95% confidence level.

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Figure 6: (Left and middle) 65-140m layer depth versus time of water temperature and annual component of heat content *Q*. Annual components are obtained via band-pass filtering using wavelet transformations. Horizontal black segments identify the depth of seasonal ventilation; dates identified by their ends are used to compute vertically integrated *Q* shown in the lower parts of panels in the right column. (Right) Vertically integrated *Q* for the beginning (warm phase) and end (cold phase) of seasonal ventilation (lower parts of the panels) and divergent heat fluxes δF_h (upper parts) for four moorings.

Figure 7: Time-averaged over M3, $M1_2$, $M1_3$, and $M1_4$ mooring records (top) vertically integrated *Q* for the beginning (warm phase, red bars, Q_{max}) and end (cold phase, blue bars, Q_{min}) of seasonal ventilation of eastern EB halocline (110-140m), (middle) divergent heat fluxes δF_h (blue bars for averages with ±1 standard error shown as black segments), and (bottom) equivalent sea ice thickness losses.

Figure 8: (a) Potential temperature, (b) annual component of heat content *Q* obtained by band-pass filtering of daily heat content using wavelet spectra; horizontal black segments identify the depth of seasonal ventilation, (c) salinity, and (d) squared buoyancy frequency for $M1_3$ mooring.

Figure 9: (left) Depth versus time diagram of potential water temperature θ (^{o}C) and (right) time series of monthly heat content *Q* for the 65-140m layer (blue) and AW core temperature (red) for four moorings. Low correlations between these time series $R_{Q\text{-}\theta}$ suggest that changes of *Q* are not related to seasonal shift of AW core relative to the slope.

Figure 10: Normalized (reduced to anomalies and divided by one standard deviation SD) annual time series of (blue) halocline salinity *S* in the eastern EB (EEB, from Polyakov et al. 2018) and (red) lagged by one year (as obtained from correlation analysis) upper ocean *S* from the northern Barents Sea (from Lind et al. 2018). Dash-dotted lines are used to fill gaps (interpolated values are *not* used for statistical estimates). Means and SDs are indicated. Trends are shown by dashed lines; all trends are statistically significant at the 95% confidence according to the Student *t* test. The break-point in 1999 separates periods with opposite trends.

Figure 11: Conceptual model of shift of the mixing regime in the eastern EB in recent years and associated suite of processes and state conditions including: 1) thinner, more mobile ice, 2) warmer surface mixed layer (SML), 3) weakening / retreat of cold halocline (HC) layer, 4) increased AW vertical heat flux (red arrows) and horizontal currents and their vertical shear

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(blue arrows), 5) shoaling of upper AW boundary, and 6) replacement of DD by shear instabilities as the fundamental mechanism of vertical flux.

Figure A1: 2013/15 mean estimates of (top) squared Brunt-Väisälä frequency N^2, and current speed |**U**| for (middle) 20-60 m and (bottom) 190-230 m depth ranges where the mooring records from $M1_3$, $M1_4$, and $M1_5$ overlap. Statistical significance of estimates for means is shown at the 95% confidence level.

Figure A2: Multiple regression reconstruction of (a,b) salinity and (c,d) current speed |**U**| at $M1_4$ mooring site using data from $M1_3$ and $M1_5$ moorings for 170-210m depth range. (a,c) Daily (dotted) and three-month running mean smoothed time series of (a) salinity and (c) |**U**| from $M1_3$, $M1_4$, and $M1_5$ moorings. (b,d) Original (blue) and reconstructed (red) time series of (b) salinity and (d) |**U**| from $M1_4$ mooring. Relatively high correlations between the original and reconstructed time series attests of good quality of reconstruction.

Figures

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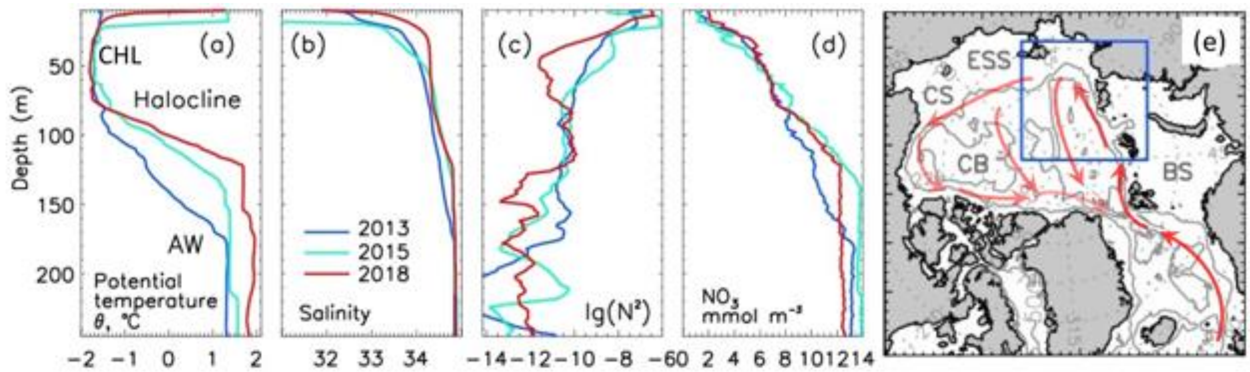

Figure 1: Vertical profiles of (a) potential temperature θ, (b) salinity S, (c) the logarithm of squared Brunt-Väisälä frequency (N^2, s^{-2}, a measure of water column stability; 5-point smoothing is applied) and (d) nutrients at $M1_4$ mooring site made on August 27, 2013, September 20, 2015, and September 2, 2018. Circulation of the intermediate Atlantic Water (AW) in the Arctic Ocean is shown schematically in (e) by red arrows. In (a), the upper part of halocline is cold halocline layer (CHL) in which salinity increases with depth while temperature remains near the freezing point. The blue box indicates the area of the Arctic Ocean with mooring positions shown in **Fig. 2**. The Canada Basin (CB), Chukchi Sea (CS), East Siberian Sea (ESS), and Barents Sea (BS) are indicated.

