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Large Scale Salinity Anomaly Has Triggered the Recent Decline of Winter Convection in the Greenland Sea

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Key Points:

- After a period of intense deep convection since 2002, convection in the Greenland Sea has weakened from 2014 to 2020
- The upper salinity and temperature in the months prior of convection are driving the deep convection variability
- The upper salinity changes are resulting from changes in large scale pattern of advection

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Abstract The Greenland Sea is a key region for open ocean convection and ventilation, which exhibit a large variability with periods of strong convection and shutdowns. After a long period of weak winter convection (from the 1970s to the early 1990s), a recovery has been reported, beginning in the 1990s and intensifying in the early 2000s until 2013. Using ISAS, an optimal interpolation product based on Argo observations, we document a recent significant weakening of deep convection between 2014 and 2020, accompanied by a continuous warming of the mixed layer but also a freshening after 2014. These hydrographic changes likely increase the ocean stratification and precondition the shutdown of winter convection. We suggest that these property changes result from a shift of the large scale atmospheric circulation, affecting the source of Atlantic Water to the Nordic seas, causing a freshening of about -0.1 g kg^{-1} that spreads into the Greenland Sea.

Plain Language Summary The Greenland Sea is a key region for the climate system. There, during winter, the ocean loses heat to the atmosphere in a process called "deep convection," that results in the formation of dense water masses that ventilate and fill the deeper layers of the ocean. The Greenland Sea exhibits a high variability in the intensity of winter deep convection, but the scientific community is still debating on which processes trigger or stop convection. After a recovery period of convection, mainly in the beginning of the 21st century, convection has stopped again after 2014 and at least until 2020. Here, we suggest that this new shutdown is mainly caused by changes in the ocean upper layer temperature and salinity that increase the upper ocean buoyancy, making it more difficult to trigger deep convection during winter. We further propose that the main mechanism driving the ocean properties changes is a shift of the large scale atmospheric circulation, which affects the amount and properties of Atlantic Water transported to the Greenland Sea.

1. Introduction

The Greenland Sea plays a key role in the climate system (Eldevik et al., 2009), as it is one of the regions where winter deep ocean convection occurs, a process triggered by surface buoyancy forcing, that can ventilate the intermediate or deep layers of the ocean (e.g., Moore et al., 2015; Strass et al., 1993; Swift & Aagaard, 1981). Previous works (Chafik & Rossby, 2019; Petit et al., 2021) have found that the main source of the AMOC lower limb is formed and transformed in the Nordic Seas, acting as a regulator for the Earth's climate (Hansen & Østerhus, 2000). Additionally, most of the dense water exported from the Nordic Seas originates in the Greenland Sea (e.g., Brakstad et al., 2023; Jeansson et al., 2008). Beyond its importance for the deep branch of the AMOC, the magnitude of the deep convection is also fundamental for the ventilation of heat and tracers to the deep ocean, such as freshwater, oxygen, and carbon dioxide (e.g., Bashmachnikov et al., 2021; Brakstad et al., 2019; Fröb et al., 2016; Lauvset et al., 2018).

Based on very limited observations collected in winter, deep convection in the Greenland Sea has been documented in the 1960s and early 1970s (Dickson et al., 1996), but has ceased after 1980 (Dickson et al., 1996; Schlosser et al., 1991) when the upper layers became warmer and saltier. More recently, observations during winter have become more abundant, particularly thanks to the development of autonomous observing platform, such as Argo floats. The return of deep convection since the 1990s, with stronger magnitudes in the beginning of the 21st century, has been reported in several studies (e.g., Bashmachnikov et al., 2021; Brakstad et al., 2019; Lauvset et al., 2018; Somavilla, 2019). However, Abot et al. (2023) have suggested a recent decrease in the intensity of winter convection through analyzing an ocean-sea ice model with data assimilation. The objective of the present study is to underpin the large scale processes responsible for the recent convection halt in the Greenland Sea.

Several processes determine the occurrence of ocean deep convection (or lack thereof): (a) in general, the ocean-atmosphere buoyancy loss is considered as one of the main triggers for convection (e.g., Marshall & Schott, 1999; Yang et al., 2016), with a stronger ocean heat loss resulting in deeper convection (Moore et al., 2015); (b) the intensity of the cyclonic circulation can help to break the stratification through Ekman pumping of the isopycnals; (c) changes of the water column density caused by advection of Arctic Water (cold and fresh) coming from Fram Strait or of warm and salty Atlantic Water flowing with the Norwegian Atlantic Current can also affect the upper ocean stratification (Lauvset et al., 2018), therefore a decrease in stratification can favor winter deep convection.

In the present work, we analyze the variability of the deep convection and its drivers over the period 2002–2020.

