A Lagrangian Climatology of Wintertime Cold Air Outbreaks in the Irminger and Nordic Seas and Their Role in Shaping Air–Sea Heat Fluxes

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ABSTRACT

Understanding the climatological characteristics of marine cold air outbreaks (CAOs) is of critical importance to constrain the processes determining the heat flux forcing of the high-latitude oceans. In this study, a comprehensive multidecadal climatology of wintertime CAO air masses is presented for the Irminger Sea and Nordic seas. To investigate the origin, transport pathways, and thermodynamic evolution of CAO air masses, a novel methodology based on kinematic trajectories is introduced.

The major conclusions are as follows: (i) The most intense CAOs occur as a result of Arctic outflows following Greenland's eastern coast from the Fram Strait southward and west of Novaya Zemlya. Weak CAOs also originate in flow across the SST gradient associated with the Arctic Front separating the Greenland and Iceland Seas from the Norwegian Sea. A substantial fraction of Irminger CAO air masses originate in the Canadian Arctic and overflow southern Greenland. (ii) CAOs account for 60%–80% of the wintertime oceanic heat loss associated with few intense CAOs west of Svalbard and in the Greenland, Iceland, and Barents Seas and frequent weak CAOs in the Norwegian and Irminger Seas. (iii) The amount of sensible heat extracted by CAO air masses is set by their intensity and their pathway over the underlying SST distribution, whereas the amount of latent heat is additionally capped by the SST. (iv) Among all CAO air masses, those in the Greenland and Iceland Seas extract the most sensible heat from the ocean and undergo the most intense diabatic warming. Irminger Sea CAO air masses experience only moderate diabatic warming but pick up more moisture than the other CAO air masses.

1. Introduction

Marine cold air outbreaks (CAOs) are cold surges occurring in association with the intermittent discharge of cold polar air masses into more temperate latitudes, where they reach comparatively warm waters and are accompanied by large air–sea heat exchanges. There are two main paths for these polar air masses in the Northern Hemisphere (Iwasaki et al. 2014). The first one is directed from the Siberian Arctic to East Asia with dissipation over the Kuroshio, and the second is from the interior

Arctic via the North American continent into the North Atlantic Ocean with dissipation over the Gulf Stream and the North Atlantic Current. These cold air mass pathways are important ingredients for the formation of hot spots with intense air–sea interactions along the warm western boundary currents (Iwasaki et al. 2014). Nevertheless, it is the air–sea heat exchanges associated with CAOs in the subpolar North Atlantic and the Nordic seas, encompassing the Greenland, Iceland, and Norwegian Seas, that play a key role in the formation of the dense waters feeding into the lower limb of the Atlantic meridional overturning circulation (e.g., Dickson et al. 1996; Talley 1996; Gebbie and Huybers 2010). Hence, CAOs in the Irminger Sea and Nordic seas are of great importance for the climate system, even though these regions do not lie among the main cold air mass routes in the Northern Hemisphere.

Surface sensible and latent heat fluxes in the Nordic seas can be in excess of 500 W m^{-2} according to measurements (Shapiro et al. 1987) and modeling studies (Wacker et al. 2005). One reason for such high fluxes are relatively warm waters brought northward by the

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FIG. 1. (a) Location map of the study domain and terrain height in ERA-Interim (shading). The domain from where CAO trajectories are started and subregions considered in this study are outlined by the solid black and colored boxes, respectively. The subregions are the Irminger Sea (IRM), Iceland Sea (ICL), Greenland Sea (GRE), Svalbard (SVA), Norwegian Sea (NOR), and eastern Barents Sea (BAE). (b) Mean sea surface temperature (shading) and mean sea ice boundary (50% sea ice concentration; gray contour) from ERA-Interim during winters (November–April) in the study period 1979–2014. Note that the sea surface temperature is masked white in regions where the mean sea ice concentration or fractional land cover exceeds 50%.

North Atlantic and the Norwegian Atlantic currents, the northeastward extensions of the Gulf Stream, leading to anomalously high sea surface temperatures (SSTs) along the Norwegian coast and as far north as Svalbard and the western Barents Sea (cf. Fig. 1b for location names). These warm, saline waters are gradually cooled by air–sea heat fluxes and freshened in the Norwegian Sea, as well as in the Arctic Ocean (Mauritzen 1996; Segtnan et al. 2011). Despite the freshening, the heat loss in the Nordic seas is sufficiently strong that some of these waters lose buoyancy and may sink to depth (Dickson et al. 1996; Spall and Pickart 2001) with the densest waters flowing across the submarine sill in the Denmark Strait located between Iceland and Greenland (Swift et al. 1980). The precise origin of these waters, however, has been a matter of ongoing debate throughout the past century owing to shifting views on the oceanic circulation (Nansen 1912; Swift and Aagaard 1981; Mauritzen 1996; Våge et al. 2013) but also a lack of understanding of the atmospheric processes setting the air–sea heat flux forcing, and thus of CAOs, north of the Denmark Strait (e.g., Våge et al. 2015; Harden et al. 2015).

Two specific examples serve to illustrate this lack of understanding. First, recent findings suggest that a substantial fraction of the Denmark Strait overflow waters are formed in the Iceland Sea gyre (Jónsson and Valdimarsson 2004; Våge et al. 2011, 2013; Harden et al. 2016). However, wintertime mean heat fluxes in the Nordic seas feature a local minimum in the Iceland Sea (Moore et al. 2012), although buoy measurements show that intermittent high heat flux events can occur

(Harden et al. 2015). Consequently, it remains unclear if the frequency and the intensity of high heat flux events are high enough to cool the ocean mixed layer by a sufficient amount to support the estimated water mass formation rates. Second, Moore et al. (2015) and Våge et al. (2015) showed a weakening of open ocean convection in the Iceland Sea due to a reduction in number and a northward shift of high heat flux events during the last two decades. They hypothesized that these changes in heat fluxes are the result of the unprecedented retreat of sea ice in the western Nordic seas (e.g., Macias Fauria et al. 2010) and an associated shift of the formation regions of CAOs. A corroboration of this hypothesis, however, requires an analysis of the spatiotemporal variability and transport pathways of CAO air masses.

Previous climatological studies of CAOs did not focus so much on the spatiotemporal variability. Instead, they were mainly motivated by the fact that CAOs are accompanied by a wealth of severe weather phenomena. These phenomena include extremely low temperatures, gale-force low-level winds, and heavy snowfall (e.g., Changnon 1979; Walsh et al. 2001), as well as polar lows (Rasmussen and Turner 2003; Claud et al. 2007; Terpstra et al. 2016). CAOs also play an important role for the maintenance of low-level baroclinicity along the North Atlantic storm track by extracting heat from the ocean (Papritz and Spengler 2015). Thus, knowledge of CAO occurrence helps to constrain where and when the aforementioned extreme weather systems can occur and gives insight into the dynamics of the storm track.

