Recent oceanic changes in the Arctic in the context of long-term observations

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Abstract: This synthesis study assesses recent changes of Arctic Ocean physical parameters using a unique collection of observations from the 2000s and places them in the context of long-term climate trends and variability. Our analysis demonstrates that the 2000s were an exceptional decade with extraordinary upper Arctic Ocean freshening and intermediate Atlantic Water warming. We note that the Arctic Ocean is characterized by large amplitude multi-decadal variability in addition to a long-term trend, making the link of observed changes to climate drivers problematic. However, the exceptional magnitude of recent high-latitude changes (not only oceanic, but also ice and atmospheric) strongly suggests that these recent changes signify a potentially irreversible shift of the Arctic Ocean to a new climate state. These changes have important implications for the Arctic Ocean's marine ecosystem, especially those components that are dependent on sea ice or that have temperature-dependent sensitivities or thresholds. Addressing these and other questions requires a carefully orchestrated combination of sustained multidisciplinary observations and advanced modeling.

Keywords: Trajectory of the Arctic, Arctic, climate, climate change

1. Introduction

Changes in the arctic climate system over the past decades were exceptional in the history of arctic observations (e.g., Lindsay et al. 2009, Belchansky et al. 2008, Meier et al. 2007, Polyakov et al. 2005, Walsh and Chapman 2001), culminating with the summer of 2007 when the arctic ice retreat broke all records (Stroeve et al. 2008, Comiso et al. 2008). Despite the fundamental importance of high-latitude changes for global climate, there are numerous gaps in our understanding of how the system functions and what forces are driving changes in the Arctic. In particular, analysis of high-latitude climate change is complicated by strong arctic intrinsic variability dominated by multidecadal fluctuations (e.g., Polyakov and Johnson 2000, Polyakov et al. 2008). The key Arctic climate parameters like the Arctic surface air temperature (SAT), Arctic Ocean freshwater content (FWC), temperature of the intermediate (depth range 150-900m) Atlantic Water (AW) and fastice thickness demonstrate a strikingly coherent pattern of multi-decadal variability (MDV, Figure 1). Anthropogenic climate change may be amplified or masked by multidecadal variations, and separating the relative contribution of anthropogenic and natural drivers is not a trivial task.

Covariability between the physical and marine biological components of the climate system suggests an important role of the oceans in shaping the marine environment (Figure 2) that is the lifeblood of the biota. Marine organisms tend to follow certain environmental conditions (particularly, water temperature, but also stratification, light and nutrient availability, e.g. Wassmann et al. 2006, Slagstad et al. 2011); it is no surprise therefore that extended warm and cold periods have led to major changes in the ecosystem. While not from the Arctic, next we present examples from the North Atlantic and North Pacific, where the links between climate and biota are better understood. The Icelandic and Greenland seas warmed considerably in the 1920-40 causing rapid northward shift of fish (e.g. Loeng 1989, Drinkwater et al. 2010, Rose 2005). The fishing industry also shifted northward following the fish population, leading to record high catches. In the Pacific sector, researchers have found a link between the pattern of atmospheric and oceanic changes and salmon production (e.g. Francis and Hare 1994, Hare and Francis 1995). Chavez et al. (2003) found variability between anchovy and sardine fisheries, linked to the Pacific Decadal Oscillation (PDO) and other climatic indices. Overland et al. (2004) analyzed 86 spatially distributed multidisciplinary time series over 1965–95 with a broad geographical coverage spanning from the Canadian Northern Territories, Siberia and Northern Europe to high-latitude Arctic. They found that the Arctic climate system (including its biological component) responds to climate change in a coherent, but very complex, way. More specifically, the locations of ice algal blooms are dependent on the extent of sea ice relative to the seasonal cycle of solar radiation (Hinzman et al., this issue). With 80% of the Arctic tundra vegetation lying within 100-km of the ocean, this maritime biome is closely linked to Arctic sea ice (Bhatt et al. 2010) variations as well as trends. Sea ice decline has triggered near coastal land surface temperature increases and consequently enhanced vegetation productivity on the tundra. Thus, the recent and long-term changes in physical component of the climate system discussed in this synthesis are relevant because of their effect on the ecosystem and consequently, ecosystem services (i.e. human benefits such as subsistence lifestyle and resource extraction).

The Arctic Ocean plays a central role in the climate of the northern high latitudes and there are numerous gaps in our knowledge, which are compounded by the harsh environment for collecting observations and by the complex interactions and feedback mechanisms involved. Near-freezing surface waters, driven by winds and ice drift, exhibit a trans-polar drift from the Siberian Arctic toward Fram Strait (Figure 3). In the eastern and central Eurasian Basin the surface flow merges with several branches coming from marginal arctic seas. The surface cap of cold fresh waters is separated from Atlantic Water by a halocline in which salinities increase to ~34.8psu (e.g. Pfirman et al. 1994, Schauer et al. 1997, 2002). The frontal boundary between water masses of Atlantic and Pacific origin is an important element of water structure [e.g. McLaughlin et al. 1996] that is roughly aligned with the Transpolar Drift; Pacific waters are essential water masses of the Canadian Basin (Figure 4). Originating in the North Atlantic, Atlantic Water is carried through the Arctic Ocean interior by the pan-Arctic boundary current following bathymetry in a cyclonic sense (Figure 3, red arrows, e.g. Aagaard 1989, Rudels et al. 1994). Two major inflows supply the polar basins with AW: the Fram Strait branch water and the Barents Sea branch water (e.g. Rudels et al. 1994). The Fram branch enters the Nansen Basin through Fram Strait and follows the slope until it encounters the Barents branch north of the Kara Sea. Near the Lomonosov Ridge the flow bifurcates, with part turning north and following the Lomonosov Ridge and another part entering the Canadian Basin (e.g McLaughlin et al. 2009).

Starting from this brief and schematic overview of extremely complex body of polar waters interacting with ice and surrounding basins, this study assesses changes in the Arctic Ocean and places them in the context of long-term climate trends and variability by linking recent and historical observations. Thus, this synthesis provides an overview of studies of recent and long-term changes in the Arctic Ocean. However, we present an interpretation of oceanic datasets updated by the most recent observations including the 2000s, providing a unique aspect to this study. Particularly, in this synthesis we connect decade long observations collected through NABOS (=Nansen and Amundsen Basins Observational System) oceanographic program, to the broader spatiotemporal context provided by other observational programs and longer datasets. The overall goal of this paper is to serve as an Arctic Ocean resource for ecologists to enrich their interpretation of changes in biological systems and to facilitate placing their recent results into the context of strong long-term Arctic trends and large-amplitude variability (including MDV).

2. Observational data

This synthesis effort builds on our previous studies that have led to an extensive collection of oceanographic data spanning the past 50–100 years. Most measurements prior the 1950s were made within areas limited by deep-basin margins (Figure 5). A few central-basin observations are however available starting from Nansen’s expedition aboard the “*Fram*” in the late 19th century (Nansen 1902). This expedition provided the first few temperature and salinity profiles from the central Eurasian Basin. In 1937, the Russian icebreaker “*Sedov*” was trapped in ice north of the Laptev Sea and was forced to drift across the central Eurasian Basin for 812 days until it was released by another icebreaker. During this drift, temperature and salinity measurements were made (Figure 5). In the 1930-40s, the Russians launched a monitoring program consisting of manned ice-drift stations and winter aircraft surveys complemented by ship-based studies during summer. Their first manned drifting station NP-1 (1937) provided several measurements from the western Eurasian Basin (Figure 5). In 1955-56 the first Russian basin-scale aircraft surveys were conducted. Manned drifting ice camps provided most high-latitude (>80°N) observations in the 1960s and 1980s. The 1970s were an exceptional period in the history of high-latitude exploration, with seven Russian winter aircraft surveys (1973-79) and 1034 oceanographic stations during this period. In the 1990s, icebreakers and submarines provided measurements covering vast areas of the central Arctic Ocean (Figure 5). Data from oceanographic CTD (Conductivity-Temperature-Depth) stations, moorings (autonomous devices moored to the seafloor) and ice-tethered profilers (ITP, http://www.whoi.edu/page.do?pid=23096) that move along a tether and sample water temperature and salinity down to the depth of 500-800m are available from the 2000s (Figure 5).

Most observations prior to the 1980s were obtained from Nansen bottle water samples and discrete temperature measurements. Typical measurement errors are 0.01°C for temperature and 0.02 for titrated salinity. CTD instruments were used in recent years which has resulted in increased accuracy and vertical resolution of at least an order of magnitude greater than that of the historical measurements. All chemical observations were carried out during the short high-latitude summers, insuring that our results are not contaminated by local seasonal variations. CTD data used for FWC analyses were de-seasoned using seasonal climatology (Steele et al. 2001), which reduces decadal biases associated with seasonal differences of sampling in different periods of time (for example, spring-time aircraft-based sampling in the 1950s and 1970s and summer ship-based surveys in the 1990s). Analysis of long-term AW changes have assumed that seasonal variations in the AW layer are small and do not affect our results (e.g. Lique and Steele 2012, Ivanov et al. 2009, Dmitrenko et al. 2009).

These deep-ocean measurements were used to study long-term changes in the intermediate AW of the Arctic Ocean (Polyakov et al. 2004) and Arctic Ocean FWC anomalies (Polyakov et al. 2008). The time distribution of measurements used in these studies is shown in Figure 6. Figure 5 demonstrates that there are numerous gaps in the early part of the record. These early measurements clearly cannot provide reliable information about the magnitude of anomalies; however, we argue that they are useful in defining in very general terms the state of the ocean (e.g., whether the Arctic Ocean was fresher or saltier, warmer or cooler). This is corroborated by coherent low-frequency fluctuations of the AW core temperature (AWCT, defined by the temperature maximum) and SAT (correlation R=0.70) and also of FWC and SAT (R=0.60); the latter time series utilizes records from several hundred meteorological stations, a quarter of which are longer than 100 years (Figure 1). Thus, despite gaps, the early parts of the oceanographic records should not be prevented from providing some useful information about the pre-conditions existing in the early part of the 20th century.

In the analysis of recent changes, distributed observations in the central Arctic Ocean made in the 2000s were complemented by measurements at several repeated oceanographic sections (series of temperature, salinity and chemical-tracer vertical profiles along a line, http://nabos.iarc.uaf.edu/data/registered/main.php, http://www.awi.de/en/research/publications_and_data/). A zonal section crossing Fram Strait at ~78°50'N has been conducted annually since 1997 (e.g. Fahrbach et al. 2001, Schauer et al. 2004). Temperature measurements from its 5–9°E segment at the 50-500m depth range (the depth associated with AW inflow into the Arctic Ocean) are used in this study. Repeated sections crossing the Siberian slope along the major AW pathway in the vicinity of Svalbard (~30^{o}E) and Severnaya Zemlya (~104^{o}E), in the central Laptev Sea (~125^{o}E) and at the junction of the Lomonosov Ridge with the Siberian slope (~140^{o}E) and a cross-section through the central Canada Basin complement the Fram Strait observations. Water samples from sections in the Laptev and East Siberian seas in 2007-08 were analyzed for dissolved oxygen, nutrients, barium, and oxygen isotopes.

