Revisiting the Causal Connection between the Great Salinity Anomaly of the 1970s and the Shutdown of Labrador Sea Deep Convection

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(Manuscript received 8 May 2020, in final form 16 October 2020)

ABSTRACT: The Great Salinity Anomaly (GSA) of the 1970s is the most pronounced decadal-scale low-salinity event observed in the subpolar North Atlantic (SPNA). Using various simulations with the Community Earth System Model, here we offer an alternative view on some aspects of the GSA. Specifically, we examine the relative roles of reduced surface heat flux associated with the negative phase of the North Atlantic Oscillation (NAO) and extreme Fram Strait sea ice export (FSSIE) in the late 1960s as possible drivers of the shutdown of Labrador Sea (LS) deep convection. Through composite analysis of a long control simulation, the individual oceanic impacts of extreme FSSIE and surface heat flux events in the LS are isolated. A dominant role for the surface heat flux events for the suppression of convection and freshening in the interior LS is found, while the FSSIE events play a surprisingly minor role. The interior freshening results from reduced mixing of fresher upper ocean with saltier deep ocean. In addition, we find that the downstream propagation of the freshwater anomaly across the SPNA is potentially induced by the persistent negative NAO forcing in the 1960s through an adjustment of thermohaline circulation, with the extreme FSSIE-induced low-salinity anomaly mostly remaining in the boundary currents in the western SPNA. Our results suggest a prominent driving role of the NAO-related heat flux forcing for key aspects of the observed GSA, including the shutdown of LS convection and transbasin propagation of low-salinity waters.

KEYWORDS: Deep convection; Meridional overturning circulation; Freshwater; Decadal variability; North Atlantic Oscillation

1. Introduction

There have been significant low-salinity signals in the upper ocean of the subpolar North Atlantic (SPNA), circulating counterclockwise along the major currents constituting the subpolar gyre (Belkin et al. 1998; Houghton and Visbeck 2002; Belkin 2004). One of the most pronounced low-salinity events occurred in the late 1960s–1970s, often referred to as the Great Salinity Anomaly (GSA; Dickson et al. 1988). The low salinities associated with the GSA first emerged in the Greenland Sea, then were successively seen along the East and West Greenland Currents (EGC and WGC, respectively), the Labrador and North Atlantic Currents (NAC), and eventually back to the Greenland Sea about a decade later (Dickson et al. 1988; Belkin et al. 1998). It is widely accepted that the GSA was initiated by an extreme Fram Strait sea ice export (FSSIE) event in the late 1960s (Dickson et al. 1988; Aagaard and Carmack 1989).

Freshwater transports (fluxes) from the Arctic Ocean to the North Atlantic, including sea ice export, could have significant implications for ocean circulation and climate, because major deep-water formation sites in the Greenland Sea and Labrador Sea (LS) are located downstream of major freshwater pathways from the Arctic Ocean. In particular, the Fram Strait is estimated as the largest outlet for the freshwater fluxes from the Arctic Ocean when liquid and solid freshwater fluxes are combined (Haine et al. 2015). A large part of the freshwater flux through the Fram Strait flows along the EGC and enters the North Atlantic via the Denmark Strait (Dickson et al. 2001; Dodd et al. 2009). It then turns westward at the southern tip of Greenland toward the LS to join the WGC. Therefore, an excess freshwater input through the Fram Strait could stabilize the water column and reduce deep-water formation in the LS. A possible ramification of such reduced deep-water formation is a weakening of the Atlantic meridional overturning circulation (AMOC), which is a vital component of the climate system as it carries a significant amount of heat to the North Atlantic and sub-Arctic seas. Hydrographic measurements at Ocean Weather Station Bravo (OWS-B) near the center of the LS indeed suggest a shutdown of deep convection associated with a substantial freshening during 1969–71 (Lazier 1980), around the time that the FSSIE-induced freshwater anomaly was expected to arrive at the LS.

This observed shutdown of convection has motivated many modeling studies that investigate the impact of an excessive FSSIE event on convection strength in the LS and thermohaline circulation (e.g., Mauritzen and Häkkinen 1997; Häkkinen 1999; Haak et al. 2003; Koenigk et al. 2006; Ionita et al. 2016). By generating an idealized FSSIE pulse in an ocean–sea ice model with its domain limited to the Arctic and Atlantic Oceans, Häkkinen (1999) finds suppressed deep convection associated with a substantial freshening in the LS as well as a weakening of AMOC in response to the pulse. Based on global ocean–sea ice model simulations, Haak et al. (2003) arrive at the same conclusion as Häkkinen (1999), although they do not find a weakening of AMOC with suppressed convection.
Through a composite analysis of strong (weak) FSSIE ($\pm$1.2 standard deviation) events from a preindustriel control simulation of a coupled model, Koenigk et al. (2006) find suppressed (enhanced) deep convection associated with a freshening (salification) in the LS, supporting the relationship between FSSIE and LS deep convection. With these strong supports from modeling studies, it is generally accepted that the excessive FSSIE in the late 1960s was the reason for the observed shutdown of deep convection in the LS during 1969–71.

Generally speaking, deep convection in the LS is predominantly driven by surface heat loss during winter (Marshall and Schott 1999) as the heat loss erodes upper-ocean buoyancy. The buoyancy is regained during the warm season through solar insolation and eddy- (Hátún et al. 2007; Chanut et al. 2008) and wind-driven (Schulze Chretien and Frajka-Williams 2018) lateral fluxes from relatively buoyant boundary current water encircling the LS (Straneo 2006; Schmidt and Send 2007). Deep convection mixes relatively fresh (and cold due to surface heat loss) upper ocean with relatively salty (and warm) deep ocean. Therefore, when surface heat loss is sufficiently weak that deep convection is suppressed, the upper ocean can be anomalously fresh. It is well established that convection strength or deep-water formation rate in the LS is highly correlated with winter surface heat fluxes associated with the North Atlantic Oscillation (NAO; Yashayaev 2007; Yashayaev and Loder 2016), which is the dominant atmospheric circulation variability mode over the North Atlantic (Hurrell 1995).

To be more precise, LS convection is directly related to the meridional sea level pressure (SLP) gradient across the LS (Kim et al. 2016). The NAO was in a negative phase during 1969–71. In particular, the NAO was a record low in 1969 based on the station-based NAO index by Hurrell (1995). Therefore, it is possible that a reduced heat loss due to anomalously warm air temperatures and weak winds in the LS associated with this negative NAO condition could have significantly contributed to, or even predominantly driven, the suppression of deep convection and, thus, the freshening in the LS during the GSA.

Such a potentially important role of the surface heat flux for the suppression of LS deep convection and associated local surface freshening is discussed in Houghton and Visbeck (2002) from analyses of historical hydrographic data. They conclude that the LS cannot simply be viewed as a conduit for freshwater from the Arctic Ocean. Indeed, eddy-permitting to eddy-resolving ocean simulations show that much of the freshwater flux due to Greenland ice sheet melt stays within the boundary currents and northern LS, so its impact on the interior LS where deep convection takes place is minimal (Marsh et al. 2010; Luo et al. 2016). In numerical experiments where a passive tracer is released along with runoff from Greenland using ocean models with three different horizontal resolutions from 1/12° to 1/2°, Dukhovskoy et al. (2016) find that only about 1% of the total released tracer reaches the interior LS while about 15% ends up in the wider LS, suggesting that most of the freshwater is restricted within the narrow boundary current region.

