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ABSTRACT

Deep convection in the Labrador Sea (LS) resumed in the winter of 2007/08 under a moderately positive North Atlantic Oscillation (NAO) state. This is in sharp contrast with the previous winter with weak convection, despite a similar positive NAO state. This disparity is explored here by analyzing reanalysis data and forced-ocean simulations. It is found that the difference in deep convection is primarily due to differences in large-scale atmospheric conditions that are not accounted for by the conventional NAO definition. Specifically, the 2007/08 winter was characterized by an atmospheric circulation anomaly centered in the western North Atlantic, rather than the eastern North Atlantic that the conventional NAO emphasizes. This anomalous circulation was also accompanied by anomalously cold conditions over northern North America.

The controlling influence of these atmospheric conditions on LS deep convection in the 2008 winter is confirmed by sensitivity experiments where surface forcing and/or initial conditions are modified. An extended analysis for the 1949–2009 period shows that about half of the winters with strong heat losses in the LS are associated with such a west-centered circulation anomaly and cold conditions over northern North America. These are found to be accompanied by La Niña–like conditions in the tropical Pacific, suggesting that the atmospheric response to La Niña may have a strong influence on LS deep convection.

1. Introduction

Located at the northwestern corner of the North Atlantic bounded by Greenland to the northeast and Labrador, Canada, to the southwest, the Labrador Sea

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Many observational and modeling studies link the strength of LS convection and properties of the LSW to the North Atlantic Oscillation (NAO; e.g., Yashayaev 2007; YD14). As the leading mode of atmospheric variability in the North Atlantic (Hurrell 1995), the NAO characterizes the in-phase variability between the Icelandic low and the Azores high, thus modulating the westerlies in the subpolar North Atlantic. The enhanced westerlies during a positive NAO phase carry cold and dry air from the Arctic/Canadian landmass to the LS, inducing substantial heat release from the ocean and thus prompting deep convection under favorable oceanic preconditioning (Marshall and Schott 1999).

Convective mixing had been weak in the LS since the mid-1990s with, on average, a neutral NAO phase (Våge et al. 2009, hereafter V09). However, data from Argo profiles (V09) and hydrographic surveys (Yashayaev and Loder 2009) show a sudden return of LS deep convection in the 2008 winter (here and in what follows, we refer to a given winter in terms of the year of the late winter months, January through March) with a moderately positive NAO state. In contrast, LS deep convection remained fairly shallow (V09) during the previous winter, despite the NAO being in a somewhat stronger positive state than in 2008. The disparity between the two winters raises questions about the connection between the NAO and LS convection.

The difference in deep convection may result from different oceanic preconditioning, essentially related to upper-ocean buoyancy characteristics imposed prior to the winter season. However, V09 report that preconditioning prior to these two winters was rather similar. Instead, the return of deep convection in the 2008 winter appears to be related to anomalously cold air temperatures and associated strong heat loss by the ocean (V09). V09 offer several possible explanations based on their analysis of observational data and reanalysis products. One particular explanation put forward is that the North Atlantic storm track was organized in the 2008 winter to more effectively bring cold air into the LS and trigger deep convection there. The possible significance of synoptic systems in driving such extremely high heat losses from the ocean and thus leading to deep convection in the LS is also discussed in Pickart et al. (2008).

In contrast, a recent modeling study by Holdsworth and Myers (2015) does not support this conjecture. Specifically, their study finds that the strength of simulated convective mixing in the 2008 winter is relatively insensitive to high-frequency atmospheric forcing—accounting for mesoscale to synoptic-scale systems—compared to other winters during their study period from 2002 to 2011. They explain this insensitivity by reasoning that the 2008 winter was already extremely cold—as determined by the mean air temperature—so that high-frequency forcing did not impact deep convection greatly because the mixed layer would have been already very deep just because of low-frequency forcing. Instead, high-frequency forcing appears to be more effective during relatively mild winters in their simulations.

In the present study, we revisit the question of what fundamentally caused the abrupt return of LS deep convection in the 2008 winter to further advance our mechanistic understanding of this event, reevaluating somewhat conflicting findings of the above studies. Specifically, we address this question by analyzing multiple reanalysis datasets and a forced-ocean (hindcast) simulation to examine the atmospheric and oceanic conditions in the 2008 winter, in comparison to those in the 2007 winter, as a case study. We then verify the results from the above analysis by performing additional sensitivity experiments where surface forcing and/or initial conditions are suitably modified to systematically investigate the change in deep convection between the two winters.

A major finding of our case study is that LS deep convection resumes in the 2008 winter under unusual atmospheric conditions in which the atmospheric circulation anomalies are centered in the western North Atlantic—in contrast to the conventional NAO whose circulation anomalies are centered in the eastern North Atlantic basin. As discussed in V09, these unusual atmospheric conditions are accompanied by a large-scale cooling across northern North America with La Niña conditions in the tropical Pacific. Following this result, we extend our investigation to the entire reanalysis period of 1949–2009 and explore the degree to which extreme heat flux events in the LS were associated with such atmospheric and tropical Pacific sea surface temperature (SST) conditions.

The paper is organized as follows. We introduce the reanalysis data, model, and numerical experiments employed in this study in section 2. A case study of the 2007 and 2008 winters using reanalysis data and numerical simulations is presented in section 3. Section 4 discusses the extended analysis for the 1949–2009 period. Summary and discussion are given in section 5.