Moorings positioned along the AW path have provided continuous information about water mass transformations caused by the warm surge of the early 2000s (e.g. Beszczynska-Möller et al. 2012). Water temperature at the Fram Strait mooring (78°50'N, 8°20'E, instrument depth ~250 m) was measured using an Aanderaa instrument. A McLane Moored Profiler (MMP) has been used at the central Laptev Sea slope moored in 2002-11 (M1 mooring, 78°27'N, 125°40'E). The AW core is located over the 2500-3000m isobaths along the Laptev Sea slope (e.g. EWG 1997 climatology). Seven CTD cross-sections (i.e. 2002, 2004–09) made as a part of NABOS (= Nansen and Amundsen Basins Observational System) captured the core within this range. The only exception was in 2003 when there were two AW cores; the deeper one was shifted to the north, and was not resolved by the section. However, the northern station of this section was located over the ~3500-m isobath. Also, there is a good correspondence between the point measurements provided by the M1 mooring at 78°27'N, 125°40'E and CTD-based estimates of the AWCT (except 2011, when the AW core was closer to the shelf and the mooring-derived temperature was cooler than CTD-based one; at the same time the cooling tendency since

2008 was well pronounced in both records, Figure 8). Thus, we argue that the M1 mooring captures the major AW changes.

Several other long-term observational records complement these oceanographic datasets. Unique observations of fast-ice thickness from 15 stations spread uniformly along the Siberian coast (Polyakov et al. 2003) are updated until 2009 and are used in this synthesis. A composite time series of SAT in the Northern Polar Area (60^{o}N) used in this study is based on monthly records from 441 meteorological stations (Bekryaev et al. 2010, see their Figure 1 showing spatial and temporal data coverage). The North Atlantic Oscillation (NAO), Atlantic Multidecadal Oscillation (AMO) and Pacific Decadal Oscillation (PDO) climatological indices are also utilized (Table 1). To broaden our synthesis and tie it more directly to marine ecosystem impacts, we also discuss diverse records such as North Atlantic herring biomass change and Pacific salmon total catch (Table 1).

3. Changes of the Arctic Ocean thermal state

a. Long-term change of Arctic Ocean temperature

The upper ocean thermal state has direct implication to the sea-ice cover. That is why long-term changes of the upper Arctic Ocean temperature have become a subject of several recent studies. Steele et al. (2008), using temperature profiles and satellite data, argued that the upper Arctic Ocean experienced changes associated with the atmospheric Arctic Oscillation (AO) index phases when the ocean cooled by ~0.5^{o}C as a result of the prevailing decrease of AO during 1930-65 and warmed as the AO rose since then. Steele et al. found substantial acceleration of warming in the recent decades. We note however, that the AO was close to neutral in the last decade. At the same time the rate of Arctic warming (including Arctic Ocean warming, see Figure 1) has accelerated suggesting that other factors not related to AO may play an important role in warming upper ocean layer. Using historical hydrographic data for the Laptev and East Siberian seas spanning from 1920 through 2009 Dmitrenko et al. (2011) demonstrated a strong warming in the bottom layer of the shallow Siberian coastal zone started in the 1960s, with particularly strong warming, up to 2.1^{o}C since the mid-1980s. They attributed this warming to reduced summer ice cover and increased absorption of atmospheric heat by the seas.

Using high-latitude hydrographic measurements Polyakov et al. (2004) analyzed long-term variability of the AW temperature. They argued that, despite gaps in the early part of the record, composite AW time series provided evidence that AW variability is dominated by a long-term warming trend superimposed on MDV with a timescale of 50-80 years (Figures 1 and 6). Associated with this variability, the AW temperature record showed two warm periods in the 1930-40s and in recent decades and two cold periods in the earlier century and in the 1960-70s. Observations from the 1990s documented positive AW temperature anomalies of up to 1°C relative to temperatures measured in the 1970s throughout vast areas of the Eurasian and Makarov basins (Quadfasel et al. 1991, Carmack et al. 1995, Swift et al. 1997, Morison et al. 1998, Steele and Boyd 1998, Polyakov et al. 2004, Figure 7). Newly available data from the 2000s demonstrate that the temperature continued its rise resulting in the decade of record-high temperatures (Figure 6).

Improved spatial data coverage over recent decades has made it possible to demonstrate that the AW warming was associated with salinification, accompanied by ~150m AW layer domelike shoaling in the 1990s and ~75-90m in the 2000s (e.g. Carmack et al. 1997, Swift et al. 1997, Polyakov et al. 2004, their Figure 5 and Polyakov et al. 2010, their Figure 7). Similarity between the spatial distributions of sea-level pressure anomalies and the pattern of the AW domelike shoaling centered in the Makarov-Canada basins suggested that the observed shoaling in the 1990s represented a dynamical response to winds driving the circulation, while temperature, salinity and heat content fluctuations may be either dynamically or thermodynamically controlled (Polyakov et al. 2004). Since the 1970s, there was a sizable weakening of the Eurasian Basin stratification above the AW core - an important finding which links changes in the Arctic Ocean interior with potentially enhanced upward oceanic heat fluxes (e.g. Polyakov et al. 2010).

The ultimate source of the observed changes in the intermediate AW of the Arctic Ocean lies in interactions between polar and sub-polar basins (e.g. Dickson et al. 2000, Schauer et al. 2002, 2004, 2008, Gerdes et al. 2003). The observed fluctuations in the AW temperature are mostly due to low-latitude ‘switchgear’ mechanisms controlling temperature and salinity inflows into the Arctic Ocean via the intensity and position of the subpolar North Atlantic gyre. However, there are probably local mechanisms for the Arctic and sub-Arctic, which may modulate the AW inflows. One of the numerous examples of such mechanisms is presented in a study by Holliday et al. (2008) who documented a rapid increase of water temperature of the Atlantic Water inflow from the North Atlantic subpolar gyre through the Nordic Seas to Fram Strait since the 1970s. Their Figure 3 clearly indicates that the warming was accelerated in the northern parts of the Nordic Seas, with the maximum warming rates at Fram Strait suggesting that local air-sea interactions modulate substantially the North Atlantic signal. Using modeling results Aksenov et al. (2011) demonstrated that the AW transports along the Arctic Ocean margins are governed by a combination of buoyancy loss and non-local wind, creating high pressure upstream in the Barents Sea. This finding is consistent with Karcher et al. (2007) who showed that buoyancy forcing in the Barents Sea is a main driver of the AW flow. According to the negative feedback mechanism proposed in (Polyakov et al. 2004), changes in polar basin density act to moderate the inflow of Atlantic Water to the Arctic Ocean, and hence provide a potential local source for fluctuations in AW inflow.

b. Change of Arctic Ocean temperatures in recent decades

Over recent decades, satellite records showed a strong 3.7% per decade decline of sea ice–extent (Parkinson and Cavalieri 2008), which culminated in a record-breaking ice minimum during the summer of 2007 (Comiso et al. 2008, Stroeve et al. 2008). Ice-mass buoys were used in the Beaufort Sea in summer 2007 to detect enhanced upper-ocean solar heating through openings in the ice and consequent bottom ice melting (Perovich et al. 2008, Toole et al. 2010). This is a manifestation of the ice-albedo feedback mechanism, in which warming leads to a reduction in ice cover and albedo, resulting in increased absorption of solar radiation, warming and further sea-ice retreat (e.g. Manabe and Stouffer 1994, Serreze and Barry 2011). Indeed, Steele et al. (2008) used 2007 satellite surface temperature observations, which covered the entire ice-free area of the Arctic Ocean to estimate solar heating of 440 JM/m^2 of the upper 25m ocean layer. Direct oceanographic measurements suggested somewhat lower estimate of ~283 MJ/m^2 (Bekryaev et al. 2010). This anomalous upper-ocean heat uptake caused by the albedo decrease is much lower than the advective annual horizontal atmospheric heat transport through 60ºN, suggesting that the solar effect of oceanic warming may be not as significant as previously thought (for details, see Bekryaev et al. 2010). At the same time, Steele et al. (2008) noted that the effect of the upper summertime ocean warming since 1965 may be evaluated in terms of an equivalent 75cm ice-thickness loss.

One more source for the Arctic Ocean heating in the 2000s was the increased influx of warm waters of Pacific origin through Bering Strait into the Chuckchi Sea and further into the Canada Basin (e.g. Woodgate et al. 2010). Observations of transports through Bering Strait showed a doubling of heat flux from 2001 through 2007, enough to explain a third of 2007 summer Arctic ice thickness loss (Woodgate et al. 2010). The sea-ice reduction in the Canadian Arctic as a result of increased influx of warm summer waters of Pacific origin clearly shows the thermodynamic coupling between the Arctic ice and the ocean interior. For example, Shimada et al. (2006) suggested a positive feedback mechanism in which enhanced inflow of warm Pacific Water into the Canada Basin weakens ice coverage which in turn causes enhanced wind-driven transport of Pacific water into the basin.

Both observations (e.g. Woodgate et al. 2001, Schauer et al. 2004, Polyakov et al. 2005) and modeling (Karcher et al. 2003) indicate that fluctuations of the intermediate AW layer in the Arctic Ocean interior are linked to the highly variable nature of the AW inflows, with abrupt cooling/warming events. The latest pulse of warm water was detected using data from Fram Strait moorings and a CTD section in 1999 (Schauer et al. 2004, Figure 8). Further observations showed the propagation of this anomaly into the polar basin interior following a shallow-to-right propagation scheme; indeed, this pulse of warm AW water was found in the eastern Eurasian Basin in 2004 (Polyakov et al. 2005, Dmitrenko et al. 2008c, Figures 8 and 9). The distinctive pattern of this warming event (compare 1997–2000 temperature increase in Fram Strait record and the 2002–05 temperature increase in the eastern Eurasian Basin, Figure 8) was used as a tracer to estimate the speed of along-slope warming propagation to higher-latitude regions. According to these estimates, it took ~5 years for the warming to reach the Laptev Sea slope from the Fram Strait region (Figure 9), suggesting an anomaly speed of ~1.5 cm/s (Polyakov et al. 2005). The pulse peaked in the polar basin interior in 2007-08 (Figure 9).

The exceptional strength of this warming was documented using extensive observations made in 2007 under the auspices of the International Polar Year. Using these data, maximum temperature anomalies of up to 1^{o}C were traced along the AW pathways in a pattern similar to that observed in the 1990s (Figure 7). Point-to-point comparison demonstrated that AW temperature from 2007 was, on average, ~0.2^{o}C higher than in the 1990s thus confirming the exceptional strength of the latest warming pulse (see also Figure 6). Sufficient spatial coverage in the 1990s (Figure 5) and 2007 (Figure 7c), with hundreds of pairs of measurements available for comparison, makes standard error associated with this estimate small (SE=0.015^{o}C). Potential contamination of the estimate by the seasonal signal is also minor because in the vast area of the Arctic Ocean interior the AW seasonal signal is negligible (Lique and Steele 2012); measurements over the slope area of the Nansen Basin where the seasonal signal was detected (Ivanov et al. 2009, Dmitrenko et al. 2009) were made in summer (August-September) so that the monthly temperature difference is of the order of 0.05^{o}C.