2. Data and Methods

We make use of the monthly ISAS Optimal Interpolated temperature and salinity fields from 2002 to 2020, which is an updated version of the product presented by Gaillard et al. (2016). ISAS includes a large amount of Argo profiles in the Greenland Sea, including during winter (Figure 1), with a mean of 42 ± 25 profiles per winter over the period. There are however no data in the region in winter 2004, so we discard that year from our analysis. To determine the intensity of convection, we estimate the maximum mixed layer depth (MLD) during winter in the center of the Greenland Sea. MLD is calculated from a potential density criterion with a difference of 0.01 kg m^{-3} from the surface.

Moreover, ISAS is used to estimate the stratification, by calculating the buoyancy content, and consequently the buoyancy that needs to be removed (B_{rem} in m^2 s^{-2}) for the mixed layer to reach a given depth z_1, following Schmidt & Send (2007):

$$B_{rem}(z_1) = \frac{-g}{\rho_0} \int_{z_1}^{0} \rho_{z_1} - \rho(z)\, dz, \tag{1}$$

where g is gravity, ρ_0 the reference seawater density, ρ_{z1} is the density at the depth z_1, and $\rho(z)$ is the density profile. Additionally, we calculate the temperature and salinity contribution to B_{rem} during the preconditioning period (September–December), as:

$$B_{\theta}(z_1) = -g \int_{z_1}^{0} \alpha \left[\theta_{z_1} - \theta(z)\right] dz, \tag{2}$$

$$B_{S}(z_1) = g \int_{z_1}^{0} \beta \left[S_{z_1} - S(z)\right] dz, \tag{3}$$

with α the thermal expansion coefficient, β the haline contraction coefficient, θ and S the temperature and salinity profiles, respectively.

We also use the ERA5 atmospheric reanalysis (Hersbach et al., 2020) to obtain monthly means of heat and freshwater air-sea fluxes, and sea level pressure (SLP), with a resolution of about 0.28°. The total heat flux (Q, in W m^{-2}) is the sum of all the radiative and turbulent heat fluxes, and the total freshwater flux (FW, in m s^{-1}) is the excess of precipitation (P) over evaporation (E). The buoyancy surface flux (B_{surf}, in m^2 s^{-3}) is estimated as (Sallée et al., 2010):

$$B_{surf} = g \frac{\alpha}{\rho_0 C_p} Q + g\beta SSS(E - P), \tag{4}$$

with ρ_0 the reference seawater density, C_p the heat capacity and SSS the sea surface salinity.

To obtain the surface geostrophic velocities from 2007 to 2020 we use the AVISO data set (https://www.aviso.oceanobs.com), that is calculated from the Absolute Dynamical Topography.

3. Results

3.1. Intensity and Variability of Winter Deep Convection

The variations in late winter (February–April) MLD in the Greenland Sea over 2002–2020 are shown in Figure 1a. We consider the maximum MLD within a box over the central Greenland Sea (red box in insert of Figure 1a).