Although the potential role of NAO-related buoyancy forcing in the shutdown of LS deep convection during 1969–71 has been acknowledged in previous studies (Hánkkinen 1999; Straneo 2006), a systematic comparison of the relative roles of extreme FSSIE and surface heat flux forcing for the shutdown of LS deep convection has been wanting. One paper that helps to fill this gap is a study by Gelderloos et al. (2012), who find nearly equal contributions of the two factors, using hydrographic data from OWS-B and a 1D mixed layer model. However, freshwater forcing in their 1D mixed layer model is applied as salinity restoring to the observed salinity profile at OWS-B. Consequently, if the observed freshening was the result of suppressed convective mixing due to a reduced heat loss, it is the response that they impose (through restoring), rather than the forcing itself. Thus, it is possible that the role of lateral freshwater transport is overestimated in their study. In the present study, we systematically re-examine the relative roles of these two drivers in the shutdown of LS deep convection associated with the GSA, using both fully coupled and forced ocean–sea ice simulations with the Community Earth System Model (CESM). In particular, we statistically isolate the relative oceanic impacts of extreme FSSIE and surface heat flux forcings through composite analyses, by taking advantage of a long CESM preindustrial coupled control simulation.

The NAO was also overall in a negative phase in the late 1950s–1960s. It has been extensively demonstrated that surface buoyancy forcing (mostly due to heat fluxes) associated with such a persistent NAO condition can modulate the thermohaline circulation in the North Atlantic (i.e., buoyancy-driven AMOC and subpolar gyre), leading to an upper-ocean heat content anomaly in the SPNA (Eden and Willebrand 2001; Lohmann et al. 2009; Robson et al. 2012; Yeager and Danabasoglu 2014; Kim et al. 2020). This anomaly tends first develop near the gyre boundary off the Newfoundland shelf and propagates northeastward following the NAC (Kim et al. 2020). While this mechanism is usually framed to explain upper-ocean temperature changes, it is expected that salinity (or freshwater) content behaves in a similar way. In fact, observed estimates of upper-ocean salinity in the SPNA show decadal fluctuations concurrent with those of upper-ocean temperature, including during the GSA period (Robson et al. 2014; Zhang 2017). Therefore, it is reasonable to hypothesize that a NAO-driven thermohaline circulation weakening might have also contributed to the advection of the low-salinity anomaly across the SPNA during the early 1970s, but this possibility has not been considered in previous studies on the GSA. This paper therefore considers two related hypotheses: 1) that anomalous NAO-related heat flux forcing was a dominant driver of the suppressed convection and upper-ocean freshening in the interior LS during the GSA and 2) that a weakening of the Atlantic thermohaline circulation played a significant role in the downstream propagation of the GSA signal.

The layout of the paper is as follows. We briefly describe the model simulations used in this study in the next section (section 2). Section 3 explores the relative roles of the extreme FSSIE- and NAO-related surface heat flux forcings for suppressed convection and freshening in the LS. Section 4 concerns the origins of the downstream propagation of the low salinity anomaly along the NAC. Potential dependency of our results on model biases and possible reasons for discrepancy with previous modeling studies are discussed in section 5, which is followed by a summary and concluding remarks in section 6.
2. Model and simulations

a. CESM1.2

Coupled simulations analyzed in this study use the CESM version 1.2 (CESM1; Hurrell et al. 2013). This is the same model version employed in the CESM1 Large Ensemble Project (Kay et al. 2015). CESM1 is a fully coupled Earth system model, which consists of the Community Atmosphere Model version 5 (CAM5; Neale et al. 2012), the Parallel Ocean Program version 2 (POP2; Danabasoglu et al., 2012), the Community Ice Code version 4 (CICE4; Holland et al. 2012), and the Community Land Model version 4 (CLM4; Lawrence et al. 2012). All components have nominal 1° horizontal resolution. CAM5 has 30 vertical levels with the model top at ~2 hPa. POP2 uses a level (z) coordinate in the vertical with 60 levels. The grid spacing increases monotonically from 10 m in the upper ocean to 250 m in the deep ocean. The grid north pole is displaced over Greenland, allowing for relatively finer grid levels. The grid spacing increases monotonically from 10 m in

b. Fully coupled simulations

We primarily use the CESM1 preindustrial control simulation (CESM1-PI) described in Kay et al. (2015) to isolate the oceanic impacts of extreme FSSIE and LS surface heat flux events, the details of which are described in the next section. CESM1-PI was run for 2200 years with the concentrations of atmospheric constituents fixed at their 1850 levels. To avoid initial transients and drifts, although the drifts mostly occur in the deep ocean (>1000 m), we discard the first 400 years and analyze the last 1800 years. Such a long integration length allows us to sample a relatively large number of extreme FSSIE and LS surface heat flux events for composite analysis.

Because of weaker-than-observed multidecadal variability of the winter NAO in CESM1-PI (Kim et al. 2018), it is hard to isolate persistent surface heat flux forcing in the LS and its potential role in freshwater advection along the NAC statistically. To overcome this difficulty, we utilize a set of fully coupled ensemble experiments with the same CESM1 version where the NAO-related surface heat flux anomaly is applied in the LS (Kim et al. 2020). In these experiments, a surface heat flux anomaly equivalent to 3 standard deviations of the observed negative NAO was imposed in the ocean component during the winter from December to March (we refer to these experiments as the ~NAO experiment). Use of such a large surface heat flux anomaly in these simulations is justified because the anomaly is applied only in a relatively small area of the LS, which is shown to be equivalent to ~1 standard deviation forcing when applied over the entire SPNA in Kim et al. (2020). The ~NAO experiment has 10 ensemble members with each member starting from different initial conditions for all components drawn from CESM1-PI. Each member was integrated for 10 years with the heat flux anomaly forcing and for another 10 years without the forcing for a total of 20 years. In this study, we consider the ensemble-mean difference from CESM1-PI (the ensemble mean of 20-yr segments that share the same initial conditions with the ~NAO experiment) as the response to the forcing in the coupled system. We refer to Kim et al. (2020) for further details of the experimental design.

Finally, to estimate the dependency of our results drawn from CESM1-PI on temperature bias in the LS region, we use a second set of fully coupled sensitivity experiments using the same CESM1 version where either ocean parameters or initial conditions were perturbed (Danabasoglu et al. 2019). In this suite, there are five experiments in which a few ocean parameters were changed within their acceptable ranges (i.e., parameters controlling the upper- and deep-ocean mesoscale eddy mixing, submesoscale eddy mixing, interior horizontal viscosity, and vertical background diffusivity) and two experiments where round-off level perturbations were applied to the initial atmospheric temperature field. Because an older version of the ozone concentration dataset was used in these experiments than the one used in CESM1-PI, a new preindustrial control simulation was also run. Each of these eight experiments was integrated for 600 years, starting from year 401 of CESM1-PI. As will be shown later, these sensitivity experiments provide a range of temperature biases in the LS region. Because we are not exploring the mechanisms that lead to different biases in each simulation in this study, we do not identify these simulations individually in our analyses. Further details of these experiments can be found in Danabasoglu et al. (2019).

c. Forced ocean–sea ice simulations

Ocean and sea ice observations are rather sparse during the GSA period (late 1960s to early to mid-1970s). Even if some data are available, they are either limited to a few locations (e.g., hydrographic data at OWS-B) or reconstructed from proxies [e.g., FSSIE reconstruction by Schmith and Hansen (2003)]. An alternative to obtain a dynamically and thermodynamically consistent picture of the ocean and sea ice states during the GSA is to use forced ocean–sea ice simulations (FOSI) constrained at the surface by our best estimates of the atmospheric states. Here we use FOSIs run at both low (~1°; FOSI-L) and high (~0.1°; FOSI-H) horizontal resolutions as benchmarks for the ocean and sea ice states during the GSA. Both FOSIs are forced with a recently developed, interannually varying forcing dataset based on the JRA-55 reanalysis (Japanese reanalysis product), called JRA55-do (Tsujino et al. 2018), which is used to drive ocean–sea ice simulations participating in the Ocean Model Intercomparison Project phase 2 (OMIP-2; Griffies et al. 2016) as a part of the Coupled Model Intercomparison Project phase 6 (CMIP6; Eyring et al. 2016).

