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Hydrographic changes in the Lincoln Sea in the Arctic Ocean with focus on an upper ocean freshwater anomaly between 2007 and 2010

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[1] Hydrographic data from the Arctic Ocean show that freshwater content in the Lincoln Sea, north of Greenland, increased significantly from 2007 to 2010, slightly lagging changes in the eastern and central Arctic. The anomaly was primarily caused by a decrease in the upper ocean salinity. In 2011 upper ocean salinities in the Lincoln Sea returned to values similar to those prior to 2007. Throughout 2008–2010, the freshest surface waters in the western Lincoln Sea show water mass properties similar to fresh Canada Basin waters north of the Canadian Arctic Archipelago. In the northeastern Lincoln Sea fresh surface waters showed a strong link with those observed in the Makarov Basin near the North Pole. The freshening in the Lincoln Sea was associated with a return of a subsurface Pacific Water temperature signal although this was not as strong as observed in the early 1990s. Comparison of repeat stations from the 2000s with the data from the 1990s at 65°W showed an increase of the Atlantic temperature maximum which was associated with the arrival of warmer Atlantic water from the Eurasian Basin. Satellite-derived dynamic ocean topography of winter 2009 showed a ridge extending parallel to the Canadian Archipelago shelf as far as the Lincoln Sea, causing a strong flow toward Nares Strait and likely Fram Strait. The total volume of anomalous freshwater observed in the Lincoln Sea and exported by 2011 was close to $1100 \pm 250 km^3$, approximately 13% of the total estimated FW increase in the Arctic in 2008.

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1. Introduction

[2] Recurring salinity anomalies in the North Atlantic subpolar gyre have been correlated with large-scale changes in the thermohaline properties and circulation of the Arctic Ocean on interannual time scales [*Dickson et al.*, 1988; *Curry and Mauritzen*, 2005]. Chemical tracers have been used to qualify the large-scale Arctic Ocean circulation and have shown that the typical residence times of fresh upper ocean Arctic waters vary from several years to decades [*Schlosser et al.*, 1994; *Bauch et al.*, 1995; *Ekwurzel et al.*, 2001]. Modeling studies have shown that anomalies in liquid freshwater flux (FWF) from the Arctic Ocean are delayed relative to changes in atmospheric circulation [*Karcher et al.*, 2005]. These liquid freshwater anomalies may have a time lag of up to 1–6 years before exiting the Arctic through the Canadian Arctic Archipelago and Fram Strait, respectively [*Jahn et al.*, 2009]. Near-surface freshwater in the Beaufort Gyre in the Canada Basin tends to accumulate through Ekman convergence associated with the intensity of the Beaufort high during an anticyclonic atmospheric state [*Proshutinsky et al.*, 2002], but not exclusively. In the 1990s, the cyclonic circulation prevailed [*Proshutinsky et al.*, 2009] and freshwater in the Beaufort Sea increased significantly during that time [*Proshutinsky et al.*, 2002] followed by a relaxation to near pre-1990 climatology in early the 2000s due to a decline in the Arctic Oscillation (AO) [*Morison et al.*, 2006]. Between 2003 and 2007, an increase of freshwater content by more than 1000 km^3 was observed in the Beaufort Gyre [*Proshutinsky et al.*, 2009]. This continued up to 2008 when it was found

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to exceed climatological values by as much as 60% [*McPhee et al*., 2009] and basin wide it had increased with $8400 \pm 2000\ \mathrm{km}^3$ relative to the 1990s [*Rabe et al*., 2011].

[3] Several explanations have been reported for the surface freshening in the Beaufort Gyre. The freshwater content increased due to declining upper-ocean salinities and regional deepening of the lower halocline related with Ekman pumping [*Rabe et al*., 2011], sea-ice melt, McKenzie River water and increased Eurasian runoff contributed to fresher surface waters of the southern Canada Basin and central Canada Basin, respectively [*Yamamoto-Kawai et al*., 2009], changes in pathways of Eurasian river runoff caused regional increases [*Guay et al*., 2009; *McLaughlin et al*., 2011; *Morison et al*., 2012], and freshwater accumulated due to strengthening of the Beaufort Gyre [*Giles et al*., 2012]. Another recent study showed that the surface waters in the Eurasian Basin between the North Pole and Fram Strait were fresher in 2010 than in 2007–2008 which was attributed to a weakening of the Beaufort Gyre's anticyclonic circulation in 2009 [*Timmermans et al*., 2011]. *Karcher et al*. [2012] combined tracer data and numerical modeling results to show that as of 2004 the upper ocean anticyclonic circulation in the Canada Basin became so strong that fresh surface flow extended as far as the Lomonosov Ridge. This strengthening was found to also influence the underlying cyclonic boundary current carrying Atlantic water around the periphery of the Arctic Ocean, through a reversal of cyclonic to anticyclonic flow at mid depth with the numerical model NAOSIM [*Karcher et al*., 2012].

[4] In our present study, annual variations of upper-ocean salinity in time and space, and hence freshwater content in the central Arctic Ocean throughout the period 2000–2011 are presented. Specific attention is given to the Lincoln Sea, located north of Greenland and just upstream of Nares and Fram straits, two important gateways through which freshwater leaves the Arctic Ocean. In 2003, the Freshwater Switchyard Program was initiated in the Lincoln Sea with the purpose of observing the variability and origin of Arctic freshwater, and relating that to variability of the large-scale Arctic Ocean circulation, e.g., related with shifts between cyclonic and anticyclonic circulation. These fresh polar water masses are eventually exported to the North Atlantic through Fram Strait, the Canadian Arctic Archipelago or Nares Strait [see e.g., *Dickson et al*., 2007]. In this paper, we investigate the observed variations in thermohaline properties in the Lincoln Sea, specifically the upper ocean salinity changes during the period 2007–2011. Second, variations in the deeper Atlantic layer are discussed and compared with the 1990s. Hydrographic data and additional satellite results from the whole Arctic are included to place the observations in the Lincoln Sea in broader context.

2. Study Area

[5] The Lincoln Sea is a shallow shelf sea (~300 m deep) with a steep continental slope and is separated from the Lomonosov Ridge by a saddle reaching approximately 1500 m depth (Figure 1a). The region is marked by a complex interplay of the cyclonic Arctic Circumpolar Boundary Current and topographically steered currents in the Amundsen and Makarov basins transporting Atlantic water into the Arctic Ocean and the Transpolar Drift Stream. Hence water masses from both the Canadian and Eurasian basins are found here which show significant interannual variability [*Newton and Sotirin*, 1997].

[6] Waters from the Canada Basin are distinct in that they contain inflow from the fresher Pacific Ocean via

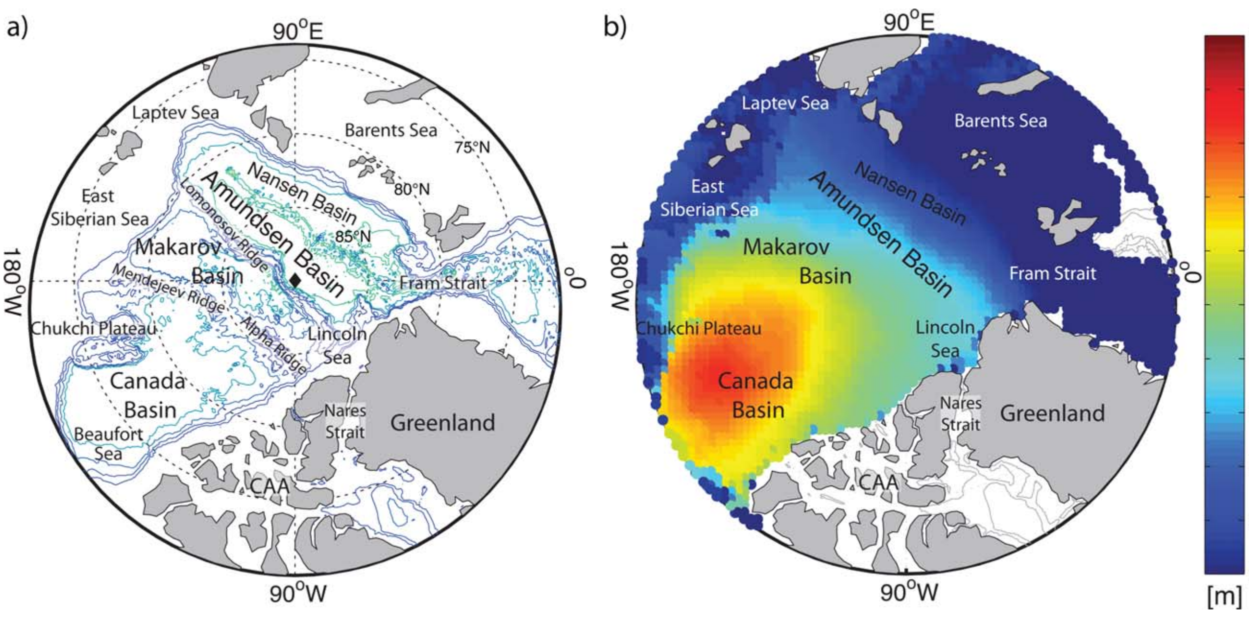

Figure 1. (a) Geographical map of the Arctic Ocean. Bathymetric contours are shown from 500 to 5500 m depth with 1000 m intervals. (b) Arctic freshwater content [m] relative to $S_{ref} = 34.8$ derived from Arctic winter climatology (1950s, 1960s, 1970s, and 1980s) from the Environmental Working Group Joint U.S. Russian Atlas of the Arctic Ocean (EWG) [*Arctic Climatology Project*, 1997].

Bering Strait: the inflow during summer is characterized by a shallow temperature maximum (Pacific T_{max}) found between 50 and 120 m at $31 < S < 32$ [*Coachman and Barnes*, 1961; *Shimada et al*., 2001; *Steele et al*., 2004] and during winter the inflow is characterized by a temperature minimum at $S = 33.1$ [*Coachman and Barnes*, 1961]. Pacific waters are also nutrient rich and the nutrient maximum is found in the winter water. Waters with a Pacific T_{max} can be subdivided into Alaskan Coastal Water (ACW) and summer Bering Sea Water (sBSW) of which the former is the freshest and shallowest of the two [*Steele et al*., 2004]. sBSW can be found in the Arctic as far east as in the Lincoln Sea, although with a reduced temperature signature due to vertical diffusion along its pathway [*Steele et al*., 2004].