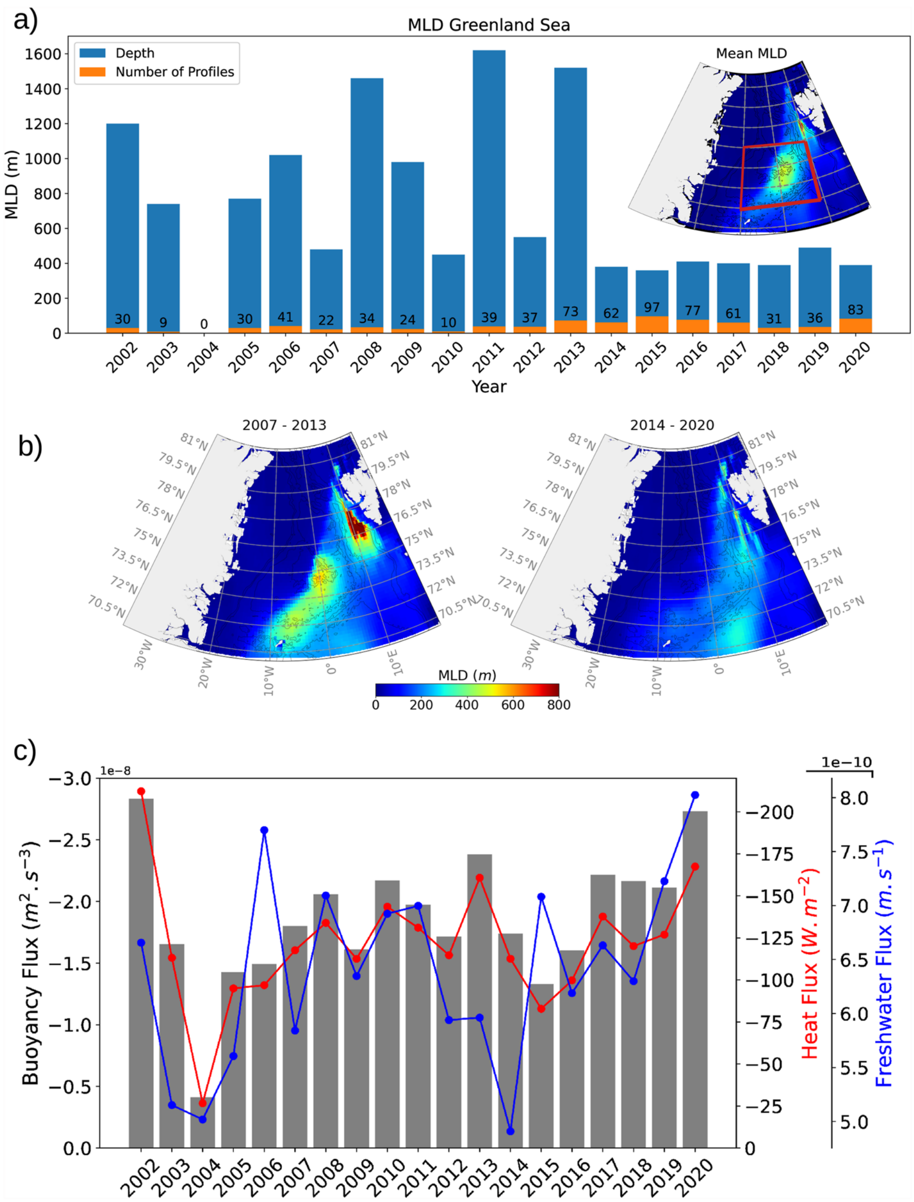

Figure 1. (a) Maximum mixed layer depth (MLD) (blue bars) per year in the Greenland Sea during late winter (February to April) in the red box area (between 10°W and 10°E and 72.5°N–76°N). The number of available Argo profiles in the box is shown in orange. The insert map shows the average (2002–2020) late winter MLD. (b) Maps of the mean MLD during the convective (left) and non-convective (right) periods. (c) Average air-sea flux in the center of the Greenland Sea (10°W to 10°E and 72°N to 76.5°N), during late winter (February–April). The total buoyancy surface flux (in $m^2\ s^{-3}$) is shown as Gray bars, and the heat (in $W\ m^{-2}$) and freshwater flux (in $m\ s^{-1}$) are shown as red and blue lines, respectively.

Since 2002, MLD has varied from about 400 m depth in 2015 (the year with the weakest convection) to about 1500 m depth in 2011. During 9 of the 19 winters, MLD was deeper than 500 m, and deeper than 1000 m during five winters (2002, 2006, 2008, 2011, and 2013). From 2002 to 2013, the MLD was deeper than 500 m during most winters, except in 2007, 2010, and 2012. However, after this period, deep convection has abruptly weakened since the winter of 2014, and MLD has not reached deeper than 500 m, with a mean value of 392 m (Figure 1). This period is also the time with the largest number of observations, making us confident in the robustness of the signal described here, including in the spatial pattern of the difference between the periods of convection (before 2014) and non-convection (after 2014, Figure 1b), that reveals a significantly shallower MLD everywhere in the Greenland Sea after 2014.