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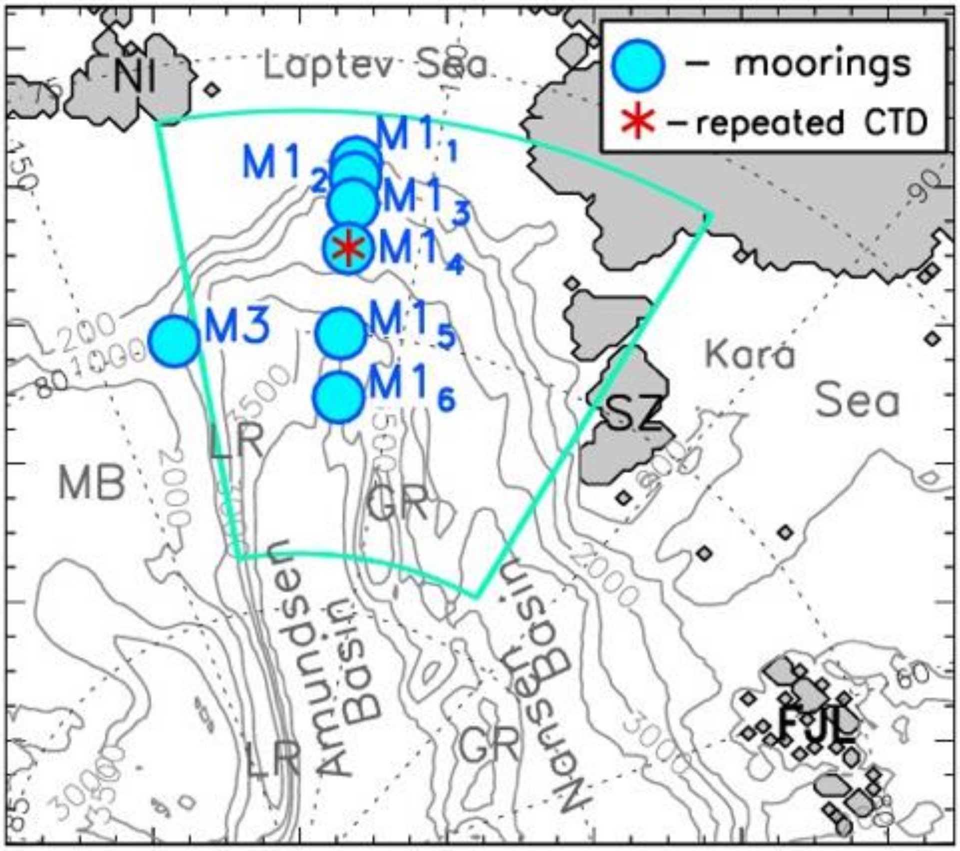

Figure 2: Map showing the focus of the study together with the positions of moorings and location of CTD (Conductivity-Temperature-Depth) profiles made in summer 2013, 2015, and 2018 reported in this study. The Gakkel Ridge (GR) divides the Eurasian Basin (EB) into the Nansen Basin and the Amundsen Basin. The Lomonosov Ridge (LR), Novosibirskiye Islands (NI), Severnaya Zemlya (SZ), Franz Joseph Land (FJL), and Makarov Basin (MB) are indicated. Grey solid lines show depth in meters. The eastern EB region used for calculation of blue time series in **Fig. 10** is identified by green line.

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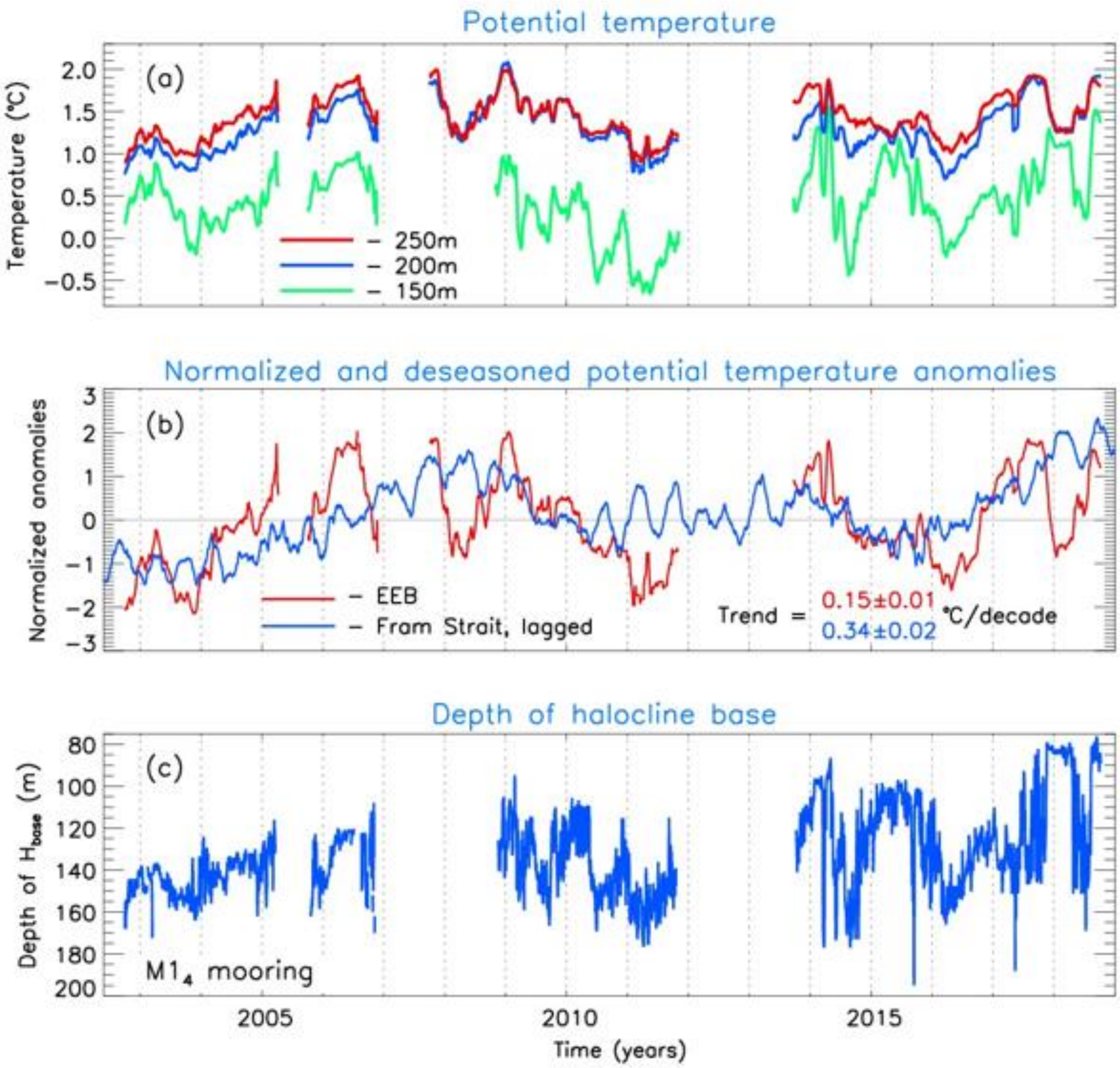

Figure 3: Composite 2002 –2018 time series of (a) monthly mean potential water temperature (θ) and (c) daily depth of the lower halocline boundary (H_{base}) defined by 0°C isotherm at M1$_4$ mooring location (for location, see **Fig. 2**). (b) Comparison of de-seasoned monthly mean time series of normalized θ anomalies from 250m of M14 mooring of the eastern EB (EEB) and lagged by 678 days (as obtained from correlation analysis) F2-F3 moorings of Fram Strait; time series are normalized by their standard deviations.