[7] The deeper lying temperature maximum of the Atlantic water (Atlantic T_{max}) can be seen throughout the Arctic between 200 and 450 m depth [*Aagaard*, 1989; *Rudels et al*., 1994; *Carmack et al*., 1997; *McLaughlin et al*., 2002]. The Atlantic T_{max} is warmest (up to 2.5°C) and found highest in the water column along the boundary of the Eurasian Basin and subsequently becomes colder and deeper as it circles around the Arctic. As this warm Atlantic water progresses along the slope into the Eurasian and Makarov basins and mixes with colder interior waters it becomes marked by thermohaline intrusions around the Atlantic T_{max} [*Walsh and Carmack*, 2003; *Woodgate et al*., 2007]. By the time the Atlantic water in the boundary current has reached the southeastern Canada Basin it has cooled to ~0.5°C and it occurs at a larger depth (<400 m). Until recently, there were no thermohaline intrusions because temperatures in the boundary current were the same as in the interior basin. However, the warm Atlantic water anomaly that entered the Arctic in the early 1990s [*Quadfasel et al*., 1991; *Carmack et al*., 1995] reached the western Canada Basin by 2002, continued to spread into the interior in 2007 [*McLaughlin et al*., 2009] increasing the local Atlantic T_{max} to ~0.7°C which led to the appearance of thermohaline intrusions in T-S space there [*Woodgate et al*., 2007; *McLaughlin et al*., 2009].

[8] The front between Atlantic and Pacific water masses has varied over time, extending somewhere between, or even passed, the Lomonosov and the Mendeleev ridges [*McLaughlin et al*., 1996; *Carmack et al*., 1997; *Morison et al*., 1998; *Swift et al*., 2005]. In the Lincoln Sea water masses of the Canada Basin with a Pacific imprint are generally found inshore of the shelf break along the Canadian slope [*Rudels et al*., 2004; *Falck et al*., 2005]. Further offshore, parallel to the slope of the Lincoln Sea, contributions from the other Arctic basins, including upper waters as well deeper Atlantic derived layers, are found. In addition, the influx and circulation of Pacific and Atlantic waters in the Arctic Ocean are variable and thus the characteristics and composition of water mass properties in the Lincoln Sea also vary in space and time.

3. Data and Methods

[9] Freshwater content (FWC) in the ocean serves as a measure of the vertically integrated salinity anomaly relative to a reference salinity. In this study, it is determined from Conductivity-Temperature-Depth (CTD) data collected through several large monitoring projects in the 2000s in the Arctic Ocean. These observational programs cover different regions, time periods and seasons (Table 1). For the Beaufort Sea, the Laptev Sea and the Nares and Fram straits, ship-based or airborne data were collected in (late) summer, whereas North Pole and the Lincoln Sea data were collected mostly in spring. Ice-Tethered Profiler (ITP) data are also included for the same spring and summer months. The FWC for each hydrographic station in the Arctic Ocean is determined according to

$$FWC = \int_{z=1\,m}^{z=D(S_{ref})} \frac{S_{ref} - S(z)}{S_{ref}} dz, \tag{1}$$

where FWC is the thickness in meters of a layer of pure freshwater required to reduce salinity to the observed value from a reference salinity S_{ref}. $S(z)$ is the salinity at depth z. The FWC is calculated by integrating from 1 m depth down to the depth D of isohaline S_{ref} such that negative freshwater thickness is not allowed. Here, a value of 34.8 for S_{ref} is used, the mean salinity of the Arctic Ocean and approximate salinity of the main ventilated Atlantic inflow branches in the Fram Strait and Barents Sea [*Aagaard and*

Table 1. Observational Programs Contributing Data Used Here: Freshwater Switchyard (SY), GreenArc (GA), North Pole Environmental Observatory (NPEO), Lomonosov Ridge off Greenland (LOMROG II), SCience ICe EXercise (SCICEX), Beaufort Gyre Exploration Project (BGEP), Nansen and Amundsen Basins Observational System (NABOS), Canadian Archipelago Throughflow Study (CATS), Fram Strait (FS), and Ice-Tethered Profiler (ITP)[a]

Program	Region	Time Period	Month	Geographic Boundaries
SY	Lincoln Sea (I)	2003–2011	Apr–May	30 – 75°W, 83 – 87°N
GA	Lincoln Sea (I)	2009	Apr–May	30 – 75°W, 83 – 87°N
NPEO	North Pole (II, III)	2000–2010	Apr	North of 87°N
LOMROG II	Amundsen Basin (III)	2009	Jul	180°W – 160°E, 86 – 90°N
NABOS	Laptev Sea (IV)	2002–2009	Aug–Oct	125 – 165°E, 78 – 82°N
SCICEX	Beaufort Gyre (V)	2000	Jul	130 – 160°W, 71 – 80°N
BGEP	Beaufort Gyre (V)	2003–2010	Aug–Sep	130 – 160°W, 71 – 80°N
CATS	Nares Strait (VI)	2003, 2006, 2009	Aug	60 – 72°W, 72 – 83°N
FS	Fram Strait (VII)	2000–2010	Sep	0 – 16°W, 77 – 81°N
ITP	Arctic Ocean	2006–2011	Apr–May	North of 74°
ITP	Arctic Ocean	2004–2011	Jul–Sep	North of 74°

[a]Roman numbers mark the regions corresponding with the upper left panel of Figure 2.

Carmack, 1989]. For reference, the FWC based on winter climatology of the Arctic Ocean obtained through the Environmental Working Group Joint U.S.-Russian Atlas of the Arctic Ocean (EWG) [*Arctic Climatology Project*, 1997] is shown relative to $S_{ref} = 34.8$ (Figure 1b). The EWG climatology is based on all available hydrographic profiles from four decades (1950–1980). One should be aware that data coverage in the Lincoln Sea was not always very good during those four decades, e.g., in the 1950s and 1960s. However, in the 1970s there was very reasonable coverage [*Polyakov et al.*, 2008].

[10] The sensitivity of FWC to the choice of S_{ref} and the integral depth was examined, e.g., results were compared with a calculation where S_{ref} varies spatially by defining it as the vertical mean salinity at each grid cell from EWG climatology. A comparison was also made with FWC integrated down to constant depth levels of 200 and 500 m. In all cases, no qualitative differences in the variability of FWC were found using these different integral boundaries, and hence, the results presented here are robust.

[11] Freshwater content variability can be either due to changes in the upper-ocean salinity within a fixed layer, or by changing the layer thickness, e.g., by downward or upward movement of the depth of the isohaline S_{ref}, or by both [*Rabe et al.*, 2011]. The relative contributions of thickness changes, salinity changes, and their combined effect (cross term) to FWC changes are determined by the following terms, respectively.

$$\Delta FWC = \Delta h\left(1 - \frac{S_1}{S_{ref}}\right) - h_1\frac{\Delta S}{S_{ref}} - \Delta h\frac{\Delta S}{S_{ref}} \tag{2}$$

where S_1 is the climatological mean salinity above the isohaline S_{ref}, h_1 is the climatological layer depth of S_{ref}, and ΔS and Δh are the differences in mean salinity and layer depth relative to climatology at a certain time.

[12] All salinity data used in this study are interpolated on 1 m depth bins. In case of data deficits in the surface layer, varying from a couple up to 10 m at times, the salinity profile is extended upward by assuming the uppermost value is constant to the surface. This conservative assumption may underestimate the FWC in summer if thin and fresh, surface layers would be present but not measured. For example, a shallow surface layer of 10 m, which is 6 salinity units fresher than the water below it due to summer sea-ice melt, amounts to an error of 1.7 m in FWC in summer. Here FWC calculated for the Laptev Sea is most prone to that error because of data gaps in the upper water column in some years.

4. Results

[13] The regional distribution of the salinity at 10 m depth is shown for 2003, 2007, 2008, 2009, 2010, and 2011 (Figure 2). Triangles and circles indicate whether the data were collected in spring or in summer/fall, respectively. All panels generally show the climatological distribution of upper-ocean salinity in the Arctic (Figure 1b) [*Aagaard and Carmack*, 1989]. Between 2003 and 2007, the salinity decreased in the Canada Basin [*Proshutinsky et al.*, 2009; *Yamamoto-Kawai et al.*, 2009; *McPhee et al.*, 2009] and a slight increase occurred in the Eurasian Basin. Very fresh surface waters (S $\sim$ 25.5 in the upper 20 m) were observed on the slope between the East Siberian Sea and the Makarov Basin in 2007. After 2008, the surface salinity started to decline in the western Lincoln Sea and minimum salinities were found over the central Arctic surrounding the North Pole and the Lincoln Sea in 2009. In 2010, the salinity north of Greenland was still anomalously low but it had increased again near the North Pole. In 2011, the salinity in all of the Lincoln Sea returned to high surface salinity values similar to those observed in 2003.

[14] The FWC anomaly relative to the EWG Arctic climatology was calculated for all salinity profiles (Figure 3). The FWC anomaly was clearly largest in the Beaufort Gyre of the Canada Basin in 2008 and 2009 in agreement with [*Proshutinsky et al.*, 2009; *Yamamoto-Kawai et al.*, 2009; *McPhee et al.*, 2009]. Here the FWC remained large in 2010. With lower data coverage in the southern Canada Basin as presented here we cannot draw strong conclusions about 2011. On the few available stations in the Lincoln Sea in 2007, the FWC showed a clear increase compared to earlier years. The largest increase in FWC occurred in the western Lincoln Sea and north of Ellesmere Island in 2009, amounting locally to 4 m of additional freshwater relative to climatology. The FWC anomaly was still large in the Lincoln Sea in 2010 but decreased sharply in 2011.

[15] To investigate the observed increase of FWC in the deep basin (deeper than 300 m) of the Lincoln Sea and the central Arctic north of 80°N we show the contribution of depth changes of S_{ref}, the first term on the right-hand side of equation (2) (Figure 4). The salinity on the shelf of the Lincoln Sea (shallower than 300 m) is generally less than S_{ref}. In general the contribution of the depth change to the FWC anomaly was small in the 2000s. Positive anomalies (i.e., deepening) were seen only at the stations on the slope in the Lincoln Sea just north of Nares Strait. In 2009, particularly an increase of isohaline depth contributed to almost 2 m FWC increase on the shelf slope. In the off-slope region of the Lincoln Sea and around the North Pole, the contribution of isohaline depth change to the FWC anomaly was close to zero. The contribution of the change of mean salinity above S_{ref}, the second term in equation (2), was large in the Makarov and northern Canada basins between 2007 and 2009 (Figure 5). In the Lincoln Sea, the salinity term increased significantly from 2007 to 2010 with the exception of the shelf slope stations where the change in isohaline depth dominated. The influence of the salinity term was largest in 2009 and returned to zero by 2011. The cross term, the last term in the right-hand side of equation (2), was in general small in the Lincoln Sea (not shown). The contribution of the cross term expressed as a percentage of the total FWC anomaly was determined for FWC anomalies larger than 1.2 m and was usually around 5% in the Lincoln Sea with exceptions up to 20% in the central Arctic.