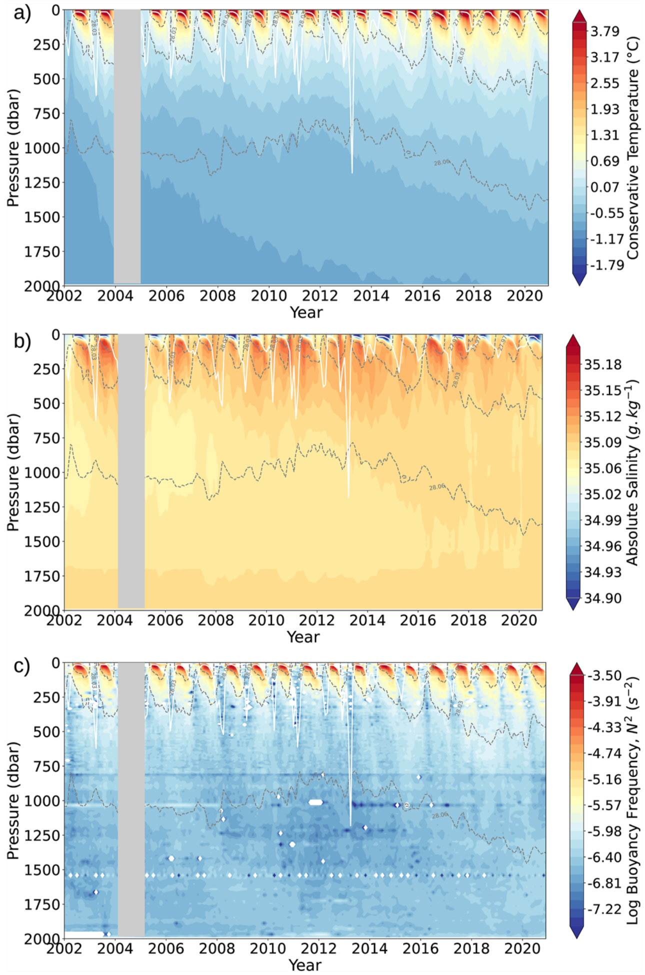

Figure 2. Time evolution of the profiles of (a) conservative temperature (in °C), (b) absolute salinity (in g kg−1), and (c) buoyancy frequency N^2 (shown in log scale, in s^{-2}), averaged between 10°W and 10°E and 72°N and 76.5°N (red box on Figure 1a). The average mixed layer depth is shown as a white contour and isopycnals as a gray dotted contours (referenced to the surface). The year 2004 is masked as there is no observation during that winter.

Previous analyses have suggested that air-sea buoyancy fluxes are one of the primary drivers of the MLD variability (e.g., Moore et al., 2015). Figure 1c shows the total atmospheric buoyancy flux during winter, as well as the heat and freshwater fluxes averaged over the Greenland Sea, estimated from the ERA5 reanalysis. We could not find a significant correlation between the variations of the freshwater flux and the MLD. Moore et al. (2015) found that MLDs in the Greenland Sea are particularly sensitive when winter heat loss exceeds 150 W m^{-2}. We also find stronger heat loss in years of strong convection, with winters characterized by deep mixed layers during 2002–2013 exhibiting heat flux exceeding 150 W m^{-2}. However, similar to Lauvset et al. (2018), some years presented a strong heat flux but not a strong convection. For example, in 2020 despite a heat loss of 170 W m^{-2}, the MLD is only 400 m. The correlation coefficient (R^2) between the total buoyancy flux and MLD is only 0.28 ($p - value < 0.05$), meaning that interannual variability in the local atmospheric heat loss alone cannot explain the variability of deep convection in the Greenland Sea, suggesting that other non local drivers are playing a leading role in the recent convection shutdown.

Additionally, we examine the recent variations of temperature, salinity and stratification (defined as N^2, the Brunt Väisälä frequency) profiles averaged over the same region (Figure 2), expanding the timeseries presented by Brakstad et al. (2019), their Figure 5, and Lauvset et al. (2018), their Figure 2, and we find similar results as Abot et al. (2023). In the upper layer (≥300 m depth), all properties exhibit seasonal variations, with higher temperature, lower salinity, and stronger stratification during summer. In winter, there is a weakening of the stratification, mainly caused by a cooling and salinization of the top layer. This seasonality is stronger during the years of strong convection (2002, 2008, 2011, 2013), when stratification is weaker, and the isopycnals shoal. For instance, the $\sigma_0 = 28.03$ kg m^{-3} isopycnal outcrops during winters 2011 and 2013, but remains mostly deeper than 250 m after 2013. Below ∼300 m depth, over the whole period, isopycnals tend to deepen. The $\sigma_0 = 28.06$ kg m^{-3} deepens from 1000 m in 2002 to 1,450 m in 2020, implying a deepening rate of about ∼−25 m year^{-1}. A similar rate was reported by Somavilla (2019) for the rim regions since at least 1980. Over the whole 0–2,000 m water column, a strong warming is also visible since 2002, with an annual mean of −0.11°C in 2002 increasing to 0.25°C in 2020. In the non-convective period after 2014, the trend is even stronger, and the surface warming visible in the top 500 m tends to intensify after 2016, accompanied by a stratification increase.

3.2. Local Ocean Conditions

In order to further examine the changes in ocean conditions associated with the change of convective regime, we look at the time evolution of the mean temperature and salinity within the mixed layer during late winter (February–March, when the mixed layer deepens) and in the upper 500 m during the preconditioning period (September–December). Results are shown in ΘS diagrams (Figure 3). The winter temperature exhibits a high variability going from −0.61°C in 2002 (the coldest winter of our record) to 0.22°C in 2020. This corresponds to a linear trend of +0.041°C/year, which is significant according to a Mann-Kendall test ($p - value < 0.01$). Mixed layer temperature reaches a maximum in 2018, during an exceptionally warm winter over the Nordic Seas (Moore et al., 2018). The absolute salinity shows a similar pattern, with an increase over most of the period. From 2002 to 2013, the mean winter salinity within the mixed layer increases from 34.89 g kg^{-1} to 34.92 g kg^{-1}, that is, at a rate of +0.0026 g kg^{-1}/year. Yet, since 2014, the salinity decreases, and reaches a minimum of 34.81 g kg^{-1} in 2020. We further estimate the contribution of temperature and salinity anomalies to density anomaly, linearizing the equation of state for small anomalies (Figure 3d). During the convective period, there is a small increasing trend in density, that reaches its highest values in years of strong convection. After 2014, the mixed layer density