Accordingly, several diagnostic indices have been developed to quantify CAO occurrence. Many of these indices are based on the air–sea potential temperature difference, $\theta_{\text{SST}} - \theta_p$, where θ_{SST} denotes potential SST¹ and θ_p is potential temperature on a specific pressure level, typically chosen between 850 and 700 hPa (Bracegirdle and Gray 2008; Kolstad and Bracegirdle 2008; Kolstad 2011; Papritz et al. 2015; Fletcher et al. 2016). Positive values of the air–sea potential temperature difference are characteristic of a CAO with a convective boundary layer. Building on such indices, the incidence of CAOs in the northeastern North Atlantic has recently been quantified by Kolstad et al. (2009), Kolstad (2011), and Fletcher et al. (2016). These authors found that CAOs are most frequent in the Irminger Sea, as well as off the sea ice edge over the Fram Strait and the Norwegian and western Barents Seas. On the other hand, these studies also indicate that CAOs are rather rare in the Iceland Sea and have a surprisingly low occurrence in the Greenland Sea—a region known for particularly intense mean heat fluxes (e.g., Moore et al. 2012).

In contrast to the incidence of CAOs, little is known about the actual pathways and characteristics of the air masses involved. Moreover, the influences of the thermodynamic properties of CAO air masses on the spectrum of air–sea exchanges of sensible and latent heat have not been systematically investigated. In this study, we aim to elucidate these aspects by addressing the following questions:

- Where do CAO air masses originate and what are their predominant pathways?
- How do the thermodynamic properties of the CAO air masses evolve?
- What is the potential of a certain CAO air mass to extract energy from the ocean along its pathway, and does this potential vary regionally?

Previously developed CAO indices derived from the air–sea potential temperature difference are unable to address these fundamental questions, as they neither take the entire depth of the air masses into account nor consider their transport. A natural way to address these questions is to adopt a Lagrangian perspective. This can be achieved by calculating kinematic trajectories from CAO air masses identified with an Eulerian criterion such as the air–sea potential temperature difference—an approach that has previously been used to investigate air masses involved in CAOs over land in the United States and Europe (Walsh et al. 2001), as well as for marine

CAOs in the South Pacific (Papritz et al. 2015). Consequently, the approach of this study is to compile a multidecadal trajectory dataset sampling the air masses leading to CAOs in the Irminger Sea and Nordic seas including the adjacent Barents Sea and to address the aforementioned questions based on this dataset.

In section 2 we outline the methodology and describe the compilation of the trajectory dataset. We analyze the pathways of CAO air masses before and after CAO formation, as well as their thermodynamic characteristics and evolution in section 3. This is followed by a quantification of the contribution of CAOs to surface sensible and latent heat fluxes and an analysis of their potential to extract energy from the ocean in section 4 and concluding remarks in section 5.

2. Methodology

The European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) provides the basis for the calculation of kinematic trajectories and the analyses presented in this study. Fields are available at 6-hourly intervals and on 60 vertical levels. In the horizontal they have been interpolated from the original T255 spectral resolution onto a $1^\circ \times 1^\circ$ latitude–longitude grid. Note that sensible and latent heat fluxes are model forecasts from ERA-Interim, which we have averaged over 6-hourly windows centered on the respective synoptic time (i.e., 0000, 0600, 1200, and 1800 UTC). To avoid spinup problems, we use only forecast steps between hours 9 and 21. We consider extended Northern Hemisphere winters (November–April), which we will generally refer to as winter, in the period 1979/80 to 2013/14.

In the spirit of previous studies on CAOs (e.g., Kolstad and Bracegirdle 2008; Papritz et al. 2015; Fletcher et al. 2016), we identify CAOs with the air–sea potential temperature difference $\theta_{\text{SKT}} - \theta$, where θ denotes potential temperature of an air parcel and $\theta_{\text{SKT}} = \text{SKT}(p_0/p_s)^{R/c_p}$ is potential skin temperature (SKT), with $p_0 = 1000$ hPa and p_s denoting surface pressure. Note that owing to practical reasons we use SKT instead of SST, as this quantity is also well defined over sea ice and land, facilitating the compilation of the trajectory dataset, as we will outline below. A marine CAO is characterized by strongly positive values of $\theta_{\text{SKT}} - \theta$ over open ocean grid points, indicating static instability associated with upward surface sensible and latent heat fluxes (cf. Papritz et al. 2015). For the sake of generality, we a priori do not impose a threshold on the magnitude of $\theta_{\text{SKT}} - \theta$, except that we require it to be positive. Consequently, we include CAOs with weakly positive $\theta_{\text{SKT}} - \theta$, which would not generally be considered a CAO but might still play

¹ $\theta_{\text{SST}} = \text{SST}(p_0/p_s)^{R/c_p}$, where $p_0 = 1000$ hPa and p_s denote reference and surface pressure, respectively.

some role in shaping mean heat fluxes—in particular in regions where more intense CAOs are not so frequent.

a. Trajectory dataset

We employ the Lagrangian analysis tool (LAGRANTO; Sprenger and Wernli 2015), which uses the three-dimensional wind on model levels from ERA-Interim to calculate kinematic trajectories forward and backward in time. The trajectory positions are output at 6-hourly intervals. Zonal and meridional winds, temperature, and specific humidity are traced along the trajectories by interpolating these to the respective trajectory positions in space, while SKT and surface sensible and latent heat fluxes are interpolated horizontally.

The procedure to obtain the CAO trajectory dataset involves the following three steps:

- 1) Every 6 h we define a set of initial grid points. This set comprises all grid points satisfying $\theta_{\text{SKT}} - \theta > 0$ at this time and where the fractional land cover and sea ice concentration are below 50%. Thereby, we choose a regular $80 \text{ km} \times 80 \text{ km}$ grid covering the starting domain outlined by the black box in Fig. 1a and equally spaced levels in the vertical with intervals of 25 hPa, ranging from 1000 to 500 hPa. The regular horizontal grid in contrast to a latitude–longitude grid ensures that every trajectory represents the same mass (i.e., $\sim 1.6 \times 10^{12} \text{ kg}$), corresponding to an air parcel of thickness 25 hPa and a base area of 6400 km^2 . Note that throughout the manuscript we will refer to the initialization time of the trajectory $t = 0 \text{ h}$ as the base time.
- 2) Starting from the initial grid points, we calculate kinematic trajectories backward in time for 6 h. If the CAO criterion is not satisfied 6 h before the base time (i.e., $\theta_{\text{SKT}} - \theta \leq 0 \text{ K}$ at $t = -6 \text{ h}$), we retain the trajectory for further analysis. Accordingly, the retained trajectories are those satisfying the CAO criterion at the base time, but not before. This approach makes it straightforward to study the CAO evolution relative to the time of CAO formation. Furthermore, it avoids a double sampling of the same CAO air mass at succeeding base times, while at the same time ensuring that the entire CAO air mass forming in the starting domain is sampled.
- 3) Finally, we extend the retained trajectories further backward (forward) in time by 186 h (144 h) such that each trajectory spans the time range $-192 \text{ h} \leq t \leq 144 \text{ h}$. The choice of an asymmetric time span relative to the time of CAO formation is justified by the fact that the CAO phase, even of the most long-lived CAOs, lasts clearly less than 144 h. After extending the trajectories in time, we trace the aforementioned variables along the trajectories. Throughout the manuscript, we

FIG. 2. Schematic illustrating the setup of trajectory calculations. Grid points on the regular starting grid are indicated by dots. The blue contour indicates the boundary where $\theta_{\text{SKT}} = \theta$ at the base time $t = 0 \text{ h}$ and grid points with $\theta_{\text{SKT}} - \theta > 0$ are highlighted in blue. Two trajectories are shown explicitly to illustrate retained (red) and discarded (orange) trajectories.

will refer to the CAO phase of a trajectory as the time period $0 \leq t \leq t_0$, where t_0 is the last time step when $\theta_{\text{SKT}} - \theta \geq 0 \text{ K}$.