These FOSI simulations use the newer CESM version 2 (CESM2) code base as described in Danabasoglu et al. (2020). As such, the ocean component, POP2, has some physics and numerical improvements. The sea ice model has been updated to CICE5 (Hunke et al. 2015). Starting from the January-mean potential temperature and salinity data from the World Ocean Atlas 2013 version 2 (WOA13; Locarnini et al. 2013; Zweng et al. 2013) as initial conditions, FOSI-L was run for five JRA55-do forcing cycles (1958–2009) with an extension to 2018 in the fifth cycle, largely following the OMIP protocol (Griffies et al. 2016). We use the fifth (last) cycle in our analyses. Sea
surface salinity was restored to climatological monthly WOA13 data with a time scale of one year over 50 m. Overall performance of the simulation is comparable to or better than that of corresponding simulations forced with the Coordinated Ocean–Ice Reference Experiments (CORE) interannual forcing dataset\(^1\) (Large and Yeager 2009), as shown in Tsujino et al. (2020) in the context of a multimodel comparison. CORE-forced FOSI has been extensively validated using available observations for interannual to decadal variability in the North Atlantic. In particular, closely related to the focus of this study, it has been shown that the observed variability in hydrography and mixed layer depth (MLD) in the LS is reasonably reproduced in the CORE-forced simulation (Yeager and Danabasoglu 2014; Kim et al. 2016). FOSI-L (JRA55-do forced) also shows very similar variability in these fields, giving us confidence that it faithfully represents processes in the LS.

FOSI-H uses the same CESM2 code base as in FOSI-L, but is configured for a higher horizontal resolution (nominal 0.1\(^\circ\)) that resolves (low latitudes) or permits (high latitudes) mesoscale eddies, which yields \(\sim 5-7\)-km grid spacing around Greenland and the LS. FOSI-H uses a tripole grid with the grid north poles over North America and Asia. The vertical grid is the same as in FOSI-L, but extends deeper to 6500 m (from 6000 m) with two additional levels for a total of 62 levels along with partial bottom cells for more accurate representation of bathymetry. The model initialization and salinity restoring are done in the same way as in FOSI-L, but it has been run for only one JRA55-do forcing cycle (1958–2018) because of the greater computational expense. It is expected that the model solutions experience some drift during the course of the first cycle. In our analysis, we use a relatively short 15-yr segment from 1965 to 1979 for FOSI-H as well as FOSI-L, and all anomalies are relative to the climatology of this 15-yr segment, unless noted otherwise. As will be shown later, overall variability in the fields of interest is very comparable between the two FOSIs during this 15-yr period. Further details of FOSI-H model setup along with a summary of its performance can be found in Chassignet et al. (2020).

### 3. Cause of the LS deep convection shutdown

We first consider the GSA signals simulated in FOSIs. Figure 1a shows the annual FSSIE time series around the GSA event. FSSIE is defined by sea ice volume transport perpendicular to the cross section between Greenland and Svalbard denoted in Fig. S1 in the online supplemental material. Both FOSIs show greatly enhanced FSSIE in 1968 and, to some extent, in 1967, but FSSIE during the preceding and following 2 years is below or close to the average of the reference period (1965–79), which is 2300 and 2030 km\(^3\) yr\(^{-1}\) in FOSI-L and FOSI-H, respectively. The FSSIE event of 1968 is also the largest in both FOSIs during the entire simulation period (Fig. S2). These characteristics of the simulated FSSIE event in the 1960s are consistent with those from other forced ocean–sea ice simulations (Haak et al. 2003; Wang et al. 2016; Ionita et al. 2016). The excess FSSIE in 1968 above the reference period average is 1470 and 1210 km\(^3\) yr\(^{-1}\) in FOSI-L and FOSI-H, respectively, which are well within the range of from several

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\(^1\) The new JRA55-do datasets are intended as updates and replacements for the CORE datasets.
hundreds up to 1700 km³ yr⁻¹ simulated in previous modeling studies (Häkkinen 1993; Jahn et al. 2012; Wang et al. 2016). The accumulated FSSIE anomalies for 1967 and 1968 are 2070 and 1960 km³ for FOSI-L and FOSI-H, respectively, which agree reasonably well with the observational estimate of 2300 km³ from Vinje (2001). Regarding the long-term average of FSSIE, observational estimates vary roughly between 2000 and 3000 km³ yr⁻¹ [see a review by Haine et al. (2015)]. Recent estimates, which employ modern remote sensing technologies, are generally around the low end of this range (e.g., Kwok et al. 2004; Spreen et al. 2009). For example, Kwok et al. (2004) estimate a mean FSSIE of 2218 ± 497 km³ yr⁻¹ for the 1991–99 period. The FOSI-L and FOSI-H averages for the same period are 2470 and 1950 km³ yr⁻¹, respectively, in good agreement with this observational estimate. Given both FOSIs use the same atmospheric forcing, the smaller FSSIE values in FOSI-H are likely due to thinner Arctic sea ice in FOSI-H (Fig. S3).

Previous studies (Dickson et al. 1988; Walsh and Chapman 1990; Häkkinen 1993; Tsukernik et al. 2010) have underscored the importance of anomalous northerly winds across the Fram Strait for the FSSIE event of 1968. Consistent with these studies, the JRA55-do datasets used to force FOSIs show SLP anomalies during the 1968 winter [January–March (JFM)] that support northerly wind anomalies across the Fram Strait (Fig. 2a) with an anomalous high (low) pressure centered over the Canadian Archipelago (northern Scandinavia). However, the sea ice thickness anomalies from FOSI-L also suggest that the FSSIE event of 1968 was preceded by anomalously thicker sea ice in the Arctic Ocean starting a few years earlier, similar to the findings of Haak et al. (2003). This thickness anomaly first emerges in the Beaufort Sea near the North American coast in 1965 and then circulates clockwise following the Beaufort Gyre and the Transpolar Drift toward the Fram Strait through 1967 (Figs. 2b–d). The anomaly is up to about 1 m at the mouth of the Fram Strait, which is a substantial increase in sea ice thickness.
considering that the mean sea ice thickness is around 1.5 m there (Fig. S3). Therefore, it appears that not only the northerly wind anomalies but also the preceding sea ice thickness anomalies in the Arctic Ocean are important for the 1968 FSSIE event.

Figures 1b–d show the time series of the annual upper 100-m salinity, March MLD, and JFM surface heat flux in the interior LS. Here, we focus on the interior LS, which is defined as the region where the ocean depth is greater than 2000 m within 51°–63°N and 42°–60°W (Fig. S1), to distinguish the convective basin from the boundary current regions. Consistent with the observed freshening in the LS (Lazier 1980; Dickson et al. 1988; Straneo 2006), FOSIs show a freshening in the upper ocean of the interior LS that starts in 1968, reaching the lowest salinity in 1970 before a sudden recovery in 1972 (Fig. 1b). Hydrographic data from Ishii and Kimoto (2009), hereafter simply called Ishii data, also show a similar freshening in the upper 100 m during 1967–71 and a recovery in 1972 (black line in Fig. 1b). Note that FOSI-H is roughly 0.1 psu saltier than FOSI-L and that both FOSIs are saltier than the Ishii data (the observed salinity is plotted with a +0.15 offset in Fig. 1b). We also note that the freshening is maximized in 1971 in Ishii data, whereas it is the largest in 1970 in both FOSIs. Associated with this freshening, March MLD in the interior LS decreases to about 300 m in FOSI-L and <200 m in FOSI-H in 1969 and 1970, and remains relatively shallow (<500 m) in 1971 in both FOSIs (Fig. 1c), indicative of suppressed deep convection in the LS. This shallow MLD is consistent with March MLD estimated from hydrographic data at OWS-B2 (Lazier 1980; black line in Fig. 1c). The MLD in FOSIs abruptly increases in 1972, suggesting the return of deep convection, and remains relatively deep during the rest of the mid-1970s. Such an abrupt MLD increase is also found at OWS-B.