[16] A time series of the mean FWC and its standard error on a north-south section in the Lincoln Sea was determined from the 1950s to1980s EWG climatology, from the available data from the 1990s [*Newton and Sotirin*, 1997], and from the data of the first decade of the 2000s (Figure 6a). This section is at approximately 83.5–84.5°N,65 (see insets in Figures 7a and 7b) and was selected based on the available data from the 1990s and the repeat stations in the

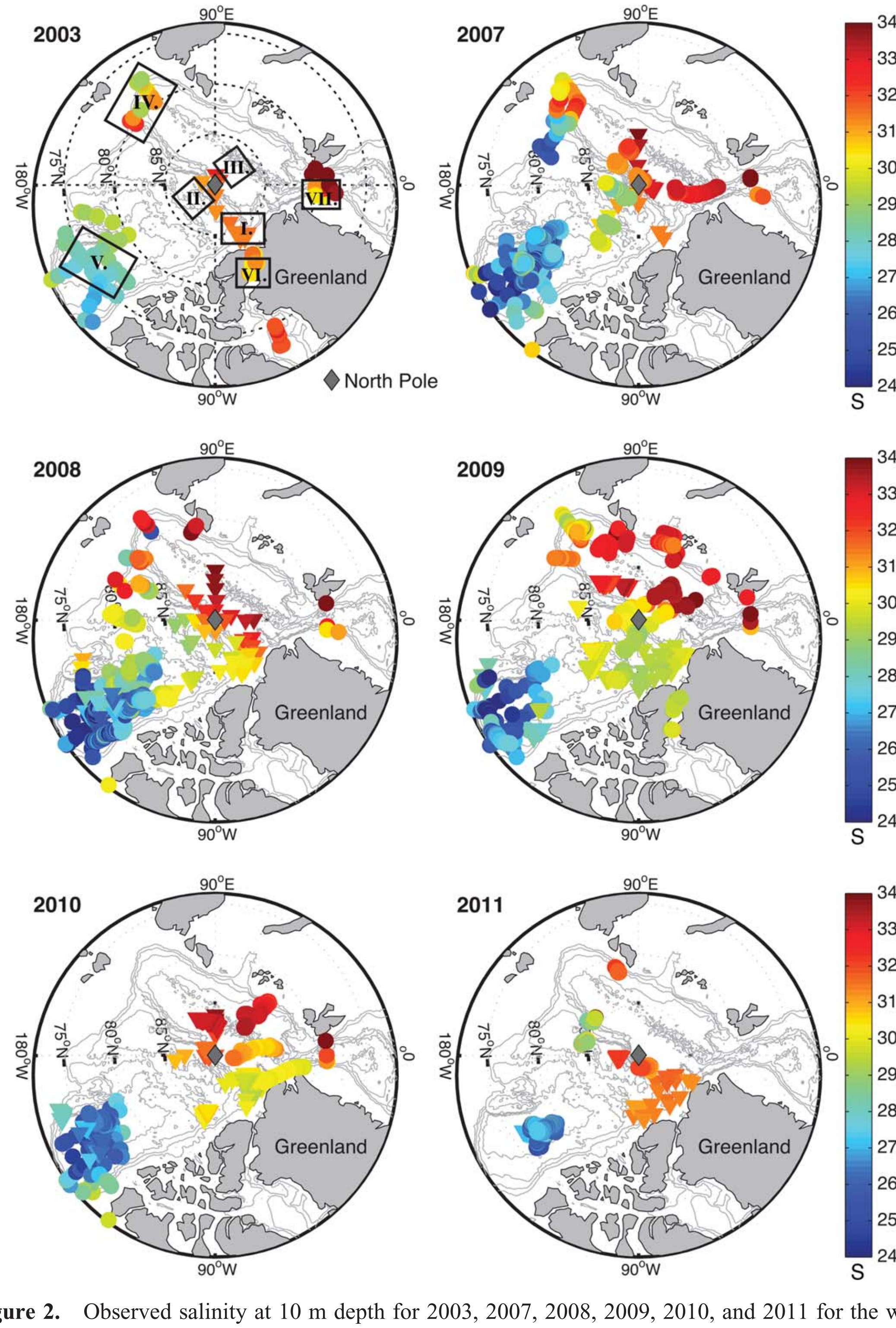

Figure 2. Observed salinity at 10 m depth for 2003, 2007, 2008, 2009, 2010, and 2011 for the whole Arctic. Triangles indicate spring data, circles are summer/fall data. In the first row separate regions are marked corresponding with the regions in Table 1. Please note that the scale of the circles and triangles does not mimic the scale of the actual observation, which is a vertical single profile.

2000s. Between 1955 and 1975, the FWC on the section was low but increased substantially in the 1980s. Interannual variability occurred both in the 1990s and the 2000s with maxima in 1994 and 2009, respectively. Even though very high FWC was seen between 2007 and 2010, the decadal mean FWC on the section (magenta lines) in the 2000s was only 0.5 m larger than the 5 year mean FWC in the 1990s. The total freshwater volume in the Lincoln Sea was also estimated (Figure 6b). For this purpose, the FWC from 1990s to 2000s data was interpolated onto the same

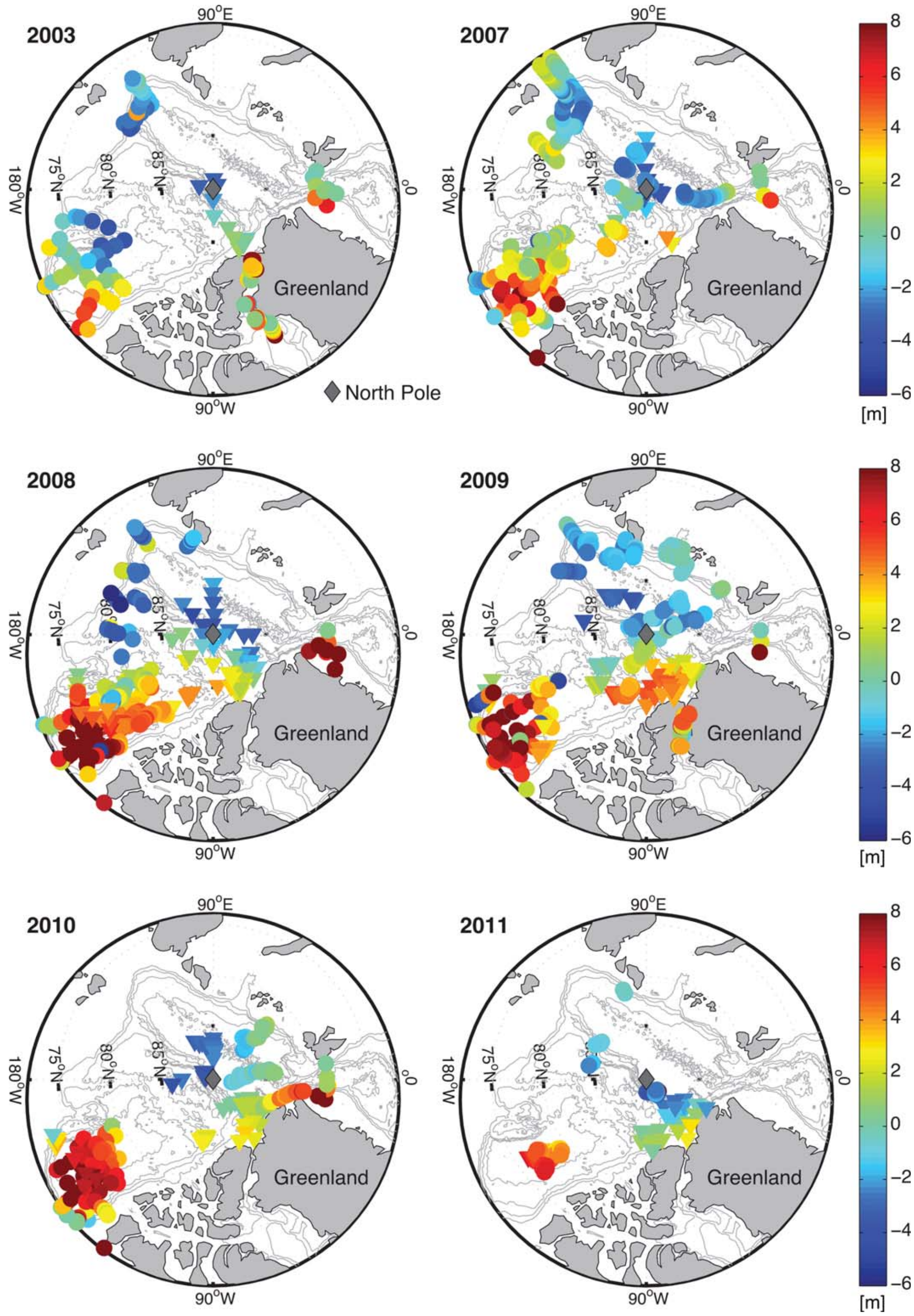

Figure 3. FWC anomaly relative to EWG climatology for 2003, 2007, 2008, 2009, 2010, and 2011 for the whole Arctic.

grid as the EWG climatology over the whole Lincoln Sea (defined within $83.5^\circ - 87.5^\circ$N, $95^\circ - 20^\circ$W). To extrapolate to grid cells that did not contain data in the 1990s and 2000s, a regression was performed onto the EWG climatology for each year. Then the FWC per grid cell was integrated over the Lincoln Sea. The total FW volume decreased from the 1950s to the 1960s and increased again throughout the 1970s and 1980s. An increase of almost 950 km^3 was found in the early 1990s, followed by decline throughout the early 2000s. The second large anomaly in FW volume of approximately $1100 \pm 250\,\text{km}^3$ occurred between 2003 and 2009. The positive FW anomaly in 2009 was fully discharged from the Lincoln Sea by spring 2011. The sensitivity of the result to this method was determined