Recent 2008–10 observations suggested that the on-going warm pulse passed its peak and the Arctic Ocean interior is in transition towards a cooler state (Polyakov et al. 2011). Stronger cooling is found in the western Nansen Basin, near the origin of AW for the

Arctic Ocean interior. Expectedly, weaker cooling was documented further downstream from Fram Strait, in the eastern Eurasian Basin (Figures 8 and 9). Time series from the Laptev Sea slope (~125^{o}E) extended by a four-year long record from recently recovered mooring demonstrates that temperature at this ocean site reached the values observed prior to when the warm pulse was detected (Figure 8). Composite pan-Arctic time series of AWCT provides further support for this cooling showing a temperature decrease since 2007 (Figure 6). A comparison of the temperature records shows an almost synchronous cooling in different Arctic Ocean regions (Figure 9). It suggests that cooling in areas remote from Fram Strait is not caused by the influx of colder AW because it would require an unrealistically rapid propagation of water from Fram Strait along the slope downstream. Enhanced shelf-basin exchanges may be one of the possible non-advective processes, which may modulate temperature changes in these regions. Diminished winter ice cover in recent years resulted in intensive sea-ice formation associated with brine rejection and probably sinking of brine-enriched dense and cold shelf water into the deep basin, thus providing intense ventilation of the basin's interior. Spatial heterogeneity of sea-ice coverage may explain substantial local differences in intensity and timing of this ventilation.

c. Upward spread of AW heat

The AW is believed to be effectively insulated from the pack ice by a cap of fresh, cold surface water bounded below by a strong pycnocline (e.g. Rudels et al. 1996) in which salinity increases from values of 33 psu or lower to around 34.5 psu at 150–300m depth. Strong stratification effectively suppresses mixing in the Arctic Ocean interior, away from the boundary and upper mixed layer. The resulting turbulent heat fluxes from the

AW layer in the Arctic Ocean interior are small, less than $1W/m^2$ (e.g. Rainville and Winsor 2008, Fer 2009). At the same time, the decrease of AW temperature with distance from Fram Strait (Figure 7) implies that AW heat must be lost as the AW spreads. Most of this heat is spread laterally by advection, eddy stirring or double diffusive processes, but some portion is lost upward, to the overlying halocline waters (e.g. Rudels et al. 1996, Steele and Boyd 1998, Martinson and Steele 2001, Polyakov et al. 2010, 2011, Walsh et al. 2007).

The extensive observations in the 2000s provide an opportunity to evaluate the upward spread of AW heat (e.g. Polyakov et al. 2010, 2012). Analysis of repeated cross-sections spanning from Svalbard to the East Siberian Sea and carried out for several years provided strong evidence of the existence of upward heat flux from the AW (Polyakov et al. 2010). For example, ten sections crossing the Siberian continental slope and spanning 43°E to 185°E taken in summer (August-September) 2007 were analyzed to quantify the along-slope change of water temperature. The potential temperature-salinity (θ-S) diagram (Figure 10, left) provides strong evidence that at low salinities (<34.3 psu, i.e. in the halocline and just below the upper layers separated from the halocline by the temperature minimum), temperatures are substantially higher at eastern sections compared with western sections. More specifically, temperatures from the eastern sections (longitudes >110°E) were 0.1–0.3°C higher than were western section temperatures. Mooring-based observations of currents at the continental slope off Svalbard, ~30°E (Ivanov et al. 2009), at the Laptev Sea slope, ~125°E (Dmitrenko et al. 2008b), and at the junction of the Lomonosov Ridge and the continental slope, ~133-150°E (Woodgate et al. 2001) suggest that the halocline waters in the Eurasian Basin travel in the same direction as the AW core. With the AW layer as the only source of heat, these observations provide strong evidence of the existence of upward heat flux from the AW.

Heat content *Q* was also used by Polyakov et al. (2010) to further quantify these along-slope changes. *Q* measures how much heat must be removed to cool the water to the in situ freezing. At each section, average *Q* was derived for two layers: an AW layer and an "overlying" layer (OL). The latter was defined to lie below the temperature minimum separating the halocline and the upper ocean layers (~30-50m) thus avoiding surface waters that are dominated by summer atmospheric heating) and 125m or the 0^{o}C isotherm, which defines the upper AW boundary point. Figure 10 (right) suggests that some heat lost from the AW is gained by the OL along the west-to-east AW spreading path. This analysis is based on the assumption that the OL in the Eurasian Basin travels in the same direction as the AW core. The strongest OL heat gain, up to 7% of the estimated AW heat loss, was found off Severnaya Zemlya (95-110^{o}E); much lower estimates were obtained for other segments of the Eurasian slope. Details of this analysis may be found in Polyakov et al. (2010).

This analysis demonstrates that the AW heat does penetrate into the overlying layers. How fast does this process occur? Polyakov et al. (2011), using visual inspection of available CTD observations (Figure 8), argued for coherent changes of AW and OL temperature as AW warming leads to immediate (within the available temporal resolution) warming of the OL at 125^{o}E. Recalculating OL *Q* to yield heat flux (assuming that the AW heat reaches the local OL in one year) yields ~3-4 W/m^2 for the early 2000s and up to ~6 W/m^2 for the peak year of 2007. The microstructure observations showed, however, that halocline mixing in the Arctic interior is very weak, ~1 W/m^2, and that mixing in the Laptev Sea is higher, with episodic peaks of ~4–8 W/m^2 (Lenn et al. 2009). How can we reconcile these estimates? Polyakov et al. (2012) analyzed high-resolution temperature and salinity vertical profiles, which resemble a staircase structure formed by layers of near-uniform water temperature and salinity interleaved with strong-gradient thin interfaces found in M1 mooring-based records from the eastern Eurasian Basin. They found strong, ~8 W/m^2, double-diffusive (i.e. ocean motion driven by different molecular viscosity of heat and salt) fluxes across several diffusive layers occupying the 150–250m depth range and overlying the AW core. Double-diffusive heat fluxes in the lower halocline of the Eurasian Basin interior based on ITP data are ~1–2W/m^2 (on-going analysis). We concluded that these fluxes provide a means for transferring AW heat upward over more than a hundred meter depth range towards the upper halocline.

4. Changes of the Arctic Ocean freshwater content

a. Long-term change of Arctic Ocean freshwater content

The Arctic Ocean is the key supplier of freshwater to subpolar basins thus, contributing to intensity of the deep convection and global thermohaline circulation (e.g. Dickson et al. 2000). Changes of the FWC of the Polar Basins are controlled by freezing and melting processes, anomalous supply of fresh shelf riverine and Pacific waters, precipitation and wind-driven redistribution of fresh water affecting ice drift and surface currents (e.g., Aagaard and Carmack 1989, Proshutinsky et al. 2002, 2009, Häkkinen and Proshutinsky 2004, Swift et al. 2005, Steele and Ermold 2005, Peterson et al. 2006; Serreze et al. 2006, Dmitrenko et al. 2008a, Newton et al. 2008, Polyakov et al. 2008, Timmermans et al. 2011, Giles et al. 2012, Morison et al. 2012). Figures 1, 6 (both updated) and 7 from

Polyakov et al. (2008) provide several examples of estimates of long-term Arctic Ocean FWC changes. These figures show that over the 20th century the central Arctic Ocean became increasingly saltier. For example, FWC anomalies averaged over 1950–1975 were estimated as -102±20 km^3; salinification led to a substantial decrease, -1478±17 km^3, of FWC over 1976–99. In contrast, long-term (1920-2003) FWC trends over the Siberian shelf were positive, 29±50 km^3 per decade, thus suggesting a general freshening tendency. The FWC temporal changes (Figures 1, 6 and 7) are consistent with the phases of MDV. Associated with this variability, the FWC record shows two periods, the 1930-40s and in the recent decade, when the central Arctic Ocean was fresher, and two periods in the earlier century and in the 1950-90s when it was saltier. Spatial pattern of FWC anomalies associated with the MDV phases are shown in Figure 5 of Polyakov et al. (2008). The latter salinification agrees with observational and modeling estimates (Steele and Boyd 1998, Häkkinen and Proshutinsky 2004, Swift et al. 2005). However, the FWC anomalies in the 2000s stand out: the freshening was dramatic, with no analogy in almost a century-long history of oceanographic observations (Figure 6).

One of the most striking features of FWC anomalies for the central basin and its shelves is that central-basin anomalies exceed those on the shelf by an order of magnitude (Figure 11, Polyakov et al. 2008). In addition, Figure 11 suggests an out-of-phase variability in the central basin and on the shelves, where sustained phases of central Arctic Ocean freshening are associated with salinification of the shelf waters and vice versa. The opposition of long-term tendencies expressed by trends showing general salinification of the central basin and freshening of shelves complement this observation. Based on this analysis, Polyakov et al. (2008) concluded that the FWC anomalies generated on arctic shelves (including river discharge inputs) *cannot* trigger the observed long-term FWC variations in the central Arctic Ocean; to the contrary, they tend to moderate long-term central-basin FWC changes.

Analysis of potential causes for the central Arctic Ocean salinification presented in Polyakov et al. (2008) suggested that the freshening/salinification of the upper ocean was not induced by the AW since the lower-layer changes were much weaker compared with the changes in the upper Arctic Ocean. Thus, ice production and sustained draining of freshwater (including ice and liquid exports) from the Arctic Ocean in response to winds are the key contributors to the salinification of the upper Arctic Ocean in the 1980–1990s. Finally, Polyakov et al. (2008) concluded that strength of the export of arctic ice and water controls the supply of Arctic fresh water to sub-polar basins while the intensity of the Arctic Ocean FWC anomalies is of less importance. In the next section we will discuss whether these conclusions still hold during the 2000s.

b. Change of Arctic Ocean freshwater content in recent decades

Observational data and modeling results provide evidence that increased arctic atmospheric cyclonicity in the 1990s resulted in a dramatic increase in the salinity in the Eurasian Basin. This salinification resulted from the increased volume of salty Atlantic-origin water entering the Eurasian Basin with a corresponding displacement towards the Canadian Basin of the Pacific-Atlantic water boundary (Carmack et al. 1995, McLaughlin et al. 1996, Morison et al. 1998). Steele and Boyd (1998) found a retreat of fresh surface waters and loss of the cold halocline layer from the Eurasian Basin, and linked this water mass change to a shift in atmospheric winds and ice motion. Steele and Boyd (1998) and Dickson (1999) argued that salinification of the upper Eurasian Basin in the late 1980s and early 1990s stemmed from the eastward diversion of Russian rivers, in response to the anomalous atmospheric circulation. Johnson and Polyakov (2001) suggested that two mechanisms account for the Eurasian Basin salinification: eastward diversion of Russian rivers, and increased brine formation due to enhanced ice production in numerous leads in the Laptev Sea ice cover. We hypothesize that these changes have probably had strong impact on the Arctic biota.

Arctic Ocean freshening in the 2000s was attributed to the strength of the Beaufort Gyre, which under anticyclonic atmospheric circulation tends to accumulate converging fresh water (Proshtinsky et al. 2002, 2009). Continuous freshening of the Beaufort Gyre was observed in 2003-07 culminating in 2008 when the FWC anomaly exceeded climatological values by as much as 60% (Proshutinsky et al. 2009, McPhee et al. 2009). Giles et al. (2012), using satellite data, reported that the dome of fresh water in the Canada Basin associated with the Beaufort Gyre continued to increase through 2010 thus suggesting a spin-up of the gyre and potentially further freshening of the Polar Basin. Recent 2010 freshening of the western Eurasian Basin was associated with the release of fresh water from the Beaufort Gyre, suggesting its important role in shaping the Arctic Ocean freshwater outflows into sub-polar seas (Timmermans et al. 2011). An alternative hypothesis explaining the FWC changes in the Arctic Ocean was proposed by Morison et al. (2012). According to this study, the observed FWC changes are driven by variations of large-scale atmospheric pattern characterized by the Arctic Oscillation index, which effectively regulates the oceanic pathways of the Siberian riverine waters into and through the central basin. Rabe et al. (2010) used observations and modeling results to attribute the strong freshening in 2006–08 to a variety of factors such as local wind-driven Ekman pumping, an increased ice melt and anomalous advection of riverine water from the Siberian shelves. Chemical observations suggested that the Beaufort Gyre freshening in the 2006 and 2007 was due to enhanced (1.3 m a^{-1}) ice melt (Yamamoto-Kawai et al. 2009).