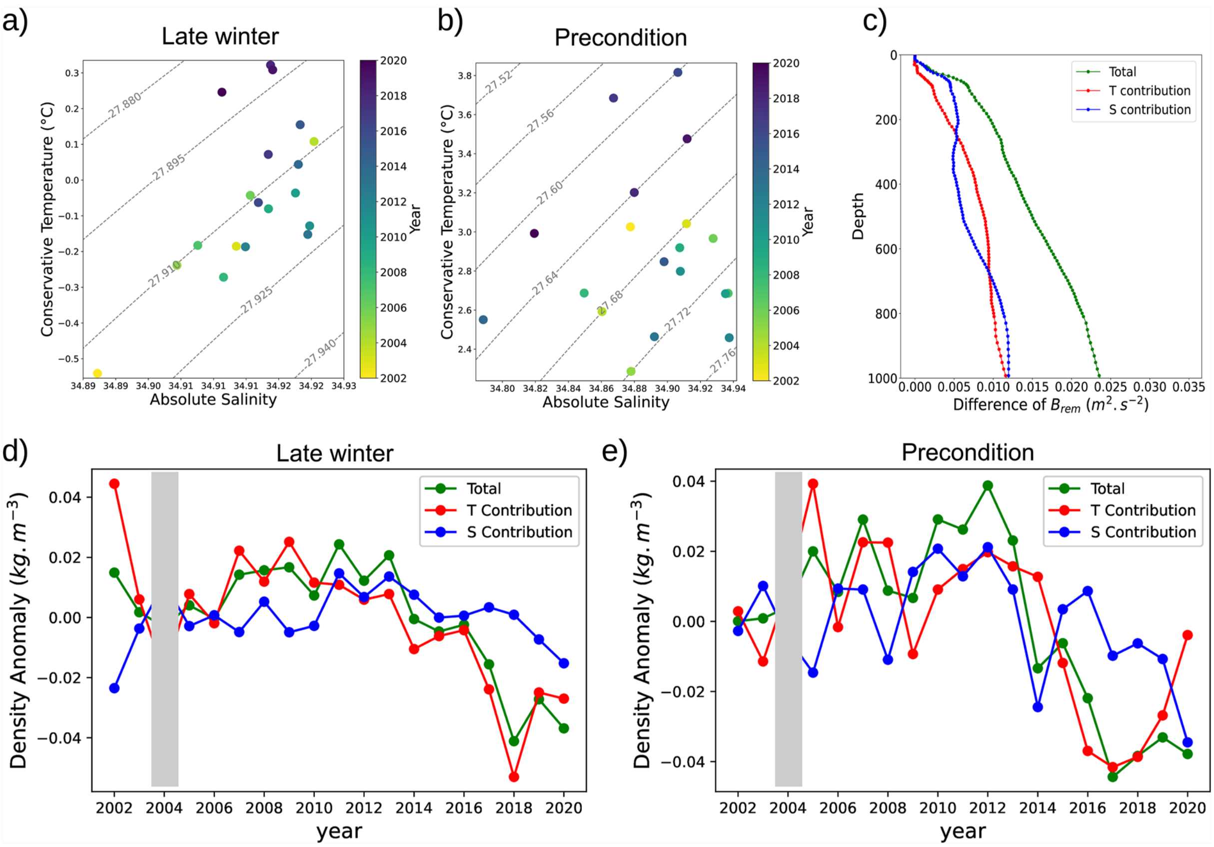

Figure 3. (a) Mean ΘS-diagram within the mixed layer during the late winter and isopycnals as gray dotted lines (surface reference), with the years represented in colors. (b) Mean ΘS-diagram above 500 m during the preconditioning period (September– December) and isopycnals (referenced to the surface) as gray dotted lines, with the years represented in colors. (c) Difference of buoyancy to be removed ($m^2\ s^{-2}$) in the preconditioning period (September–December) between a non-convective (2014–2020) and a convective (2007–2013) periods. (d) Time series of the late winter density anomalies in the mixed layer. (e) Time series of the fall density anomalies in the top 500 m. For (c–e), the total is in green, and the salinity and temperature contributions are in blue and red, respectively.

decreases drastically as a response to both the warming and freshening, causing an increase in stratification in the region that halts deep convection.