2. Data and model simulations

a. Reanalysis data

To examine the atmospheric conditions, we mainly use the Coordinated Ocean-Ice Reference Experiments
phase 2 (CORE-II), interannually varying atmospheric dataset. This dataset is based on the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996) for surface state variables, which are adjusted using more reliable sources of data (e.g., satellite and in situ measurements) so that they can be used to force historical ocean–sea ice hindcast simulations (Large and Yeager 2009). The CORE-II dataset has been used extensively in forcing hindcast simulations whose results are analyzed in a series of publications (e.g., Danabasoglu et al. 2014; Griffies et al. 2014; Danabasoglu et al. 2016). We also employ the CORE-II dataset in this study to force the hindcast simulations described below.

One possible drawback of the use of the CORE-II dataset in the present study is that this dataset does not resolve mesoscale atmospheric phenomena, such as Polar lows and topographic jets, because of its low temporal and spatial resolutions (6-hourly and 2.5°, respectively), which have been suggested to be an important forcing for deep water formation in the subpolar North Atlantic, especially in the Irminger Sea (Pickart et al. 2003) and the Nordic seas (Condon and Renfrew 2013). To verify and supplement the CORE-II data, particularly over the LS region, we also consider the Arctic System Reanalysis product (ASR; Bromwich et al. 2016) in our analysis. ASR has been produced using the polar Weather Research and Forecasting (WRF) numerical weather prediction system, a version optimized for use in Polar regions (Bromwich et al. 2016). ASR has both a higher temporal (3-hourly) and spatial resolution (30km) than CORE-II, and thus is expected to better resolve Polar lows whose spatial and temporal scales are approximately 250 km and 1–2 days, respectively (Condon and Renfrew 2013). However, as shown below, the ASR and CORE-II data are very similar in the LS region for the fields of interest in the current study, indicating that the conclusions drawn using only the CORE-II data are reliable despite their low resolution. This in turn bolsters the fidelity of the numerical experiments performed here using the CORE-II dataset as forcing fields.

For both CORE-II and ASR, we analyze the surface fields that are directly related to turbulent (latent plus sensible) heat flux computations (i.e., near-surface atmospheric temperature, specific humidity, and zonal and meridional winds) as well as sea level pressure (SLP). The daily turbulent heat fluxes for CORE-II are computed with the bulk formulas given in Large and Yeager (2009) in conjunction with the observational sea surface temperature from Hurrell et al. (2008). To complement the CORE-II-derived heat flux, we also make use of the objectively analyzed air–sea fluxes dataset (OAFlux; Yu and Weller 2007). The daily wind speed is the daily average of the wind speed computed at every time step of the reanalysis data (i.e., every 6 and 3 h for CORE-II and ASR, respectively).

To quantify the intensity of synoptic storms, we choose to use the variance of 2–8-day bandpass-filtered meridional wind $\nu^2$ for each winter from December to March (DJFM). Tracking a cyclones’ pressure center and gathering bandpass-filtered SLP variance statistics are widely used approaches to depict surface storm tracks. We note, however, that the turbulent heat fluxes associated with synoptic storms are tightly related to storm’s meridional velocity, as it advects relatively cold (warm) air in the rear (forward) area of migrating cyclones (Zolina and Gulev 2003). Therefore, as far as the turbulent heat fluxes associated with storms are concerned, our choice here of $\nu^2$ might be the more relevant variable to be monitored.

Finally, we note that we also examined the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-I; Dee et al. 2011). The results from this product are found to be very similar to those from CORE-II and ASR and are not included in this paper.

b. Numerical experiments

The ocean model used in this study is the Parallel Ocean Program, version 2 (POP2; Smith et al. 2010), at a nominal 1° horizontal resolution with 60 vertical levels. It is coupled to the Los Alamos Sea Ice Model, version 4 (CICE4; Holland et al. 2012), which shares the same horizontal grid as POP2. Detailed description of the ocean model configuration and experimental design for the hindcast simulations can be found in Danabasoglu et al. (2012), YD14, and references therein.

The control hindcast simulation (CTRL) is identical to the one described in YD14, except that it is extended to the end of year 2009 to accommodate the needs of our present study. It is forced with the CORE-II dataset described above for the 1948–2009 period, following the CORE-II protocol (Griffies et al. 2012; Danabasoglu et al. 2014). CTRL shows good skill in reproducing observed variability in the subpolar North Atlantic (Yeager et al. 2012; YD14). Specifically relevant to the topic of interest here, CTRL captures the return of LS deep convection in the 2008 winter and reproduces well the observed decadal variability of LSW properties (YD14).

To demonstrate this agreement, we present in Fig. 1 the March-mean mixed layer depth (MLD) time series from CTRL averaged over a central LS region bounded by 56°–60°N and 56°–48°W in comparison with the observed MLD estimates from Gelderloos et al. (2013).
We note that this central LS region is intended to match the region with available observations and differs slightly from the LS regions used in this study (see below). Gelderloos et al. (2013) consider several sources of direct MLD observations for the 1993–2009 period and categorize LS MLDs into shallow (<1000 m), intermediate (between 1000 and 1500 m), or deep (>1500 m) regimes as well as two intervening regimes when observed MLD is within 50 m of the transition values between these three major regimes. To extend the observational MLD estimates back in time, we apply the same regime definitions to directly measured MLDs at the Ocean Weather Station Bravo located in the central LS for the 1964–74 period (56°8′ N, 51°8′ W; Gelderloos et al. 2012).