The procedure to obtain the trajectory dataset is schematically illustrated in Fig. 2. Grid points satisfying the CAO criterion $\theta_{\text{SKT}} - \theta > 0 \text{ K}$ at the base time, from which 6-hourly backward trajectories are started, are highlighted in blue. Two backward trajectories inside the cold air mass are sketched: one deep inside the CAO air mass (orange) and another one close to its boundary (red). The orange trajectory is discarded, as it already satisfied the CAO criterion at $t = -6 \text{ h}$, whereas the red trajectory did not satisfy the criterion and is therefore retained and extended backward and forward in time.

It is important to note that by this approach we restrict the analysis to CAO air masses forming within the starting domain, whereas CAO air masses forming elsewhere but being advected into the study region are not sampled. Specifically, CAO air masses forming in the Labrador Sea and being advected around Cape Farewell, the southern tip of Greenland, into the Irminger Sea do not enter our analysis, as the starting domain comprises only the Irminger Sea and Nordic seas.

b. Selection of trajectory subsets

A major benefit of the Lagrangian approach is that it allows the isolation of subsets of CAO air masses that satisfy a specific criterion at a certain time during their evolution. The dynamic and thermodynamic evolution of air masses represented by these subsets can then be studied independently from the other CAO air masses.

Specifically, we divide the trajectories into four intensity classes defined by the air–sea potential temperature difference at the time of CAO formation ($t = 0 \text{ h}$): weak ($0 \text{ K} < \theta_{\text{SKT}} - \theta \leq 4 \text{ K}$), moderate ($4 \text{ K} < \theta_{\text{SKT}} - \theta \leq 8 \text{ K}$), strong ($8 \text{ K} < \theta_{\text{SKT}} - \theta \leq 12 \text{ K}$), and very strong ($12 \text{ K} < \theta_{\text{SKT}} - \theta$) CAOs. These classes contain 74.80%,

17.38%, 5.83%, and 1.99%, respectively, of the total of more than 26 million trajectories.

One of the major goals of this study is to quantify the contribution of CAOs of different intensities to winter surface sensible and latent heat fluxes. Hence, a separation of CAO trajectories using equidistant thresholds, instead of, for example, a separation by quartiles, is justified by the fact that the associated surface heat fluxes largely scale with the air–sea potential temperature difference (Papritz et al. 2015). It is important to note that the air–sea potential temperature difference typically decreases after $t = 0$ h owing to surface sensible heat fluxes and the release of latent heat. Thus, the air–sea potential temperature difference at $t > 0$ h can be less than the lower threshold of the intensity class to which the trajectory belongs.

In addition to classifying trajectories according to intensity, we also extract trajectories located at $t = 0$ h in the subregions outlined in Fig. 1a. Specifically, the subregions comprise the Irminger, Iceland, Greenland, and Norwegian Seas, as well as the region west of Svalbard and the eastern Barents Sea.

c. Aggregation of trajectories

To obtain frequency maps of CAOs, we aggregate all trajectory positions for trajectories in their CAO phase (i.e., for $0 \leq t \leq t_0$) on a $1^\circ \times 1^\circ$ latitude–longitude grid using the nearest-neighbor method. This way we obtain a binary field for each intensity class every 6 h, which indicates if a CAO trajectory belonging to the respective intensity class is present at this time step and specific grid point. Time averaging of these binary fields yields CAO frequencies for each intensity class. In the case of several collocated trajectories, the presence of a trajectory is only marked in the binary field corresponding to the most intense trajectory but not in the other binary fields. This ensures that CAO frequencies are additive between intensity classes. Accordingly, the sum of all four classes yields the total frequency of weak to very strong CAOs. Surface sensible and latent heat fluxes associated with CAO trajectories in a specific intensity class are then determined based on collocation with these binary CAO fields. By construction of the binary CAO fields, the obtained heat fluxes are additive between intensity classes as well.

For each time step, we define the formation rate of CAO air masses as the number of trajectories that satisfy the CAO criterion for the first time within the preceding 6-hourly window.² To obtain spatial maps of the formation rate, we aggregate the trajectory positions at

$t = 0$ h on a $1^\circ \times 1^\circ$ latitude–longitude grid using a conservative gridding procedure with linearly decreasing weights in a radius of 160 km. As every trajectory represents the same mass, we express the formation rate in units of hPa day^{-1} , allowing for the concrete interpretation of the formation rate as the rate at which the depth of the CAO air mass changes. In the same way, we aggregate horizontal mass fluxes $M \cdot \mathbf{u}$, where M is the mass of a trajectory and $\mathbf{u}$ its horizontal velocity vector, for all trajectories in the time windows $-96 \leq t < 0$ h and $0 \text{ h} < t \leq 96$ h. Consistent with the formation rate, we express the mass fluxes in units of hPa m s^{-1} , noting that these are not strictly speaking fluxes of mass.

3. Cold air outbreak climatology

a. Frequency of cold air outbreaks and formation rate

Figure 3 shows the frequency of CAOs, obtained according to the procedure outlined in the previous section. For example, Fig. 3a shows maxima of weak CAOs occurring up to 45% of the time during winter. Generally, weak CAOs are most frequent over the warm waters of the Norwegian Sea, with local maxima southwest of the Lofoten Archipelago, and over the southern Irminger Sea. It is noteworthy that these maxima are not located near the sea ice edge but are collocated with elevated SSTs associated with the North Atlantic and the Norwegian Atlantic Currents instead (Fig. 1b). The cold waters in the Greenland and Iceland Seas have the opposite effect, reducing the frequency of weak CAOs as illustrated by the tongue of low frequency east of Iceland as well as along the sea ice edge.

In contrast to weak CAOs, the spatial patterns of the frequency of moderate and strong CAOs are clearly less influenced by the SST distribution but rather dictated by the distribution of land and the location of the sea ice edge. They most frequently occur in three distinct regions (Figs. 3b,c): in the Irminger Sea immediately off Greenland's coast and over the Denmark Strait, in an elongated band along the sea ice edge in the Iceland and Greenland Seas with a maximum over Fram Strait, and in the eastern Barents Sea off Novaya Zemlya, as well as off the Kola Peninsula. Furthermore, a less intense frequency maximum is located immediately off the Lofoten Archipelago extending farther north and south along the Norwegian coast. This region is well known for frequent polar low development (Bracegirdle and Gray 2008; Noer et al. 2011).

The occurrence of very strong CAOs (Fig. 3d) is essentially restricted to the regions off the sea ice edge extending from the Greenland Sea to Svalbard, where they are most frequent over the ice-free portion of the Fram Strait. The Barents Sea is another major region with

² This is simply the number of trajectories with base time ($t = 0$ h) at the time step for which the formation rate is calculated.

FIG. 3. Mean frequencies of CAOs with trajectories binned according to CAO intensity. Note that each time step contributes toward the frequency in the bin with the highest intensity for which a trajectory is present such that the frequencies in the four bins are additive. The mean sea ice boundary (50% sea ice concentration) is shown by the gray contour.

frequent very strong CAOs, where frequency maxima are located off Novaya Zemlya and, a bit weaker, in the western Barents Sea.