It has been previously suggested that interannual variations in convection intensity are mainly governed by the variations of the upper salinity and density in the Greenland Sea during the previous season (September– December), which result from changes in the advection of re-circulating Atlantic Water (Bashmachnikov et al., 2021). Here we find that temperature in the upper 500 m during the preconditioning period exhibits a larger range of variability, going from 3°C to 4.4°C, and a warming after 2014 (Figure 3b). The correlation coefficient (R^2) of the fall temperature and MLD during the following winter is 0.73 ($p - value < 0.05$). Salinity during the preconditioning period also shows some large variations, ranging between 34.94 g kg^{-1} in 2012 and 34.80 g kg^{-1} in 2014, with a correlation with winter MLD of 0.76 ($p - value < 0.05$), slightly higher than with temperature. Combining both physical properties, the correlation between density and MLD reaches 0.75, strongly suggesting that the variability in the intensity of convection variability is determined by the upper layer density in the months before convection starts.

Interestingly, we find that, for all years characterized by deep convection, the mean salinity over 0–500 m during the preconditioning period is higher than 34.92 g kg^{-1}, meaning that saltier conditions are probably a key ingredient for deep convection to occur. This is in agreement with Brakstad et al. (2019) who showed that mean near-surface (0–50 m) practical salinity lower than 34.71 (equivalent to 34.88 g kg^{-1}) during the preconditioning period summer generally results in MLD not exceeding 300 m the following winter. Yet, our results suggest that temperature also plays an important role. There are years when salinity reaches high values, but the convection is weak in the following winter. This is for example, the case in 2006, characterized by salinity as high as

34.93 g kg^{-1} during the preconditioning period, but a MLD shallower than 400 m during the next winter, most likely because of the high temperatures found during the preconditioning months (with a mean value of 3.7°C). These property changes are strongly affecting the local stratification. Figure 3c shows the difference between the non-convective and convective periods of B_{rem} during the preconditioning time. It reveals that the water column is more stable during the most recent period, largely due to the stabilizing effect of the change in salinity down to 250 m. From 250 to 650 m, the temperature contribution dominates. Since 2014, salinity during the preconditioning months remains below 34.92 g kg^{-1}, with a strong negative anomaly in 2014 (Figure 3e). After 2014, there is a strong warming affecting the upper layer of the Greenland Sea, which, together with the salinity decrease, results in a lower surface density and a subsequent increase of the water column stability (Figures 3d and 3e), most likely driving the weakening of deep convection in this region during this period.

3.3. Large Scale Hydrographic Changes

In the previous section, we have shown that the density of the top 500 m in the center of the Greenland Sea preceding a given winter is the main factor determining the occurrence of deep convection. However, the processes setting up the density profile and its recent changes in the Greenland Sea still need clarification. We further investigate if these changes are mostly driven locally or if they are the local signature of widespread large scale changes. We compute the mean SLP anomaly for the convective and non-convective periods against the 2002–2020 average (Figure 4a). During the convective period, there is a clear widespread positive SLP anomaly over Greenland contrasting with negative values over the southeast regions. In contrast, during the non-convective period, the SLP anomaly over Greenland is strongly negative, and expands toward the southeast, down to the European Coast. The SLP anomaly patterns resemble largely the anomalies associated with the two phases of the North Atlantic Oscillation (NAO), which is not surprising as the NAO index was largely negative between 2007 and 2015 and positive afterward.

The SLP anomaly pattern over the two periods is most likely associated with changes of the large scale circulation, as suggested by Holliday et al. (2020) and Kenigson & Timmermans (2021). The non-convective period is preceded by a period with a strong wind stress curl in the subpolar region that increases the freshwater convergence through three mechanisms: (a) re-routing the Labrador Current (and thus Arctic Water) eastward off the Newfoundland shelf, (b) shifting the baroclinic subpolar front to the southern branch of the North Atlantic Current (NAC), and (c) extending the southern branch further to the east (Holliday et al., 2020). Therefore, during the convective period, the contraction of the anomaly toward Greenland allows for a more direct path of warm and salty subtropical water flowing with the NAC, combined with a weaker Subpolar Gyre (SPG). This is also visible in the velocity anomaly (Figure 4b, the standard deviation is about 0.001 m s^{-1}), with negative values within the SPG and a higher transport associated with the NAC, resulting in an increase in salinity of $\sim$+0.15 g kg^{-1} in the upper water column (Figure 4c, standard deviations about 0.05 g kg^{-1}). In contrast, during the non-convective period, the eastward expansion of the SLP anomaly may favor the advection of fresher water masses circulating within a stronger SPG (Figure 4b, right panel), and thus results in a decrease of salinity in the Nordic Seas (Figure 4c), with a mean amplitude of ~ -0.15 g kg^{-1}. We track the salinity changes from the subpolar North Atlantic to the Nordic Seas (Figure 4d). The mean salinity over 2002–2011 is about 35.26 g kg^{-1} but decreases to 35.17 g kg^{-1} after 2012, which is a similar change to the one documented by Mork et al. (2019). The NAC freshens significantly after 2011, and the trend intensifies after 2016. Note that we do not see a similar advective pattern for temperature (not shown). The signal takes about three to 4 years to propagate from the subpolar region to the interior of the Greenland Sea. The strong negative salinity anomaly during the non-convective period is also most likely a consequence of the new "great salinity anomaly," detected first in 2010 in the Atlantic Ocean and propagating northeast in the Nordic Seas (Holliday et al., 2020).