During the low-salinity event, the winter surface heat loss in FOSIs is also anomalously weak in the interior LS (Fig. 1d) associated with negative NAO conditions (black line in Fig. 1d). In particular, the surface heat loss is more than 100 W m⁻² lower in the 1969 winter (JFM) than the reference period average of 210 and 250 W m⁻² in FOSI-L and FOSI-H, respectively, associated with the record low NAO. The sudden return of deep convection in 1972 also appears to be associated with a strong heat loss during the 1972 winter (Gelderloos et al. 2012). The winter NAO conditions strongly control the surface heat fluxes in the LS in FOSI-L with a (simultaneous) correlation coefficient of 0.75 for the entire simulation period. The surface heat fluxes, in turn, are the primary driver of deep mixing in the LS with a correlation coefficient of 0.82 with March MLD in the interior LS. These relationships are somewhat weaker in FOSI-H with 0.66 and 0.72 as the respective correlation coefficients, but these are also statistically significant at the 95% confidence level.

These results suggest that oceanographic conditions associated with the GSA in the LS appear to be reasonably reproduced in FOSIs and demonstrate that there are two potential driving factors for the suppression of deep convection: FSSIE and winter surface heat fluxes in the LS. However, because their respective influence on the LS is almost concurrent and there are other oceanic and atmospheric processes that may contribute to year-to-year variability in the LS, further obscuring the intrinsic role of these two factors, it may not be easy to disentangle the relative role of FSSIE and the surface heat flux for the suppression of LS deep convection without sensitivity experiments. Instead, we follow an alternate path and perform composite analyses for extreme FSSIE and LS surface heat flux events by taking advantage of the long integration length of CESM1-PI.

To sample FSSIE events from CESM1-PI as close as possible to the extreme FSSIE event of the late 1960s in FOSIs, we first inspect the monthly anomalies of FSSIE from FOSIs. Because FOSI-L has the same model resolution as CESM1-PI, we focus here on FOSI-L (Fig. 3), but similar monthly anomalies are also found in FOSI-H (Fig. S4). FOSI-L shows substantially enhanced FSSIE during the 1968 winter (starting from October 1967) with positive anomalies persisting throughout the summer (Fig. 3a). Some positive anomalies are also found during the first half of 1967. To find similar (analogous) events from CESM1-PI, we start with searching years (t = 0) when the FSSIE anomalies averaged between January and June are greater than 1.8 standard deviation, which is equivalent to exceed monthly average FSSIE anomaly of 166 km³ month⁻¹ above the 1968–69 baseline. From those years, we next find subsequent years (t = −1) when January to June FSSIE anomalies are greater than the long-term average. We identify 39 events that satisfy these criteria. The composite time series of monthly FSSIE anomalies from a year before and a year after the main event year (t = 0), which correspond to the calendar years 1967–69, shows a good similarity to the monthly FSSIE anomalies from FOSI-L (Fig. 3a). The accumulated composite FSSIE anomalies from t = −1 to 0 are 2440 km³, which is slightly greater than 2070 km³ in FOSI-L during 1967–68.

The impact of extreme surface heat flux events in the interior LS is isolated using a similar method. The monthly anomalies from FOSI-L (Fig. 3b) show substantially reduced surface heat loss (positive anomalies) in the winter of 1969, especially for JFM, and, to a lesser degree, in the winter of 1969–70. To mimic such events, we first search years (t = 0) when the LS surface heat flux anomalies averaged over JFM are greater than 1.8 standard deviations above the long-term JFM average of −309 W m⁻². This is equivalent to exceed monthly average positive anomalies that reduce the heat loss from the ocean. From those years, we next find subsequent years (t = 1) when the JFM surface heat flux anomalies in the interior LS are greater than the

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2 The observed MLD is a rough and subjective estimate based on hydrography at OWS-B, as stated in Lazier (1980), while the modeled MLD is computed based on a buoyancy gradient criterion (Large et al. 1997) and averaged over a larger interior LS area. We present the observed estimate for reference purposes, not for a direct comparison to the modeled MLD.

3 This threshold is chosen to match the anomalies of FOSI-L as close as possible while also sampling as many events as possible. However, the results are not very sensitive to the threshold values.
long-term average. We identify 37 events that satisfy these criteria. The composite time series of monthly LS surface heat flux anomalies from a year before and a year after the main event year, which correspond to the calendar years 1968–70, also exhibit a good similarity to the monthly LS surface heat flux anomalies from FOSI-L. The heat flux composite anomalies averaged over the cold season (November–March) are 78 W m$^{-2}$ for the main event ($t=0$) and 60 W m$^{-2}$ for the following year ($t=+1$), which are comparable to the corresponding anomalies of 67 and 69 W m$^{-2}$, respectively, in FOSI-L. The (one standard deviation) spread of both FSSIE and surface heat flux composite time series is shown in Fig. S4. We note that the two events are almost mutually exclusive: out of about 40 events, there are only two surface heat flux events that overlap with FSSIE events (with the former lagging the latter by 1 and 2 years) and there is no simultaneous overlap.

Figure 4 shows the composite maps of the annual upper 100-m salinity and March MLD anomalies for the FSSIE events from the main event year ($t=0$) to the following two years ($t=+1$) (Figs. 4a–c) and for the LS surface heat flux events from a year before ($t=-1$) to a year after ($t=+1$) the main event ($t=0$) (Figs. 4d–f). Note that for both event composites, these years correspond to the calendar years 1968–70, and the corresponding anomalies from FOSI-L for these years are also shown (Figs. 4g–i). In the main FSSIE event year (Fig. 4a), the enhanced FSSIE appears to lead to a freshening in the Greenland and Irminger Seas with the strongest signal occurring to the northeast of Iceland. This anomaly pattern bears a resemblance to that of FOSI-L in 1968 (Fig. 4g), although the magnitude is much weaker and the freshening is more extensive toward the LS in the FSSIE event composite. In the following year (Fig. 4b), the freshening further advances to the LS with the strongest freshening along the southern coast of Greenland. However, the freshening is still confined to the northern LS, concentrated on the Greenland shelf, and a MLD decrease is noticeable only in a limited region near the southern tip of Greenland. In contrast, FOSI-L shows a substantial decrease in MLD in the interior LS, up to about 1000 m, in 1969 (Fig. 4h) with a freshening in the western and interior LS. In particular, the freshening is the largest to the immediate south of the MLD anomaly. In the following year (Fig. 4e), the freshening moves farther along the Labrador shelf following the Labrador Current in the FSSIE event composite, but no significant MLD reduction is found. In FOSI-L, the freshening grows stronger in 1970 and the MLD decrease in the interior LS persists (Fig. 4i). These composite results suggest that an extreme FSSIE event could lead to some freshening and MLD decrease in the northern LS and Irminger Sea, but only in a limited region close to the shelf, leaving the interior LS largely unaffected.

Figure 5 shows the salinity and MLD anomalies from FOSI-H, corresponding to Figs. 4g–i from FOSI-L. Although more confined to narrower boundary currents and showing more complex structures, the salinity anomalies in FOSI-H are largely consistent with those in FOSI-L: a freshening along the EGC and around the Greenland Sea in 1968 (cf. Figs. 5a and 4g), continued freshening in the northeastern LS and suppression of convection in the interior LS with an accompanied freshening in 1969 (cf. Figs. 5b and 4h), and a propagation of freshening along the Labrador Current with sustained suppression of convection and freshening in the interior LS in 1970 (cf. Figs. 5c and 4i). A noteworthy difference is that the propagation of the low-salinity anomaly along the boundary currents is more evident in FOSI-H, which appears to be separated from the anomaly in the interior. This may be indicative of different driving mechanisms for the freshening in the interior and boundary currents.