While the spatial patterns of elevated CAO frequencies in the Irminger Sea, west of Svalbard, and in the Norwegian and western Barents Seas agree with previous studies (e.g., Kolstad et al. 2009; Kolstad 2011; Fletcher et al. 2016), this is not the case for the Greenland and Iceland Seas, as well as the eastern Barents Sea, where the previous studies show rather low frequencies. This discrepancy is likely an artifact of the method employed for the identification of CAOs in these previous studies. Typically, the depth of CAOs that form as off-ice flows increases with fetch from the sea ice edge from an initially shallow layer of often less than 200-hPa depth (e.g., Brümmner 1996, 1997; Renfrew and Moore 1999). Accordingly, CAOs remain undetected close to the sea ice edge if the 800- or 700-hPa levels are used for the calculation of the air-sea potential temperature difference.

We obtain a detailed picture of the geographic locations where CAO air masses form from CAO formation

rates (Fig. 4). Two strikingly different behaviors are apparent for the classes of weak and moderate to very strong CAOs, respectively. Topographic boundaries, including the coasts of southeast Greenland, Iceland, and Novaya Zemlya, as well as and most prominently the sea ice edge, constitute the main formation regions for moderate to very strong CAOs (Figs. 4b–d). Furthermore, owing to the warm SSTs, the Norwegian coast is an important formation region for moderate and strong CAOs. Significant formation rates of very strong CAOs are restricted to the northernmost portion of the sea ice edge in the Greenland and Barents Seas, where very cold Arctic air masses can be accessed. The Fram Strait is particularly conducive for the export of Arctic air masses—a circumstance that is reflected in the collocated maximum of the formation rate across all intensity classes. In contrast to moderate to very strong CAOs, weak CAOs also form within the ocean basins (Fig. 4a), whereby the spatial distribution of the SST plays a key role. Specifically, the high formation rate

FIG. 4. Mean formation rate of CAO air mass in units of hPa day^{-1} binned according to CAO intensity. The formation rate in each bin is obtained from gridding the locations of all trajectories with $t = 0$ h for the respective intensity range at every 6-hourly time step in the study period. The mean sea ice boundary (50% sea ice concentration) is shown by the gray contour.

along the SST gradient associated with the Arctic Front separating the Greenland and Iceland Seas from the Norwegian Sea (cf. Fig. 1b) suggests that flow across the SST front contributes to the high frequency of weak CAOs over the warmer waters of the Norwegian Sea.

b. Horizontal mass fluxes

Insight into the pathways of the CAO air masses is obtained from maps of horizontal mass fluxes associated with moderate to very strong CAO trajectories during the four days prior to ($-96 \text{ h} \leq t < 0 \text{ h}$; Fig. 5) and after ($0 \text{ h} < t \leq 96 \text{ h}$; Fig. 6) CAO formation in specific subregions, respectively. It is important to note that the maps of mean CAO mass fluxes displayed in Figs. 5 and 6 include contributions from contrasting synoptic configurations that evolve in time. Therefore, the maps do not represent instantaneous flow patterns. Trajectories become less coherent further away from the time of CAO formation such that the contributions to the mass fluxes decline for increasing $|t|$. This justifies

the restriction to mass fluxes during the four days prior and after CAO formation.

Distortion of the flow by topography leads to very high local wind speeds due to barrier effects (e.g., Petersen et al. 2009) and a remarkable directional constancy along Greenland's eastern coast (Moore and Renfrew 2005). Thus, southward flow from the Fram Strait along the eastern coast of Greenland provides the most important pathway for Arctic air masses leading to CAOs in the Nordic seas (Figs. 5a–c). A substantial fraction of Irminger Sea CAO trajectories flow southward until they reach the Denmark Strait (Fig. 5a). Some of these trajectories form a CAO in the Greenland or Iceland Seas previously but subsequently flow over Iceland before they induce a new CAO along the western coast of Iceland.

Trajectories in the Iceland and Greenland Seas follow similar paths parallel to the barrier imposed by Greenland's topography (Figs. 5b,c). In contrast to Irminger Sea CAO trajectories, they cross the sea ice edge before reaching the Denmark Strait. Katabatic downslope

a) Irminger Sea **b) Iceland Sea** **c) Greenland Sea** **d) Svalbard** **e) Norwegian Sea** **f) Barents Sea East**

FIG. 5. Mean horizontal mass fluxes (vectors) and their magnitude (shading) associated with CAO trajectories for $-96 \text{ h} \leq t < 0 \text{ h}$ (i.e., four days before CAO formation). Only trajectories with an intensity of $\theta_{\text{SKT}} - \theta > 4 \text{ K}$ are included. The trajectories are divided into subsets according to their location at $t = 0 \text{ h}$ in subregions (cf. Fig. 1a and green boxes): (a) Irminger Sea, (b) Iceland Sea, (c) Greenland Sea, (d) Svalbard, (e) Norwegian Sea, and (f) eastern Barents Sea. The mean sea ice boundary (50% sea ice concentration) is shown by the gray contour.

FIG. 6. As in Fig. 5, but for $0\text{ h} < t \leq 96\text{ h}$ (i.e., four days after CAO formation).

flows from the ice sheet (Heinemann and Klein 2002) enhance these CAO airmass fluxes, especially along the coast of northeastern Greenland. Greenland's topography plays a less important role for shaping the airmass fluxes associated with CAOs west of the Svalbard Archipelago

(Fig. 5d). The main cause for the strong cold airmass fluxes and the associated peak in CAO formation rates observed here lies in the northern position of the sea ice edge and the strong air–sea temperature contrast associated with the warm waters of the West Spitsbergen

Current. However, topographic steering over and around Svalbard likely has a strong influence on the detailed regional patterns of the CAO airmass fluxes (Skeie and Grønås 2000; Kolstad 2008).

CAO formation in the Irminger Sea is particularly interesting, as these air masses include off-ice flows from the north leading to CAOs over Denmark Strait, as well as downslope flows from southeast Greenland leading to CAOs off the coast of southeastern Greenland. In fact, a pathway of equal importance to the one directed along Greenland's eastern coast is directed from the Canadian Arctic over the Labrador Sea to the plateau of southern Greenland, from where the air masses descend into the Irminger Sea (Fig. 5a). The horizontal mass fluxes associated with this pathway peak near Ammassalik and immediately north of Cape Farewell. Ammassalik is well known for events of converging, intense, and cold downslope flows from the Greenland ice sheet, which contribute substantially to the winter heat loss of the ocean to the atmosphere in that part of the Irminger Sea (Oltmanns et al. 2014).

Some of the trajectories following the pathway over southern Greenland may already form a CAO over the Labrador Sea. However, the evolution of pressure (supplementary Fig. 1) reveals that more than 50% of the trajectories do not ascend substantially before reaching the plateau of Greenland. This indicates that these trajectories are located in the free troposphere above the inversion layer of a CAO air mass in the Labrador Sea, which itself is blocked by Greenland's topography. This is akin to the so-called Austrian type of foehn flows, where air parcels do not ascend substantially on the upstream side and consequently latent heat release is weak (e.g., Hann 1866; Richner and Hächler 2013; Elvidge and Renfrew 2016). Such downslope flow typically occurs in association with westerly tip-jet events (Doyle and Shapiro 1999; Petersen et al. 2003; Moore and Renfrew 2005; Våge et al. 2009).