Local wind conditions make the analysis of the freshwater pathways even more complex. For example, using oxygen isotope samples ($\delta^{18}O$) from stations north of the New Siberian Islands, Abrahamsen et al. (2009) calculated the freshwater composition in the upper 50 m of the water column, deriving a balance between meteoric water (primarily river runoff), sea ice meltwater, and, for 2007 and 2008, when phosphate and dissolved oxygen data were available, determining the split between Atlantic and Pacific water masses. Oxygen isotope data from 1993 and 1995 are from Schmidt et al. (1999, http://data.giss.nasa.gov/o18data). While winds in 1993 were cyclonic, causing the freshwater plume from the Lena to remain on the shelf, winds in 1995 were offshore, causing a wider spread of riverine waters over the shelf break and into the Amundsen Basin. In 2007, winds were cyclonic around the New Siberian Islands; average summer winds north of the New Siberian Islands were easterly, turning northerly over much of the Laptev Sea. This caused much of the outflow from the Lena to remain on the shelf, or to be forced to flow eastward over Lomonosov Ridge and into the Makarov Basin. This can be seen in Figure 12, where large amounts of meteoric freshwater can be found in the central and eastern sections. Figure 13 shows a significant presence of Pacific water in the easternmost stations in 2007, where it accounts for up to 40% of the surface layer. Some Pacific water was also measured in the eastern section in 2008, but in smaller quantities.

4. Discussion and concluding remarks

Does the host of recent Arctic Ocean changes represent an irreversible climate shift or can the polar basins recover (at least partially) to their previous state? For example, was the Arctic Ocean cooling after the warming of the 1930–40s accompanied by enhanced shelf-basin interactions as suggested by the recent synchronous cooling of the Arctic Ocean interior? There is much yet to understand, explanations for which remain obscure and will require further investigation. Advances in modeling and theory as well as continued observations are required in order to develop a deeper understanding of the mechanisms of high-latitude climate change. This will be a nontrivial task due largely to the poorly defined character of high-latitude variability and the changing relationship with large-scale climate parameters like the North Atlantic Oscillation (NAO, where positive values are characterized by a stronger north-south pressure gradient in the North Atlantic and stronger westerly winds) (Polyakova et al. 2006). The validity of extrapolating trends of the Arctic climate system into the future is impacted by the existence of large-amplitude MDV. Therefore it is imperative to understand how to separate these two processes and understand the underlying climate mechanisms. However, the exceptional decay of Arctic ice and anomalously strong upper Arctic Ocean freshening and high-latitude atmospheric and oceanic warming suggest that at least some of the observed Arctic Ocean changes are irreversible.

A comprehensive overview of the footprints of climate changte in Arctic ecosystems was given by Wassmann et al. (2011). They provided compelling evidence that all components of the high-latitude marine ecosystem are impacted by global warming as reflected in a wide range of changes including demography of Arctic species, their abundance, mortality and growth. Wassmann et al. emphasized that most reports considered large mammals and birds and the number of reports related to plankton and benthic species was surprisingly low. Despite uneven spatiotemporal coverage, this overview delivered an important message about the potentially alarming fate of Arctic species in a changing climate. The processes that give rise to ecosysten changes that are reflected in demography, growth and mobility are all determined by changes in temperature and stratification.

The trends and expectations for the carbon flux in a warming Arctic Ocean caused by climate change are manifold. The largest changes will take place in the northern sections of today's seasonal ice zone, which will in decades to come expand to cover the entire Arctic Ocean. Primary production will increase. The stratified and nutrient-poor surface waters prevent further increases in new production that would otherwise be expected as light availability increases. In regions subjected to large-scale advection or at shelf breaks additional nutrients can be supplies. Whether the new production of the central Arctic Ocean will remain low depends obviously upon the physical oceanography. Due to the thinning of the ice, the significance of ice algae for the total primary production of the Arctic Ocean may increase in the central Arctic Ocean, but will decrease in the outer seasonal ice zone. The blooms of ice and plankton algae will stretch over longer periods of time. Again, these processes depend upon insights and understanding of ice melt and surface freshening. Freshening of the Arctic Ocean, nutrient limitation and a prolonged growing season will change the community composition and carbon flux. To improve the estimates of primary production and carbon flux in the Arctic Ocean, attempts have to be made to increase our basic knowledge, in particular concerning the central Arctic Ocean basins and the entire Siberian shelf, which are poorly investigated (Wassmann et al. 2011).

While the ecosystem response to a warming climate may not distinguish between MDV and a long-term trend, the long-term predictability of the system is impacted by the nature of the forcing of the warming. If we want to develop a thoughtful response for a sustainable society, continued monitoring and understanding of the ocean, ice, atmosphere, terrestrial and biological components of the Arctic system must be a priority. However, we need to improve our understanding of key processes such as disspation of energy across density gradients, nutrient limitation and new production and responses of key organisms to changes in food, light and temperature.

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Table 1. Data used in this synthesis study

Climate system component	**Variable/Source**	**Region**
Ocean [historical]	Temperature/salinity, oceanographic stations Temperature/salinity, oceanographic stations	Arctic Ocean Siberian marginal seas
Ocean [recent]	CTD surveys, snapshot observations Mooring observations, time series Geochemical snapshot observations Ice-tethered profiler snapshot observations	central Arctic Ocean Arctic continental slope Arctic Ocean central Arctic Ocean
Atmosphere	SAT (monthly), meteorological stations	Arctic/sub-Arctic (>60^{o}N)
Ice	Fastice thickness, 15 coastal stations, AARI	Siberian marginal seas
Climate indices	NAO climate index, *Portis et al*. [2001] AMO climate index, *Enfield et al*. [2001] PDO climate index, *Mantua et al*. [1997]	North Atlantic North Atlantic North Pacific
Other	Herring biomass, *Toresen and Ostvedt* [2000] Salmon catch, *Klyashtorin and Lyubushin* [2007]	North Atlantic Pacific Ocean

Figure legends

Figure 1. Comparative long-term evolution of key components of the Arctic climate system. Composite time series of 7-year running mean anomalies of (from top to bottom) the Arctic surface air temperature (Bekryaev et al. 2010), upper 150~m Arctic Ocean freshwater content (*fresher* Arctic Ocean is associated with positive FWC anomalies, Polyakov et al., 2008), fast ice thickness and intermediate Atlantic Water core temperature of the Arctic Ocean (Polyakov et al. 2004). All records are updated using data from the 2000s.

Figure 2. Time series of major climatological indices AMO and PDO and lagged North Atlantic spring-spawning herring biomass anomalies (10^3 tons, Toresen and Ostvedt 2000) and Pacific salmon total catch anomalies (10^3 tons, Klyashtorin and Lyubushin 2007).

Figure 3: Circulation of the surface water (blue) and intermediate Atlantic Water (AW, red) of the Arctic Ocean.

Figure 4: Vertical profiles of water temperature (left) and salinity (right) collected in the Eurasian Basin (blue) and Canadian Basin (red) in 1974 showing Arctic Ocean water mass structure. Intermediate AW is identified by water temperatures >0^{o}C whereas Pacific Water is associated with the temperature maximum above the AW layer and below the upper mixed layer.

Figure 5. Maps of the Arctic Ocean showing the locations of deep-basin and shelf oceanographic stations used in this study (red dots).

Figure 6. The Arctic Ocean normalized AWCT anomalies and upper σ_{θ}□layer FWC anomalies (km^3). Annual anomalies are shown by blue dotted lines, 7-yr running means are shown by blue thick lines (dashed segments represent gaps in the records), and red dotted

lines show their confidence intervals defined by standard errors. Numbers at the bottom denote the 5-yr averaged number of stations used in the data analysis.

Figure 7. (a) Mean AW temperature (°C) averaged over the 1970s; (b and c) AW temperature anomalies (°C) averaged over the 1990s and for data from 2007. Anomalies are computed relative to climatology shown in (a). Isolines 0.05, 0.1 and 0.2 in a and b show standard errors (^{o}C); they are small (<0.05^{o}C) in the Canadian Basin and higher, up to 0.1^{o}C, at some places 0.2^{o}C, in the Eurasian Basin.

Figure 8. Time series of Atlantic Water (AW) temperature anomalies (^{o}C) relative to the time-series means from oceanographic sections (blue), mooring observations (red) and heat content density of the layer overlying the AW (~50–125m depth range, MJ/m^3, green). The mooring records were de-seasoned; CTD data are collected in summer so that the seasonal signal does not preclude meaningful interpretation of CTD records. From Polyakov et al. 2011, updated with 2009–11 data.

Figure 9. Vertical cross-sections of water temperature (^{o}C) from the Arctic Ocean. The five series of cascaded plots show temperatures measured at the five locations shown by yellow lines on the map. In each section, the horizontal axis shows distance from the southern end of the section (km) and the vertical axis shows depth (m). Note that the horizontal scale and temperature scale vary from one cascaded section to another. Warming in the Eurasian Basin is associated with the warm AW pulse, which was found in Fram Strait, the gateway to the Arctic Ocean, in 1999. This pulse peaked in the Eurasian Basin in 2007–08. In contrast, the warm anomaly in the Canada Basin is related to an earlier pulse of warm water, which entered the Arctic Ocean interior through Fram Strait in the early 1990s. Note that not all available sections are shown for the Fram Strait region. From

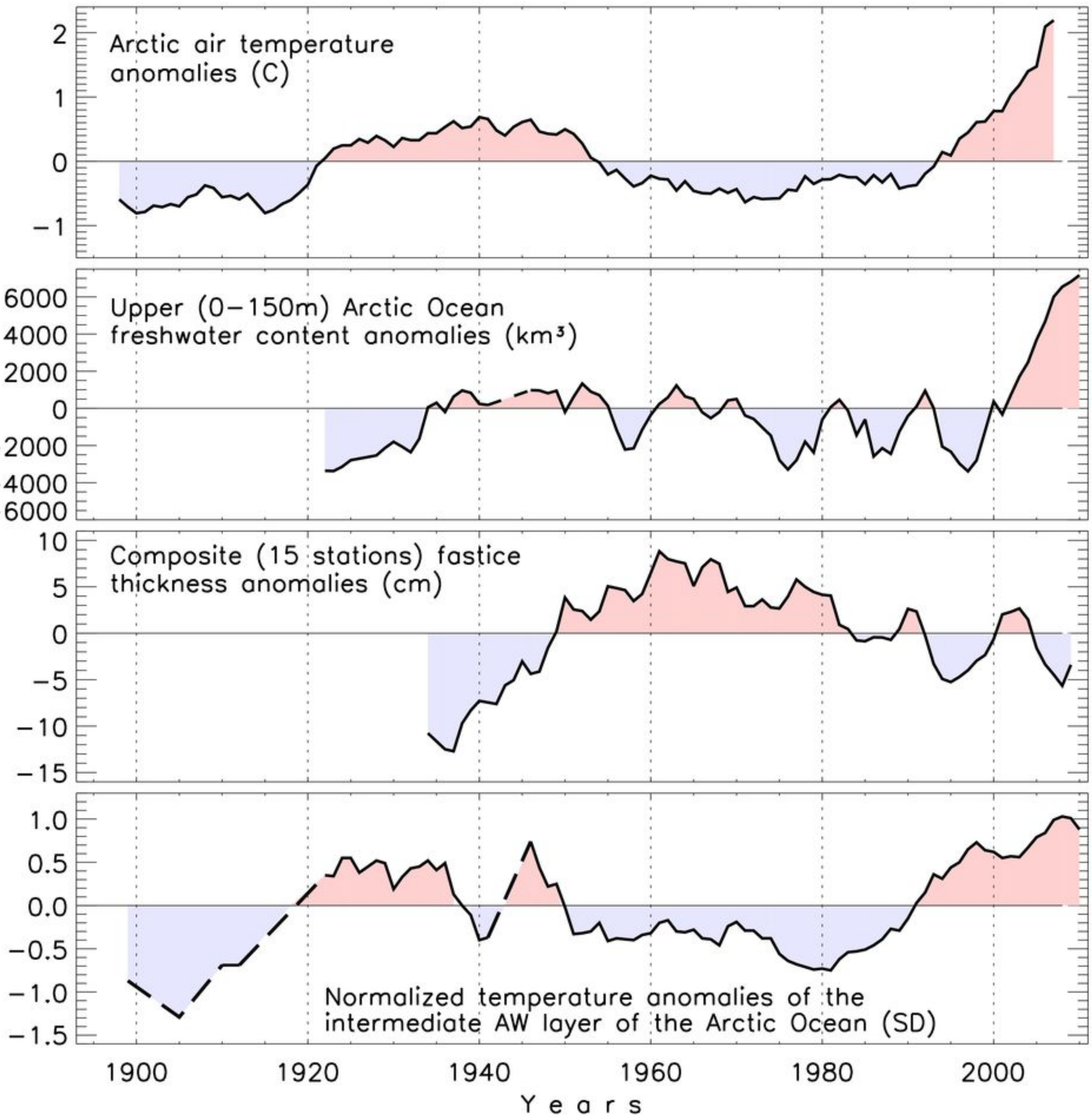
Polyakov et al. 2011.