As discussed above, the initial negative salinity anomaly is weaker in the FSSIE composite than FOSI-L, particularly in the Iceland and Irminger Seas (cf. Figs. 4a and 4g), which may contribute to a weaker MLD response in the following years in the FSSIE composite. This contrasting magnitude could be related to differences in atmospheric circulation. As hinted by the winter SLP anomalies shown in Fig. 2a, there is an
FIG. 4. Annual upper 100-m salinity (colors) and March MLD (purple contours) anomalies from the (a)–(c) FSSIE event composite from \( t = 0 \) (main event year) to \( t = 2 \), (d)–(f) LS SHF event composite for a year before \( (t = -1) \) and after \( (t = 1) \) the main event year \( (t = 0) \), and (g)–(i) FOSI-L for the 1968–70 period. Negative March MLD anomalies are shown in purple contours at intervals of 300 m (there are no positive MLD anomalies greater than 300 m) and the zero contour is omitted. The black contours in (a)–(f) indicate statistically significant salinity composite anomalies at the 95% confidence level. The anomalies of FOSI-L are relative to the 1965–79 averages.
anomalous high pressure ridge located at the center of the SPNA in 1968 in observations. Together with the low pressure anomaly over Scandinavia, this generates anomalous northwesterlies over the Iceland and Irminger Seas that could divert fresh surface water and sea ice in the EGC to the Iceland Sea. On the other hand, the FSSIE composite largely averages out anomalous atmospheric circulations, so that there is less diversion of freshwater into the Iceland Sea. If the FSSIE events are subsampled for those showing a greater freshening in the Iceland Sea, the resulting SLP composite anomalies indeed show an anomalous ridge in the SPNA (Fig. S5), supporting our interpretation. Despite an initially stronger accumulation of freshwater north of Iceland in this subsampled composite, however, the MLD weakening in the interior LS is still absent. This is also supported by low-salinity anomalies in Baffin Bay in FOSIs, most notably in FOSI-L, which is absent in the FSSIE composite. However, the anomalous freshwater transport through the Davis Strait is small (only positive into the LS for 1968 and 1969 with the magnitude of about 100 km$^3$ yr$^{-1}$). Thus, we expect that the contribution of the low-salinity anomaly in Baffin Bay to the downstream freshening is small.

The LS surface heat flux event composites, in contrast, suggest that an extreme surface heat flux event in the interior LS is quite effective at suppressing convection in this region. In the main event year (Fig. 4e), there is a substantial shoaling of MLD, up to 900 m, in the southern interior LS. This decrease is comparable in magnitude to that in 1969 in FOSI-L (Fig. 4h), but takes place farther south and over a larger area. This spatial discrepancy is likely due to the spatial pattern of the mean March MLD, which extends farther south in the LS in CESM1-PI than in FOSI-L (Fig. S6). Also, a freshening to the immediate south of the MLD anomaly is apparent in the main event year, similar to the freshwater anomaly of 1969 in FOSI-L. We will discuss later the reason for this freshening that appears to be related to the MLD anomalies. There is also a hint of a MLD decrease and freshening in the following year in the surface heat flux event composite (Fig. 4f) as seen in FOSI-L (Fig. 4i), but the magnitude of the signal is much weaker than in FOSI-L, partly because of the weaker winter surface heat flux anomaly in the SHF composite (Fig. 3b).

The spatial patterns of the JFM surface heat flux anomalies associated with the suppressed convection in the surface heat flux event composite (Figs. 6e,f) agree reasonably well with those in FOSI-L (Figs. 6h,i) with reduced surface heat loss in the LS. In particular, the reduced heat loss is accompanied by a negative NAO-like SLP pattern in the surface heat flux event composite, with a strong anomalous high pressure around Greenland (Fig. 6e), reasonably capturing the observed negative NAO pattern in the winter of 1969 (Fig. 6h). As mentioned earlier, the JFM SLP as well as the surface heat flux anomalies are negligible in the FSSIE event composites (Figs. 6a–c), except for some flux anomalies around the sea ice edge and a low pressure anomaly over Scandinavia, which drives extreme FSSIE as found in FOSIs (Fig. 2a). This suggests that the surface ocean and sea ice conditions induced by the FSSIE events do not lead to a particular anomalous atmospheric circulation.

The FSSIE events identified from CESM1-PI are already, by construction, unusually strong. However, it is still possible that the response of convective mixing in the LS is nonlinear so that it is only significant when FSSIE is the most extreme. To investigate if there is such a nonlinearity, we show in Fig. 7a the March MLD anomalies in the interior LS as a function of the FSSIE anomalies from individual FSSIE events. Here the MLD anomalies are the average over the next two years after the main event year (from $t = 1$ to 2) and the FSSIE anomalies are accumulated over the main event and the preceding years (from $t = -1$ to 0). No general relationship is found between the two quantities, as indicated by a low correlation coefficient of −0.26, and even for the most extreme FSSIE anomalies (e.g., $>3000$ km$^3$), MLD anomalies are found across a wide range from $-600$ to +400 m (the latter indicating MLD deepening), suggesting that a nonlinear response is unlikely.

![Fig. 5](image-url)
The MLD anomalies from the FSSIE events are instead strongly related to concurrent surface heat flux anomalies in the interior LS with a correlation coefficient of $0.81$ (Fig. 7b), regardless of the strength of FSSIE, further underpinning the primary role of the local surface heat flux in controlling convective mixing in the LS.

Our composite analyses suggest that much reduced surface heat loss is more effective than extreme FSSIE at driving a
How can surface heat flux forcing create such a freshening? An upper-ocean low-salinity anomaly can be generated when winter surface heat loss is sufficiently weak so that mixing of relatively fresh upper ocean with salty deeper ocean is prevented. Figures 8a–c show the monthly anomalous salinity profiles from the LS surface heat flux event composite during the main event winter (starting from November of \( t = -1 \)) and from both FOSIs during the 1969 winter (starting from November 1968). The profiles are averages over respective regions where the local freshening is collocated with the suppressed convection in the LS (each domain is outlined in Fig. S7) in order to maximize the signal associated with reduced convective activity, which takes place at somewhat different locations in each of the simulations (Figs. 4 and 5). However, a similar result, but with a weaker signal, is also found when the common interior LS domain used earlier is considered (Fig. S8). In the surface heat flux event composite, a freshening occurs in the upper ocean gradually from November to March with a slight increase in the depth of the fresh layer. Meanwhile, a gradual salinification also takes place in the deeper ocean, which starts from a relatively thin layer immediately below the fresh layer in January and penetrates into the deeper ocean (>1000 m) by March (Fig. 8a). A similar evolution of freshening (salinification) in the upper ocean (deeper ocean) is found in both FOSIs (Figs. 8b,c), although the upper-ocean freshening is smaller, but penetrating deeper, and salinification at depth starts earlier in time with slightly higher magnitudes.

Such a freshening (salinification) in the upper (deep) ocean can also be generated when deep mixing is suppressed due to an excess FSSIE-induced freshwater input. In such a case, however, an initial freshening is expected during the restratification season (May–October). In the case of GSA, this restratification would have started in the summer of 1968 when the associated fresh anomaly arrived at the LS via the EGC (Dickson et al. 1988). Indeed, the freshening in the FSSIE composite is evident during the restratification season (Fig. S8a). Yet, both FOSIs exhibit a rather abrupt freshening at the beginning of the 1969 winter (Figs. 8e,f) when the anomalous heat fluxes start to increase (Fig. 3b and Fig. S4b). The Ishii data also reveal a freshening mostly taking place during the 1969 winter (Fig. S8e). This picture of the freshening is largely consistent with that of the surface heat flux event composite (Fig. 8d), suggesting that the surface freshening in the interior LS observed during the GSA could have primarily resulted from the suppressed deep mixing induced by greatly reduced surface heat loss. In other words, the freshening in the interior LS could be a response rather than a forcing for the suppression of convection.