After CAO formation, most of the air masses are advected southward and eastward toward the Norwegian coast from all of the western subregions. Especially for the Irminger Sea, airmass fluxes are directed toward Scotland and Norway (Fig. 6a), whereas Iceland Sea, Greenland Sea, and Svalbard air masses, in addition to those along the Norwegian coast, also reach into the Barents Sea (Figs. 6b–d). The mass flux vectors also indicate that some of the Irminger Sea air masses flow westward around Cape Farewell into the Labrador Sea (Fig. 6a), likely associated with easterly tip jet events (Renfrew et al. 2009a; Outten et al. 2009). Furthermore, a splitting of the mass fluxes associated with Iceland Sea and Greenland Sea CAO air masses occurs north of Iceland, where some of the air masses are diverted to the Denmark Strait and subsequently into the Irminger Sea.

The air masses leading to CAOs over the Norwegian Sea are mostly of continental origin, though Fig. 5e additionally points at the occurrence of airmass fluxes across the Barents Sea, potentially with antecedent CAO formation there. Subsequent to CAO formation, the air masses are carried westward (Fig. 6e) where they quickly encounter colder waters.

Finally, CAO air masses from the eastern Barents Sea originate largely in the Siberian Arctic with the dominant pathways leading across the northern sea ice edge and Novaya Zemlya (Fig. 5f). Despite the two clear-cut pathways prior to CAO formation, these air masses leave the eastern Barents Sea in virtually all directions, with a substantial fraction returning into the Arctic (Fig. 6f). The returning flow into the Arctic allows for the possibility of recurring CAO formation involving the identical air mass, provided that the air mass is exposed to sufficiently strong intermittent radiative cooling or turbulent mixing with the ambient cold air masses in the Arctic.

c. Lagrangian characteristics

After identifying the major pathways of CAO air masses, we characterize the evolution of thermodynamic properties along CAO trajectories in the various subregions and quantify diabatic cooling and heating rates before and after CAO formation. The goal is to improve our understanding of the mechanisms that constrain the potential for intense CAO air masses in a specific subregion to extract energy from the ocean via air–sea heat fluxes. We focus—as we did for the mass fluxes—on moderate to very strong CAO air masses, which have the highest potential to extract energy from the ocean.

1) DIABATIC COOLING AND HEATING RATES

Generally, the evolution of the median potential temperature (black line in Fig. 7) is characterized by a steady decrease prior to and a rapid increase up to 48 h after CAO formation. In all subregions the 10th–90th-percentile range (light shading in Fig. 7) remains surprisingly narrow during the period of major diabatic heating immediately after CAO formation.

Trajectories in the Iceland and Greenland Seas, as well as west of Svalbard, originate predominantly in the interior Arctic. They have been exposed to continued diabatic cooling before CAO formation, reflecting their typically long residence time in the interior Arctic (not shown). The median cooling rate of these trajectories averaged over the period before CAO formation amounts to about 1 K day^{-1} , whereby 10% of the trajectories experience a cooling rate in excess of 2 K day^{-1} (cf. Table 1), leading to particularly low potential temperatures and large air–sea potential temperature differences at the time of CAO formation. Furthermore,

FIG. 7. Evolution of potential temperature (color) and potential skin temperature (gray; only shown for $t \geq 0$) along CAO trajectories for CAOs forming in subregions as defined in Fig. 1a: (a) Irminger Sea, (b) Iceland Sea, (c) Greenland Sea, (d) Norwegian Sea, (e) Svalbard, and (f) eastern Barents Sea. The median, the interquartile range, and the 10th–90th-percentile range are shown by a solid line and dark and light shading, respectively. Only moderate to very strong CAO trajectories are considered with $\theta_{\text{SKT}} - \theta > 4$ K.

trajectories in the Iceland and Greenland Seas encounter increasingly warmer waters, implying that potential SKT increases at a similar rate as potential temperature for a duration of more than 24 h after CAO formation (Figs. 7b,c). As a result, the air–sea potential temperature differences are maintained against diabatic warming of the air masses by surface sensible heat fluxes and the release of latent heat for about 24–30 h.

Since surface sensible (and latent) heat fluxes in CAOs scale with the air–sea potential temperature difference (cf. Papritz et al. 2015), the surface energy input along these trajectories remains high as well. Consequently, CAO trajectories are subject to the most intense diabatic heating rates in these subregions (Table 1). In particular, trajectories in the Greenland Sea stand out with the most intense cooling rates prior to CAO formation. Furthermore, they have the lowest potential temperature and consequently the largest air–sea potential temperature difference at the time of CAO formation, which is consistent with the large formation rates of strong and very strong CAOs along the sea ice edge in this region. Hence, the diabatic heating rate of 50% of the

Greenland Sea trajectories exceeds 9 K day^{-1} , whereby 10% are heated at a rate in excess of 16 K day^{-1} .

As a fraction of Svalbard trajectories (Fig. 7e) flow over the colder waters in the Barents Sea, as well as over Scandinavia (cf. Fig. 6d), these trajectories are exposed to surface fluxes for a short period only and the median heating rate is lower than in the Greenland and Iceland Seas. However, owing to trajectories moving over warmer waters, the 90th percentile of the heating rate of Svalbard trajectories is as large as that of Iceland Sea trajectories ($\sim 14 \text{ K day}^{-1}$).

Trajectories in the eastern Barents Sea also experience typical Arctic cooling rates prior to CAO formation (Table 1 and Fig. 7f), leading to a low potential temperature and a pronounced air–sea potential temperature difference at $t = 0$ h. However, because of the geographical location of the Barents Sea, many of these trajectories reach land or return into the Arctic soon after CAO formation, which is reflected in a rapid reduction of the median potential SKT. This limits their capability to extract energy from the ocean by surface fluxes and thus to be heated diabatically. Nevertheless,

TABLE 1. Median and 10th percentile of the average cooling rates prior to CAO formation ($-192\text{ h} \leq t \leq -12\text{ h}$), as well as median and 90th percentile of the average heating rates after CAO formation ($0\text{ h} \leq t \leq 24\text{ h}$). Only moderate to very strong CAO trajectories are considered (i.e., with $\theta_{\text{SKT}} - \theta > 4\text{ K}$). For the cooling and heating rates the column minima and maxima, respectively, are highlighted in bold.

Region	Cooling rate (K day^{-1})		Heating rate (K day^{-1})	
	$-192\text{ h} \leq t \leq -12\text{ h}$		$0\text{ h} \leq t \leq 24\text{ h}$	
	Median	10th percentile	Median	90th percentile
Irminger Sea	-0.6	-1.7	5.3	9.6
Iceland Sea	-1.0	-2.0	7.8	13.7
Greenland Sea	-1.2	-2.2	9.1	16.9
Norwegian Sea	-0.5	-1.6	3.5	7.3
Svalbard	-1.1	-2.0	6.0	13.9
Eastern Barents Sea	-1.1	-2.0	4.1	11.1

potential SKT in fact increases along the subset of trajectories leaving the Barents Sea into the Norwegian and Greenland Seas such that the 10% most intensely heated trajectories experience heating rates of more than 11 K day^{-1} .