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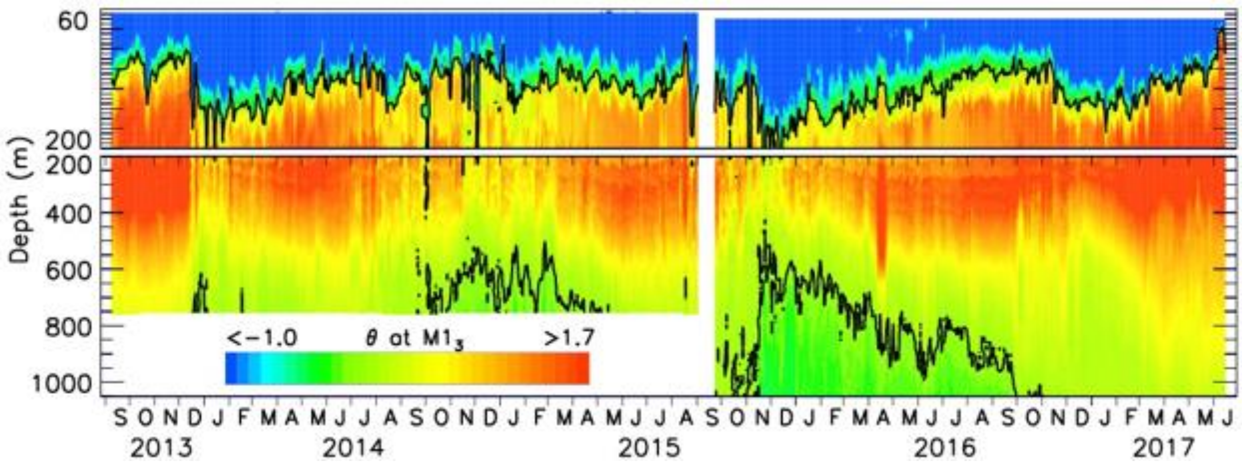

Figure 4: Depth–time diagram of potential temperature θ (°C) from $M1_3$ mooring. Black lines show the depth of the halocline base and lower Atlantic Water boundary both defined by 0°C isotherms.

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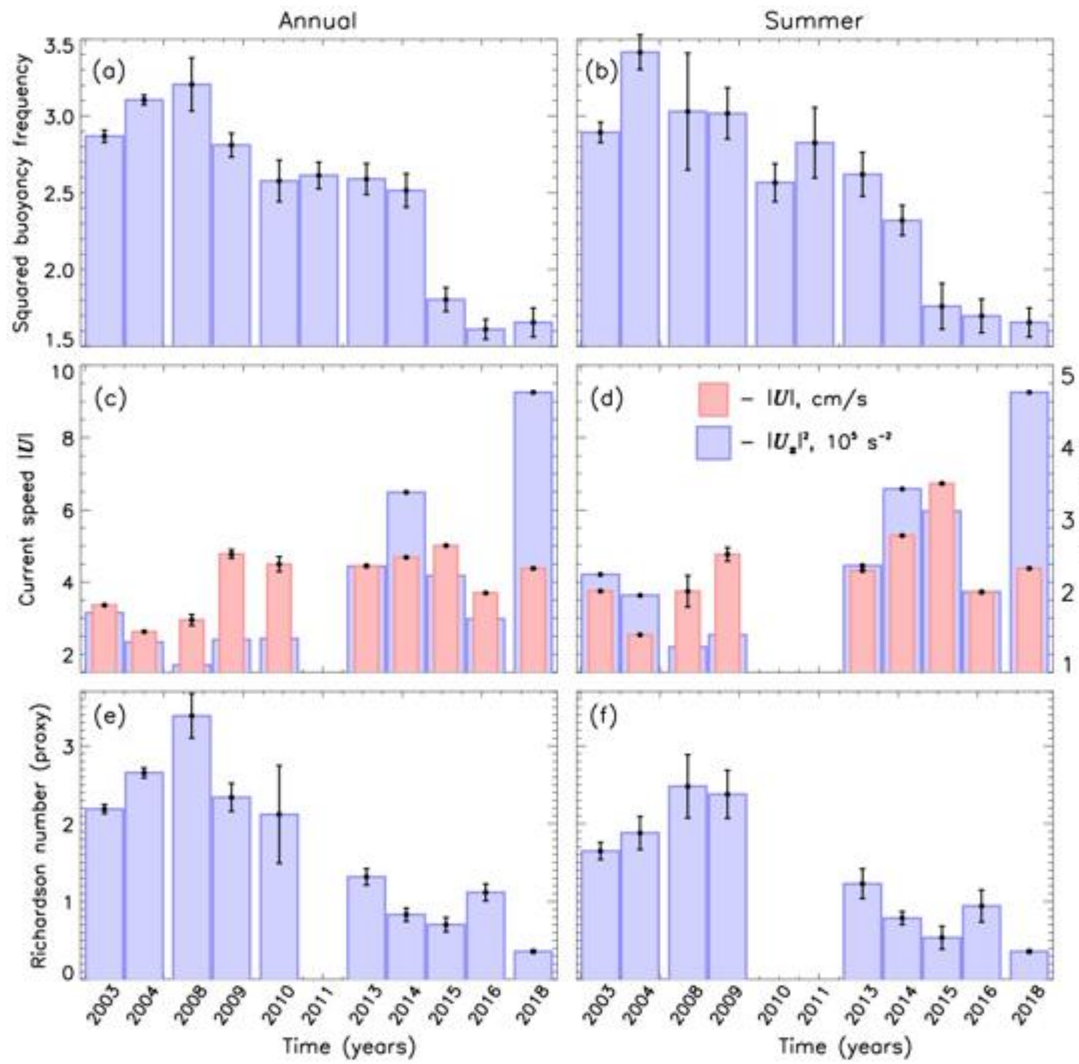

Figure 5: Estimates of (left) annual and (right) summer mean (a,b) squared buoyancy frequency N^2 (10^5 s^{-2}), (c,d) current magnitude $|\mathbf{U}|$ and squared vertical shear of horizontal currents $|\mathbf{U}_z|^2$, and (e,f) proxy of Richardson number Ri for the 110-140 m depth range for the M1$_4$ mooring location. Statistical significance of means is shown at the 95% confidence level.