We note that some mismatches between the modeled and observed MLDs are expected because differing MLD definitions used: While we use a buoyancy gradient criterion as described in Large et al. (1997) to determine modeled MLDs, the definitions of MLD for the observational data used in Gelderloos et al. (2013) vary but are essentially based on potential density profiles. Figure 1 shows good qualitative agreement between the modeled and observed MLDs. Specifically, for the later period, CTRL reproduces the observed regime shift from a deep convection phase in the early-to-mid 1990s to a shallower intermediate phase during the late 1990s and mid-2000s, followed by the resumed deep convection in 2008. In addition, CTRL successfully simulates the observed abrupt return of deep convection in the 1972–74 winters from suppressed convection for the 1969–71 winters. These agreements give us confidence that the hindcast simulations can indeed be used to explore the origins of the deep convection event in the 2008 winter.

We augment CTRL by several sensitivity experiments, summarized in Table 1, to identify both the dominant contributors from among the various atmospheric forcing fields and the role of oceanic preconditioning to the 2008 deep convection event. The atmospheric forcing impact is decomposed in terms of flux components and frequency band (i.e., synoptic vs longer frequency). We isolate the most important processes by integrating the model with various combinations of atmospheric forcing variables and initial conditions of the 2007 and 2008 winters. We note that the purpose of the sensitivity experiments in which different forcing variables from the two winters are combined is to heuristically identify the most important atmospheric variable responsible for the deep convection in the 2008 winter, and not to rigorously quantify their relative contributions. The details of the experimental setups for these sensitivity experiments are given in section 3c, together with results from these experiments.

3. A case study of the 2007 and 2008 winters

a. Atmospheric conditions

We first show, in Fig. 2, the daily time series of the CORE-II-derived and OAFlux turbulent heat fluxes, and near-surface (10 m) air temperature (SAT), zonal and meridional winds, and wind speed from CORE-II and ASR (except for zonal and meridional winds) for the winters of 2007 and 2008. All time series are averages over the central LS region defined by 56°–62° N and 59°–46° W (boxed region in Fig. 3b). The mean, variance, and correlation values discussed below are based on the CORE-II-derived data, but similar values are obtained for other products.

The turbulent heat fluxes (positive upward; i.e., heat losses from the ocean) in the LS agree remarkably well between the CORE-II-derived product and OAFlux. As discussed earlier, greater winter-mean (December through March) heat release from the ocean in the 2008 winter than in the 2007 winter (241 vs 173 W m⁻²) is accompanied by much colder 2008 winter-mean SAT (–6.0° vs –1.2°C). The daily variability of these fluxes is primarily dictated also by SAT in both winters with turbulent heat flux–SAT correlation coefficients of −0.87 and −0.74 for the winters of 2007 and 2008, respectively (>99% confidence level for both winters). If the colder average SAT in the 2008 winter is due to a direct influence of storms, one would expect a stronger wind variance in this winter compared to that of 2007. However, $u^2$ in the 2008 winter is only about one-half of that in the 2007 winter (6.4
Table 1. Summary of sensitivity experiments. The first column shows the experiment names. Columns 2 through 4 list the forcing years of the winds $U$, air temperature $T$, and specific humidity $Q$ (all at 10 m), respectively. The last column indicates the month and year of the initial conditions used in the experiments. The initial conditions are obtained from CTRL. The experiment names reflect the year of the initial conditions (IC), main forcing $F$, and perturbed forcing, i.e., $U$, $T$, and/or $Q$. Note that the year of forcing includes November and December of the previous year. For example, 7IC-7F8U denotes an experiment initialized in November 2007 and forced with 2008 winds and 2007 air temperature and specific humidity data. LF and HF indicate low- and high-frequency parts of forcing (see text for details). We note that $T_{10m}$ is the same as SAT.

<table>
<thead>
<tr>
<th>Expt name</th>
<th>$U_{10m}$</th>
<th>$T_{10m}$</th>
<th>$Q_{10m}$</th>
<th>IC</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL (7IC-8F)</td>
<td>2008</td>
<td>2008</td>
<td>2008</td>
<td>November 2007</td>
</tr>
<tr>
<td>7IC-7F</td>
<td>2007</td>
<td>2007</td>
<td>2007</td>
<td>November 2007</td>
</tr>
<tr>
<td>7IC-7F8U</td>
<td>2008</td>
<td>2008</td>
<td>2008</td>
<td>November 2007</td>
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<tr>
<td>7IC-7F8T</td>
<td>2007</td>
<td>2008</td>
<td>2007</td>
<td>November 2007</td>
</tr>
<tr>
<td>7IC-7F8TQ</td>
<td>2007</td>
<td>2008</td>
<td>2008</td>
<td>November 2007</td>
</tr>
</tbody>
</table>

vs 13.4 m$^2$s$^{-2}$). Also, both the mean and variance of the wind speed, which is the variable actually used for the turbulent heat flux computation, are weaker or similar in the 2008 winter compared to those of the 2007 winter. Therefore, the colder SAT in the LS in the 2008 winter appears to be accompanied by reduced, not increased, storm activity.