The diabatic cooling and heating rates are weaker in magnitude for trajectories in the Irminger Sea (Fig. 7a). The combination of relatively high initial potential temperature of Irminger Sea trajectories with moderate cooling rates prior to CAO formation (Table 1) restricts the air–sea potential temperature difference attained at the time of CAO formation. Accordingly, their potential to extract sensible heat from the ocean remains limited. In consequence, the median heating rate after CAO formation lies slightly above 5 K day^{-1} , which is barely more than half the heating rate to which trajectories in the Greenland Sea are exposed.

Weak cooling rates prior to CAO formation are also found for the Norwegian Sea trajectories (Fig. 7d), which is partly owed to the fact that some of these trajectories have previously been exposed to diabatic heating in the Barents Sea. Furthermore, the diabatic temperature increase of Norwegian Sea trajectories is essentially limited by the decreasing air–sea potential temperature difference as trajectories move away from the Norwegian coast over colder waters, imposing a limit on the amount of heat that can be extracted from the ocean.

As a consequence of the maintenance of the air–sea potential temperature difference by rising SKT along the trajectories, the median CAO phase lasts about 42 h for the Greenland Sea and 36 h for the Iceland Sea and Svalbard CAO trajectories. Because of the rapid decrease of the SKT along Norwegian Sea CAO trajectories, in

contrast, the median CAO phase lasts for 24 h for those, and likewise in the eastern Barents Sea it is limited to 18 h only. The comparatively low heating rates of Irminger Sea CAO trajectories, finally, imply a rather long CAO phase of 36 h in the median.

2) $\theta - T$ CHARACTERISTICS

The material rate of change of temperature is given by the following (e.g., Holton and Hakim 2012):

$$\frac{DT}{Dt} = \frac{\kappa T \omega}{p} + \left(\frac{p}{p_0}\right)^\kappa \frac{D\theta}{Dt}, \quad (1)$$

with $\kappa = R/c_p = 0.286$ and the reference pressure $p_0 = 1000\text{ hPa}$. Accordingly, each subset of CAO trajectories has a specific signature in the potential temperature θ –temperature T space, determined by $\theta(t = 0\text{ h})$ and $T(t = 0\text{ h})$, diabatic heating/cooling processes, and adiabatic temperature changes associated with ascent and descent (cf. Bieli et al. 2015).

Figure 8a shows the median $\theta - T$ curves for each of the subregions. Trajectories of Arctic origin undergoing CAO formation along the sea ice edge (i.e., the subsets of Greenland Sea, Svalbard, and eastern Barents Sea trajectories) are characterized by a steady and almost purely diabatic decrease of temperature prior to CAO formation. In addition, they have low (potential) temperature at $t = 0\text{ h}$, followed by an increase of temperature due to intense diabatic heating. Thereafter, diabatic heating decreases and changes in temperature are dominated by adiabatic cooling associated with ascent (cf. supplementary Fig. 1). Note that the putative ascent does not arise from a coherent rise of all trajectories but instead reflects that CAO trajectories become vertically more dispersed after they have been concentrated in the lower troposphere at $t = 0\text{ h}$.

The $\theta - T$ curves of trajectories in the Irminger Sea and in the Norwegian Sea differ from those of trajectories with Arctic origin. They undergo less diabatic cooling before CAO formation, while instead they are subject to strong descent (cf. supplementary Fig. 1) and thus adiabatic warming. In particular, the Irminger Sea receives significant contributions from trajectories flowing over Greenland, with strong descent immediately before CAO formation. However, the median temperature at $t = -192\text{ h}$ is below that of trajectories in the other subsets, whereas potential temperature is comparatively high. Similar to the Norwegian Sea, the Iceland Sea also receives contributions from descending trajectories such that their median $\theta - T$ curve does not show purely Arctic characteristics but also a period of adiabatic warming.

Over the period of 144 h after CAO formation, potential temperature increases the most along trajectories

FIG. 8. The $\theta - T$ diagram for CAO trajectories at $t = 0$ h located in one of the subregions (cf. Fig. 1a): namely, IRM, ICL, GRE, NOR, SVA, and BAE. (a) Absolute values and (b) values relative to the time of CAO formation ($t = 0$ h). Lines represent the medians of θ and T from $t = -192$ h to $t = 144$ h, and dots indicate the values at 24-hourly intervals. The values at $t = -192$ h are marked by triangles and the time of CAO formation ($t = 0$ h) by black circles. Only moderate to very strong CAO trajectories are considered with $\theta_{\text{SKT}} - \theta > 4$ K.

in the Greenland and Iceland Seas, followed by Svalbard and the Irminger Sea. This can best be seen from Fig. 8b, where the curves are shown relative to the conditions at $t = 0$ h. In contrast, the heating is weakest for the Norwegian Sea because of the limiting effect of decreasing SST toward the west. Furthermore, some of the eastern Barents Sea trajectories circulate back into the Arctic, where they are diabatically cooled again. Therefore, the $\theta - T$ curve hints at the existence of a closed loop for some of the trajectories with sequential oceanic heat loss to the same CAO air mass in the Barents Sea.

3) MOISTURE GAIN

A characteristic property of CAO air masses is their absolute dryness prior to CAO formation, which is reflected in low median specific humidity ($< 0.5 \text{ g kg}^{-1}$) with a narrow spread (Fig. 9). In contrast, these air masses feature a rapid gain of moisture due to evaporation from the ocean surface during the CAO phase with a large spread around the median. As the trajectories become more dispersed in the vertical after the CAO phase, specific humidity declines again. In contrast to specific humidity, the median relative humidity remains strikingly constant along CAO trajectories, with the exception of an increase of 10% to 20% to nearly 80% at the time of CAO formation. Furthermore, a substantial reduction of relative humidity occurs in association with descending trajectories prior to CAO formation. As pointed out by Papritz et al. (2015), the reduced relative humidity of descending CAO trajectories amplifies evaporation at the time of CAO formation, but this effect weakens rapidly as relative humidity is restored to a value of about 80%.

The near constancy of relative humidity during the CAO phase implies that specific humidity is essentially slaved to the temperature evolution by virtue of the Clausius–Clapeyron relation. Consequently, the moisture gain is limited by the maximum temperature attained during the CAO phase, which is ultimately determined by the underlying SST. Hence, the largest median increase of specific humidity occurs in the Iceland Sea, where the maximum temperature reached along the trajectories is highest among all subregions (cf. Fig. 8a), followed by the Irminger and Greenland Seas. Because of the lower SSTs and, therefore, the lower maximum temperatures achieved along the Svalbard or eastern Barents Sea trajectories, their moisture gain is considerably weaker in comparison. Finally, the Norwegian Sea stands out with more than 50% of the trajectories exceeding 1 g kg^{-1} in specific humidity at $t = -24$ h, which is due to trajectories that were moistened while residing in the Barents Sea before. Furthermore, their moisture gain is limited by the early decrease of temperature as the trajectories move over colder waters shortly after CAO formation.

4. Surface sensible and latent heat fluxes associated with cold air outbreaks

a. Fractional contributions of cold air outbreaks

In the following we quantify the contributions of CAO trajectories in each intensity class to the upward sensible and latent heat fluxes, hereafter referred to as the upward turbulent heat flux. The upward turbulent heat flux

FIG. 9. As in Fig. 7, but for specific humidity (color) and relative humidity (gray). Note the shifted axis for relative humidity on the right.

at a given location is attributed to a specific CAO intensity class if (i) the most intense trajectory with $\theta_{\text{SKT}} - \theta > 0$ present at this location belongs to this intensity class and (ii) the turbulent heat flux itself is directed upward.³ If, for example, a moderate and a strong trajectory are present at the same time and location, the upward turbulent heat flux contributes to the class of strong trajectories but not the class of moderate ones. Time averaging of the so obtained attributed upward turbulent heat flux fields yields the climatological contributions of each class to the mean upward turbulent heat flux. By construction, the fractional contributions of each class are additive, and they add up to about 60% to 80% of the climatological fluxes (Fig. 10a), with low values in regions with strong advection of CAO air masses forming outside of the study area, such as along the southern boundary of the study domain.