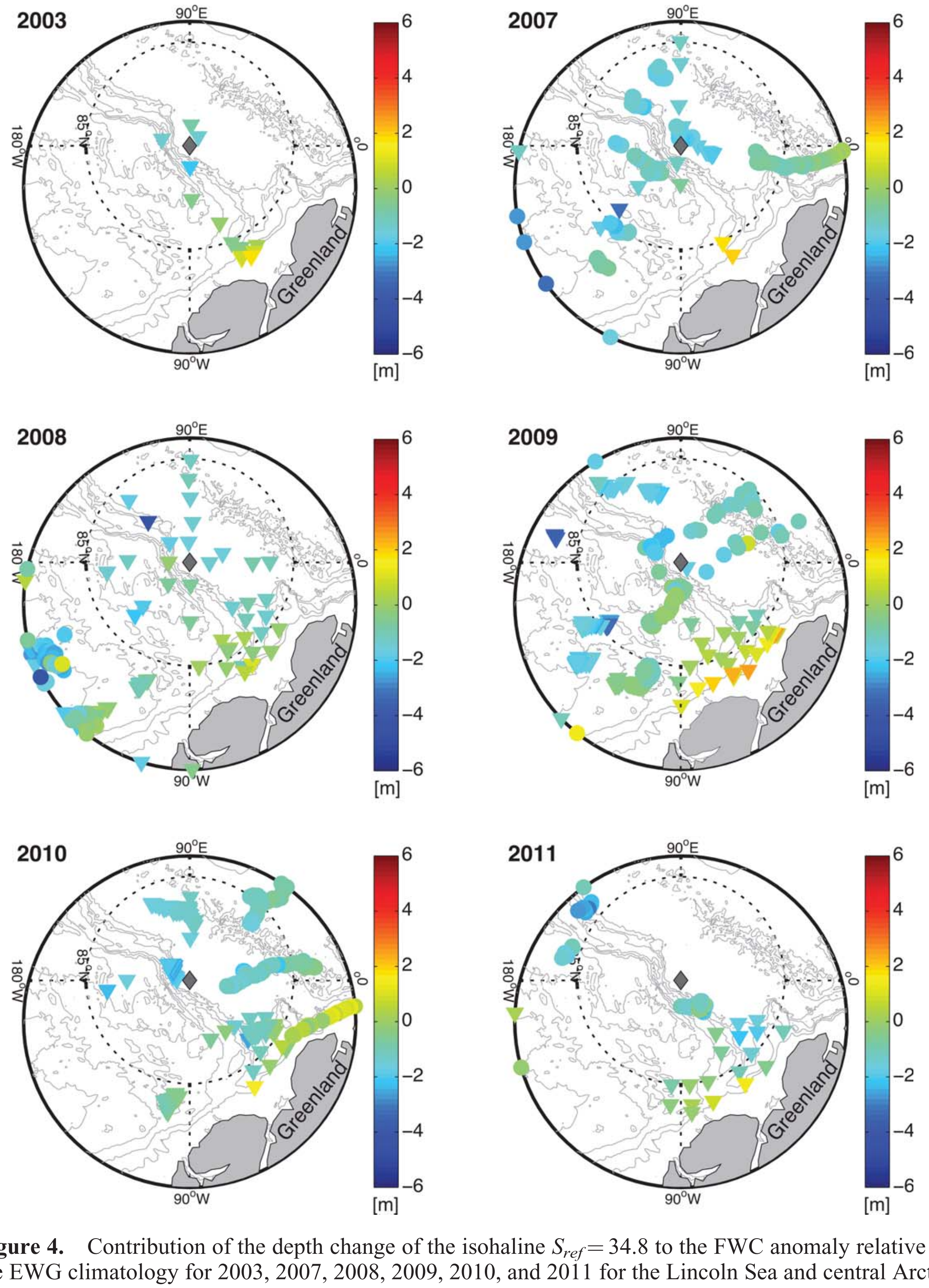

Figure 4. Contribution of the depth change of the isohaline $S_{ref} = 34.8$ to the FWC anomaly relative to the EWG climatology for 2003, 2007, 2008, 2009, 2010, and 2011 for the Lincoln Sea and central Arctic north of 80°N for profiles deeper than 300 m.

by a comparison of the results of three different subsets where we arbitrarily left out station data. This showed that the uncertainty in the volume estimate related to the extrapolation to grid cells without data is $\pm 150 km^3$. If we calculate the standard error of the calculated mean FWC for each year and multiply that with the area of the Lincoln Sea we obtain an error of $\pm 250\,km^3$. Even though the largest FWC was observed in 2009, the decadal mean for the 2000s was smaller than that calculated for the mid 1990s. It should be kept in mind that data from the 1990s was limited in space and time (just 5 years) so the estimate of a mean FW volume for the 1990s is likely biased. The same holds for the 1950s and 1960s when the data coverage in the Lincoln Sea was sparse.

4.1. Observed Water Mass Properties and Variability

[17] The observed freshening in the Lincoln Sea is likely caused by one or a combination of the following possible

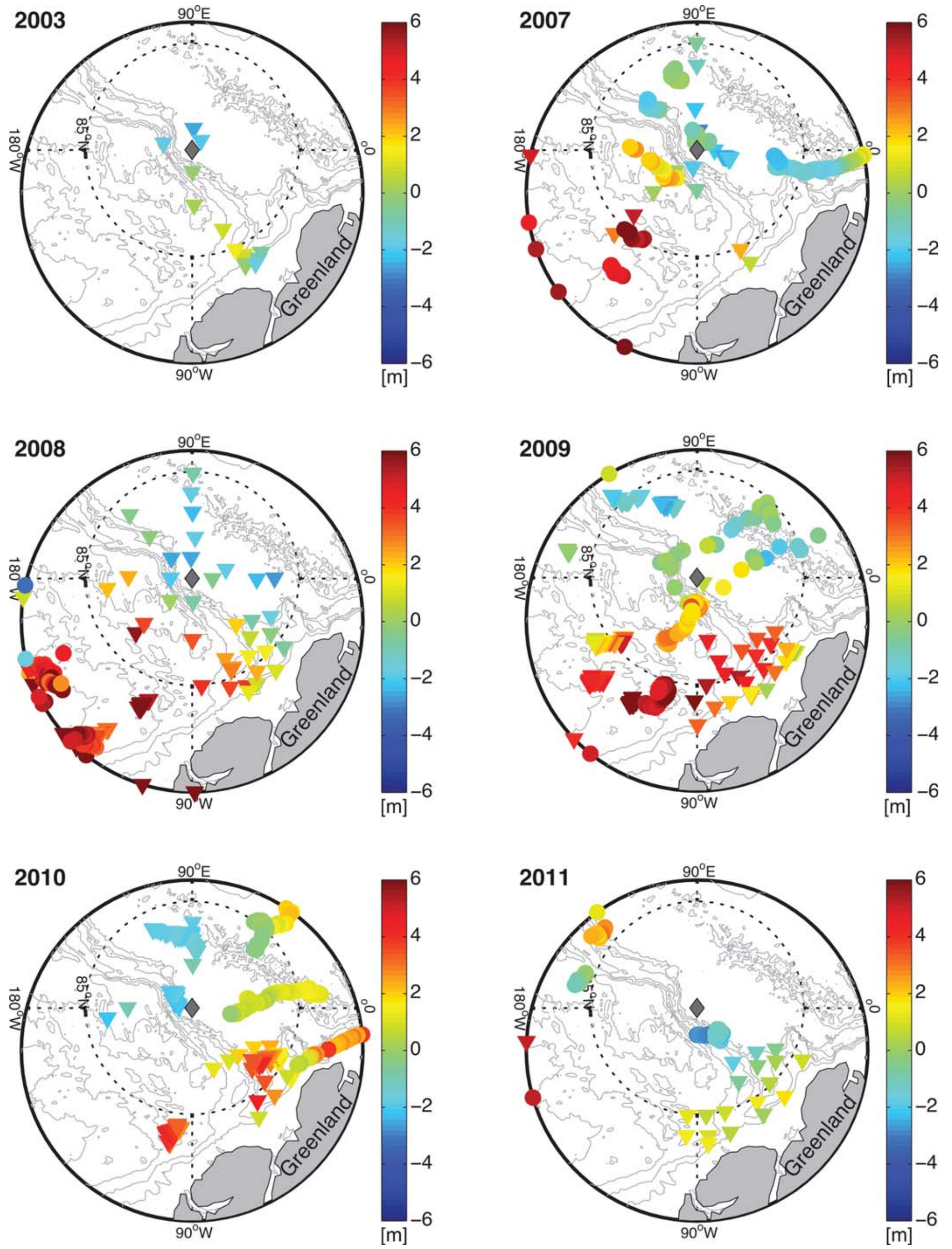

Figure 5. Contribution of mean salinity change above the isohaline $S_{ref} = 34.8$ to the FWC anomaly relative to the EWG climatology for 2003, 2007, 2008, 2009, 2010, and 2011 for the Lincoln Sea and central Arctic north of 80°N for profiles deeper than 300 m.

advective processes: a release of freshwater from the Beaufort Gyre, an increase of transport of freshwater from the Siberian shelves by the Transpolar Drift Stream, and a change in pathways of freshwater. Here we examine the variability of the upper ocean water mass properties in the Lincoln Sea, with specific attention to the presence/absence of a Pacific T_{max} and silicate concentrations in the upper ocean to identify the source region of the observed freshwater anomaly in the 2000s. Second, observed variations in the Atlantic T_{max} below 2∼50 m are discussed.

[18] The Pacific and Atlantic T_{max} in T-S space in the Lincoln Sea on the section at ∼65°W are compared using spring profiles from the 1990s to 2003, 2008, and 2009 (Figure 7). Low surface salinity waters with $S = 30.5$ were observed in 1993, 1994, and 1996 but all 5 years were associated with a well-defined warm Pacific T_{max} between −1.4 and −1.2°C [*Newton and Sotirin*, 1997] (Figure 7a). The deeper-lying Atlantic T_{max} was close to 0.5°C. Data on the same section in the 2000s show that surface salinities were more saline ($S = 31.5$) in 2003, about one salinity

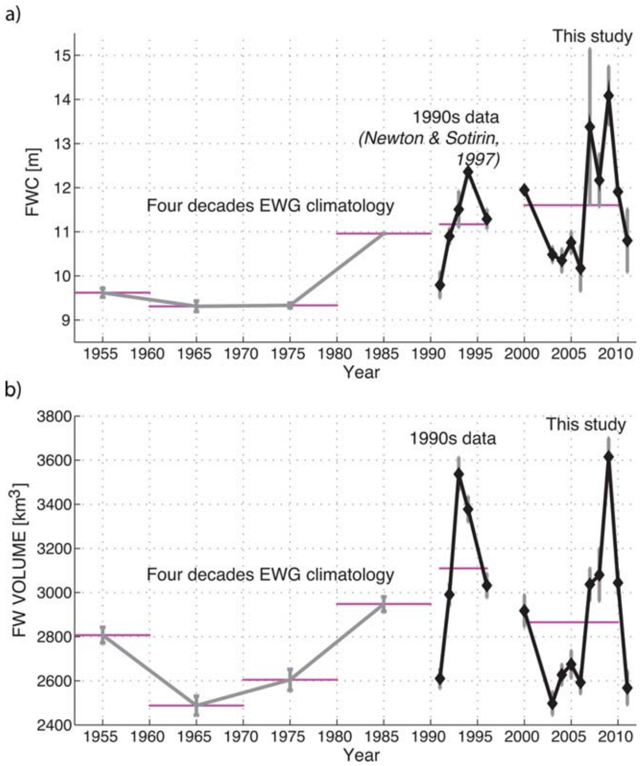

Figure 6. (a) Time series of the mean FWC [m] and the standard error (gray error bars) on a north-south section at 83.5 – 84.4°N, 65°W, in the Lincoln Sea including four decades EWG climatology, the available data from the 1990s, and from the 2000s Switchyard program (see Table 1). The magenta lines illustrate the decadal mean or 5 year mean in the case of the 1990s. (b) Time series of the mean freshwater volume (km^3) and the standard error within the Lincoln Sea bounded by 83.5° – 87.5°N, 95° – 20°W.