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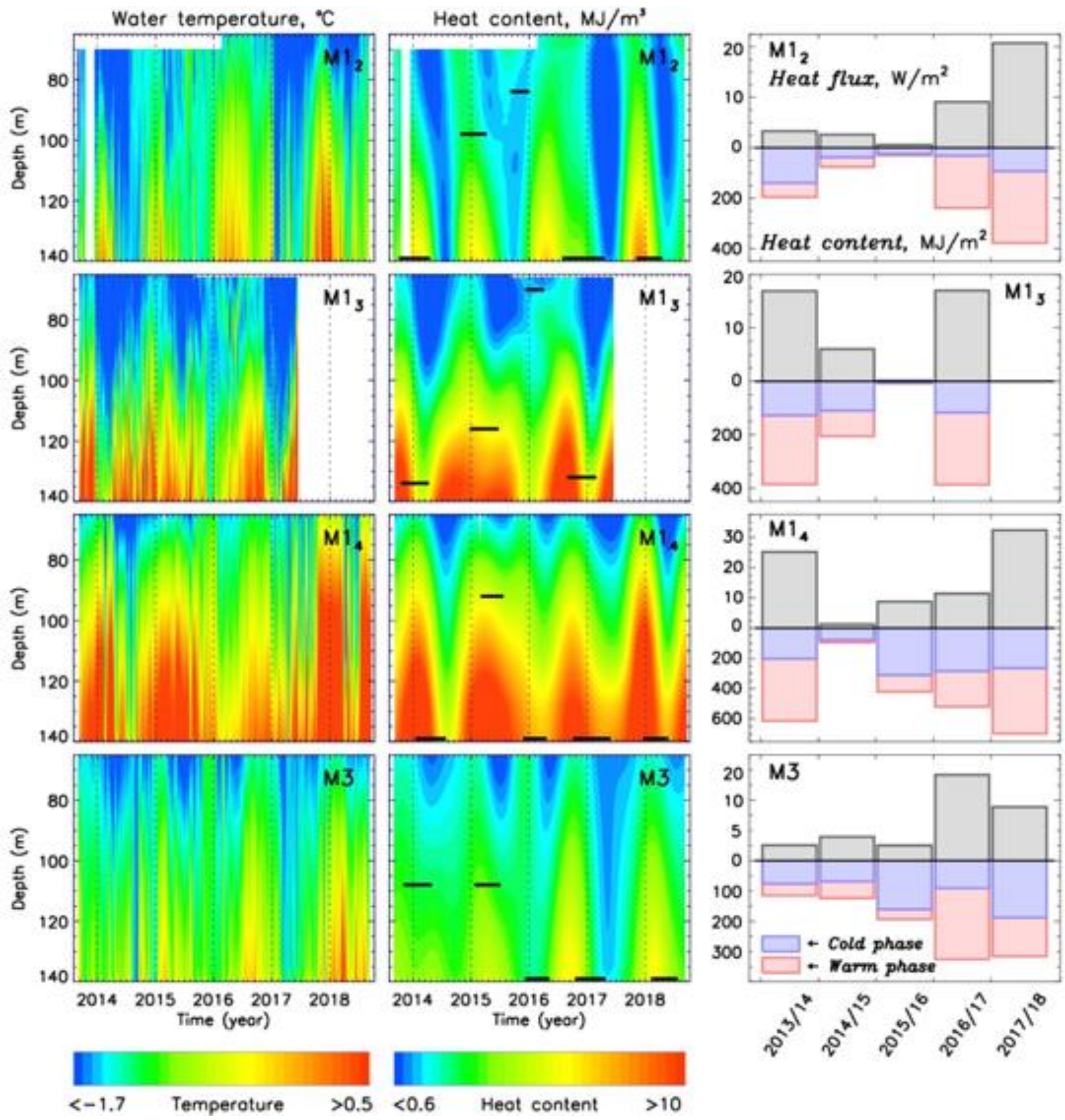

Figure 6: (Left and middle) 65-140m layer depth versus time of water temperature and annual component of heat content *Q*. Annual components are obtained via band-pass filtering using wavelet transformations. Horizontal black segments identify the depth of seasonal ventilation; dates identified by their ends are used to compute vertically integrated *Q* shown in the lower parts of panels in the right column. (Right) Vertically integrated *Q* for the beginning (warm phase) and end (cold phase) of seasonal ventilation (lower parts of the panels) and divergent heat fluxes δF_h (upper parts) for four moorings.