The spatial distribution of the fractional contributions is to first order determined by the frequency of occurrence of CAOs of the respective intensity class. For most

intensity classes and regions, the fractional contributions of CAOs strongly exceed their frequencies of occurrence (cf. Fig. 3). Accordingly, they have a strong impact on the climatological upward turbulent heat flux and its spatial distribution.

Weak CAO trajectories are the dominant contributor to the upward turbulent heat flux in the southern Irminger Sea and over the warm waters of the Norwegian Sea (not shown), whereas along the sea ice edge—the preferential formation regions of strong and very strong CAOs—the presence of a weak CAO trajectory is often concomitant with a more intense trajectory. Moderate CAOs account for up to 30% of the upward turbulent heat flux in the Iceland Sea, over the Denmark Strait, and off south-eastern Greenland (Fig. 10b), and the relative importance of strong and very strong CAO trajectories increases toward the north (Figs. 10c,d), at the expense of moderate CAOs. Along the sea ice edge, moderate to very strong CAO trajectories contribute about 20% each, with the largest shares of very strong CAO trajectories west of Svalbard, as well as in the eastern Barents Sea off Novaya Zemlya. Furthermore, moderate and strong CAOs

³ This is normally the case if $\theta_{\text{SKT}} - \theta > 0$.

FIG. 10. Percentage of positive (upward) surface turbulent heat flux (sensible and latent) associated with CAO trajectories (a) for all trajectories and (b)–(d) for moderate to very strong trajectories. Note that percentages in (b)–(d) are additive. The mean sea ice boundary (50% sea ice concentration) is shown by the gray contour.

contribute substantially to the upward turbulent heat flux in a narrow band along the Norwegian coast.

These findings imply that off the sea ice edge in the Iceland and Greenland Seas, west of Svalbard, and the eastern Barents Sea a small number of intense CAOs has a tremendous impact on the winter upward turbulent heat flux, whereas the upward turbulent heat flux in the Irminger and Norwegian Seas is predominantly determined by frequent weak and moderate CAOs. This suggests that the winter mean upward turbulent heat flux in the former regions is more prone to interannual variations in synoptic activity than in the latter.

b. Lagrangian evolution of surface turbulent heat fluxes

The previous considerations of the thermodynamic evolution of CAOs suggest that the amount of energy extracted from the ocean by a CAO air mass depends not only on its initial air–sea potential temperature difference but also on the evolution of the SST along its pathway. Thus, the question arises whether very strong

CAOs that form over colder waters, as is the case along the sea ice edge in the Greenland Sea, are more efficient in extracting energy than moderate CAOs that form over warmer waters, for example in the Irminger Sea. To address this inherently Lagrangian question, we first assess the temporal evolution of the surface turbulent heat flux along CAO trajectories. Second, we quantify the sensible and latent heat fluxes extracted along CAO trajectories during their CAO phase (i.e., $0 \leq t \leq t_0$) for each subregion and intensity class.

Figure 11 shows the temporal evolution of surface sensible and latent heat fluxes for each subregion. Sensible heat fluxes scale approximately linearly with the air–sea potential temperature difference, whereas latent heat fluxes saturate toward the limit imposed by saturation specific humidity at the SST (Papritz et al. 2015). For CAOs forming over warm waters, namely in the Irminger and Norwegian Seas, the latent heat flux exceeds the sensible heat flux at all times. In contrast, sufficiently intense CAOs forming over the cold waters off the sea ice edge extract sensible heat more efficiently

FIG. 11. As in Fig. 7, but for surface turbulent (sensible and latent) heat fluxes in the period $-48 \text{ h} \leq t \leq 96 \text{ h}$. In addition, the median sensible and latent heat fluxes are shown by gray dashed and dotted lines, respectively.

than latent heat. However, the relative importance of latent over sensible heat fluxes increases as these CAO air masses are advected away from the sea ice edge over increasingly warmer waters, where the latent heat fluxes can exceed the sensible heat fluxes in the later stage of the CAO evolution. A similar increase of the relative importance of the surface latent heat flux over the sensible heat flux with fetch from the sea ice edge has been documented by in situ observations of CAOs west of Svalbard (Brümmer 1997).

The large spread in turbulent heat fluxes during the 48 h prior to CAO formation shows that 25% to 50% of the trajectories in the Irminger (Fig. 11a), Iceland (Fig. 11b), and Norwegian (Fig. 11d) Seas have been exposed to surface fluxes, most likely associated with antecedent CAO formation in the Labrador and Iceland Seas, Greenland Sea, and Barents Sea, respectively (see section 3b). Such a preconditioning of the CAO air masses is absent in the case of purely Arctic air masses leading to CAOs in the Greenland Sea (Fig. 11c) and west of Svalbard (Fig. 11e), as well as the eastern Barents Sea (Fig. 11f). Subsequent to CAO formation, turbulent heat loss from the ocean such that the extracted sensible heat

fluxes peak within 6 h in all subregions, followed by a decay to 50% of the peak value during 18 to 36 h after CAO formation. Thereby, the decay is most rapid for rather short-lived CAOs in the Norwegian Sea and the eastern Barents Sea.

The integrated sensible heat flux in each subregion scales approximately linearly with the initial air–sea potential temperature difference (Fig. 12a), supporting the previous scaling arguments. Comparing the same intensity classes between subregions reveals further that the integrated sensible heat flux for moderate and strong CAOs is largest in the Greenland Sea, followed by the Iceland Sea and the Svalbard subregion. The shorter lifetimes of CAOs in the Norwegian Sea, as well as in the eastern Barents Sea, limit the integrated sensible heat flux. The effect of increasing SST along trajectories in the Greenland and Iceland Seas carries less weight for very strong CAOs, as the increase of the SST is smaller there than the gain of potential temperature due to diabatic heating. It is interesting to note that very strong CAOs are often accompanied by intense winds in the Irminger Sea (median $\approx 17 \text{ m s}^{-1}$; 90th percentile $\approx 28 \text{ m s}^{-1}$), which considerably enhance the sensible heat loss from the ocean such that the extracted sensible heat is comparable in magnitude to that of the Iceland and Greenland Seas trajectories.

FIG. 12. Boxplots of integrated (a) sensible, (b) latent, and (c) sensible and latent heat flux during the CAO phase (i.e., for $0 \leq t < t_0$) for trajectories located in one of the subregions at $t = 0$ h (cf. Fig. 1a): namely, IRM, ICL, GRE, NOR, SV), and BAE. The trajectories are binned according to intensity, with increasing intensity from left to right. The whiskers indicate the 10th–90th-percentile range.