unit fresher ($S = 30.4$) in 2008, and freshest ($S = 29.4$) in 2009 (Figure 7b). The Pacific T_{max} was very weak in 2003, i.e., barely above the freezing point, and was significantly warmer in 2008. Even though surface waters observed in 2009 where fresher than in 2008 they showed a slightly colder Pacific T_{max}. The Atlantic T_{max} in 2008 was clearly warmer than in the 1990s. A time series constructed from stations between 83.8 – 84.5°N shows that Pacific T_{max} in the 2000s was significantly lower than between 1992 and 1994 (Figure 7c). However, from 2004 onward the Pacific T_{max} started to increase on the section. The deeper lying Atlantic T_{max} was nearly constant (close to 0.44°C) in the 1990s, while in the 2000s an increase of the Atlantic T_{max} was observed as of 2004.

[19] Vertical S and T profiles and T-S diagrams are shown for the hydrographic stations in the whole Lincoln Sea for 2003, 2008, 2009, 2010, and 2011 (Figures 8 and 9). The profiles are color coded by longitude going from west (blue) to east (red). In 2003, the stations were confined to a small portion of the Lincoln Sea (Figure 8, top). S profiles and T-S diagrams were all very similar with the freshest water found to the west and a weak Pacific T_{max} at $32 < S < 33$ including sBSW. T profiles showed small differences in the Atlantic T_{max} varying from smooth profiles with T_{max} around 450 m to profiles that show intrusions at warmer T_{max} between 250 and 350 m depth. From 2008 onward, there was a much larger coverage of the Lincoln Sea (Figure 8, middle). In 2008, the profiles taken at a similar location as in 2003 were fresher by one salinity unit in the upper 50 m and show a much warmer Pacific T_{max}. In general, the S, T, and T-S profiles showed a very clear dependency on longitude; from east to west surface waters become more saline, the Pacific T_{max} of the sBSW at approximately 80–100 m depth disappears, and a strong increase in the underlying Atlantic T_{max} is seen. Large differences were seen in 2009 (Figure 8, bottom): all surface waters were fresher by two salinity units and the Pacific T_{max} was weaker than in 2008. Profiles on the shelf slope further east (red profiles) had a lower Atlantic T_{max} of around 0.6°C than off-slope stations which had $T_{max} \sim 1.0$°C.

[20] In 2010 (Figure 9, top), there was little coverage of the western Lincoln Sea. The surface waters in the eastern Lincoln Sea were still as fresh as in 2009. Now an upper ocean T_{max} at low salinity ($S < 31$) was observed; too fresh and too shallow to be identified as a Pacific T_{max}. This near surface T_{max} was likely a remnant of warming during the previous summer, identified by *Jackson et al.* [2011] as a near-surface temperature maximum (NSTM). All the stations east of 60°W showed clear intrusions in T profiles around the Atlantic T_{max}. Finally, in 2011, the upper-ocean salinity in the whole Lincoln Sea had increased again to values similar to 2003 (Figure 9, bottom). Now both western Lincoln Sea stations (dark blue profiles) and eastern Lincoln stations (red profiles) showed an upper-ocean T_{max} at higher salinity values ($S \sim 33$) than in 2010. In the eastern sector, it occurred between 50 and 100 m depth and in the western sector at 120–140 m depth and above that a fresher Pacific T_{max} was present.

4.2. Upstream Hydrographic Properties

[21] Hydrographic properties further upstream from the Lincoln Sea are investigated here with focus on the potential pathways of the different upper-ocean Pacific and deeper-lying Atlantic signatures.

4.2.1. Pacific Signatures

[22] The observations presented here show that the freshening in the Lincoln Sea between 2003 and 2008 coincided with an increase of Pacific T_{max} from −1.6°C to −1.4°C in the western sector of the Lincoln Sea (Figure 7). The Pacific T_{max} was not as strong as was seen in the mid-1990s and in the eastern Lincoln Sea it remained generally below −1.5°C or was nonexistent. This suggests that a different source or different circulation of upper ocean fresh waters had also contributed to the observed freshening there.

[23] Vertical S and T profiles from ITP and CTD data from the central Arctic are compared for 2008–2010 (Figure 10). The profiles are from different locations each year but we can identify some general features in the northern Canada and Makarov basins. Low salinities were seen in 2007 north of the Alpha Ridge in the Makarov basin (see Figure 2) and in the upper 50 m near the North Pole in 2008 (blue profiles in Figure 10, top left). These surface waters were fresher than observed in the northern and northeastern Canada Basin (magenta and cyan profiles Figure 10. There was no clear Pacific T_{max} present near the North Pole (blue profiles), while the fresh 100 m layer in

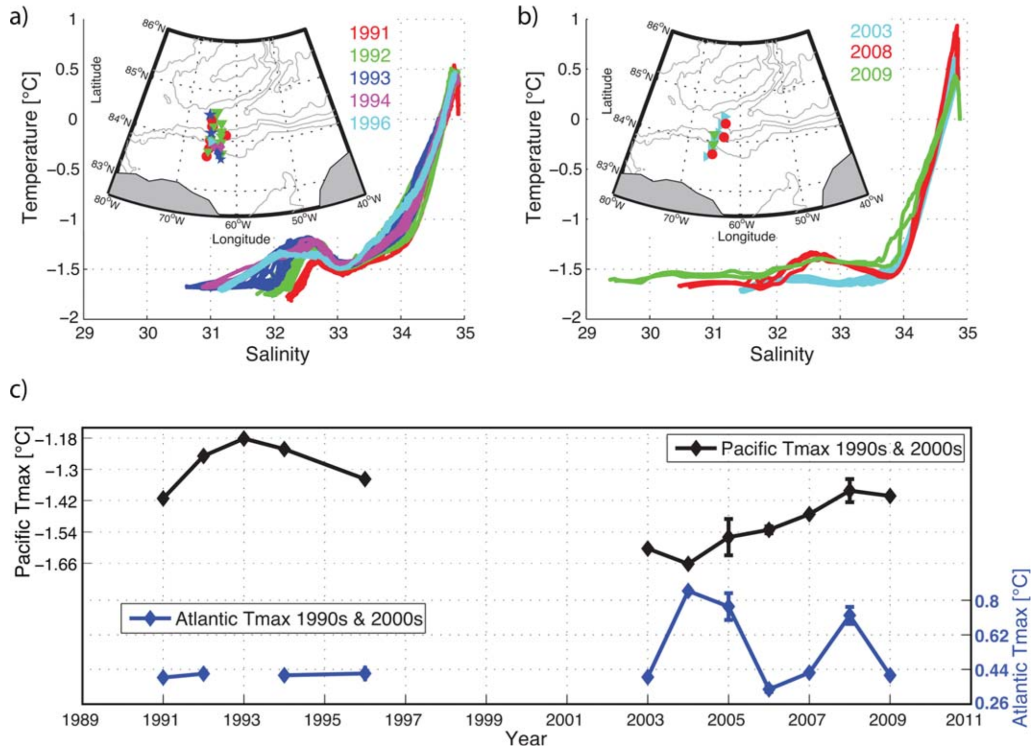

Figure 7. T-S diagrams and station locations (inset) in (a) the 1990s, (b) the 2000s, and (c) The Pacific T_{max} and the standard error (black line) and the Atlantic T_{max} and its standard error for repeat stations on the north-south section at 65°W between 83.8 − 84.5°N in the western Lincoln Sea.

the northern Canada Basin (magenta profiles) show a very warm Pacific T_{max} in 2008. In the northeastern Canada Basin (cyan profiles), two local T_{max} are seen in the upper 150 m: a weak NSTM near 40 m, a remnant from previous summer solar heating [*Jackson et al*., 2011], and a larger Pacific T_{max} at 120 m.

[24] In 2009, the blue profiles in the southern Makarov Basin show very fresh upper ocean water with a NSTM at 40–50 m. The magenta profiles from the Makarov Basin have a relatively cold Pacific T_{max} around 100 m depth. The latter were also seen in the Lincoln Sea in 2010 (see orange profiles in Figure 9). In 2010, the upper ocean salinity in the central Arctic increased again while the eastern Canada Basin surface salinity remained low. In 2010, profiles near the North Pole show a Pacific T_{max} between 50 and 100 m at a relative high salinity of S ~33. Identical profiles with this T_{max} were seen in the eastern Lincoln Sea in spring 2011, illustrating the strong linkage between these two regions. By then the upper ocean salinity (<50 m) in the Lincoln Sea had increased again to pre-2008 values. For comparison profiles from the Amundsen Basin (green profiles) are included which in general were less fresh than from the Makarov and Canadian basins.

[25] Vertical silicate profiles and silicate-salinity plots based on bottle data are shown for the western Lincoln Sea and North Pole for 2008, 2009, and 2010 to confirm the presence or absence of Pacific water (Figure 11). Large silicate values were observed at 100 m depth in the western Lincoln Sea in all 3 years implying a strong connection with Pacific waters from the southern Canada Basin [*McLaughlin et al*., 2004]. Near the North Pole elevated silicate values indicative of Pacific origin waters were seen in 2009 and 2010 but not in 2008. The latter is in agreement with *Alkire et al*. [2010] who showed that the Pacific freshwater fraction was very small in the northern Makarov Basin in 2008. Large freshwater inventories in the eastern Canada Basin were associated with river water from the Siberian shelves [*Yamamoto-Kawai et al*., 2009; *Alkire et al*., 2010; *Morison et al*., 2012]. Although Pacific water with high silicate values in the northern Makarov Basin in 2009 and 2010 occurred at shallower depth than in the Lincoln Sea the maximum was associated with salinities S ~ 33.1, evident in the few square shape profiles. In 2010, one profile with high silicate showed that even though Pacific Water was present the salinity had increased relative to 2008 and 2009 (see also Figure 10).