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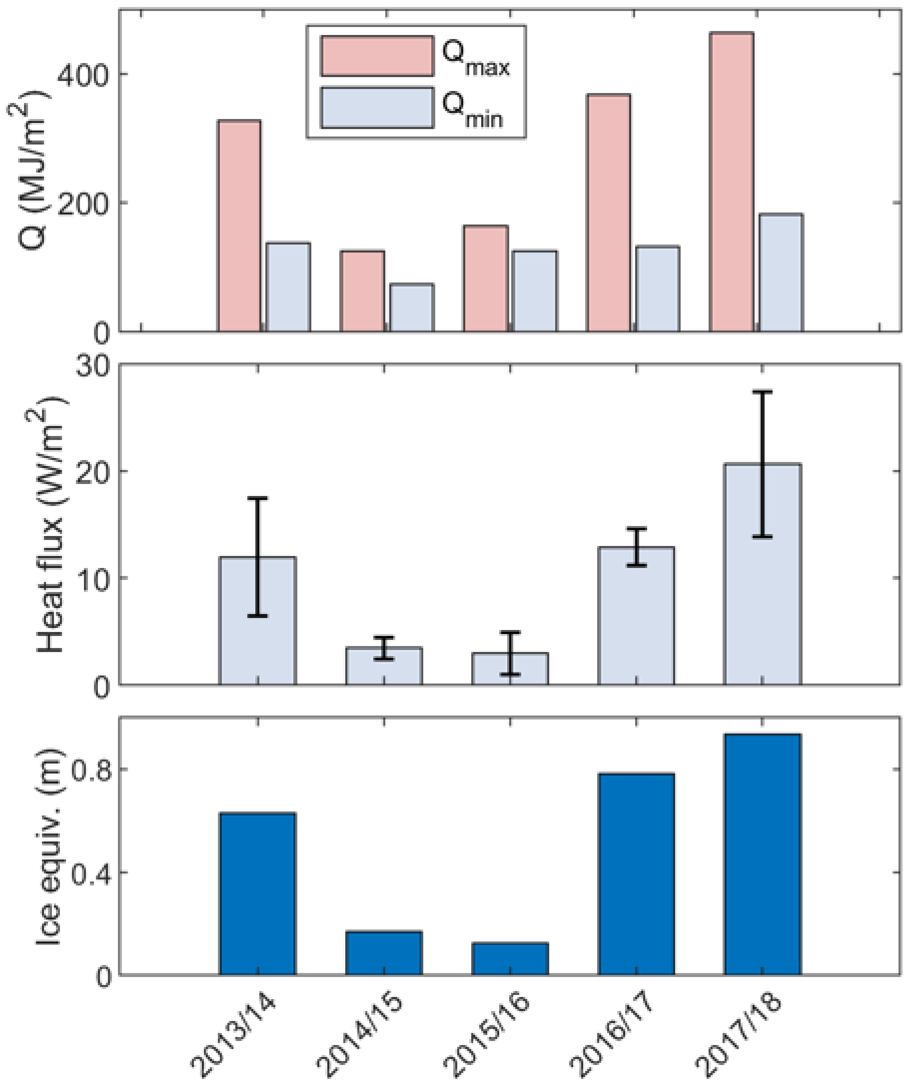

Figure 7: Time-averaged over M3, $M1_2$, $M1_3$, and $M1_4$ mooring records (top) vertically integrated Q for the beginning (warm phase, red bars, Q_{max}) and end (cold phase, blue bars, Q_{min}) of seasonal ventilation of eastern EB halocline (110-140m), (middle) divergent heat fluxes δF_h (blue bars for averages with ±1 standard error shown as black segments), and (bottom) equivalent sea ice thickness losses.

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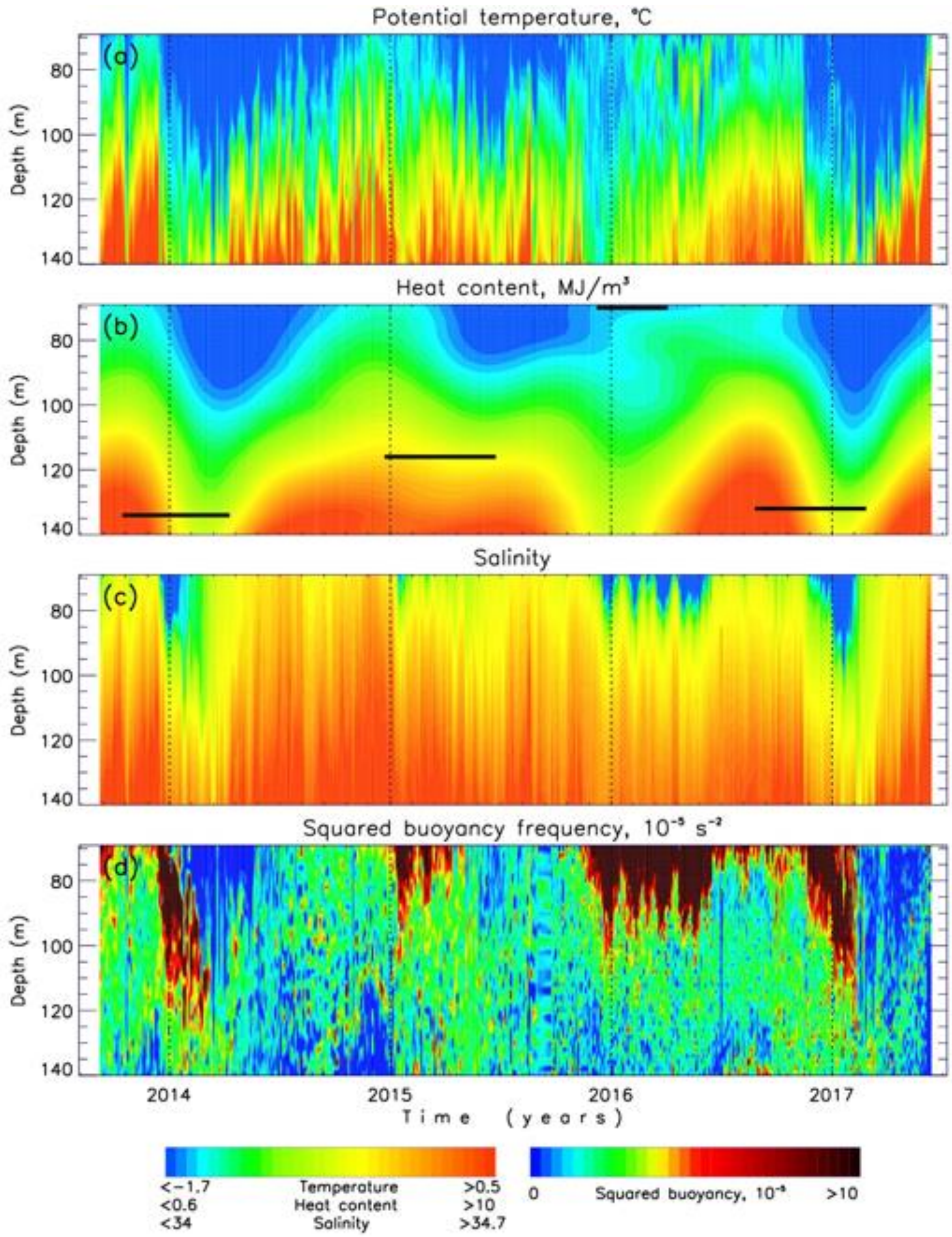

Figure 8: (a) Potential temperature, (b) annual component of heat content *Q* obtained by band-pass filtering of daily heat content using wavelet spectra; horizontal black segments identify the depth of seasonal ventilation, (c) salinity, and (d) squared buoyancy frequency for $M1_3$ mooring.