4. Downstream propagation of the low-salinity anomaly

Another key aspect of the observed GSA is the downstream advection of a low-salinity signal around the SPNA, which eventually returns to the Nordic seas about a decade later (Dickson et al. 1988). It is generally accepted that this propagation is an advection of the freshwater anomaly originating from the excess FSSIE in the late 1960s (Dickson et al. 1988; Belkin et al. 1998). Figures 9e–l show the 2-yr averages of the upper 100-m salinity anomalies from FOSIs for the 1971–78 period when the low-salinity anomaly was observed at successive locations across the SPNA. This propagation of the low-salinity anomaly in FOSIs is generally consistent with the observation-based description: a negative salinity anomaly near the Labrador Shelf and Newfoundland Basin in 1971–72 (Figs. 9e,i) moves toward the central SPNA around 1974...
reaching south of Iceland in 1975–76 (Figs. 9g,k) and the Norwegian Sea in 1977–78 (Figs. 9h,l). The salinity anomaly from the FSSIE event composite during the corresponding years, i.e., 3–10 years after the main FSSIE event year (Figs. 9a–d), also show a propagation of low-salinity water along the Labrador Current and into the Newfoundland Basin 3–6 years after the FSSIE events (Figs. 9a,b). However, the low-salinity anomaly subsequently dissipates near the gyre boundary without any significant downstream propagation into the eastern SPNA (Figs. 9c,d). Even for the most extreme FSSIE events (nine cases in which the accumulated FSSIE anomalies are greater than 3000 km³ as shown in Fig. 7a), the low-salinity anomaly does not advance farther north from the gyre boundary and dissipates there (Fig. S9). Therefore, these results suggest that an injection of freshwater from the extreme FSSIE event as found in CESM1-PI may not be strong enough to maintain a low-salinity anomaly that propagates all the way to the northeastern SPNA during the late 1960s.

Another possible explanation is that the downstream propagation of the low-salinity anomaly is attributable to an adjustment of the thermohaline circulation system in the North Atlantic in response to surface heat flux forcing in the LS associated with a decade-long negative NAO condition during the 1960s (Robson et al. 2014). If this is the case, one would expect a similar advection of low-salinity water from the gyre boundary to the northeastern SPNA in the subsurface because of the vertical extent of the upper limb of the AMOC. In contrast, the low-salinity anomalies related to FSSIE would be surface-intensified. The Ishii data indeed show evidence of subsurface (275–450 m) salinity anomalies (Figs. 10a–d) that could be related to changes in the large-scale ocean circulation. There is a notable absence of low-salinity anomaly in the western basin in 1971–72 followed by a northward propagation of a low-salinity water across the SPNA. Both FOSIs show a spatiotemporal evolution of subsurface salinity that is broadly consistent with the Ishii data (Figs. 10e–l). FOSI-H, in particular, shows a clear propagation of low-salinity water from the

**Fig. 8.** (a)–(c) Monthly anomalies of the salinity profile in respective LS domains, defined in Fig. S7, from (a) the LS surface heat flux event composite from November of a year before the main event year \( t = -1 \) to March of the main event year \( t = 0 \), and from (b) FOSI-L and (c) FOSI-H from November 1968 to March 1969. For clarity, only November, January, and March anomaly profiles are shown. (d)–(f) Close-up contour plots of the respective monthly salinity anomalies in the upper 200 m for every month from \( t = -1 \) to 0 and 1968–69 for (d) the surface heat flux event composite and (c),(f) FOSIs, respectively. The anomalies of FOSIs are relative to the respective monthly 1965–79 averages.
Newfoundland shelf to the Norwegian Sea via the eastern SPNA that appears unrelated to any signals in the Labrador Current (Figs. 10i–l). All three products show a positive salinity anomaly on the northern flank of the Gulf Stream. Together with the negative salinity anomaly in the SPNA, this resembles the pattern suggested as a fingerprint of AMOC weakening (Zhang 2008).

As described in section 2, we utilize the coupled ensemble simulations where a persistent surface heat flux forcing associated with the observed negative NAO was applied in the LS (−NAO experiment) to overcome the weak multidecadal NAO variability in CESM1-PI. Figures 10m–p show the 2-yr averages of the ensemble mean subsurface salinity anomalies from this experiment. In this experiment, the propagation speed appears to be slower than that in FOSIs, so 2-yr averages are plotted from simulation years from 3 to 13 skipping a year in between. A few years after the forcing starts, a low-salinity anomaly emerges near the gyre boundary (Fig. 10m). The anomaly then moves toward the eastern SPNA over the next few years (Figs. 10n,o) and is eventually found south of Iceland (Fig. 10p). This propagation is similar to that seen in FOSIs, but the amplitude is weaker. The stronger and faster signal of the anomaly in FOSIs could be aided by the positive NAO conditions during 1972–76 (Fig. 1d), which could itself generate a low-salinity anomaly in the SPNA via its momentum forcing and associated spinup of the subpolar gyre (Mignot and Frankignoul 2003; Holliday et al. 2020). We note that no large-scale freshening signal and subsequent
advection in the SPNA are found in the surface heat flux composite (Fig. S10), bolstering the importance of persistent NAO-related surface buoyancy forcing for a significant thermohaline circulation response, the generation of a large-scale subsurface salinity anomaly, and northeastward advection across the SPNA.

The subsurface low-salinity anomalies examined above correspond to anomalously weak AMOC in both FOSIs (in the

![Fig. 10](image-url)
early 1970s) and the −NAO experiment, as a result of preceding persistent negative NAO forcing (Fig. 11). The −NAO experiment shows more persistent negative AMOC anomalies with magnitudes generally larger than FOSIs. This is because the AMOC from FOSIs (Fig. 11b) includes both buoyancy- and wind-driven anomalies, while ensemble averaging in the −NAO experiment isolates the buoyancy-forced component. It is shown in Yeager and Danabasoglu (2014) that the AMOC strength is predominantly weak from the mid-1960s to mid-1970s when forced with interannually varying surface buoyancy fluxes only (i.e., Fig. 6 of Yeager and Danabasoglu 2014). In contrast, the AMOC anomalies from the FSSIE event composite are very small, less than a few tenths of a Sverdrup (1 Sv = 10⁶ m³ s⁻¹; Fig. 11a), indicating that the impact of extreme FSSIE on AMOC is rather minor.

In summary, the results presented in this section lend support to the hypothesis that persistent negative NAO forcing during the 1960s probably contributed substantially to, or might well have dominated, the observed advection of anomalously low salinity from the gyre boundary to northeastern SPNA during the 1970s. This scenario could hold even if the advection of low-salinity water along the boundary currents in the western basin had its origin in the extreme FSSIE in the 1960s, as is widely accepted and not disputed here.

5. Discussion

a. Dependency on temperature and salinity biases

Similar to many other climate models (e.g., Menary et al. 2015), CESM1-PI suffers from upper-ocean temperature and salinity biases in the LS, which is warmer and saltier compared to observed climatology. Because of the nonlinearity of the seawater equation of state, density change due to a given salinity change is larger at low temperatures. In other words, the MLD decrease induced by the extreme FSSIE events in CESM1-PI could be greater if the temperature bias is reduced. Also, a salty bias could exaggerate the influence of surface heat flux on deep convection due to the fact that excess temperature (but not excess salinity) can be easily removed by winter surface fluxes, making the upper ocean denser than it would otherwise be. To evaluate the dependency of our results from the composite analyses on temperature and salinity biases in the LS, we utilize, as described in section 2b, a suite of CESM1 sensitivity experiments (Danabasoglu et al. 2019), which show a range of temperature and salinity biases in the LS. We perform the same extreme FSSIE and LS SHF events composite analyses for the individual simulations. Because of the shorter length of these simulations (600 years each) compared to CESM1-PI, the threshold of 1.8 standard deviation used above to define events gives a small number of samples (generally <10) per simulation. Therefore, we reduce the threshold to 1.25 standard deviation for both extreme FSSIE and LS SHF events to obtain sample sizes comparable to those obtained from CESM1-PI. We also divide CESM1-PI into three 600-yr segments as in Danabasoglu et al. (2019) and repeat the composite analyses for each segment with this reduced threshold. The sample sizes of these composites are generally between 30 and 40 and the characteristics of the upper-ocean properties as well as MLD responses are largely consistent with those found from the composite analyses with CESM1-PI.