The 2008 winter is characterized by an overall northwesterly wind regime in the LS that supplies cold, dry air to the convection regions (Fig. 2f). This predominant northwesterly wind is interrupted by episodic changes to easterly and southeasterly winds, which coincide with days with warmer SAT (e.g., day 62 and days 111–114; Figs. 2d,f). On the other hand, in the 2007 winter, there is a period dominated by easterlies lasting longer than a month (roughly from day 50 to 90), which coincides with anomalously warm conditions (Figs. 2c,e). Although it appears that storms can cool the SAT even further and thus substantially enhance surface heat loss (e.g., day 53 in the 2008 winter), they can also bring warm air into the LS region and cause anomalous heat gain by the ocean in the middle of the winter (e.g., days 62 in the same winter). Furthermore, the most extreme wind events in the 2007 winter (days 52–53 and day 66 in Fig. 2g) occur during the anomalously warm period discussed above and barely affect the surface heat fluxes. These results lead us to conclude that the higher heat flux in the 2008 winter is associated with a change in persistent, large-scale atmospheric conditions, not with a change in storm activity. We note that the time evolution of specific humidity follows that of SAT very closely; as will be shown later, it contributes to the return of deep convection in the 2008 winter through the latent heat flux.

The daily time series from ASR, including the zonal and meridional winds (not shown in Fig. 2 for clarity), are in remarkable agreement with those from CORE-II. To make sure that these agreements between the CORE-II and ASR datasets are not an artifact of spatial averaging, we also perform gridpoint comparisons between the two considering nearest locations, confirming the excellent agreement between the CORE-II and ASR datasets (not shown).

Why are the large-scale atmospheric conditions so different between these two winters, despite similar, positive NAO states? Figure 3 shows the spatial patterns of the DJFM SLP, SAT, and $v^2$ anomalies in these two winters relative to the 2000–07 mean from CORE-II. In the 2007 winter, the anomalous SLP pattern is similar to the conventional positive NAO pattern (e.g., Fig. 1 of Hurrell 1995). However, compared to the conventional positive NAO pattern, the Icelandic low extends farther southwestward into the subpolar North Atlantic, thus weakening the meridional pressure gradient across the LS (Fig. 3a). In contrast, in the 2008 winter, the Azores high is
shifted farther west to the western North Atlantic (−40°N, 40°W) and stretched meridionally, creating a strong pressure gradient across the LS together with the Icelandic low (Fig. 3b).

The anomalous SLP pattern of the 2008 winter is accompanied by anomalously cold conditions over a large part of northern North America and subpolar North Atlantic, as pointed out by V09, with an intense cold anomaly over the LS (Fig. 3d). Therefore, the persistent northwesterlies in this winter (Fig. 2f) can effectively bring the cold SAT from the northern North America into the LS. In contrast, the SAT is, in general, anomalously warm over the LS and surrounding areas in the 2007 winter (Fig. 3c).

The $\nu^2$ anomalies indicate that the storm tracks are better defined in the 2008 winter than in the 2007 winter with an enhanced $\nu^2$ along the path across the North Atlantic from the east coast of the United States to the British Isles (Fig. 3f), again consistent with the results of V09. However, this intensification is accompanied by a weaker $\nu^2$ in a large area of the western subpolar North Atlantic, centered in the LS. Although not explicitly discussed in V09, this is also consistent with their result showing reduced cyclone frequency north of 55°N in the western subpolar North Atlantic in 2008 (see their Fig. 5).

A consistent spatial pattern of $\nu^2$ anomalies is also found in ASR (Fig. 4) although the mean $\nu^2$ is somewhat weaker than in CORE-II [note that the base period of ASR is one year shorter (2001–07) than that of CORE-II because ASR is not available for 2000]. With its higher resolution, the 2008 storm tracks appear to be more

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**Fig. 2.** Daily time series of (a),(b) turbulent heat fluxes (latent plus sensible), (c),(d) 10-m air temperature (SAT), (e),(f) zonal (black) and meridional (gray) winds, and (g),(h) wind speed for the (left) 2007 and (right) 2008 winters (December–March), averaged over the central LS (56°–62°N, 59°–46°W) from CORE-II. Also shown are the turbulent heat fluxes from OAFlux in (a),(b), and SAT and wind speed from ASR in (c),(d) and (g),(h), respectively. The turbulent heat flux is positive upward (heat release from the ocean). Positive zonal (meridional) winds denote westerlies (southerlies). The climatological cycle (1949–2009) of the CORE-II turbulent heat flux in (a),(b) and SAT in (c),(d) is also displayed (dashed line).
enhanced in ASR. The negative $v^2$ anomalies in the 2008 winter over the LS are clearly evident also in ASR, confirming that storm activity is weaker there than in the 2007 winter.

b. Oceanic conditions

In the interior LS, lateral heat and freshwater fluxes from relatively warmer and fresher boundary currents increase upper ocean buoyancy largely during warm seasons (Straneo 2006). For an onset of deep convection, this stratification must be eroded by surface buoyancy fluxes during winter, which are dominated by turbulent heat fluxes in the LS (Marshall and Schott 1999). Therefore, the upper ocean buoyancy (preconditioning) needs to be considered, together with surface heat flux, in order to explain the origins of the deep convection event in the 2008 winter. Figure 5 shows the upper ocean buoyancy anomalies prior to the 2007 and 2008 winters as simulated in CTRL. These upper ocean buoyancy anomalies before the convection season are calculated as vertically integrated November-mean density differences relative to the density at 700-m depth (Bailey et al. 2005), and the anomalies are relative to the 1999–2006 November mean. Also shown in Fig. 5 are the DJFM turbulent heat flux and February–April (FMA)-mean MLD anomalies for the winters of 2007 and 2008 in CTRL.