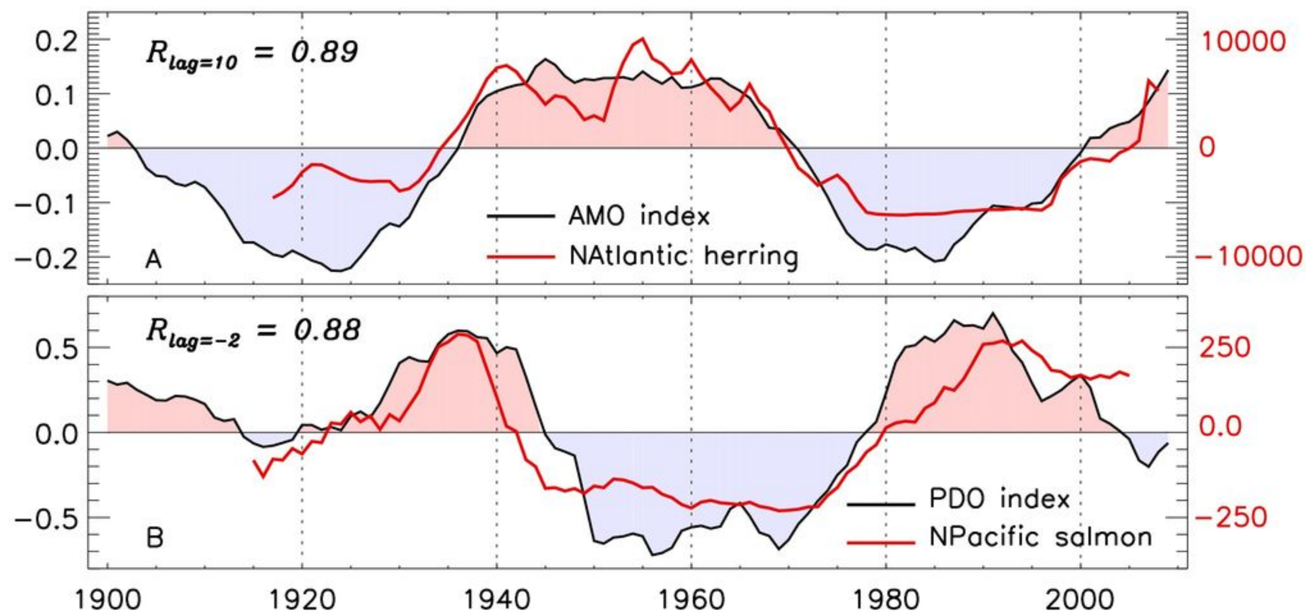
Figure 10. (Left) Potential temperature-salinity plot for the ten cross-sections carried out in 2007. All temperature (^{o}C) and salinity (psu) profiles for each cross-section are shown. At low salinities (<34.3 psu), temperatures are substantially higher at eastern sections (orange) compared with western sections (green). Water masses shown are lower-halocline water (LHW) and Atlantic Water (AW). (Right) Anomalous heat content (GJ/m^2) in the AW and overlying (OL) layers. Black triangles show positions of cross-sections that provided observational data; linear interpolation is used in between. Insert shows along-slope OL thickness change. These two panels provide evidence of the upward spread of AW heat along the AW path in the basin interior. From Polyakov et al. 2010.

Figure 11. (Top) Decadal (except for the last two years) freshwater content (FWC) anomalies and their standard errors for the central Arctic Ocean and Greenland and Barents seas. (Middle) Decadal FWC for the Siberian marginal, Barents, and Greenland seas. (Bottom) Pentadal freshwater input anomalies of the P-E over the Arctic Ocean (red) and river discharge (blue; adopted from Peterson et al. 2006). Linear trends over 1955–2002 are shown by dotted lines. All anomalies are in km^3. Positive anomalies represent *fresher* basin or input leading to *freshening*. This figure suggests that the FWC anomalies generated on arctic shelves (including anomalies resulting from river discharge inputs) and those caused by net atmospheric precipitation were too small to trigger long-term FWC variations in the central Arctic Ocean; to the contrary, they tend to moderate the observed long-term central-basin FWC changes. From Polyakov et al. 2008.

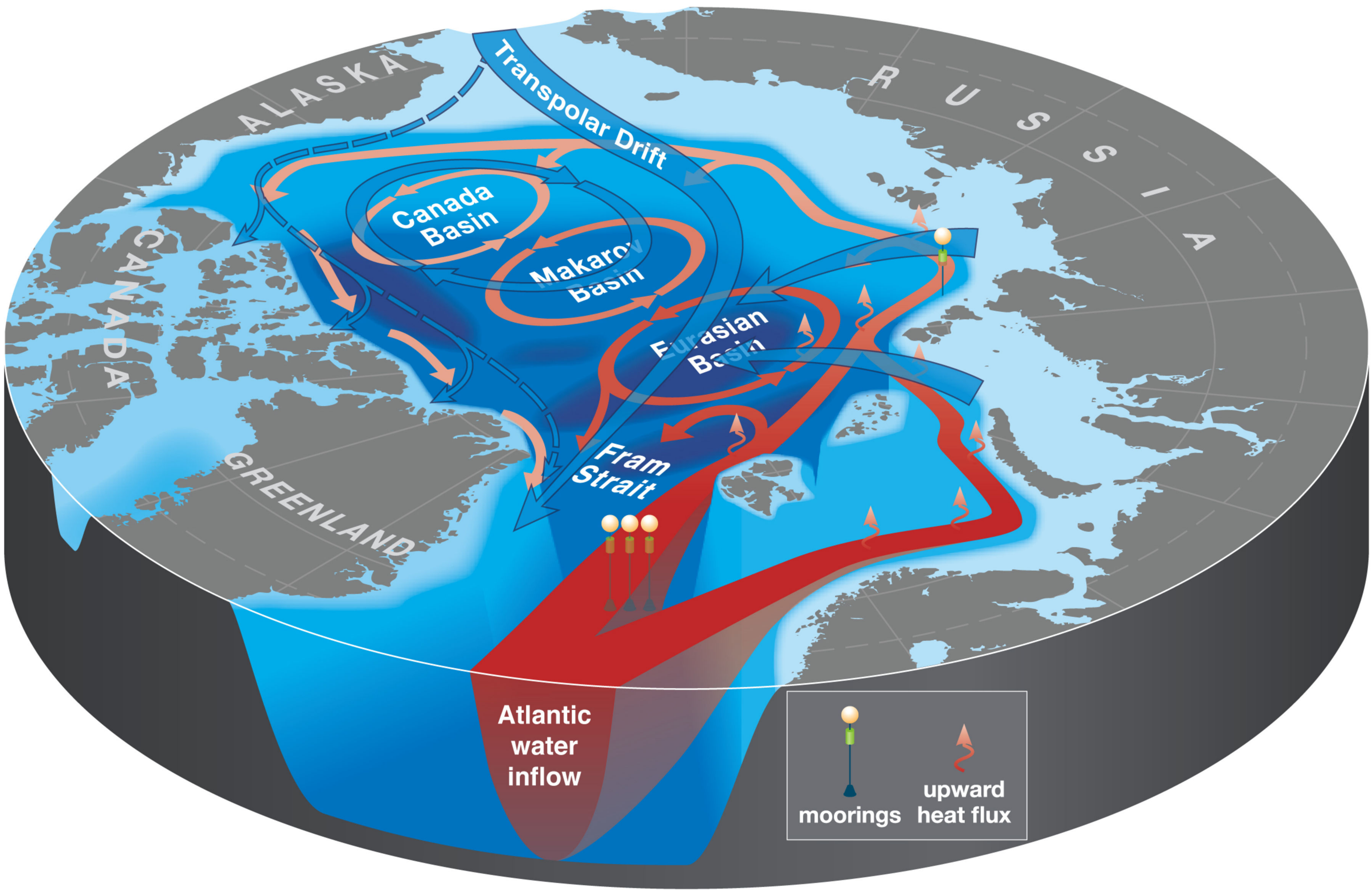
Figure 12. Maps showing the integrated (0-50m) water mass fractions for meteoric water in 1993, 1995, 2007 and 2008. From Abrahamsen et al. 2009, updated.

Figure 13. (a,b) Average dissolved barium concentration, (c,d) integrated water mass fractions for sea-ice melt and (e,f) integrated water mass fractions for the Pacific Water. All of these plots are integrated or averaged from the surface to 50 m depth. From Abrahamsen et al. 2009, updated.