Contrary to the sensible heat flux, the latent heat flux clearly levels off at higher CAO intensities despite scaling linearly at low CAO intensities (Fig. 12b). As such, the SST plays a crucial role in setting the maximum amount of moisture the air mass can gain through

evaporation. Accordingly, Irminger Sea CAOs extract the most latent heat from the ocean, followed by Iceland Sea and Greenland Sea as well as Svalbard CAOs. The strong impact of the SST on the integrated latent heat flux is emphasized by the comparatively high values for

very strong CAOs in the relatively warm Norwegian Sea. Based on this analysis, we conclude that a small number of fairly intense CAOs forming in the Irminger, Iceland, and Greenland Seas, as well as west of Svalbard—in descending order—extract the most heat and thus have the strongest cooling impact on the ocean (Fig. 12c).

5. Concluding remarks

We developed a novel method to sample CAO air masses using kinematic trajectories and presented a multidecadal, Lagrangian dataset of wintertime CAOs over the Irminger Sea, Nordic seas, and Barents Sea (i.e., regions of central importance for the densification of surface waters and thus the oceanic meridional overturning circulation). Based on this dataset we analyzed the origin, pathways, and thermodynamic evolution of these CAO air masses and pinpointed their role in shaping the spatial variability of mean and extreme upward air–sea heat fluxes.

The formation of the most intense CAOs occurs along the sea ice edge in the Greenland Sea and Fram Strait, and to a lesser extent along the northern sea ice edge in the Barents Sea and off the west coast of Novaya Zemlya. While intense CAOs exclusively form along the sea ice edge and along topographic boundaries, the flow across the SST gradient in the interior of the Norwegian Sea plays an important role for the formation of weak CAOs.

We identified three distinct pathways of CAO air masses:

- 1) The most prominent airmass stream leaves the interior Arctic via the Fram Strait diverting into two main branches. One branch constitutes the air masses leading to the frequency maximum of very strong CAOs over the warm waters west of Svalbard. The other branch is guided south over the sea ice–covered ocean along the eastern coast of Greenland, feeding CAOs in the Greenland and Iceland Seas, as well as the northern Irminger Sea via the Denmark Strait and Iceland.
- 2) In the eastern Barents Sea, air masses prevail that cross the sea ice edge from the north or flow across Novaya Zemlya. These air masses quickly return to the interior Arctic or move over northern Scandinavia, from where they can reach the warm waters of the Norwegian Sea, subsequently forming a secondary CAO there.
- 3) In addition to the southbound stream along Greenland's eastern coast, a second airmass stream leads into the Irminger Sea via southern Greenland. These air masses originate in the Canadian Arctic, flow over southern Greenland, and descend along Greenland's eastern slope.

CAOs forming in the Iceland and Greenland Seas are the CAO subsets extracting the most sensible heat from the ocean and experiencing the strongest diabatic temperature increase. In contrast, the largest moisture gain via evaporation occurs along Irminger Sea CAO trajectories, owing to the high SST to which these trajectories are exposed. However, the diabatic temperature increase along these trajectories is smaller than that of trajectories originating in the Iceland and Greenland Seas because not all of the latent heat is released.

The CAO trajectories account for 60% to 80% of the mean winter oceanic heat loss by upward sensible and latent heat fluxes in the studied regions. Furthermore, the spatial variations in the intensity of mean and extreme sensible heat fluxes are set by the formation rate of CAOs, the spectrum of their intensity, and their pathways over the underlying SST distribution. In particular, the mean upward sensible and latent heat fluxes are set by a few intense CAOs along the sea ice edge, whereas the high frequency of weak CAOs sets the heat fluxes in the interior basin of the Norwegian Sea, as well as the southern part of the Irminger Sea.

As the analyses presented in this study depend on the quality and reliability of ERA-Interim, we refer to the extensive literature documenting the skill of ERA-Interim in representing the atmospheric conditions at high latitudes. Compared to other reanalyses ERA-Interim is outstanding in terms of small biases with respect to in situ observations (Lindsay et al. 2014), features surface fluxes within the range of observational uncertainty (Renfrew et al. 2009b), and represents the near-surface atmospheric conditions in the Iceland Sea as observed from a 2-year-long deployment of a meteorological buoy to a very high degree of accuracy (Harden et al. 2015). Based on these assessments we believe that ERA-Interim provides an adequate basis for the present kind of large-scale climatological studies of the wintertime atmospheric flow over the Nordic seas. However, the 6-hourly temporal and the coarse spatial resolution of ERA-Interim present caveats to this study by limiting the accuracy of the trajectory calculations (e.g., owing to noise from gravity waves). Furthermore, the spatial resolution might be a critical factor for correctly representing topographic flow distortions around Greenland, as well as katabatically driven downslope flows feeding into the southbound cold airmass stream along Greenland's eastern coast. However, when considering the large-scale features of the flow over and around Greenland, ERA-Interim represents major topographic effects reasonably well, such as intense downslope winds (Oltmanns et al. 2014), barrier flows, and tip jets (Harden et al. 2011). As we focus on the large scale rather than on the mesoscale, we believe that our

main conclusions should not be severely affected by these limitations.

Dense water formation in the Greenland and Iceland Seas is primarily the result of cooling of the ocean mixed layer by turbulent heat fluxes, which, as shown in this study, is brought about by a few intermittent but intense CAOs. The favorable conditions for the formation of the densest overflow waters in these basins rely to some extent on the circumstance that salinity is not substantially reduced by atmospheric freshwater fluxes, which is also a consequence of the fact that evaporation partly balances precipitation (e.g., Mauritzen 1996; Segtnan et al. 2011). Therefore, we conclude that latent heat fluxes associated with CAOs potentially play a twofold role in dense water formation in the Greenland and Iceland Seas: first and most importantly, they contribute to the cooling of the oceanic mixed layer, and second, they balance part of the freshwater input by precipitation.

Changes in the air–sea heat flux forcing of the subpolar North Atlantic and the Nordic seas can have global climatic implications owing to their importance for open ocean convection contributing to the Atlantic meridional overturning circulation. In addition, the Iceland Sea has recently been suspected to play a more fundamental role for open ocean convection than previously thought (e.g., Våge et al. 2011, 2013; Harden et al. 2016). While mean upward sensible and latent heat fluxes in the Iceland Sea are weaker than those farther north in the Greenland Sea, our results suggest that the strongest CAOs are accompanied by heat losses comparable in magnitude in these two regions. This is in line with buoy observations showing intermittent high heat flux events with periods of near-zero heat flux in between (Harden et al. 2015). These findings suggest that the oceanic mixed layer in the Iceland Sea deepens as a result of a few intermittent but intense CAOs instead of continuous heat losses during weak CAOs.

Recently, Moore et al. (2015) found that the air–sea heat fluxes in the gyre circulations of the Iceland and Greenland Seas declined by about 20% during the past decades, which they attributed primarily to a reduction of sea ice extent over the same period. As intense CAOs preferentially form along the sea ice edge, it is clear that variations in sea ice extent are translated into changes of the oceanic heat loss via associated shifts of the formation regions of CAOs. However, we showed further that not only the location and characteristics of CAO air masses at the time of CAO formation are crucial in dictating the potential of CAOs to extract energy from the ocean but also the pathways of CAO air masses over the underlying SST distribution. Accordingly, changes in the large-scale circulation (e.g., modulations of the intensity and location of the storm track) can have a

profound impact on the air–sea heat flux forcing imparted by CAOs on the Nordic seas. We aim to analyze the consequences of this implication in future work based on the CAO trajectory dataset developed in this study and suitable storm-track measures.

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