The simulated upper ocean buoyancy in the western subpolar North Atlantic shifts rapidly from an anomalously positive to negative state between the two years, most notably near the eastern end of the LS (Figs. 5a,b). However, in the western to southern LS, the buoyancy changes between the two winters are relatively small, in
line with the findings of V09 from observations. In contrast, while similar to the mean turbulent heat flux of the base period in the 2007 winter (Fig. 5c), the simulated turbulent heat flux in the 2008 winter is substantially stronger over the LS (Fig. 5d), especially in the western LS, consistent with the results based on re-analysis data (Fig. 2). Therefore, it is expected that in the western LS, the surface heat flux is the dominant factor over preconditioning of convective mixing in this winter.

The MLD anomaly pattern indeed shows that the deepening of MLD in the 2008 winter occurs most intensively in the southwestern LS (up to 700 m) with a secondary maximum in the western LS (Fig. 5f). The positive MLD anomalies also extend from the southwestern LS, along a band, into the Irminger Sea where the upper ocean buoyancy is anomalously weak, forming a check-shaped MLD anomaly pattern (Fig. 5f). This spatial pattern and magnitudes of the simulated MLD anomalies in the 2008 winter are largely consistent with those in observations shown in V09 based on the Argo data (see their Fig. 2), underscoring the fidelity of the simulated variability in the western North Atlantic in CTRL. The return of deep convection in the 2008 winter in CTRL is also accompanied by an advance of the sea ice edge (Fig. 5f), again in agreement with observations (V09). In contrast, MLD and sea ice distributions in the 2007 winter do not deviate much from the 2000–07 mean (Fig. 5e).

c. Sensitivity experiments

We perform a set of sensitivity experiments (Table 1) to verify that the turbulent heat fluxes associated with persistently colder and drier air conditions in the 2008 winter are indeed primarily responsible for the return of deep convection in the winter of 2008. Figure 6 shows the FMA-mean MLD anomalies from the sensitivity experiments relative to the 2000–07 FMA mean in CTRL, which can be directly compared to the 2008 MLD anomaly in CTRL (Fig. 6a). The first sensitivity experiment (7IC-7F; Fig. 6b) starts from the initial conditions of 1 November 2007, as in CTRL, but is forced with the atmospheric state of the previous year (i.e., November 2006 through October 2007) for those fields used in the turbulent heat flux computations (i.e., winds, SAT, and specific humidity). If oceanic preconditioning were the critical factor in the 2008 deep convection, then this experiment would largely recover the 2008 MLD deepening seen in CTRL. However, the strongest MLD deepening occurring in the western to southern LS of CTRL is largely absent in this experiment. The MLD deepening in the eastern LS coincides with the weaker buoyancy region (Fig. 5b), suggesting only a limited role of oceanic preconditioning in the return of deep convection in the 2008 winter.

To complement this experiment, a reverse experiment is also performed: that is, starting from the initial condition of 1 November 2006 and forced with the atmospheric state of 2008 (i.e., November 2007 through October 2008; 6IC-8F, Fig. 6c). Even in the presence of a positive upper-ocean buoyancy anomaly in the LS (Fig. 5a), 2008 forcing is strong enough to reproduce, to some extent, the MLD deepening of CTRL, confirming that contemporaneous atmospheric forcing, and not ocean preconditioning, was primarily responsible for the return of simulated deep convection in the 2008 winter.

In the next set of experiments, we investigate which atmospheric fields are playing a prominent role in the creation of the 2008 MLD anomaly. Starting with the 7IC-7F setup, one or more atmospheric variables are returned to their original 2008 values used in CTRL in a systematic way. When the wind fields are returned to the 2008 values (7IC-7F8U; Fig. 6d), the MLD difference is only marginally different from that in 7IC-7F, but there is an advance of the LS sea ice edge and likely consequent MLD shoaling seen in CTRL. Therefore, while confirming our earlier finding that winds were not a key element in giving rise to the deep convection event in the 2008 winter, this result suggests that
the enhanced westerlies may have played a role in the advance of the sea ice edge during this winter.

When 7IC-7F is altered by using 2008 values only for SAT (7IC-7F8T; Fig. 6e), the MLD anomaly of CTRL is largely reproduced, although the intense MLD deepening in the southwestern corner is still lacking. When specific humidity is also returned to its 2008 values in addition to SAT (7IC-7F8TQ; Fig. 6f), the MLD anomaly resembles more closely to that of CTRL, with somewhat excessive deepening in the southeastern LS. The large MLD differences between 7IC-7F8U and 7IC-7F8T (or 7IC-7F8TQ) confirm that the cold and dry air conditions identified in Fig. 3 were the key atmospheric changes that explain the MLD anomaly in the 2008 winter.

Finally, we explore the impacts of synoptic storms on the MLD deepening in the 2008 winter by decomposing the surface forcing fields $F$ into their low- and high-frequency components using a 30-day low-pass filter $F_L$ and taking the residual $F_H = F - F_L$, respectively. We use this fairly low cutoff frequency to ensure the suppression of all the synoptic variability. We repeat experiment 7IC-7F, except that either the low- (7IC-7F8F_L; Fig. 6g) or high-frequency forcing (7IC-7F8F_H; Fig. 6h) is returned to the 2008 values, while the other frequency component is retained as in their 2007 values. The simulated MLD anomalies in these two sensitivity experiments are strikingly distinct: the return of deep convection in the 2008 winter is fully evident only in

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**Fig. 5.** Anomalies of (a),(b) November-mean vertically integrated density differences with respect to 700 m (buoyancy; kg m$^{-2}$), (c),(d) DJFM turbulent heat flux ($Q_{he}$; W m$^{-2}$), and (e),(f) FMA MLD (m) in (left) 2007 and (right) 2008 as simulated in CTRL. For November buoyancy, years 2006 and 2007 are used. The anomalies are relative to the respective 2000–07 means (1999–2006 for buoyancy), which are shown as contours in (a),(c), and (e) with intervals of 50 kg m$^{-2}$, 80 W m$^{-2}$, and 300 m, respectively. Also displayed in (f) is the FMA sea ice extent, defined as the 15% sea ice concentration line, for the 2008 winter (magenta) and for the 2000–07 mean (black). The sea ice extent in the 2007 winter is almost identical to that of the base period, and is not displayed for clarity.
7IC-7FH8FL. These series of sensitivity experiments clearly demonstrate that the origin of the LS deep convection in the 2008 winter is the large-scale, low-frequency atmospheric forcing that results in persistent cold and dry air conditions over the LS.