Figure 12 shows the scatterplots of March interior LS MLD composite anomalies for the FSSIE and LS surface heat flux events as a function of the mean upper 400-m temperature and salinity biases relative to WOA13 from the CESM1 sensitivity.
experiments and the three segments of CESM1-PI. The March LS MLD composite anomalies are the averages over one to two years after the main event year ($t = 1$ and $2$) for the FSSIE event composite and over the main event year and the next year ($t = 0$ and $1$) for the LS surface heat flux event composite (both corresponding to the 1969–70 average). These simulations show positive temperature and salinity biases with a range of $1.8$–$2.8$°C and $0.27$–$0.44$ psu, respectively. The scatterplot for the FSSIE event composite shows that the LS MLD composite anomalies are indeed positively correlated with the temperature and salinity biases (Figs. 12a,c), although the correlations are moderate and statistically significant only at the 75% confidence level, suggesting that the MLD reduction could be greater if the biases are reduced. A regression analysis indicates that the MLD composite anomaly could be from around −140 to −220 m if the biases are absent, while the MLD composite anomaly is about −40 m in CESM1-PI with the temperature and salinity biases of −1.3°C and 0.35 psu, respectively. However, this projected MLD reduction with zero bias is still much smaller than that of the original surface heat flux composite using the full length of CESM-PI (−2350 m). In addition, the MLD anomalies in the LS surface heat flux event composite show even higher positive correlations with the biases with a projected MLD decrease with zero bias of −450 m (Figs. 12b,d). These results suggest that our conclusions drawn from the composite analyses using CESM1-PI are not likely affected by the temperature and salinity biases in the LS. A caveat here is that the bias range across the CESM1 sensitivity experiments is much smaller than CMIP5 range from −3° to 4°C and from −0.5 to 1 psu (Menary et al. 2015), so the dependency
needs to be further confirmed considering a wider range of the biases.

Another possible caveat of our conclusions is that the MLD response in the SHF composite takes place farther south of the LS convection region seen in observations and is perhaps too extensive (cf. Fig. 4e and Fig. S6d). This is likely because the stratification in the deep ocean of the LS in CESM1-PI as well as FOSI-L is too weak (Fig. S11), in particular in the northern LS. This leads to low (high) sensitivity of the MLD to surface buoyancy forcing in the northern (southern) LS. However, observed LS stratification is reproduced very well in FOSI-H (Fig. S11). Given the very consistent year-to-year MLD variability in both FOSIs during the GSA period (Fig. 1c), to a degree that would be hard to explain if the variability were driven by different processes, we think that the location bias of MLD in the SHF composite does not invalidate our conclusions. In addition, we note that in a series of studies where nearly 20 ocean–sea ice models are forced with identical atmospheric fields, Danabasoglu et al. (2016) show that LS MLD variability is fairly consistent between the models, despite a wide range of upper LS temperature and salinity biases across the models (Danabasoglu et al. 2014).

b. Discrepancy with previous studies

Our results underscoring local surface heat flux forcing as a potentially dominant driver for the suppression of deep convection in the LS during the GSA are in conflict with several previous modeling studies that suggest the extreme FSSIE event in the 1960s was the primary driver (e.g., Hakkinen 1999; Koenigk et al. 2006). There could be many factors contributing to this discrepancy. The simulations in these previous studies were carried out more than a decade or two ago when the model resolutions were generally coarser and parameterizations for unresolved processes were not as advanced as used in more recent state-of-the-art models. For example, the horizontal resolution of the ocean component of the coupled model used by Koenigk et al. (2006) was 2.8°. At such a resolution, the boundary currents such as the EGC and WGC would be broader than those in the models used in this study and thus, when there is enhanced FSSIE, it is likely easier for excess freshwater to enter the interior LS and more effectively suppress deep convection there [cf. Figs. 8 and 9 of Koenigk et al. (2006) and Fig. 4 herein]. Even at a similar horizontal resolution, the sharpness of the boundary currents can be considerably different depending on the choice of horizontal viscosity parameterization and its local magnitude (Jochum et al. 2008). In CESM1-PI and FOSI-L, the horizontal viscosity is 600 and 1200 m² s⁻¹ in the zonal and meridional directions, respectively, around the LS and Greenland, which is much smaller than the value of 3 × 10⁵ m² s⁻¹ used in Hakkinen (1999). The larger value in the latter study is expected to produce broader (more diffuse) boundary currents around the LS. Needless to say, the horizontal resolution of CESM1-PI and FOSI-L around the LS and Greenland is not fine enough to resolve all the details of the circulation systems there, including eddies and frontal jets. However, the overall consistency of the results between FOSI-L and FOSI-H suggests that the nominal 1° horizontal resolution that telescopes to approximately 35- to 50-km resolution near the LS region, with proper choice of subgrid-scale parameterizations and associated parameters, can capture the first-order processes involved in the downstream propagation of FSSIE and related freshwater to the LS. We note that even 0.1° horizontal resolution is likely insufficient for realistic representation of all the processes that govern downstream propagation of freshwater transport through Fram Strait.

The impact of an extreme FSSIE on LS convection may also depend on how much freshwater export actually enters the SPNA through the Denmark Strait. In Hakkinen (1999), the enhanced FSSIE is associated with an enhanced sea ice export through the Denmark Strait (DSSIE) in the next winter, whose magnitude is comparable to that of FSSIE (Fig. 4 of Hakkinen 1999). This is because a large portion of sea ice exported through the Fram Strait accumulates in the Greenland Sea, survives during the melting season, and then exits through the Denmark Strait during the next cold season. This does not appear to be the case in the FSSIE event composite in our study. The composite time series of the monthly DSSIE anomalies, defined as the sea ice volume transport perpendicularly to the section denoted in Fig. S1, for the extreme FSSIE events (Fig. 13a) shows that the increase in DSSIE in the following cold season is very small. Such weak DSSIE anomalies are found consistently in both FOSIs during the following cold season after the FSSIE event in 1968 (Figs. 13b-c). The small downstream propagation of excess FSSIE in our simulations occurs because much of the excess FSSIE melts within the Greenland Sea. While this meltwater could also propagate downstream, the anomalous total (liquid plus solid) freshwater export through the Denmark Strait is still much smaller than that through the Fram Strait in the FSSIE event composite and both FOSIs (Fig. S12), likely because of the accumulation and recirculation of freshwater in the Iceland Sea, which is clearly visible in the salinity anomalies (Figs. 4 and 5). We note that total freshwater transport at both the Fram and Denmark Straits is somewhat larger in FOSI-H than in FOSI-L (Fig. S12).

While much of liquid freshwater from the Fram Strait is exported to the SPNA via the EGC (Dickson et al. 2007; Dodd et al. 2009), sea ice does not stay within the EGC (Dodd et al. 2009), but expands eastward during the winter, as the sea ice extent indicates (Fig. S3). Therefore, some accumulation and recirculation of freshwater anomalies in the Nordic seas does not seem to be unrealistic, as some sea ice would melt in the interior away from the EGC during the melting season. Nevertheless, the strong negative salinity anomaly in the Iceland Sea (Figs. 4 and 5) may be an overestimate due to model bias. Ypma et al. (2019) show, using an older version of the ocean component of FOSI-H (POF2), that simulated near-surface salinity is fresher in the Iceland and Norwegian Seas compared to observations due to the fact that warm and salty Atlantic water stays confined too close to the Scandinavian coast, and unrealistic recirculations in the Iceland and Norwegian Seas (Figs. 2g and 2i of Ypma et al. 2019) bring fresher water from the EGC to these seas. However, none of the simulations used in the present study appears to suffer from such a spurious recirculation (Fig. S13), as inferred from the
mean sea surface height, although the inflow of Atlantic water appears to be still too confined to the Scandinavian coast. As discussed earlier, the strong negative anomaly in the Iceland Sea could also be contributed by offshore transports induced by anomalous atmospheric circulation.