Moreover, we note that the anomaly in salinity could in principle also be advected from the Arctic rather than from the North Atlantic. This is however unlikely as the observed freshwater transport through Fram Strait entering the Nordic Seas has been declining in the recent period (Karpouzoglou et al., 2022), and the sea ice export did not increase significantly recently (Ricker et al., 2018; Sumata et al., 2022). Yet, we can not fully rule out the hypothesis that changes of the large scale atmospheric circulation may have modulated the amount of freshwater transferred from the East Greenland Current to the interior of the Nordic seas. This would be somewhat at odds with the lack of salinity anomaly that extends up north of 70°N in the Greenland Coast (Figure 4c), but we acknowledge that ISAS may not be able to capture the conditions close to the coast as this region is largely ice-covered and shallow, preventing largely the implementation of Argo floats.

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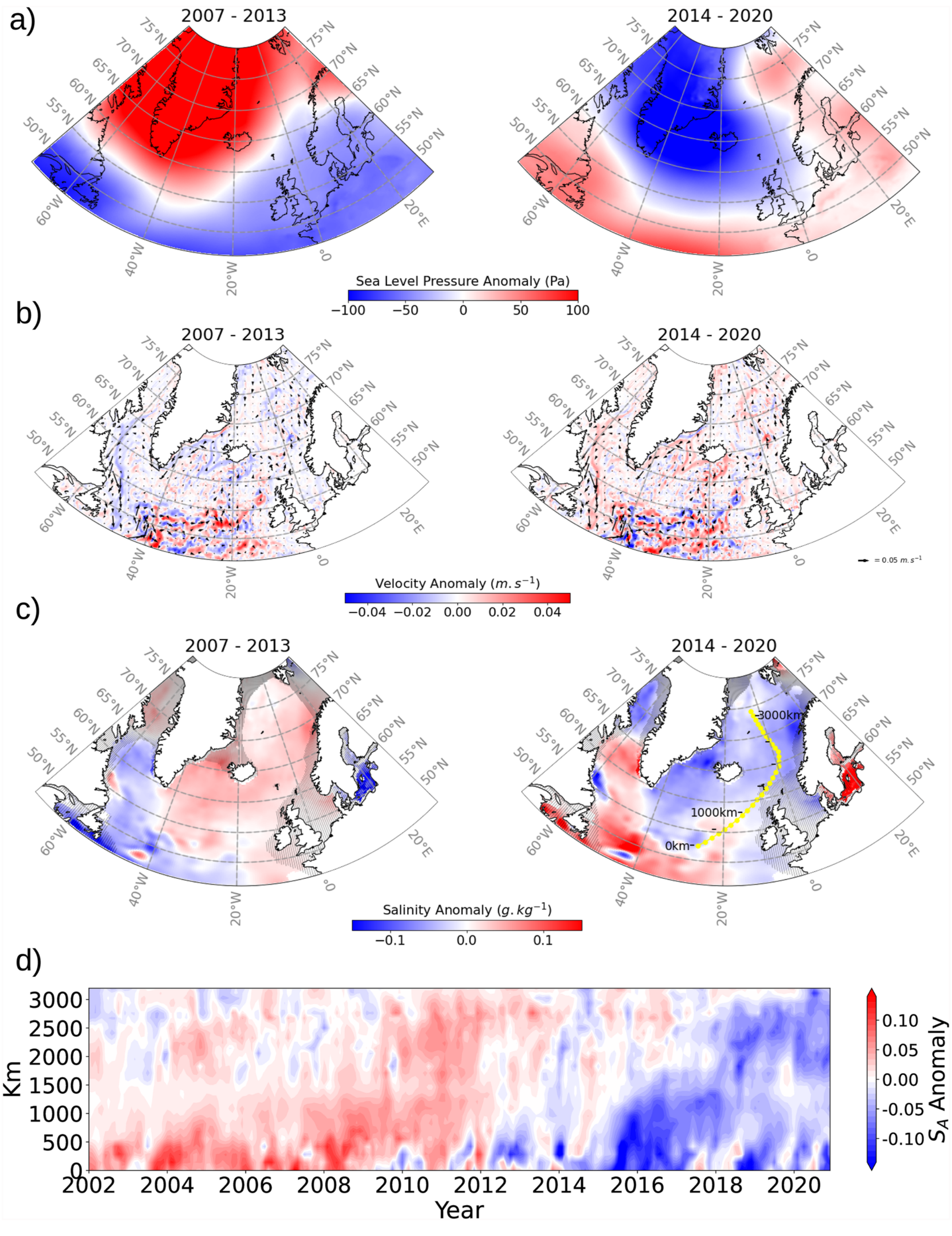