4. Generalization to the winters of the 1949–2009 period

One major finding of our analysis thus far is that the atmospheric conditions between the winters of 2007 and 2008 are quite different despite similar NAO states. Furthermore, the atmospheric conditions only in the 2008 winter are optimal for deep convection in the LS. Given these findings, we now explore whether such atmospheric conditions that are associated with elevated turbulent heat fluxes in the LS exist in other years over the 1949–2009 period spanned by the CORE-II dataset. Figure 7a shows the normalized time series of the DJFM CORE-II-derived turbulent heat fluxes, averaged over the LS (52°–62°N, 58°–44°W), and the station-based NAO index (Hurrell 1995) for the 1949–2009 period. We note that the LS domain used here is slightly wider than one used for Fig. 2 in order to capture the interannual movement of the elevated turbulent heat flux region around the LS. As reported in many previous studies (e.g., Pickart et al. 2002), the turbulent heat fluxes in the LS are highly correlated with the NAO index ($r = 0.76$, > 99% confidence level). However, it is also found that the turbulent heat flux time series deviates substantially from the NAO index in some winters. For example, the 1972 winter is marked by the
second strongest turbulent heat flux during the period considered here after the 1989 winter, but the NAO in the 1972 winter is almost neutral.

Consistent with the case study discussed in the previous section, the interannual variability of the turbulent heat fluxes in the LS is mostly controlled by that of the SAT. Figure 7b shows the normalized time series of the DJFM SAT, 10-m zonal wind, and $\nu^2$ from CORE-II, averaged over the same LS region, along with the turbulent heat flux time series from Fig. 7a. The interannual variability of the turbulent heat fluxes is most highly correlated with that of the SAT ($r = -0.90$, >99% confidence level). The turbulent heat fluxes are also moderately correlated with the zonal wind ($r = 0.43$, >99% confidence level), but not with the wind speed ($r = 0.17$, <95% confidence level; not shown). That is, the winds only influence the interannual variability of the turbulent heat fluxes in the LS through the zonal wind variability that modulates the advection of cold SATs into the LS region (the correlation coefficient between the zonal wind and SAT is $-0.45$, >99% confidence level), but the winds themselves do not directly drive the turbulent heat flux variability.

That is, the winds only influence the interannual variability of the turbulent heat fluxes in the LS through the zonal wind variability that modulates the advection of cold SATs into the LS region (the correlation coefficient between the zonal wind and SAT is $-0.45$, >99% confidence level), but the winds themselves do not directly drive the turbulent heat flux variability. This mechanism (i.e., zonal wind–driven SAT control on the turbulent heat fluxes) is consistent with that operating on the intra-seasonal time scales found in the case study of 2007 and 2008 winters.

Surprisingly, the correlation of either $\nu^2$ ($r = 0.23$, <95% confidence level) or the variance of same bandpass-filtered wind speed ($r = -0.02$, <95% confidence level; not shown) with the turbulent heat fluxes is very weak. Namely, there seems to be no direct relationship between synoptic wind variability and turbulent heat fluxes in the LS. Given the dominance of the SAT on the interannual variability of the turbulent heat fluxes, in the following we focus on the SLP and SAT patterns associated with years of strong heat losses in the LS.

Figures 8a and 8b compare the DJFM SLP and SAT anomaly composites constructed using the years with standard deviation (SD) greater than one (SD > 1) for the normalized NAO index (10 winters) and LS turbulent heat flux (11 winters) time series of Fig. 7. The NAO-based composites (Fig. 8a) show the well-known meridional dipole structure in SLP, centered in the eastern North Atlantic, with the zonal dipole structure of SAT (cold air over the LS and warm air over Europe). The SAT over northern North America is generally anomalously warm during a positive NAO phase. The composites based on the LS turbulent heat fluxes (Fig. 8b) are not very different from the NAO composites, but show a westward shift of both the low and high pressure centers, accompanied by even colder SAT over the LS, but close to neutral SAT over northern North America.