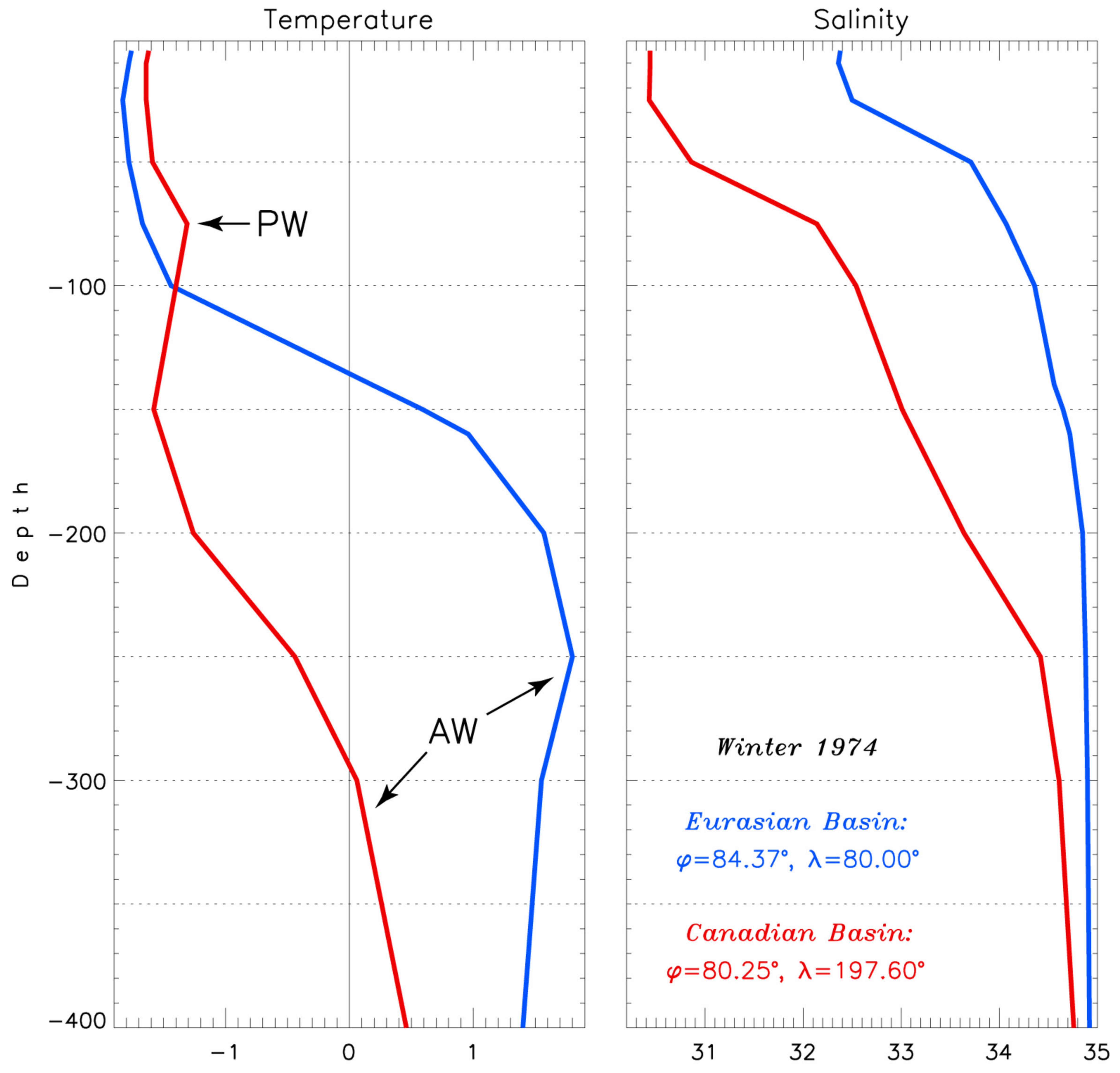

Arctic air temperature anomalies (C)
Upper (0–150m) Arctic Ocean freshwater content anomalies (km³)
Composite (15 stations) fastice thickness anomalies (cm)
Normalized temperature anomalies of the intermediate AW layer of the Arctic Ocean (SD)
Years
1900
1920
1940
1960
1980
2000



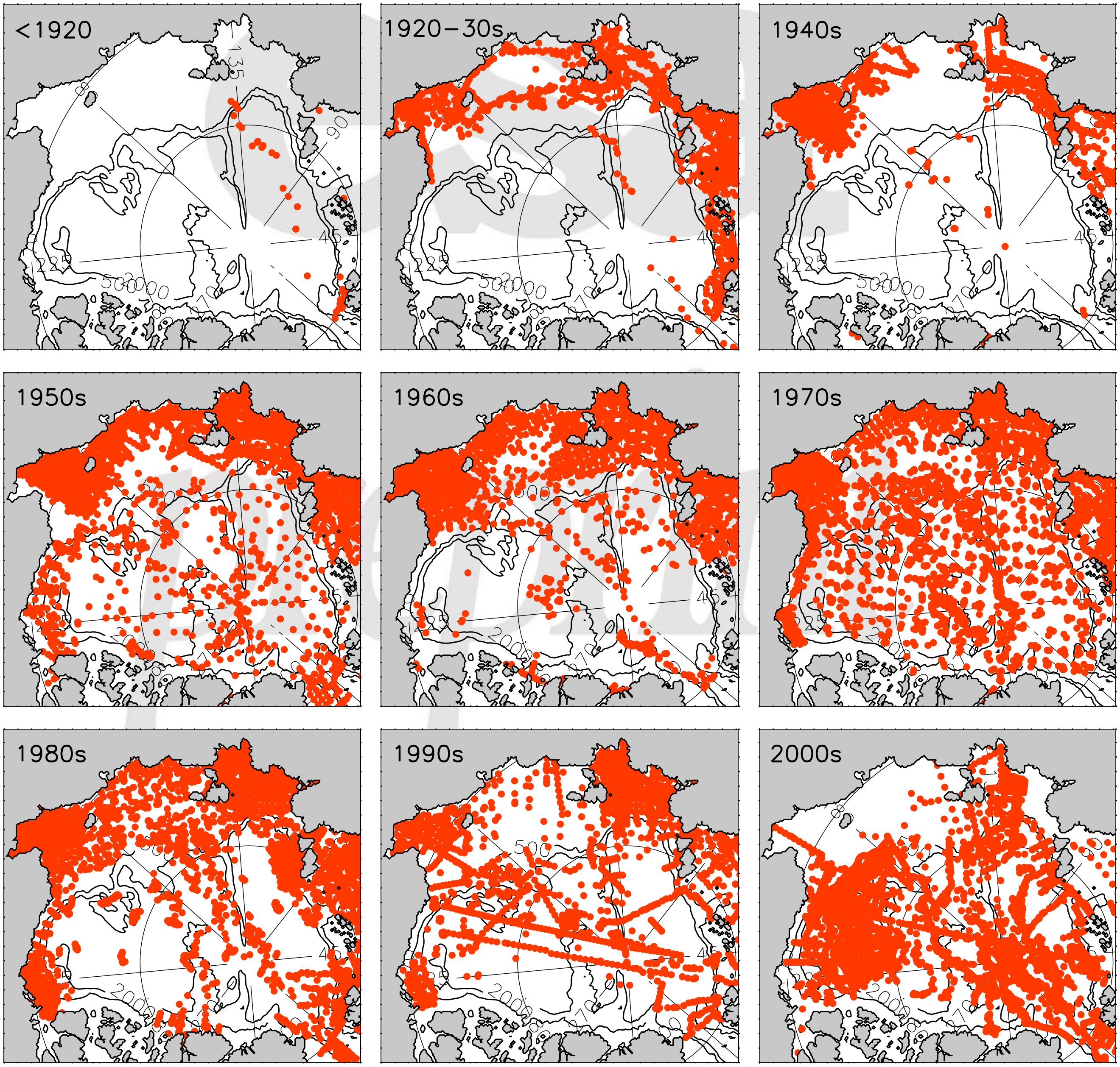
$R_{lag=10}$ = 0.89
AMO index
NAtlantic herring
A
0.2
0.1
0.0
−0.1
−0.2
10000
0
−10000
$R_{lag=-2}$ = 0.88
PDO index
NPacific salmon
B
0.5
0.0
−0.5
250
0.0
−250
1900
1920
1940
1960
1980
2000



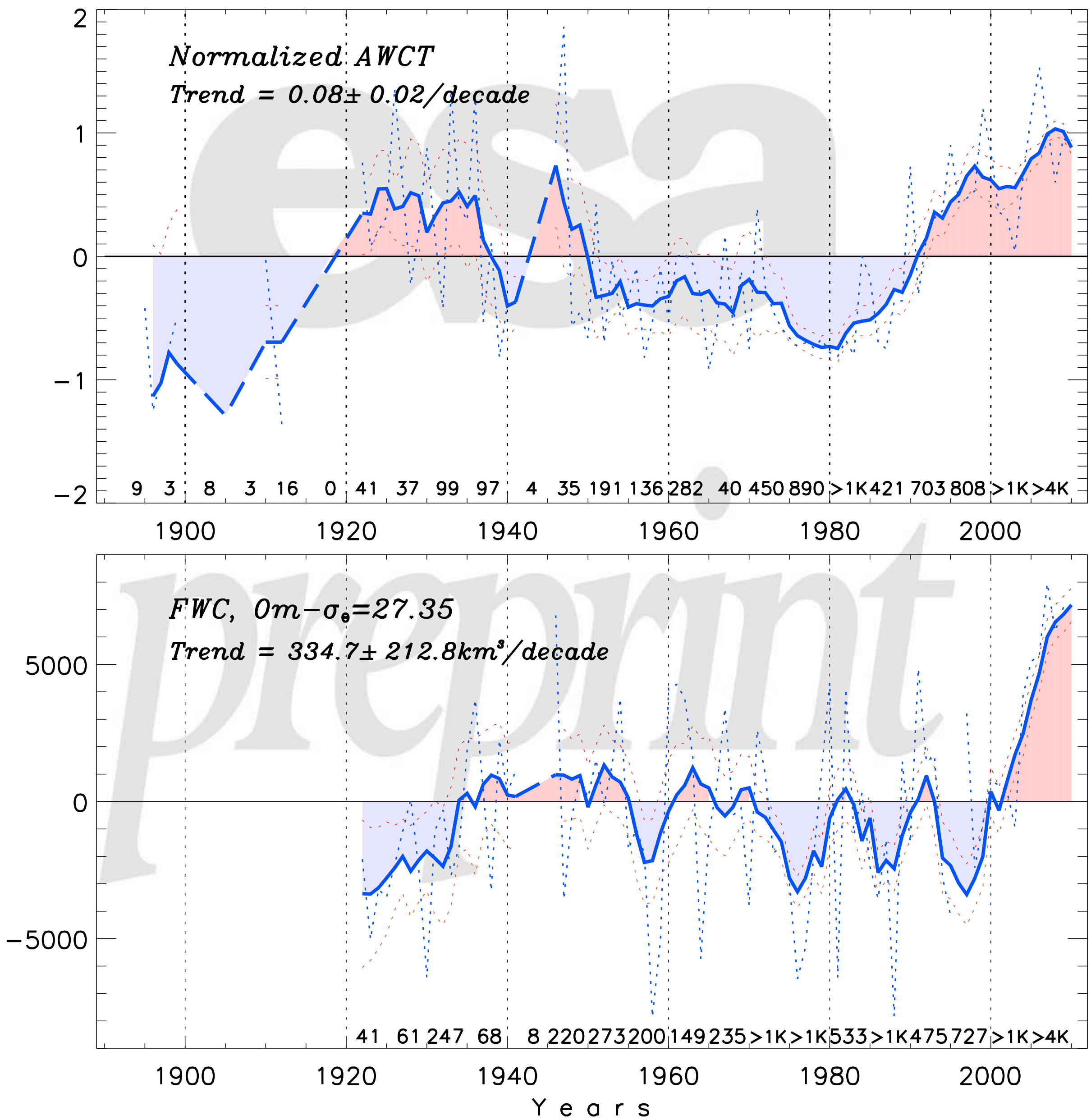
ALASKA
RUSSIA
CANADA
GREENLAND
Transpolar Drift
Canada Basin
Makarov Basin
Eurasian Basin
Fram Strait
Atlantic water inflow
moorings
upward heat flux