Another way to verify whether the downstream propagation of the FSSIE-induced freshwater anomaly is realistic is to compare the simulated upper ocean salinity with observations at the Fylla bank section (Fig. 14), which is a relatively well-sampled hydrographic section in the northern LS (Fig. S1). Observations show a decrease of about 0.6 psu for 1969–71 relative to the early to mid 1960s at the Fylla Bank (Belkin et al. 1998). FOSI-H shows a salinity decrease that compares reasonably well with these observations. While FOSI-L also shows a similar salinity decrease, the amplitude is much weaker, about one-half of that seen in observations and FOSI-H. The composite upper ocean salinity for the FSSIE event at this section also shows negative salinity anomalies 1–3 years after the main event year (corresponding to 1969–71), but similar to FOSI-L, the amplitude is weak. This suggest that the downstream propagation of FSSIE-induced freshwater anomaly may be underestimated in the coarse resolution simulation due to a recirculation in the Iceland Sea and/or excessive lateral mixing en route to the LS. Note also that the total freshwater transport through the Denmark Strait is somewhat larger in FOSI-H than in FOSI-L. However, when the nine largest FSSIE events (as for Fig. S9) are considered, the amplitude of the negative salinity anomaly is comparable to both observations and FOSI-H in 1969, but as shown earlier (Fig. 7) the MLD decrease is not significantly different from the full FSSIE event composite.

6. Summary and concluding remarks

In this study, the origins of the GSA of the 1970s are revisited using multiple model simulations with CESM1. A particular motivation for this revisit is the recognition that the atmospheric conditions characterized by a negative winter NAO state were as favorable for GSA-like oceanic responses as the excessive solid freshwater input from the Arctic through the Fram Strait (i.e., FSSIE). To our knowledge, a systematic comparison of these two factors has not been made using comprehensive general circulation models. Given sparse observations during the GSA event, we use FOSIs at both nominal 1° and 0.1° horizontal resolutions, constrained at the surface by realistic atmospheric states, as benchmarks for the ocean and sea ice states during the GSA. Both FOSIs simulate reasonably well the observed characteristics of the GSA: greatly enhanced FSSIE, suppressed deep convection associated with a
freshening in the LS, and downstream propagation of the low-
salinity anomaly along the boundary currents around the LS,
into the NAC, and across the SPNA to the Nordic seas.

We investigate the relative contributions of extreme FSSIE
and local surface heat flux events to suppressed convection in
the LS by statistically isolating each impact through composite
analysis of the 1800-yr-long preindustrial control simulation of
CESM1 (CESM1-PI), and compare the composite anomalies
to those from FOSIs. The results reveal, surprisingly, a minor
response of convective mixing strength in the LS to the ex-
treme FSSIE events, while showing greatly suppressed con-
vective mixing in response to extreme anomalous surface heat
flux into the ocean, that is, reduced heat loss by the ocean.

Another interesting finding is that a substantial freshening in
the interior LS is found in the extreme surface heat flux event
composite, as a result of reduced vertical mixing of relatively
fresher surface waters with saltier deeper ocean (Houghton
and Visbeck 2002). In contrast, the freshwater anomaly is
largely found only along the boundary currents encircling the
LS in the extreme FSSIE event composite. These results sug-
gest that the observed suppression of convection and fresh-
ing in the interior LS associated with the GSA could, in fact,
be substantially contributed or even dominated by significantly
reduced surface heat loss in the LS related to the negative
NAO conditions at the time of the event. Our results stressing
the critical role of anomalous surface heat flux forcing for the
shutdown of LS convection go further than Gelderloos et al.
(2012), who argued for a roughly equal contribution of fresh-
water and heat flux forcing based on a 1D mixed layer model.
The key factor explaining this difference seems to be the
freshening induced by reduced convective mixing in the inte-
rior LS, as the forcing of the 1D mixed layer model of
Gelderloos et al. (2012) potentially contains this mixing-
induced freshening (through restoring to observed salinity
profiles at OWS-B).

The FSSIE event composite also suggests that the freshwa-
ter anomaly induced by the FSSIE event during the late 1960s
might not be strong enough to sustain the observed low-salinity
signal propagating downstream along the NAC to the north-
eastern SPNA (Dickson et al. 1988), as the low-salinity signal
dissipates along the gyre boundary. Instead, the subsurface
signature of salinity anomalies from both observations and
simulations suggest the downstream propagation of the low-
salinity anomaly could have been caused by an adjustment of
thermohaline circulation in response to a persistent negative
NAO-related forcing. This mechanism is identical to what is
used to explain the multidecadal upper ocean heat content
variations in the SPNA (e.g., Kim et al. 2020). Observations
indeed show the upper ocean salinity evolving concurrently,
on decadal time scales, with the upper ocean temperature in
the SPNA (Robson et al. 2014; Zhang 2017), further supporting
the conclusions of the present study.

Our study suggests that the interior LS, specifically deep
convection, is relatively insensitive to the freshwater input
through the Fram Strait. This is likely because the freshwater
anomaly is largely confined within narrow boundary currents
encircling the LS. This is consistent with recent modeling studies
that simulate a relatively small accumulation of freshwater from
Greenland runoff in the interior LS, as inferred from exper-
iments in which passive tracers are released simultaneously with
This might be the reason why LS convection and AMOC responses
are also rather small in response to a recent increase in fresh-
water fluxes from the Greenland ice sheet in an eddy-resolving
ocean simulation (Böning et al. 2016). However, these findings do
not necessarily mean that the freshwater fluxes from Greenland
ice sheet melting and Arctic sea ice melting in the future will not
impact deep-water formation in the LS and thermohaline circu-
lation. While aforementioned studies, including ours, suggest that
the immediate impact of excess freshwater on the LS is small,
freshwater input can accumulate in the LS over a long time after
propagating around the SPNA (Dukhovskoy et al. 2016; Böning
et al. 2016; Robson et al. 2016). Therefore, given an anticipated
increase in freshwater input from Greenland and Arctic due to a
warming climate in the future, deep-water formation in the LS
and AMOC could eventually weaken.

The results of the present study offer an alternative per-
spective on two aspects of the GSA story: 1) the surface heat
flux forcing associated with the extreme negative NAO con-
ditions in the late 1960s and early 1970s could have substan-
tially contributed to or even dominated the suppression of
deep convection and associated freshening in the interior LS;
and 2) the advective propagation of the negative salinity
anomaly from the gyre boundary to the Nordic seas across the
SPNA could have been due to an adjustment of ocean ther-
mothaline circulation in response to the persistent negative
NAO condition during the 1960s. Although our results are
robust within the framework of the models we used (i.e.,
CESM1), they admittedly suffer some biases, particularly in the
1° models, that could affect the findings of the study.

Acknowledgments. We thank Dr. Igor Belkin, two anony-
umous reviewers, and the editor, Rong Zhang, for their con-
structive comments that helped to improve this manuscript.
This research was supported by the National Oceanic and Atmosphe-
ric Administration (NOAA) Climate Program Office under Climate
Variability and Predictability Program Grant NA16OAR4310170
and the Regional and Global Modeling Program (RGCM) of the
U.S. Department of Energy’s Office of Science (BER) through its support of the HiLAT project. This material is based upon work supported by the National Center for Atmospheric
Research, which is a major facility sponsored by the National
Science Foundation under Cooperative Agreement 1852977. The
availability of the datasets used in this study is as follows: CESM1-
PI at https://www.earthsystemgrid.org; FOSI-L at https://esgf-
node.llnl.gov; FOSI-H, −NAO, and perturbed CESM1 experi-
ments available upon request; Ishii hydrography data at https://
rdac.cci.climate-dataguide.ucar.edu/climate-data/hurrell-north-atlantic-
ocean-oscillation-nao-index-station-based.