4.2.2. Atlantic Signatures

[26] The Atlantic water in the Lincoln Sea off the shelf slope between about 84°N – 84.7°N and 55°W – 65°W (e.g., green and yellow profiles in 2010 in Figure 9) showed mostly southeastern Canada Basin characteristics, i.e., a cold Atlantic T_{max} around 450 m. Profiles with warmer Atlantic T_{max} and with thermohaline intrusions higher in the water column were seen further north and northeast of the shelf.

[27] Moving from the Eurasian Basin into the Makarov and the Canada basins in the central Arctic the Atlantic T_{max} decreases and the intrusions become smoother (Figure 10). The cyan profiles east of ~135°W in general show a colder Atlantic signature T_{max} ~ 0.5° below 350 m depth.

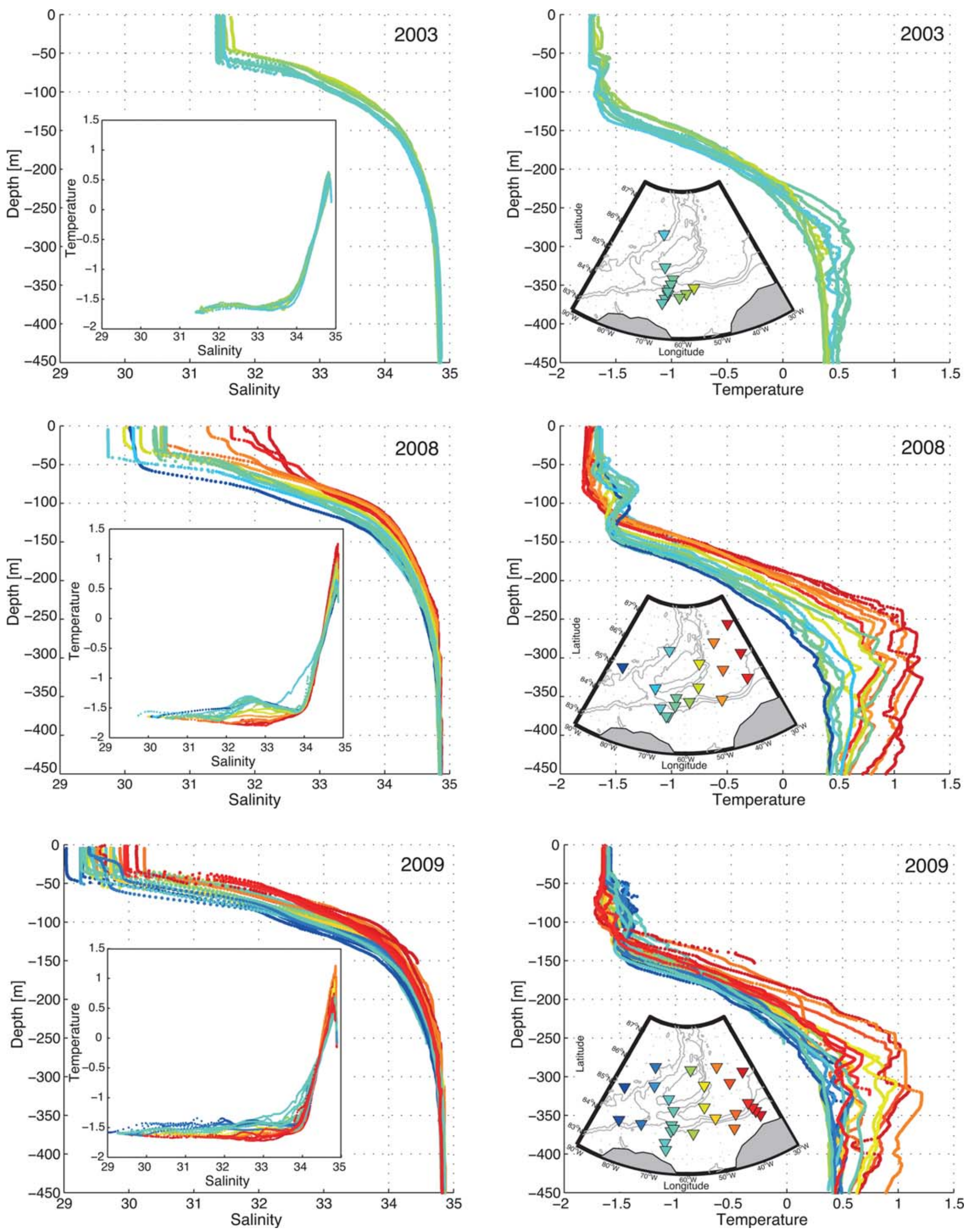

Figure 8. CTD profiles of salinity (left) and temperature (right) with insets of the T-S diagram (left column) and the station locations in the Lincoln Sea (right column) for 2003, 2008, and 2009. The profiles are color coded according to longitude.

The difference between the magenta profiles which have $T_{max} \sim 0.5 - 0.8°C$ and the blue profiles with Atlantic $T_{max} \sim 0.8 - 1°C$ north of the Mendejeev Ridge illustrates that the former are older and less ventilated while the latter arrive from the boundary current near the Eastern Siberian Shelves, potentially from the Chukchi Plain, into the interior along the Mendeleev Ridge. The Atlantic T_{max} in 2008 and 2010 in the northern Makarov Basin is much warmer compared to data from the early 1990s in the central Arctic when Atlantic T_{max} in the Makarov basin was below 0.5°C.

4.3. Upper Ocean Circulation From Dynamic Ocean Topography

[28] Dynamic ocean topography (DOT) determined from ICESat elevation data has been shown to agree very well with geostrophic velocities obtained from hydrography

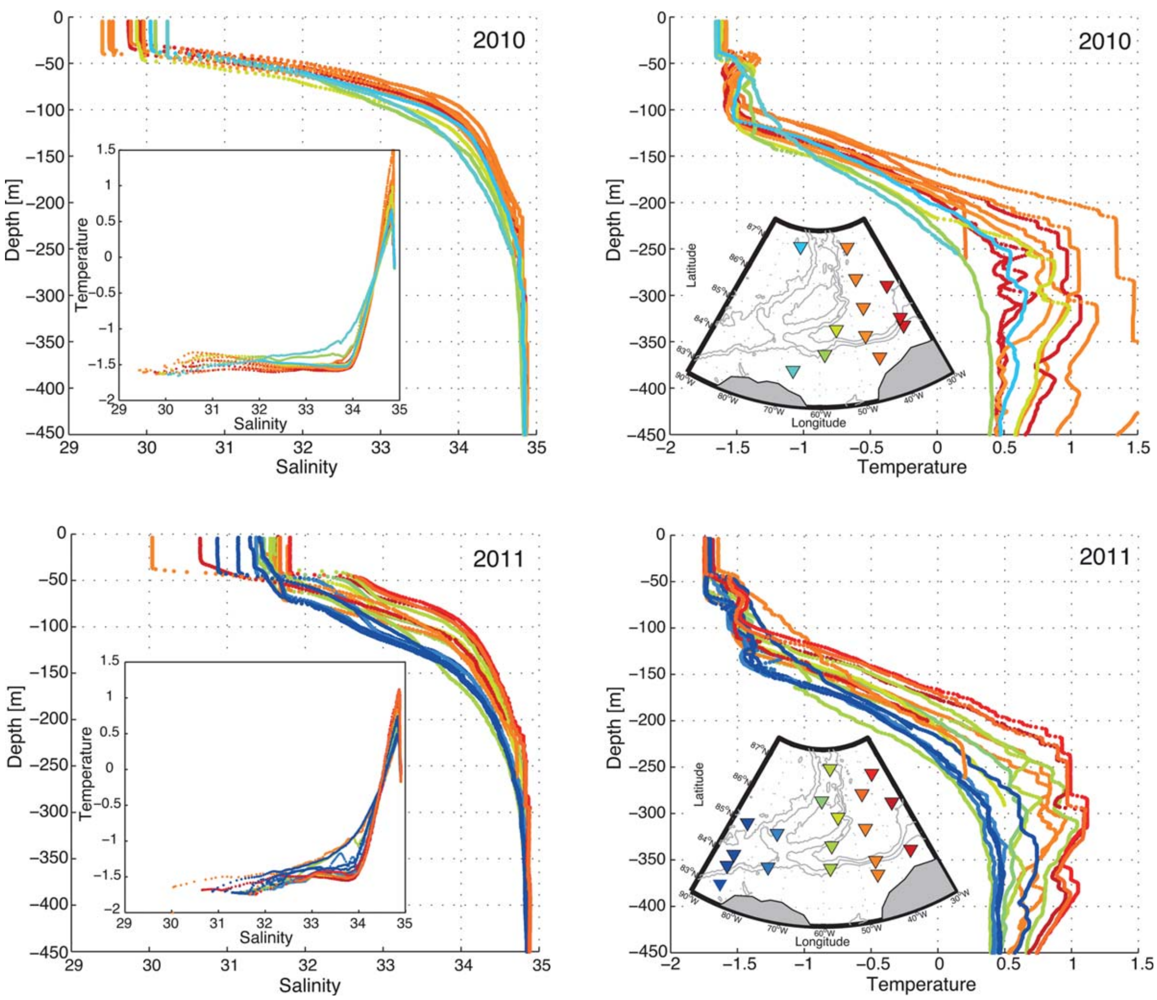

Figure 9. CTD profiles of salinity (left) and temperature (right) with insets of the T-S diagram (left column) and the station locations in the Lincoln Sea (right column) for 2010 and 2011. The profiles are color coded according to longitude.

[*Kwok and Morison*, 2011], where DOT is the sea-surface height relative to the geoid. We use it here to examine the difference in large-scale circulation in the Arctic Ocean between February and March 2008 and 2009 (Figure 12). The ICESat mission ended in 2010 so it is not possible to show 2010 or beyond. *Morison et al*. [2012] discuss the Arctic Ocean DOT and show how increasingly cyclonic circulation from 2005 to 2008 on the Russian side of the Arctic Ocean, associated with a positive AO, freshened the Canada Basin with Eurasian runoff. Here we compare 2008 DOT with 2009 DOT to investigate the FWC anomaly in the Lincoln Sea.