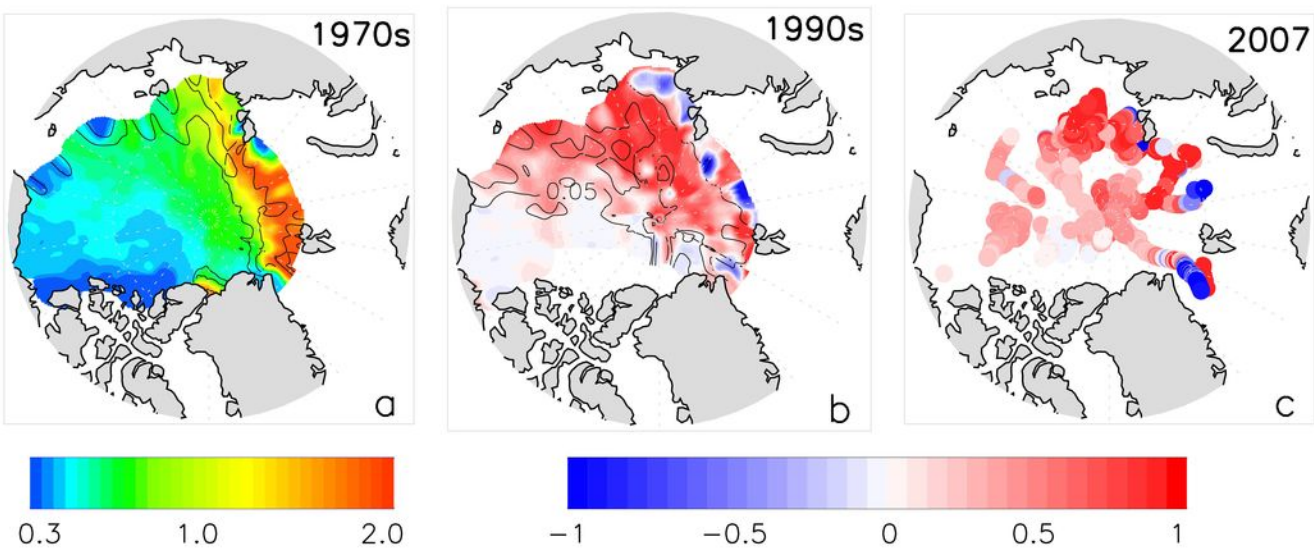


Temperature
Salinity
Depth
PW
AW
−100
−200
−300
−400
−1
0
1
31
32
33
34
35
Winter 1974
Eurasian Basin:
φ=84.37°, λ=80.00°
Canadian Basin:
φ=80.25°, λ=197.60°

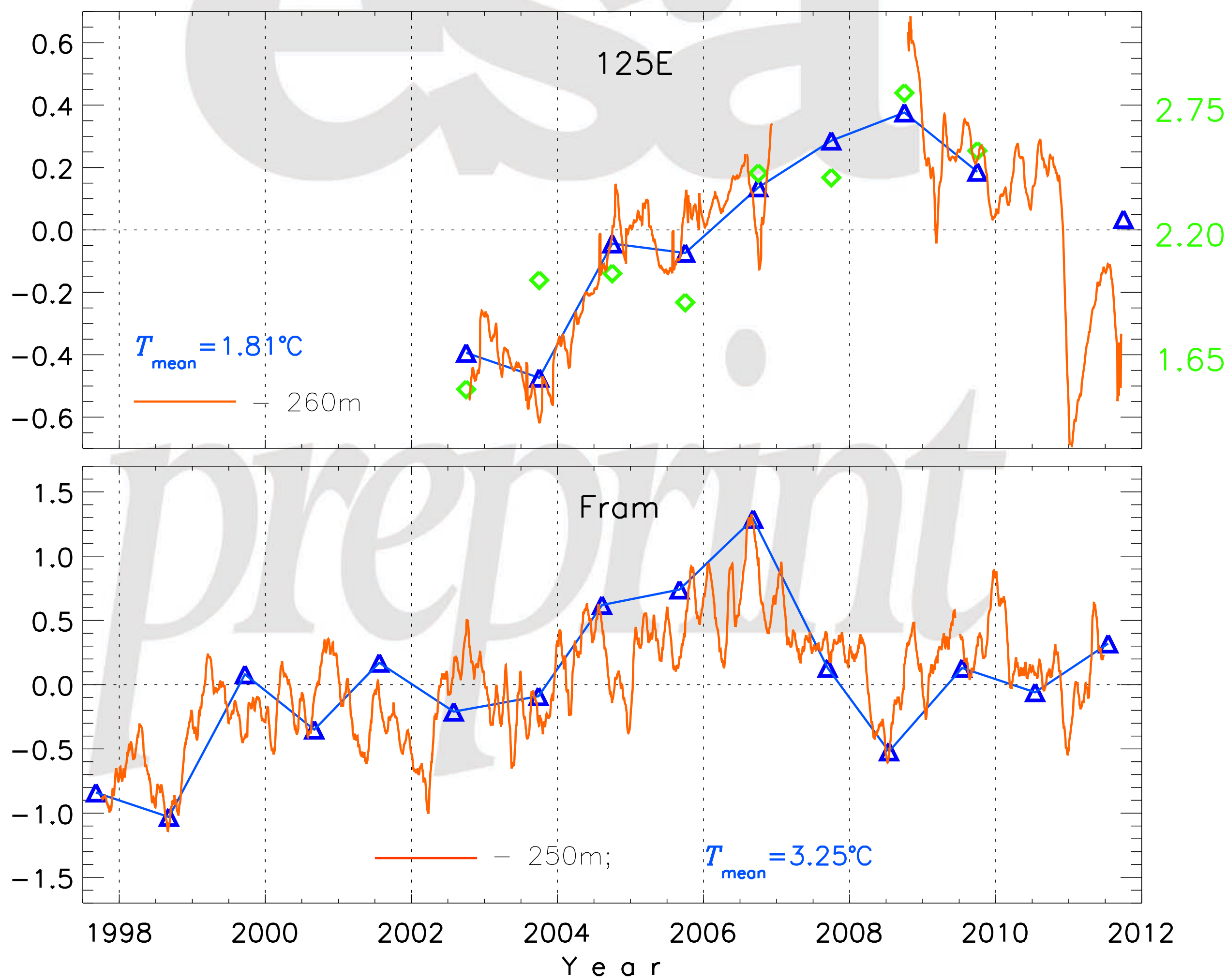


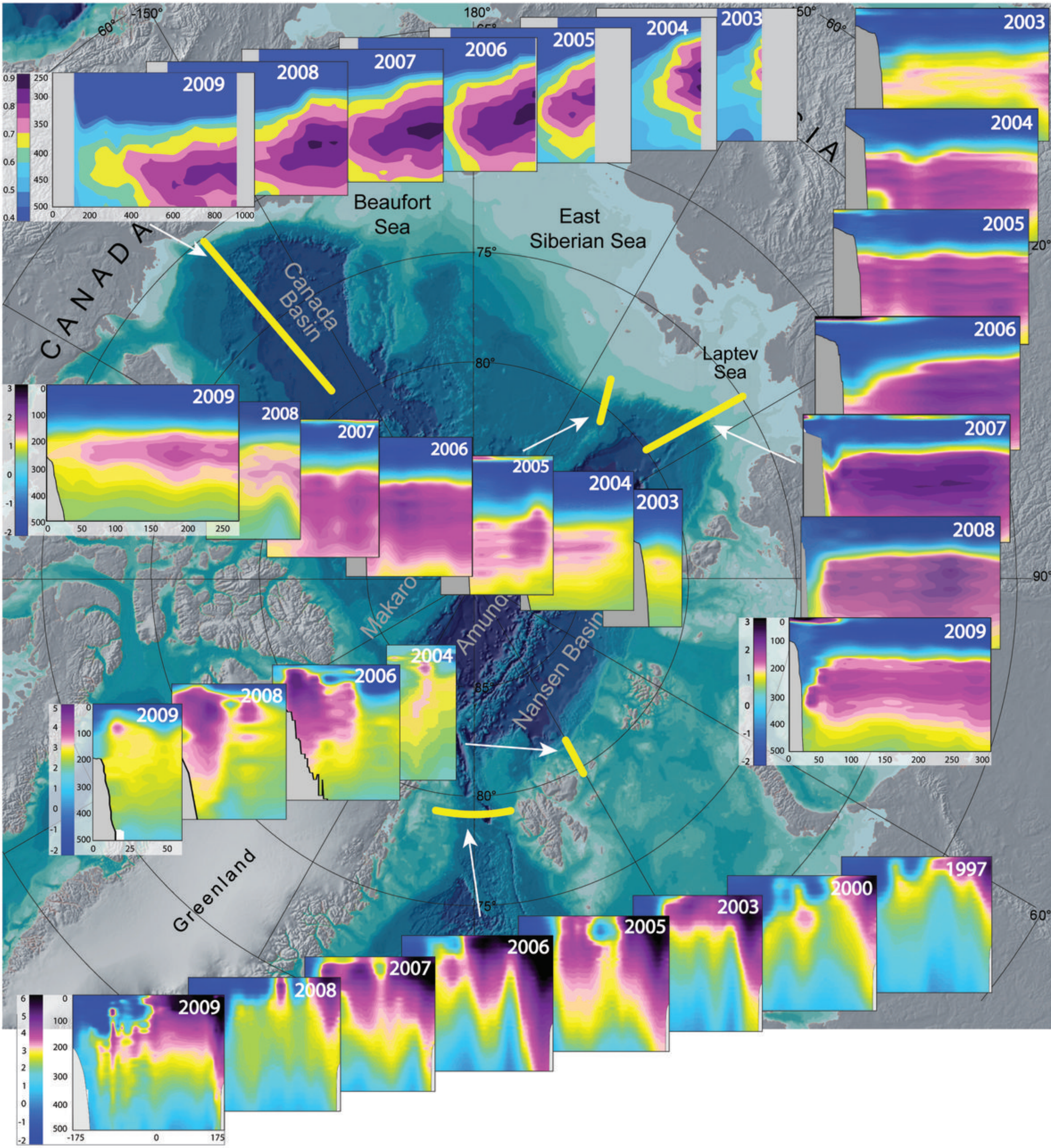
<1920
1920–30s
1940s
1950s
1960s
1970s
1980s
1990s
2000s


Normalized AWCT
Trend = 0.08± 0.02/decade
9 3 8 3 16 0 41 37 99 97 4 35 191 136 282 40 450 890 >1K 421 703 808 >1K >4K
FWC, 0m−σθ=27.35
Trend = 334.7± 212.8km³/decade
41 61 247 68 8 220 273 200 149 235 >1K >1K 533 >1K 475 727 >1K >4K
Years

