Arctic Lower-Tropospheric Warm and Cold Extremes: Horizontal and Vertical Transport, Diabatic Processes, and Linkage to Synoptic Circulation Features

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ABSTRACT

The thermodynamic processes and synoptic circulation features driving lower-tropospheric temperature extremes in the high Arctic ($>80^{\circ}\text{N}$) are investigated. Based on 10-day kinematic backward trajectories from the 5% most intense potential temperature anomalies, the contributions of horizontal and vertical transport, subsidence-induced warming, and diabatic processes to the generation of the Arctic temperature anomaly are quantified. Cold extremes are mainly the result of sustained radiative cooling due to a sheltering of the Arctic from meridional airmass exchanges. This is linked to a strengthening of the tropospheric polar vortex, a reduced frequency of high-latitude blocking, and in winter also a southward shift of the North Atlantic storm track. The temperature anomaly of 60% of wintertime extremely warm air masses (90% in summer) is due to transport from a potentially warmer region. Subsidence from the Arctic midtroposphere in blocking anticyclones is the most important warming process with the largest contribution in summer (70% of extremely warm air masses). In both seasons, poleward transport of already warm air masses contributes around 20% and is favored by a poleward shift of the North Atlantic storm track. Finally, about 40% of the air masses in winter are of an Arctic origin and experience diabatic heating by surface heat fluxes in marine cold air outbreaks. Our study emphasizes the importance of processes in the Arctic and the relevance of anomalous blocking—in winter in the Barents, Kara, and Laptev Seas and in summer in the high Arctic—for the formation of warm extremes.

1. Introduction

The Arctic climate system has been subject to dramatic changes throughout the past decades, most prominently manifest by a bottom amplified warming that is far more rapid than the global mean temperature increase. This phenomenon is known as Arctic amplification (e.g., Serreze and Barry 2011; Cohen et al. 2014). Another dramatic change is the fast decline of sea ice and, in particular, the loss of multiyear sea ice (Stroeve et al. 2012; Simmonds 2015). Arctic amplification has been attributed to a number of mechanisms, namely, sea ice albedo feedback (Arrhenius 1896; Screen and Simmonds 2010a), enhanced surface sensible and latent heat fluxes (Screen and Simmonds 2010b; Boisvert et al. 2015), changes to the radiation balance due to

temperature feedbacks and the nonlinear increase of water vapor with temperature following the Clausius–Clapeyron relationship (Francis and Hunter 2006; Pithan and Mauritsen 2014; Woods and Caballero 2016; Graversen and Burtu 2016), and circulation changes (Graversen 2006; Sorteberg and Walsh 2008; Graversen and Wang 2009; Ding et al. 2017; Mewes and Jacobi 2019).

Transient, synoptic-scale events play an especially important role for the energy balance of the Arctic and in particular sea ice melting. More specifically, it has been recognized that the bulk of the seasonal horizontal heat and moisture transports into the high Arctic (in this study defined as the Arctic poleward of 80°N) is due to a few episodic but very intense injections of warm and humid air masses from midlatitudes into the high Arctic (Woods et al. 2013; Laliberté and Kushner 2014; Liu and Barnes 2015; Dufour et al. 2016; Naakka et al. 2019). Such injections induce a transition of the atmosphere from a cold and clear to a warm and opaque state, where the trapping of longwave radiation provides a positive warming feedback with a strong impact on the surface

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energy balance and sea ice (D.-S. R. Park et al. 2015; H.-S. Park et al. 2015; Woods and Caballero 2016). A number of weather systems have been found to contribute to their occurrence, namely, cyclones in the midlatitudes and in the Arctic (Sorteberg and Walsh 2008; Rinke et al. 2017; Messori et al. 2018), blocking and Rossby wave breaking (Woods et al. 2013; Liu and Barnes 2015), as well as tropically excited Rossby wave trains and their interference with synoptic-scale waves (Baggett et al. 2016; Fournoy et al. 2016). In addition to airmass injections, near-surface warm temperature and surface longwave, as well as in summer also shortwave, radiation anomalies in the Arctic have been linked to collocated anticyclonic flow anomalies (Pfahl and Wernli 2012; Knudsen et al. 2015; Ding et al. 2017; Wernli and Papritz 2018). While many studies attempted to quantify the contributions of such flow features to the surface energy balance and their importance for sea ice melting, comparatively little attention has been given to the thermodynamic mechanisms that lead to the formation of warm extremes in the Arctic lower troposphere and how they relate to atmospheric dynamical processes.