[29] The DOT and derived geostrophic velocities in 2008 show a clear transpolar drift stream along the Mendeleev Ridge, west of the Chukchi Plateau, into the Makarov Basin and along the Alpha Ridge northwest of the Lincoln Sea. DOT shows a ridge extending from the Canada Basin northeastward along the coast of Ellesmere Island. The flow is diverted eastward toward the Lomonosov Ridge and then anti-cyclonically around the DOT ridge toward northern Greenland. The observed hydrographic properties of a fresh surface layer with no or a small Pacific signal in the northern Makarov Basin in 2008 (Figure 10) and in the central to eastern Lincoln Sea are in line with this pathway from the Siberian shelves. The water mass with a clear Pacific T_{max} appeared to underlie the DOT ridge. A second surface flow pattern from the eastern Canada Basin near the Chukchi Plateau is found near 76°N. This stream splits near 80°N, one part extending northeast to 83°N and then turns south toward the southeastern Canada Basin, shaping the outermost edge of the Beaufort gyre. The other part immediately turns southeast to form part of a smaller closed gyre in the Beaufort Sea between 72°N and 78°N. Further east, branches are deflected from the Siberian Shelves into the Eurasian and Makarov basins, respectively [*Rudels et al*., 1994; *Carmack et al*., 1997], and these waters enter the central to eastern Lincoln Sea from the northwest and north.

[30] In 2009, the northwest boundary of anticyclonic circulation in the Canada Basin shifted counterclockwise relative to conditions in 2008. As a part of this shift, the ridge in DOT just northwest of the Lincoln Sea is stronger and is shifted further eastward forcing the upper ocean currents from the northern Makarov Basin near the Alpha Ridge toward the Lomonosov Ridge. The DOT ridge also forces strong southward flow toward Nares Strait in 2009 and little

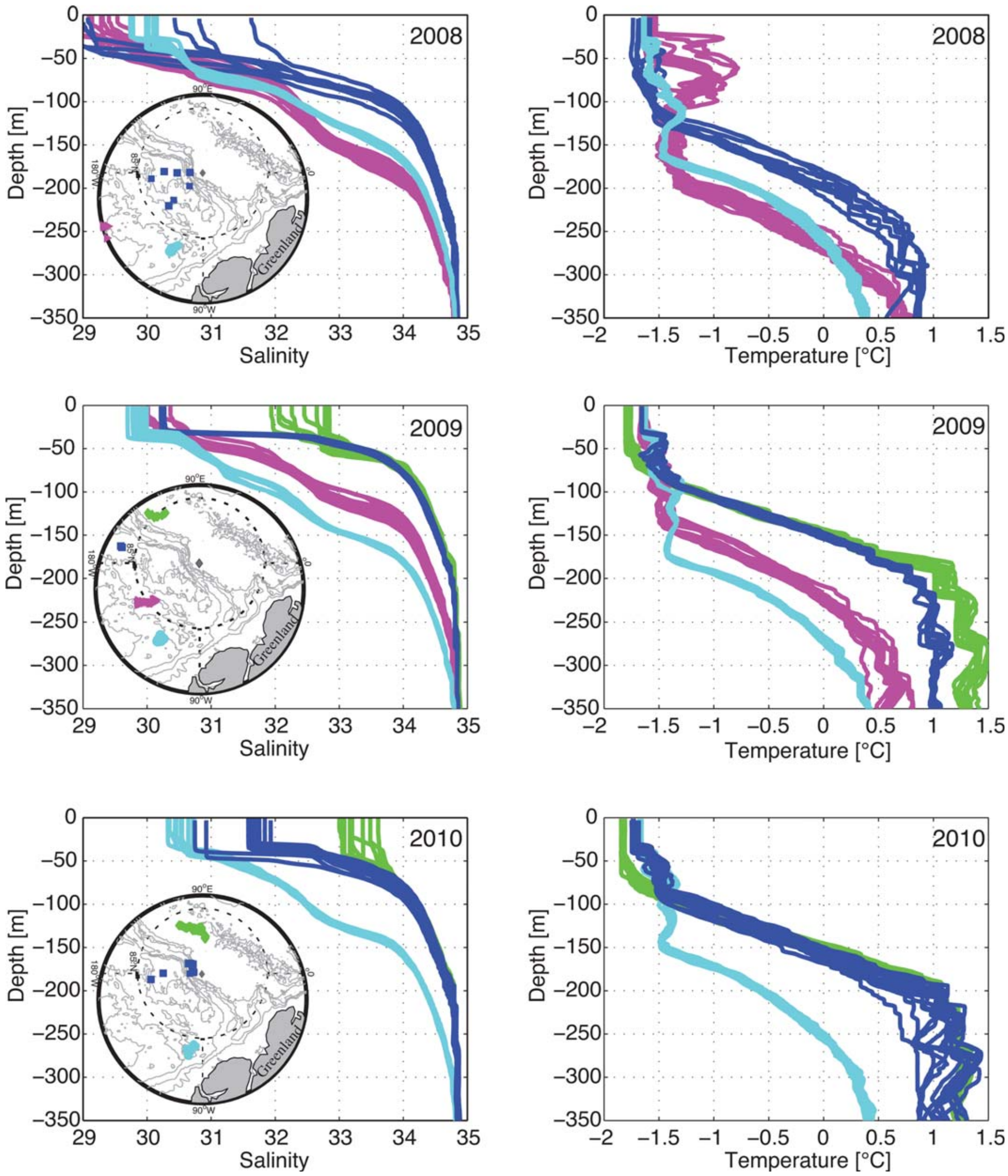

Figure 10. Vertical profiles of salinity (left) and temperature (right) in the central Arctic and Makarov Basin for 2008, 2009, and 2010 with insets of the station locations. The dark blue profiles in 2008 and 2010 were CTD stations from the NPEO program. The other profiles are ITP data.

toward Fram Strait. In 2009, the flow from the Chukchi Plateau is directed much more eastward, feeding the interior of the Canada Basin, whereas it was more northward in 2008. The central core of the Beaufort Gyre was oriented more east-west in 2009 than in 2008. From the DOT maps, one can see that an unusually strong flow was directed toward Greenland and Nares Strait in 2009, carrying anomalously fresh water across the central Arctic to the Lincoln Sea.

5. Discussion

[31] The lateral and temporal variability of the Pacific T_{max} was large in both the Makarov Basin and the Lincoln Sea in recent years. The absence of a Pacific signal in the northern Makarov Basin in 2007 and 2008 implies that the freshening there originated from sea-ice melt or unusually large Eurasian river input. The latter source has been confirmed by chemical analysis of samples from the Makarov Basin in 2007 and 2008 which showed that the largest contribution to freshwater in the upper 150 m was riverine water [*Alkire et al*., 2010; *Bauch et al*., 2011]. The sea-ice meltwater fraction was in fact negative in 2007. These fresh waters hence originated from the East Siberian Sea and took a direct route into the Makarov Basin, along the Mendeleev Ridge across to the North Pole and Lincoln Sea augmented by strong northeastward flow from the western Canada Basin toward the Lincoln Sea in spring 2009. In 2009, summer hydrographic data showed low salinity waters with a weak Pacific T_{max} in the northern part of the Makarov Basin near the North Pole. Waters with two salinity units lower than normal were also seen in the

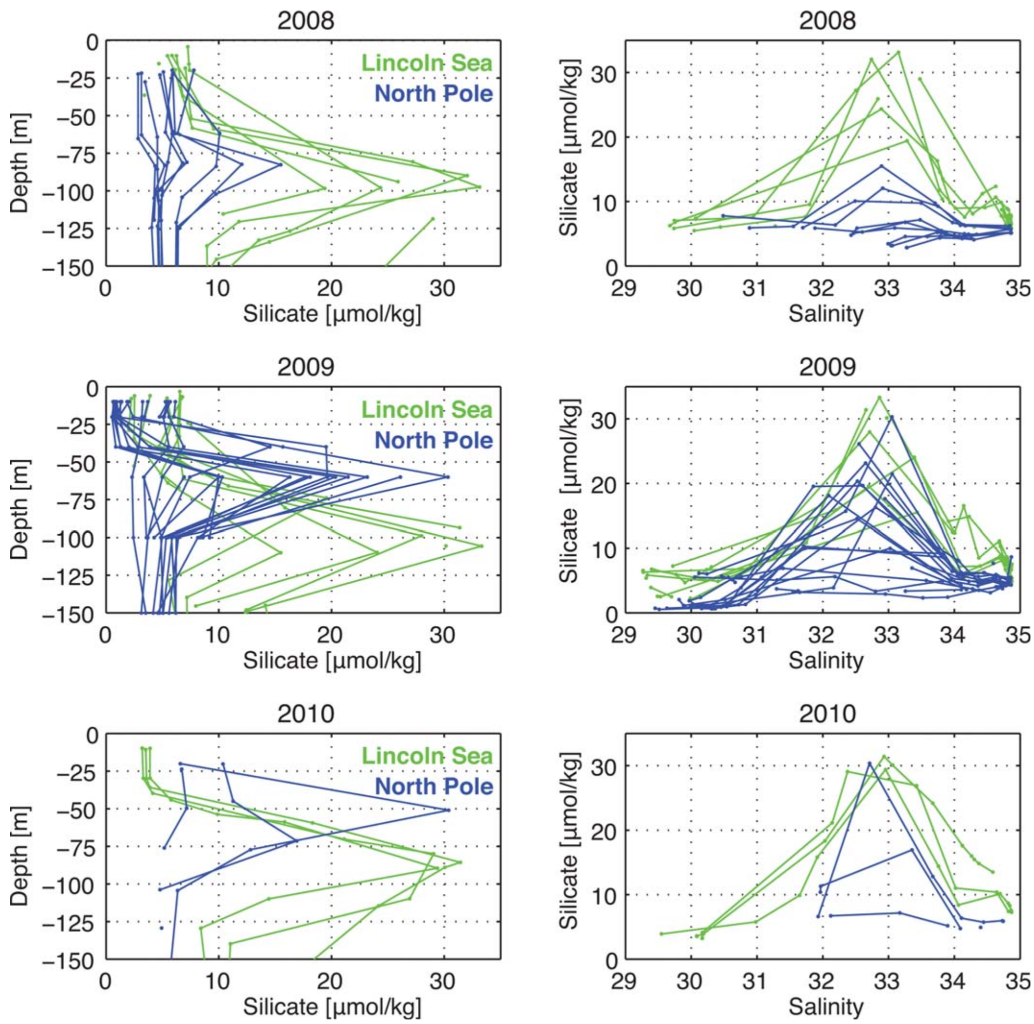

Figure 11. Vertical silicate profiles (left) and silicate-salinity diagrams (right) from the Lincoln Sea and around the North Pole north of 87.5°N for 2008, 2009, and 2010. The silicate data from the Lincoln Sea are mostly west of 60°W.

continuous salinity record from the uppermost instrument (at ~40 m) of the deep ocean mooring at the North Pole during winter 2009–2010 (not shown). These fresh waters with a small Pacific signature arrived in the eastern Lincoln Sea in spring 2010. Salinity increased again in the northern Makarov Basin and North Pole by 2010 and in the Lincoln Sea by 2011.