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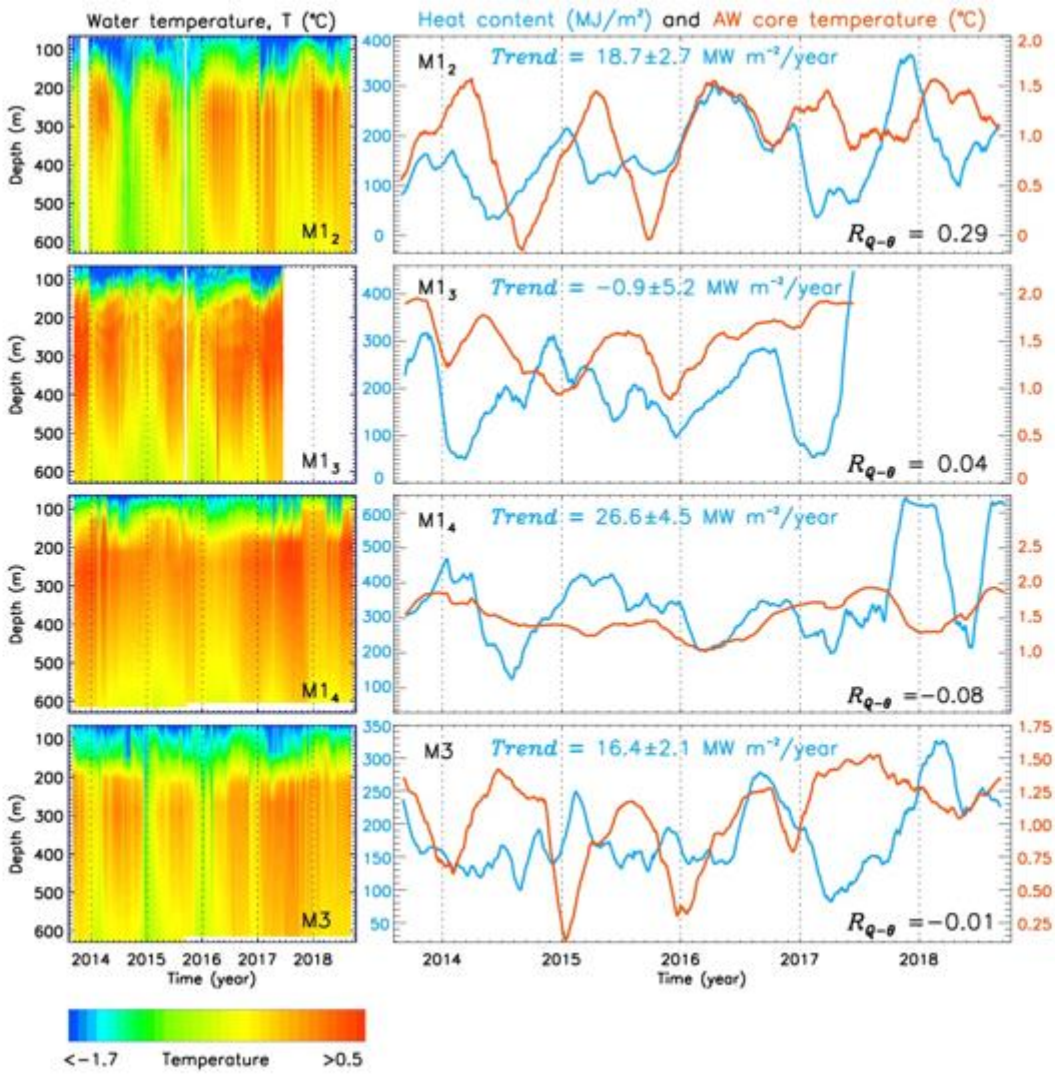

Figure 9: (left) Depth versus time diagram of potential water temperature θ (^{o}C) and (right) time series of monthly heat content Q for the 65-140m layer (blue) and AW core temperature (red) for four moorings. Low correlations between these time series $R_{Q\text{-}\theta}$ suggest that changes of Q are not related to seasonal shift of AW core relative to the slope.

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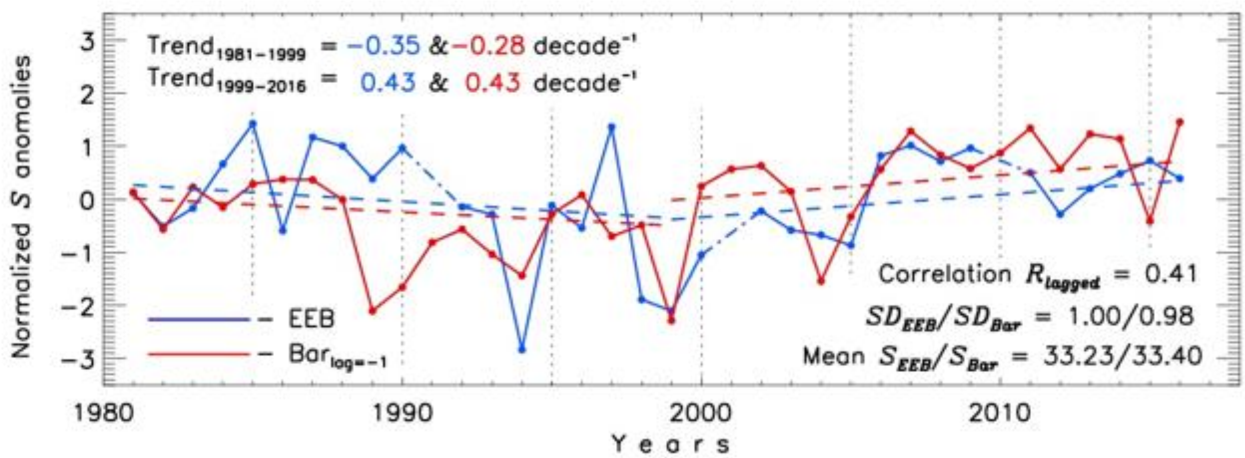

Figure 10: Normalized (reduced to anomalies and divided by one standard deviation SD) annual time series of (blue) halocline salinity *S* in the eastern EB (EEB, from Polyakov et al. 2018) and (red) lagged by one year (as obtained from correlation analysis) upper ocean *S* from the northern Barents Sea (from Lind et al. 2018). Dash-dotted lines are used to fill gaps (interpolated values are *not* used for statistical estimates). Means and SDs are indicated. Trends are shown by dashed lines; all trends are statistically significant at the 95% confidence according to the Student *t* test. The break-point in 1999 separates periods with opposite trends.