Figure 4. (a) Sea level pressure anomalies (in Pa) during the convective (left) and non-convective (right) periods. (b) Surface geostrophic velocity anomalies (in m s^{-1}) provided by AVISO (http://www.aviso.oceanobs.com) during the convective (left) and non-convective (right) periods, the arrows are the mean velocity and direction for each period. (c) Absolute salinity anomalies (in g kg^{-1}) in the top 300 m layer of the Nordic seas during the convective (left) and non-convective (right) periods, the hatched regions presented a percentage of variance higher than 75% and it is less trustworthy in these analyses. The anomalies are calculated in reference to the mean over the full period of analysis (2002–2020). (d) Time-distance plot of the monthly absolute salinity anomaly (in g kg^{-1}, referenced to the monthly mean over 2002–2020) averaged over 0–500 m in the path of the North Atlantic Current (yellow dots in panel c).

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4. Conclusion

There is an extensive body of literature documenting convection in the Greenland Sea (e.g., Böning et al., 2016; Dickson et al., 1996; Moore et al., 2015). After a long period of shutdown (from the late 1970s to the early 1990s), MLD has exceeded 500 m again after 1993. The convection was pronounced from 2000 onward (e.g., Brakstad et al., 2019; Lauvset et al., 2018), but decreased again after 2015 (Abot et al., 2023). Here, we document a similar reduction of the convection since 2014. In addition, we reveal the role of the large scale ocean circulation in the recent convection's decrease. Changes of the hydrographic conditions in the Greenland Sea also occurred during this period: a deepening of isopycnals, a continuous warming, and a salinification until 2013, similar to what was previously documented (e.g., Abot et al., 2023; Bashmachnikov et al., 2021; Brakstad et al., 2019; Lauvset et al., 2018; Somavilla, 2019). Additionally, we observe a new tendency of surface freshening after 2013, which, combined with the previous observed warming, has resulted in an increased upper layer buoyancy and stratification.

We further elucidate that the major driver of convection is the change in upper layer density, specifically temperature and salinity during the months preceding convection (September – December), which is again consistent with the results of Somavilla (2019) and Bashmachnikov et al. (2021). The top layer salinity (upper 500 m) shows a strong correlation with winter MLD, although temperature also plays an important role (despite a weaker correlation with MLD), as a strong warming could stop convection by increasing upper ocean stability. However, the recent freshening trend is most likely the main explanation of the convection shutdown in the Greenland Sea, possibly associated with an advection of the "great salinity anomaly" in the North Atlantic Ocean. We suggest that the changes in water column salinity are resulting from changes of the water masses advected to the region rather than the local buoyancy fluxes, in accordance with previous studies (Böning et al., 2016; Lauvset et al., 2018). In our analysis, we did not take into account for the contribution of melt-water and sea-ice formation to the surface buoyancy forcing, which can also contribute to modulating the MLD (Pellichero et al., 2017). Yet, given that sea ice did not cover a large portion of the Greenland Sea during the 2000's (Schmitt & Lüpkes, 2022), it is unlikely that the change in local sea ice condition would be the main driver of the mixed layer variations.

Dickson et al. (1996) suggested that convection in the Greenland Sea occurred during the negative phase of the NAO; the first well-documented deep mixed layer in the region occurred in the late 1960s and early 1970s, followed by a convection shutdown for several decades. Deep convection returned in the early 2000s (e.g., Bashmachnikov et al., 2021; Brakstad et al., 2019; Lauvset et al., 2018). In our study, we present evidence of a recent weakening of convection, initially influenced by the advection of fresher surface waters and sustained by a warmer mixed layer in subsequent years. We showed that convection in this region is strongly influenced by the pattern of large scale atmospheric circulation that determines the large ocean circulation in the subpolar North Atlantic and consequently impacts the waters exported to the Nordic Seas. A prolonged convection shutdown in the Greenland Sea can severely affect the ocean heat uptake and impact the properties of the water masses feeding the AMOC (Buckley & Marshall, 2016).

Our analysis stops in 2020, constrained by the availability of the ISAS product. Given that the NAO (and associated SLP anomaly) has shifted to a negative state after 2020, it is likely that deep convection could have returned after 2020. Monitoring surface salinity in the SPG could help to forecast deep convection in the Greenland Sea, with a 3–4 years lag.

Data Availability Statement

Data sets for this research are available in these in-text data citation references: the ocean data set is from ISAS (Kolodziejczyk et al., 2021); the air-sea fluxes are from ERA5 reanalyses data set (Hersbach et al., 2020); Argo float data and metadata from Global Data Assembly Centre (Argo GDAC) (Notarstefano, 2020). The geostrophic velocities was provided by AVISO (CMEMS, 2019).

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