[32] Relative to the 1990s, there was substantially more freshwater present in the upper ocean (<500 m depth) over the deep Arctic Ocean basins, mostly in the Canada Basin [*Rabe et al*., 2011]. It has been shown that the positive AO from 2005 to 2008 diverted Eurasian river runoff into the Canada Basin and shifted the Transpolar Drift and the front between Atlantic and Pacific-derived upper ocean waters counterclockwise [*Morison et al*., 2012]. The positive AO continued and resulted in a northward extension of the Canada Basin dome in DOT along the coast of the Canadian Archipelago to the north end of Greenland in spring 2009. This brought Canada Basin freshwater as far as the mouth of Nares Strait, and potentially as far as Fram Strait. The freshening of the eastern Lincoln Sea continued up to 2010 and was accompanied with the arrival of fresh waters with a shallow (~50m) T_{max} at $S < 31$. This T_{max} is too fresh to be defined as a pure Pacific T_{max} but could rather be characterized as a NSTM. Although there was not much data from the western Lincoln Sea in that year, spring profiles in the Makarov Basin and summer profiles from as far north as the Lomonosov Ridge, showed similar NSTM characteristics in 2009.

[33] The 1990s fresh upper ocean in the Lincoln Sea was associated with a strong Pacific Water signal [*Newton and Sotirin*, 1997]. The observed decrease of the Pacific T_{max} in the Lincoln Sea from the 1990s to 2000 and 2001 was found to be related with changes in the extent of the Beaufort Gyre and path of the Transpolar Drift Stream [*Steele et al*., 2004]. The decline of the Pacific T_{max} correlated with a positive Arctic Oscillation index with a 3 year lag [*Steele et al*., 2004]. Over the extended time series from 1991 to 2009, we found smaller but significant correlations at 3 year lag but also at zero lag (correlation R ~0.5 to 0.7 and p values ~0.05 to 0.01, respectively). This change relative to the 1990s is possibly related to the absence of a very strong AO index for several years as was the case in the mid-1990s or due to a more rapid response of the upper

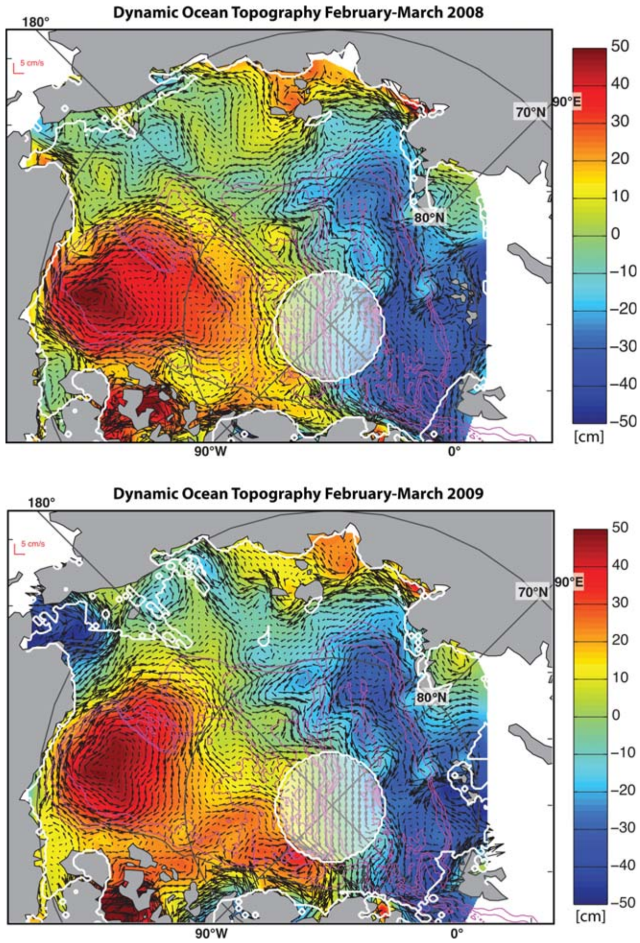

Figure 12. Dynamic ocean topography (color) and geostrophic velocity (vectors) for February to March in 2008 and 2009. The opaque disc around the North Pole (north of 86.5°N) shows the region where interpolation is used since there are no data available there.

ocean associated with reduced ice coverage, ice thickness, and strength.

[34] The observations of DOT agree very well with Figure 11 of *McLaughlin et al*. [2009] in which the northernmost branch of the boundary current going around the Chukchi Plateau debouches into the central Canada Basin. It brings Russian shelf waters rich in Eurasian river water across the northern Canada Basin and northeast to the Lincoln Sea. A recent study argued that the strengthening of the Beaufort Gyre post-2004 may have caused a reversal in circulation of the Atlantic water in the Canada Basin [*Karcher et al*., 2012]. Relative to the 1990s, a significant increase in Atlantic water temperature off the shelf slope at the section at 65°W in the western Lincoln Sea was seen. The increase of the Atlantic T_{max} from 0.44°C to approximately 0.8°C suggests that the warm Atlantic temperature anomaly that entered the Arctic through Fram Strait in the early 1990s [*Quadfasel et al*., 1991, 1993] arrived in the western Lincoln Sea by 2004. The warm Atlantic signature in the Lincoln Sea could be indicative of Atlantic water arriving from the southern Makarov Basin guided along the Mendeleev and Alpha ridges as the Atlantic temperatures there are of similar magnitude (Figure 10). Individual profiles (Figures 8 and 9) however show that the Atlantic T_{max} northwest of the Lincoln Sea was in general colder than 0.8°C. We explain the increase of the Atlantic T_{max} on the section at 65°W by arrival of Atlantic water from the Eurasian Basin following the bathymetry which veers westward where the Lomonosov Ridge meets the shelf break of the Lincoln Sea. As of yet we do not find evidence for increased influx of Atlantic water from the northern Canada or Makarov basins as suggested by *Karcher et al*. [2012].

[35] The reduction of freshwater in the Lincoln Sea by 2011 suggests the anomaly has exited the Arctic, likely through Nares or Fram Strait or both. In general, both the flow and salinity in Nares Strait are highly variable, and hence fluxes through the strait are difficult to attain. Integrated results from a moored instrument array (2003–2006) near 80.5°N gives at present the best estimates for the pre-2009 ambient fluxes [*Rabe et al*., 2010, 2012]. The moored record between 2006 and 2009 suggests that the formation of land-fast ice (usually starting late winter, early spring) slows the flux through Nares Strait from April 2008 onward. However, geostrophic freshwater flux is 20% greater when sea-ice is mobile than when it is land-fast [*Rabe et al*., 2012], and in 2009 land-fast ice did not form which implies there was a strong southward flux in spring 2009. The ice-free periods of greater mass flux were associated with concurrent fresh events. A first estimate of the total 2009 FW flux in Nares Strait suggests that it was 20% larger than over the period 2003–2006 (H. Melling et al., manuscript in preparation, 2013).

[36] Comparing the salinity at 10 m in Figure 2, we see that the values measured by an ITP drifting from the Lincoln Sea to Fram Strait in 2010 are similar to those found in Fram Strait. This implies that part of the Lincoln Sea fresh surface anomaly spread eastward and left the Arctic through Fram Strait. Fresh surface waters were reported in the Amundsen Basin in 2010, between the North Pole and Fram Strait. This low salinity anomaly was suggested to have been released from the Beaufort Gyre as a result of changing wind patterns [*Timmermans et al*., 2011]; however, they were still less fresh than observed in the Lincoln Sea or Fram Strait. It is more likely that these surface salinities seen in the Amundsen Basin also originated from fresh waters in the Makarov Basin, crossing the North Pole in 2009, and were certainly part of the general freshening of the central Arctic and Lincoln Sea from 2007 to 2010. The salinity record from the westernmost mooring from the mooring array in the East Greenland Current in Fram Strait indicated a relative fresh event occurred in winter 2010 (not shown) and the salinity at this depth was comparable to that in the Lincoln Sea. In addition, two recent studies on the variability of freshwater composition in Fram Strait showed that a significant fraction of Pacific water was first observed again in September 2011 [*Dodd et al*., 2012; *Rabe et al*., 2013] confirming that at least a portion of the Lincoln Sea fresh anomaly with a Pacific signature exited the Arctic through Fram Strait.

6. Conclusions

[37] A freshwater anomaly was observed in the Lincoln Sea during 2008–2010. This was caused by a large decrease

in the mean upper ocean salinity rather than by changes in the depth of the isohaline $S = 34.8$. Only along the shelf slope north of Nares Strait isohaline depth changes were significant. The total increase in FW volume was estimated to be $1100 \pm 250\,\mathrm{km}^3$ which was exported completely from the Lincoln Sea again by 2011. The anomaly was characterized by a combination of fresh waters from the northern Canada Basin extending to the western Lincoln Sea, and from the Makarov Basin north of the Mendejeev Ridge in the central to eastern Lincoln Sea. During this time, an increase of the Pacific T_{max} relative to the early 2000s was seen in the western Lincoln Sea, however, it was not as strong as observed in the mid-1990s.

[38] Watermass characteristics observed in the central Arctic around the North Pole appeared about a year later in the Lincoln Sea. Fresh upper ocean waters with a NSTM observed in the eastern Lincoln Sea in 2010 could also be identified near the North Pole in 2009, illustrating the advective nature of it. The fresh upper ocean anomaly in the Lincoln Sea covered Atlantic water with signatures from the eastern Canada Basin as well as the Makarov and Eurasian basins. This was general case for 2008–2011 when the Lincoln Sea was sampled more widely than pre-2008. The increase of the Atlantic temperature on repeat stations in the western Lincoln Sea relative to the 1990s can be explained by lateral spreading of warm Atlantic water from the Eurasian Basin and not from the Makarov Basin.

[39] The freshwater anomaly has likely exited the Lincoln Sea through Nares and Fram straits and analysis of the moored instrumentation records post-2009 has yet to reveal this. The majority of freshwater that has accumulated in the Canada Basin up to 2010 has not been released from the Arctic Ocean. A close intercomparison of the freshwater fluxes coming from Fram Strait, the Canadian Arctic Archipelago as well as Davis Strait is recommended for the next years. If a prolonged period of positive wind-stress curl anomaly occurs, a release of the anomalous FWC from the Beaufort Gyre toward the Lincoln Sea, and subsequently into the subpolar gyres, can take place, modifying upper ocean salinity there.

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