Figures 8c show the conditional composites where the LS turbulent heat flux SD > 1, but the NAO SD < 0.7. This is done to isolate years of strong heat flux that occur when NAO conditions are not extreme. We chose this NAO threshold to maximize the number of winters in the composite while minimizing the NAO influence. This threshold is just low enough to exclude the 2008 winter. These composites (Fig. 8c) clearly demonstrate why the pressure centers shift to the west and the SAT is colder in the LS turbulent heat flux–based composites. An isolated high pressure center emerges in the western North Atlantic with a weaker low pressure center over Greenland, accompanied by colder SAT casting over northern North America and the subpolar North Atlantic, when the NAO influence is minimized. This setting is very similar to that of the 2008 winter (Fig. 3b). This conditional composite includes 4 winters (1972, 1984, 1991, and 1997), and including the 2008 winter, they account for 5 of the 11 winters with the LS turbulent heat flux SD > 1. In other words, in roughly one-half of the
winters with extreme turbulent heat flux forcing in the LS, the atmospheric circulation anomalies are centered in the western North Atlantic and the SAT is colder over northern North America, in contrast to the conditions associated with the strongly positive conventional NAO. As discussed in V09, the cold conditions over northern North America in the 2008 winter coincided with a La Niña event in the tropical Pacific. Figure 9 shows the November–January mean SST and DJFM SLP composites for the Northern Hemisphere and the southern tropics obtained using the same criteria as in Fig. 8. The conditional SST composite (strong LS heat flux, but weak NAO) indeed reveals a La Niña–like condition in the tropical Pacific (Fig. 9c), in sharp contrast to the strong NAO composite showing an El Niño–like condition. Associated with the cold SST anomaly in the eastern tropical Pacific, the conditional composite shows a wavenumber-3 SLP pattern in the Northern Hemisphere extratropics, including the ridge centered in the western North Atlantic. An almost identical SLP
response is found by Zhang et al. (2015) in atmospheric general circulation model simulations where a La Niña SST condition is imposed as boundary conditions with a climatological SST elsewhere. Therefore, the conditional composites suggest that both the cold SAT in northern North America and the SLP pattern in the western North Atlantic are the ramifications of remote influence of La Niña events.

The SST composite for the strong positive NAO (Fig. 9a) may differ from some previous studies that suggest a relationship between El Niño and negative NAO phase (e.g., Li and Lau 2012). However, this spatial pattern, including warm SST anomalies in the tropical Indian Ocean, is consistent with the findings of Hoerling et al. (2001), who emphasize that the positive trend in the NAO during the 1980s and early 1990s is related to a warming in the tropical Indian and Pacific Oceans. We note that 7 out of the 10 years included in the strong positive NAO composite indeed are found during the 1980s through the mid-1990s period. It is interesting to note that SST anomalies in the tropical Pacific are almost absent in the composite for the strong heat flux years (Fig. 9b), possibly caused by the compensation of the above two patterns.
The results of the above composite analysis suggest that the atmospheric variability over the LS can be better represented if an index is constructed based on the SLP anomalies in the western subpolar North Atlantic, rather than those in the eastern side (i.e., the conventional NAO). We generate such an index by computing the pressure difference between the region around Nova Scotia–Newfoundland and southern Greenland, indicated by the boxes in the middle-left panel of Fig. 8, and display it in Fig. 7. We refer to this new index as the western North Atlantic (WNA) index.

Table 2 compares the correlations of the conventional NAO and the WNA index with the CORE-II-derived turbulent heat flux (Fig. 7) as well as the LS MLD (Fig. 1) and AMOC strength at 45°N from CTRL [for the AMOC time series, see Fig. 2 of Danabasoglu et al. (2016)]. Although these time series are already quite highly correlated with the conventional NAO index (all correlations are significant at >99% confidence level), the correlations become even higher, explaining 10%–25% more variance with the WNA index. In particular, while the conventional NAO explains slightly more than one-half of the turbulent heat flux variance in the LS (58%), the WNA index explains considerably more LS turbulent heat flux variance (83%), as one can easily deduce from Fig. 7. Similarly, the explained variance of the simulated LS MLD (AMOC at 45°N) increases from 48% (27%) to 62% (37%). This stresses the importance of atmospheric conditions in the western North Atlantic—which appear to be remotely influenced by SST conditions in the tropical Pacific—in understanding temporal variations of air–sea interactions over the LS region, which in turn impact large-scale ocean circulations over the North Atlantic.

### 5. Summary and discussion

Motivated by the study of V09, we have systematically examined the atmospheric and oceanic conditions that led to the abrupt resumption of LS deep convection in the winter of 2008. This event is particularly intriguing because the NAO index was higher in the previous winter with no enhanced deep convection. By analyzing multiple reanalysis products, we find that the 2008 winter is characterized by large-scale atmospheric circulation anomalies centered in the western North Atlantic, in contrast to the conventional NAO pattern centered in the eastern North Atlantic. These western North Atlantic circulation anomalies are accompanied by anomalously cold SATs over a large part of northern North America and the subpolar North Atlantic. Furthermore, they set a strong pressure gradient over the LS and, with the associated northwestlies, appear to effectively and persistently bring the cold and dry air from the Arctic/Canadian landmass into the LS throughout the 2008 winter.

A better-defined North Atlantic storm track in the 2008 winter than in the previous winter, as identified in V09, is also found in the present study (Figs. 3e,f and 4a,b). However, over the LS, storm activity appears to be weaker in 2008 than in 2007. This would likely rule out synoptic storm forcing as the explanation for the return of LS convection in the 2008 winter. The extremely cold conditions over the LS in the 2008 winter reported in V09 thus appear to result from the large-scale anomalous atmospheric conditions described above, which are likely the reason behind the return of LS deep convection in this winter.

The results from the reanalysis products are further supported by a forced ocean hindcast simulation (CTRL) and a series of sensitivity experiments, in which surface forcing and/or oceanic initial conditions are suitably modified in order to find the primary drivers for the return of deep convection in the 2008 winter. The results from these experiments indicate that oceanic preconditioning prior to the 2008 winter is, to some extent, more favorable for deep convection (weaker buoyancy in the upper ocean), but it appears that its effect is secondary, consistent with the conclusions of V09 and Yashayaev and Loder (2009).