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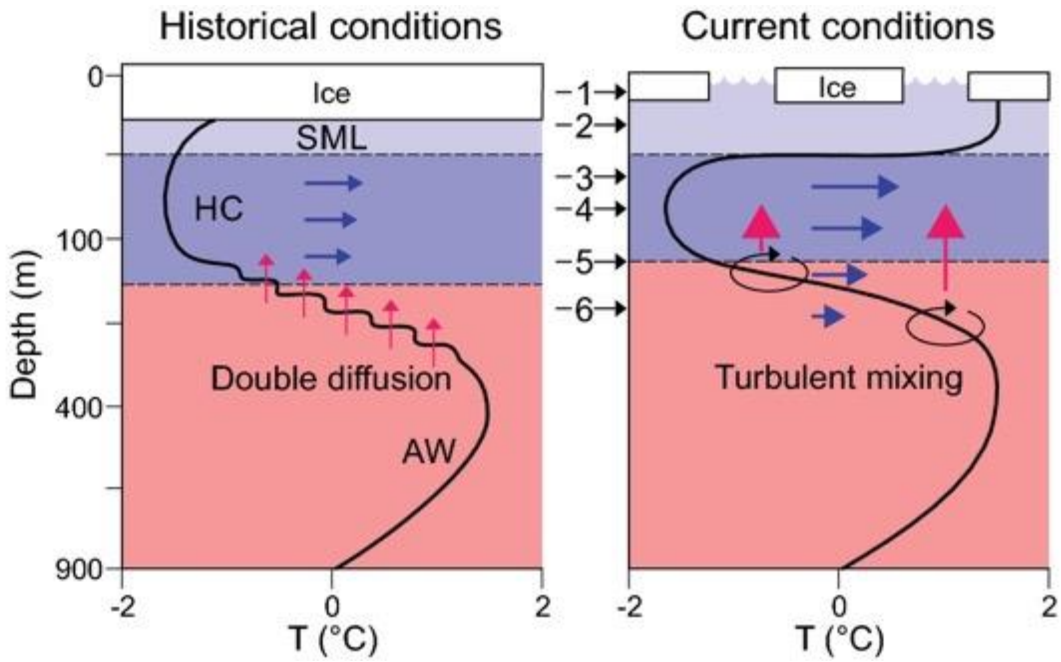

Figure 11: Conceptual model of shift of the mixing regime in the eastern EB in recent years and associated suite of processes and state conditions including: 1) thinner, more mobile ice, 2) warmer surface mixed layer (SML), 3) weakening / retreat of cold halocline (HC) layer, 4) increased AW vertical heat flux (red arrows) and horizontal currents and their vertical shear (blue arrows), 5) shoaling of upper AW boundary, and 6) replacement of DD by shear instabilities as the fundamental mechanism of vertical flux.

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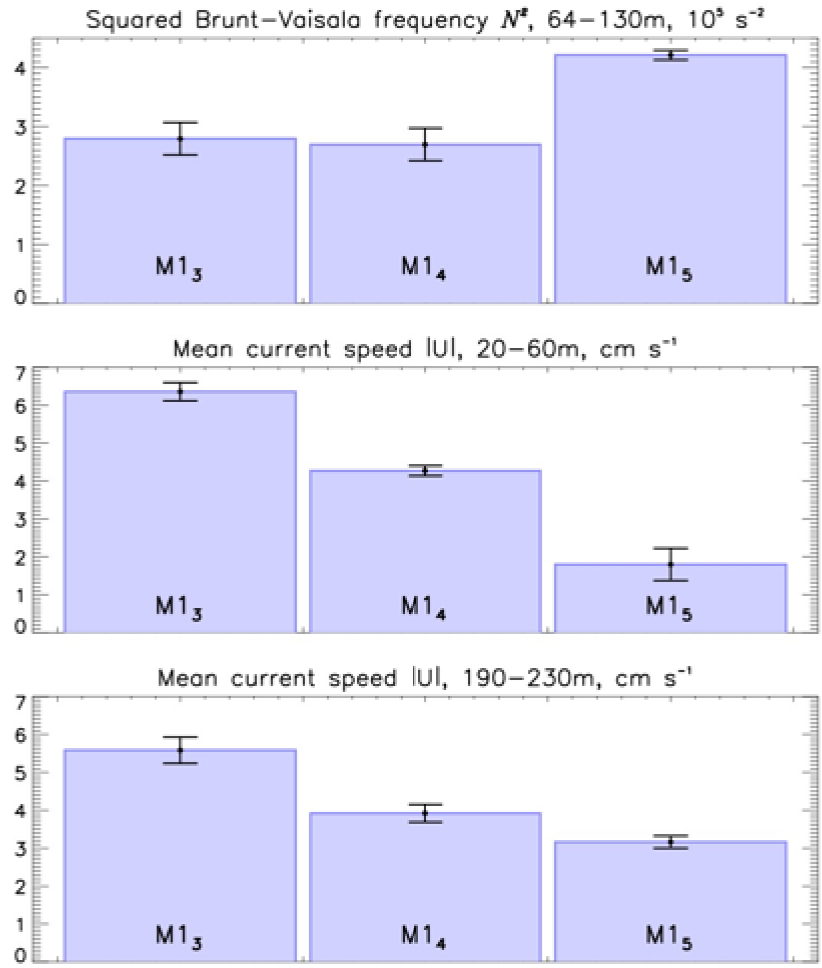

Figure A1: 2013/15 mean estimates of (top) squared Brunt-Väisälä frequency N^2, and current speed $|\mathbf{U}|$ for (middle) 20-60 m and (bottom) 190-230 m depth ranges where the mooring records from $M1_3$, $M1_4$, and $M1_5$ overlap. Statistical significance of estimates for means is shown at the 95% confidence level.