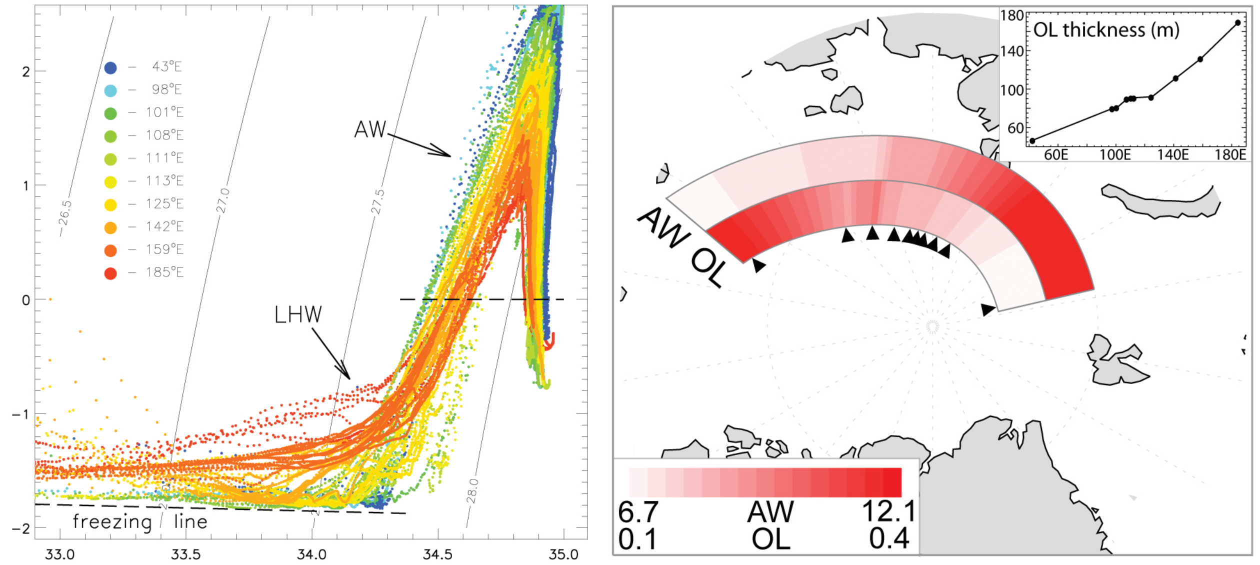


1970s
a
1990s
0.05
b
2007
c
0.3
1.0
2.0
−1
−0.5
0
0.5
1

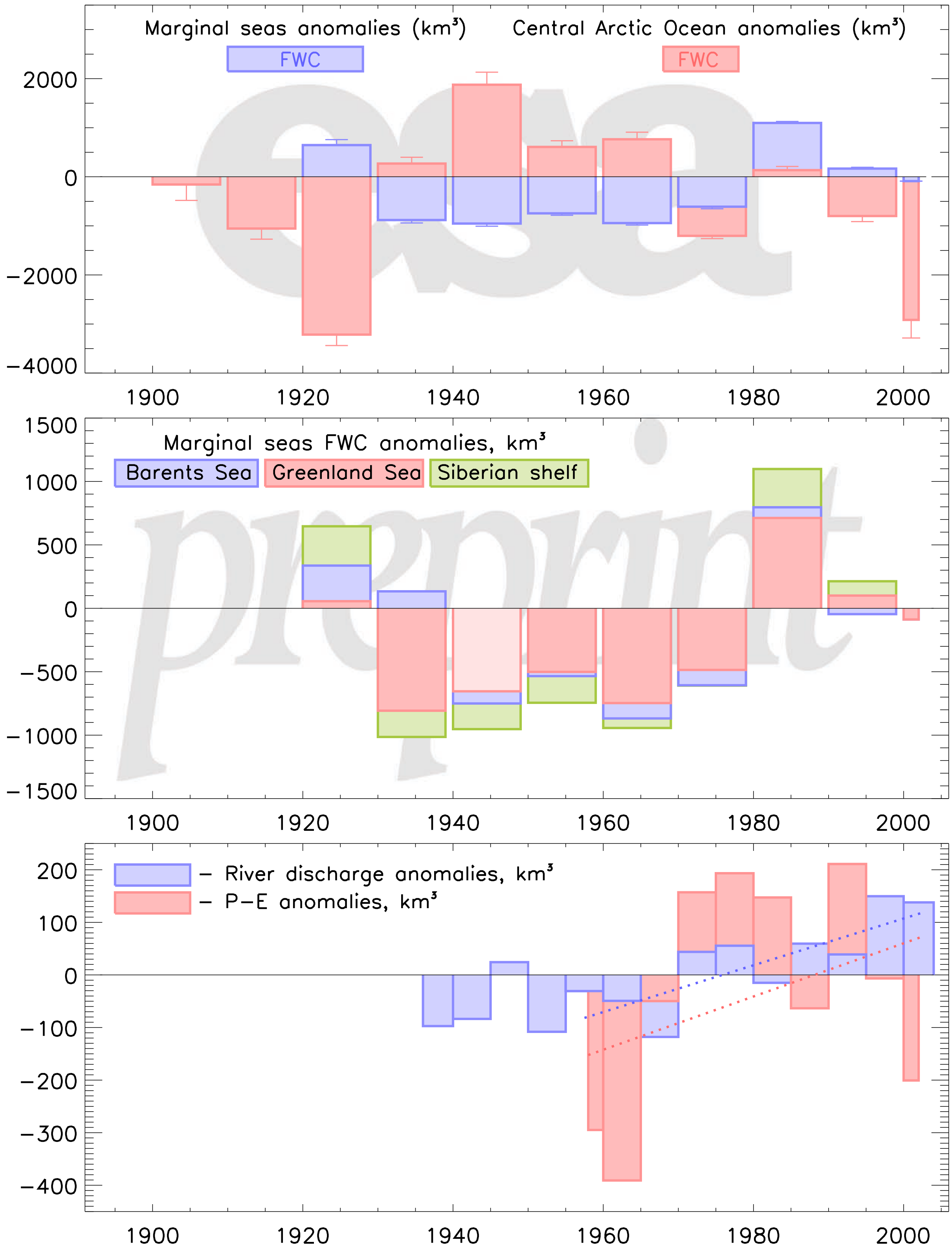

125E
T_{mean}=1.81°C
— 260m
2.75
2.20
1.65
0.6
0.4
0.2
0.0
−0.2
−0.4
−0.6
Fram
— 250m;
T_{mean}=3.25°C
1.5
1.0
0.5
0.0
−0.5
−1.0
−1.5
1998
2000
2002
2004
2006
2008
2010
2012
Y e a r

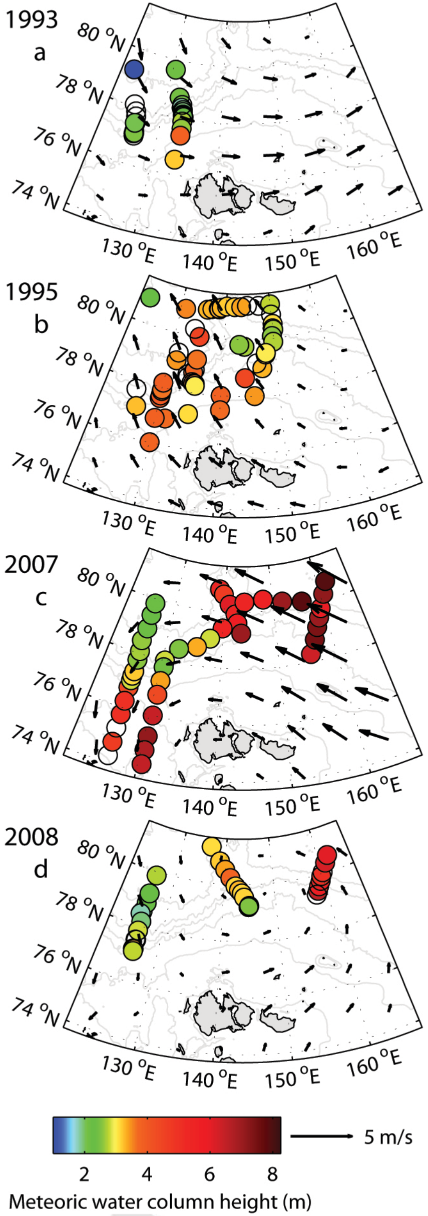


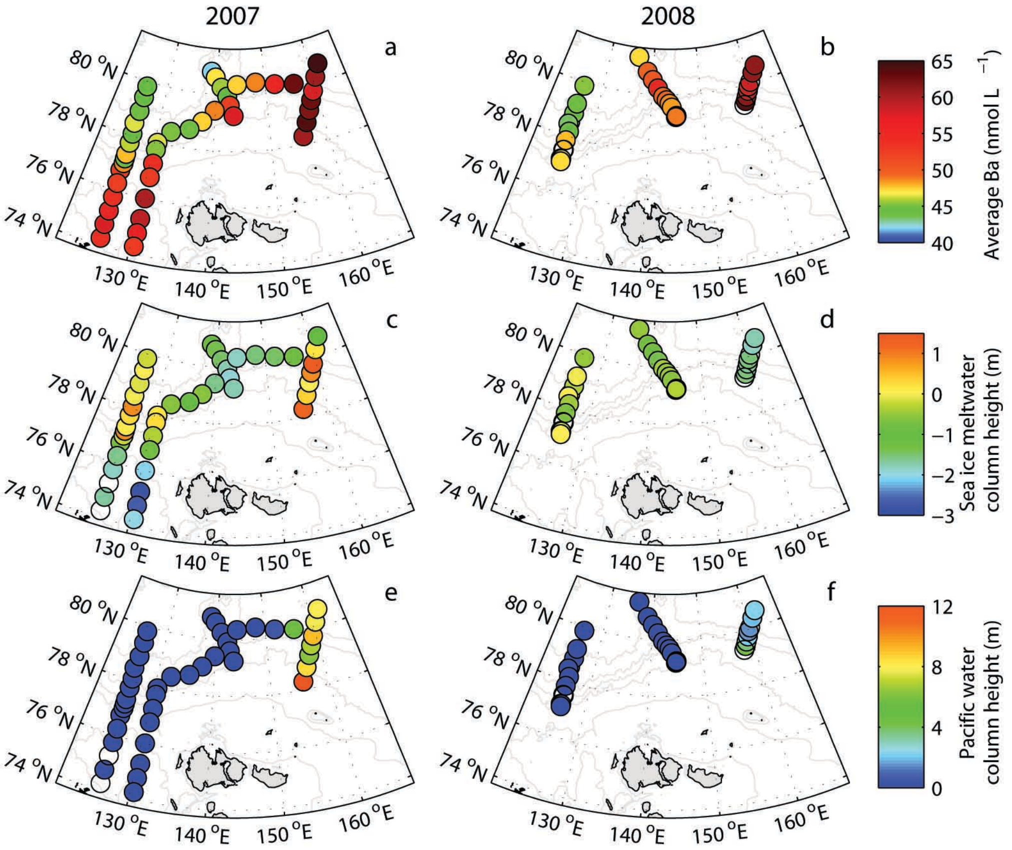
CANADA
Beaufort Sea
East Siberian Sea
Laptev Sea
Canada Basin
Makarov
Amundsen
Nansen Basin
Greenland
2009
2008
2007
2006
2005
2004
2003
2000
1997
75°
80°
85°
90°
180°
-150°
60°



43°E
98°E
101°E
108°E
111°E
113°E
125°E
142°E
159°E
185°E
AW
LHW
freezing line
26.5
27.0
27.5
28.0
2
1
0
−1
−2
33.0
33.5
34.0
34.5
35.0
OL thickness (m)
180
140
100
60
60E
100E
140E
180E
AW OL
6.7
AW
12.1
0.1
OL
0.4


Marginal seas anomalies (km³)
FWC
Central Arctic Ocean anomalies (km³)
FWC
2000
0
−2000
−4000
1900
1920
1940
1960
1980
2000
1500
1000
500
0
−500
−1000
−1500
Marginal seas FWC anomalies, km³
Barents Sea
Greenland Sea
Siberian shelf
200
100
0
−100
−200
−300
−400
– River discharge anomalies, km³
– P−E anomalies, km³


1993
a
80 °N
78 °N
76 °N
74 °N
130 °E
140 °E
150 °E
160 °E
1995
b
2007
c
2008
d
2
4
6
8
5 m/s
Meteoric water column height (m)



2007
2008
a
b
c
d
e
f
80 °N
78 °N
76 °N
74 °N
130 °E
140 °E
150 °E
160 °E
65
60
55
50
45
40
Average Ba (nmol L −1)
1
0
−1
−2
−3
Sea ice meltwater column height (m)
12
8
4
0
Pacific water column height (m)