We extend our analysis to the 1949–2009 period to further investigate the atmospheric conditions in years with strong wintertime surface heat losses in the LS (>1 SD above the mean). A composite analysis shows that about one-half of the strong turbulent heat flux winters in the LS (5 out of 11, including the 2008 winter) reveal anomalous atmospheric conditions that closely resemble those of the 2008 winter (Fig. 8c). This indicates that such atmospheric conditions are not uncommon and may play an important role in explaining the development of extreme North Atlantic buoyancy forcing.

Consistent with the case study of the 2007 and 2008 winters, we find very low correlations of the turbulent heat fluxes with synoptic wind variability over the LS for the

### Table 2. Correlations of the CORE-II-derived turbulent heat fluxes in the LS, LS MLD, and AMOC at 45°N from CTRL with the DJFM NAO index and the new index defined in the WNA. The correlations for AMOC are for when the indices lead AMOC by 1 yr. All correlations are statistically significant at a 99% confidence level.

<table>
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<tr>
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<th>NAO index</th>
<th>WNA index</th>
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<tbody>
<tr>
<td>LS turbulent heat flux</td>
<td>0.76</td>
<td>0.91</td>
</tr>
<tr>
<td>LS MLD</td>
<td>0.69</td>
<td>0.79</td>
</tr>
<tr>
<td>AMOC at 45°N (lag +1)</td>
<td>0.52</td>
<td>0.61</td>
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</table>
1949–2008 period (Fig. 7b). This finding indicates a weak link between the interannual variability of the heat fluxes and storm activity over the LS, at least based on the measure of storm activity considered in this study (i.e., $v^2$).

Our findings identify the west-centered atmospheric conditions in the North Atlantic as an important factor for influencing air–sea interactions in the LS, which are believed to be vital forcings for the decadal variability of both the North Atlantic subpolar gyre and AMOC. We show that using an index based on SLP anomalies in the western North Atlantic (i.e., WNA) results in substantial improvements in explaining the variance of the turbulent heat fluxes in the LS, compared to using the conventional NAO index. The use of WNA index also leads to better correlations with the simulated LS MLD and AMOC variability.

We further present evidence that both the above anomalous atmospheric circulation and cold SAT patterns are associated with La Niña events in the tropical Pacific. While the cold SAT anomalies over the northern North America during La Niña winters are well known (e.g., Halpert and Ropelewski 1992), there is no firm consensus on the circulation response over the North Atlantic, largely because of nonlinear effects in the responses of the surface pressure field to varying magnitudes and sign of the tropical Pacific SST anomalies (Greatbatch et al. 2004; Toniazzo and Scaife 2006). The short record of observations and thus a small number of events that we identify in this study preclude us from firmly establishing a link between La Niña events and atmospheric conditions that favor strong heat release, and thus deep convection in the LS. Nevertheless, we consider the results presented here to be an important step toward improving our understanding of underlying dynamics controlling LS deep convection and its connection to AMOC.

There are a few possible caveats with our numerical experiments. First, as in many other studies using ocean-only models, our simulations are exposed to issues arising from the lack of air–sea interactions, which are extensively discussed in Griffies et al. (2009). However, it is difficult to quantify the impacts of such missing air–sea coupling on our results using ocean-only modeling approach, and we hope to assess these effects by using a coupled atmosphere–ocean model in a future study.

A second possible caveat is that we combine the forcing variables from the two winters in the sensitivity experiments in order to identify the dominant variable responsible for the 2008 convection event. One may question this approach because the forcing fields are intimately connected (e.g., strong westerlies are linked to cold SAT in the LS) and thus mixing one variable from one year with other variables from another year is unphysical. However, if the primary focus is to answer the question of which variable is the most dominant factor in controlling strong surface heat losses, such a forcing separation approach represents a very powerful numerical framework that can provide a qualitative answer. This is especially the case of our study: the differences between the sensitivity experiments (i.e., 7IC-7F8U and 7IC-7F8T in Fig. 6) are large to provide a qualitative assessment of the relative roles of surface forcing fields. We note that similar decompositions of surface datasets in forcing ocean models have been successfully used in many previous modeling studies (e.g., Biastoch et al. 2008; Robson et al. 2012; YD14).

The third possible caveat is related to the use of a coarse-resolution ocean model in which subgrid-scale processes, likely important for the simulation of deep convection events in the LS, are not resolved, but parameterized. For example, mesoscale eddies shedding at the tip of Greenland are known to carry relatively warm waters into the LS, playing an important role for restratification in the interior LS (Lilly et al. 2003). The absence of explicit representation of this process may be partly responsible for too-deep MLD in the northern LS, west of southern Greenland, compared to limited observational estimates. However, as illustrated in Fig. 1, the interannual variability of simulated LS MLD agrees reasonably well with observational estimates, and seems to be less affected by such missing processes. This is likely because the interannual MLD variability is more dominantly controlled by buoyancy fluxes through the surface than those through the lateral boundaries.

Last, another potential issue related to the model’s resolution is that such a resolution may be too coarse to represent full effects of high-frequency forcing in our sensitivity experiments. However, as discussed in section 1, Holdsworth and Myers (2015) find only minor MLD changes in the 2008 winter when they suppress high-frequency forcing in their high-resolution (0.25°) simulations. In addition, Luo et al. (2014) show that the variability of convective mixing in the LS is insensitive to whether monthly mean or high-frequency (6 hourly) forcing fields are used in their high-resolution (7.5 km) regional model simulations. Therefore, it appears that the resolution is not a crucial factor to capture the MLD response to the high-frequency forcing changes.

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