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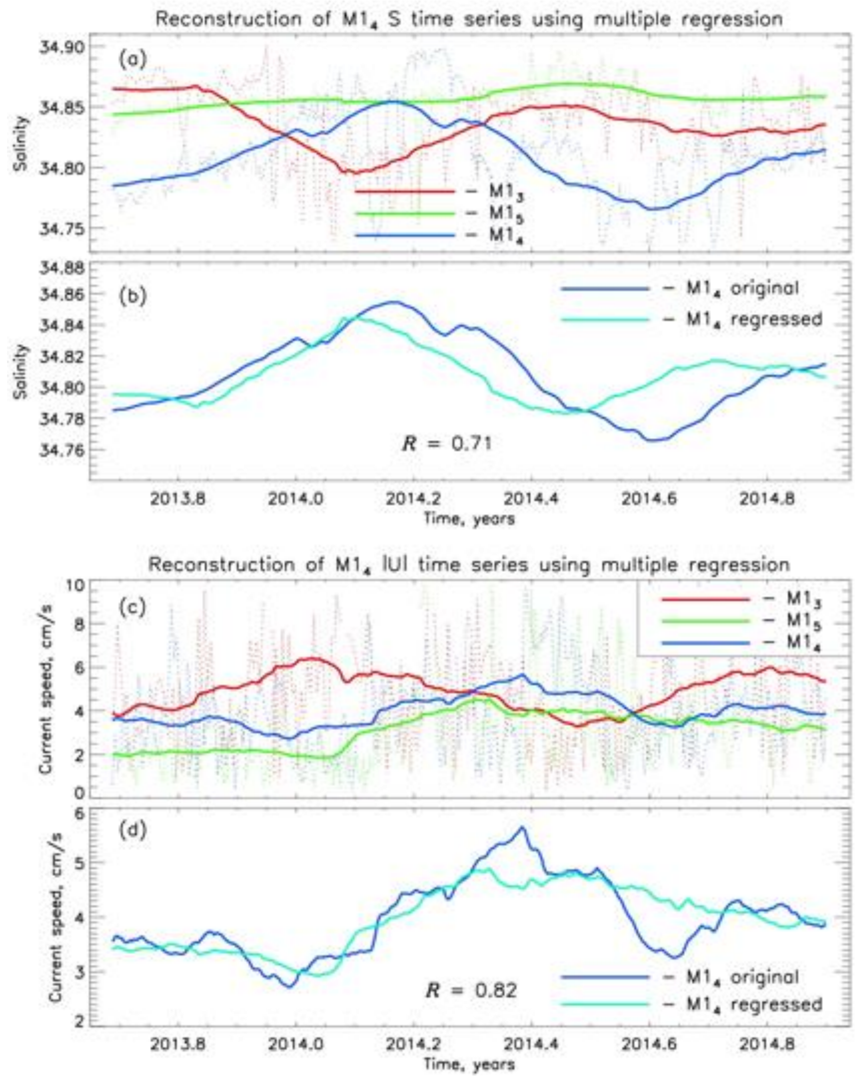

Figure A2: Multiple regression reconstruction of (a,b) salinity and (c,d) current speed |**U**| at M1$_4$ mooring site using data from M1$_3$ and M1$_5$ moorings for 170-210m depth range. (a,c) Daily (dotted) and three-month running mean smoothed time series of (a) salinity and (c) |**U**| from M1$_3$, M1$_4$, and M1$_5$ moorings. (b,d) Original (blue) and reconstructed (red) time series of (b) salinity and (d) |**U**| from M1$_4$ mooring. Relatively high correlations between the original and reconstructed time series attests of good quality of reconstruction.

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Table 1: Summary of deep-water moorings used in this study (only those instruments are shown which records have been used here). For mooring locations, see **Fig. 2**.

Mooring	Latitude (N) Longitude (E)	Depth (m)	Instrument	Depth range (m)	Beginning of record	End of record
Moorings deployed in 2002/09 and collocated with $M1_4$ mooring						
M1a	78 27.360 125 40.440	2680	MMP SBE37	164 –2598 57, 136	09/02/2002	09/01/2003
M1b	78 26.637 125 40.194	2686	MMP	104 –1484	09/08/2003	09/09/2004
M1c	78 26.637 125 40.194	2690	ADCP MMP	5 – 50 72 – 900	09/14/2004 09/15/2004	09/15/2005 07/16/2005
M1e	78 25.940 125 43.419	2692	ADCP MMP	5 – 57 70 – 900	09/02/2006	09/18/2007 10/11/2006
M1g	78 25.735 125 28.527	2765	ADCP SBE37	20 – 130 110, 116, 132, 339	10/18/2008 10/19/2008	06/16/2010 09/22/2011
Mooring section $M1_1$ – $M1_6$, 2013/15						
$M1_1$	77 04.252 125 48.288	250	ADCP	20 – 250	08/26/2013	09/10/2015
$M1_2$	77 10.376 125 47.516	787	ADCP MMP	5 – 63 70 – 754	10/27/2013 08/26/2013	09/01/2015 08/31/2015
$M1_3$	77 39.286 125 48.401	1849	ADCP MMP	5 – 56 64 – 750	09/06/2013 09/07/2013	09/02/2015 09/03/2015
$M1_4$	78 27.543 125 53.758	2721	ADCP ADCP SBE37	5 – 55 193 – 463 62, 129, 214, 265, 617	09/05/2013	09/19/2015
$M1_5$	80 00.199 125 59.673	3443	ADCP MMP	23 – 83 88 – 754	08/28/2013	06/16/2014 08/21/2015
$M1_6$	81 08.182 125 42.673	3900	ADCP MMP	5 – 55 60 – 754	08/29/2013	09/04/2015 08/22/2015
Mooring section $M1_1$ – $M1_5$, 2015/18						
$M1_1$	77 04.221 125 49.577	252	ADCP	200 – 232	09/21/2015	09/03/2018
$M1_2$	77 10.373 125 47.974	783	ADCP SBE	5 – 60 31, 44, 67, 138, 213, 266, 628	09/21/2015	09/03/2018
$M1_3$	77 39.234 125 48.686	1866	ADCP MMP	5 – 55 70 – 1056	09/21/2015 09/22/2015	09/03/2018 06/15/2017
$M1_4$	78 28.084 125 57.679	2700	ADCP ADCP SBE37	5 – 30 155 – 430 38, 107, 188, 240, 604	09/21/2015	09/18/2018
$M1_5$	79 56.194 126 01.228	3443	ADCP MMP	5 – 61 172 – 806	09/21/2015 09/24/2015	08/31/2018 08/29/2018

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Mooring M1$_{4-short}$ (September 2 –20, 2018)						
M1$_{4\text{-short}}$	78 30.833 125 58.924	2700	MMP	30 – 194	09/02/2018	09/20/2018
Moorings M3						
M3e	79 56.136 142 14.887	1335	ADCP SBE	5 – 61 41, 45, 57, 64, 130, 270 600	08/31/2013	09/07/2015
M3f	79 56.194 142 15.216	1357	ADCP SBE	5 – 44 30, 50, 133, 217, 268, 614	09/07/2015	09/06/2018

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