The temperature evolution of an air mass is described by the thermodynamic energy equation (cf. Holton and Hakim 2012)

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

where $\kappa = R/c_p = 0.286$ and $p_0 = 1000$ hPa denotes reference pressure. It relates the material rate of change of temperature (DT/Dt) to adiabatic temperature changes associated with expansion and compression due to vertical motion ($\omega = Dp/Dt$), and diabatic heating and cooling ($\dot{\theta}$). Hence, considering the trajectory of an air mass in θ - T phase space yields quantitative insight into the contributions of vertical motion and diabatic processes to its temperature evolution (Bieli et al. 2015). According to Eq. (1), lower-tropospheric warm extremes can form as the result of the poleward advection of already warm air masses from lower latitudes, adiabatic warming associated with subsidence from mid-tropospheric levels, as well as diabatic heating. In particular, subsidence-induced adiabatic warming has been recognized as the key warming mechanism for the formation of midlatitude heat waves (Bieli et al. 2015; Quinting and Reeder 2017; Zschenderlein et al. 2018, 2019). The relatively large fraction of Arctic near-surface temperature extremes collocated in space and time with atmospheric blocking (Pfahl and Wernli 2012), and the importance of Arctic anticyclones for sea ice melting (Wernli and Papritz 2018), suggest that this

mechanism may also be particularly important for Arctic warm extremes.

An illustrative case of an extreme Arctic warm episode occurred at the turn of 2015 to 2016 with above freezing temperature near the North Pole (Moore 2016; Cullather et al. 2016) and significant sea ice melting in the Barents and Kara Seas (Boisvert et al. 2016). This event occurred during a sequence of exceptionally intense meridional airmass exchanges over the Nordic seas, namely, first the export of a large volume of Arctic cold air masses over the Nordic seas leading to one of the most severe cold air outbreaks during the past decades (Papritz and Sodemann 2018), followed by the injection of anomalously warm and humid air masses. This injection was, in the first place, attributed to the poleward transport of warm air masses from the south in the warm sectors of two cyclones—an intense Atlantic storm propagating poleward along the coast of Greenland (Boisvert et al. 2016; Kim et al. 2017) and a cyclone that developed in the Arctic (Moore 2016). This interplay of a sequence of midlatitude and Arctic cyclones was confirmed by Messori et al. (2018) as an important mechanism for the injection of warm air masses all the way into the high Arctic. As Binder et al. (2017) showed, however, the origin and thermodynamic history of the air masses involved in the 2015/16 event was diverse. Specifically, they identified three distinct airstreams: a warm subtropical air mass moving poleward, a mid-tropospheric air mass that subsided in an anticyclone over Scandinavia and warmed adiabatically, and an originally Arctic air mass that gained heat via surface heat fluxes over the Labrador Sea and was subsequently advected cyclonically around Greenland.

The above example indicates that lower-tropospheric warm extremes in the high Arctic may not simply be the result of the poleward transport of already warm air masses. As we will show here, such temperature extremes rather emerge from the complex interplay of various thermodynamic warming mechanisms and horizontal and vertical airmass transport associated with specific weather systems. Hence, our study is guided by the following three specific questions:

- To what extent are warm extremes the result of poleward transport of already warm air masses?
- How much do subsidence-induced adiabatic warming and diabatic heating contribute?
- Can these thermodynamic and transport processes be linked to specific synoptic circulation features?

In addition to warm extremes, we will also address cold extremes as a contrasting case. This is further motivated by the fact that their formation has not obtained much attention in the literature, although they are an

important prerequisite for the occurrence of most intense cold air outbreaks over the subpolar seas. To address these questions, we compute kinematic backward trajectories from the lower troposphere in the high Arctic, which we classify according to their evolution in θ – T space. Based on this classification, we then develop a framework for quantifying the contributions of the initial temperature anomaly of an air mass, transport, and diabatic processes for the formation of the final temperature anomaly. The identification of synoptic circulation features will then allow us to link these processes to their atmospheric dynamical drivers.

2. Methodology

We base this study on the ERA-Interim reanalysis dataset (Dee et al. 2011) from the European Centre for Medium-Range Weather Forecasts (ECMWF). We use 6-hourly analyzed fields on 60 model levels, spatially interpolated from the native T255 spectral resolution to a $1^\circ \times 1^\circ$ longitude–latitude grid. The study period includes winters [December–February (DJF)] from 1979/80 to 2017/18 and summers [June–August (JJA)] from 1979 to 2017.

For identifying anomalously warm and cold air masses, we define a transient potential temperature climatology θ_c as follows. First, potential temperature θ is temporally smoothed with a 21-day running mean filter. Then, for each calendar day the 9-yr running mean is computed. For the five years at the start and at the end of the study period, instead the mean over the first and last 9 years is taken, respectively. Note that this approach for defining the temperature climatology is similar to that employed by Messori et al. (2018), but it provides a smoother seasonal cycle as short-term fluctuations are removed by the 21-day running mean filter. It results in a climatology that retains the seasonal cycle with smooth day-to-day variations, while taking the overall warming trend in the high Arctic into account (Cohen et al. 2014). With anomalies defined as the deviation from climatology, that is, $\theta^* = \theta - \theta_c$, the definition of warm and cold anomalies is independent from intraseasonal variations of the climatological background state as well as from the general warming trend in the study period. Consequently, the distribution of extreme warm and cold anomalies remains fairly uniform within seasons and throughout the study period.

a. Trajectory calculation

Trajectory starting positions are defined on a regular $80 \text{ km} \times 80 \text{ km}$ grid poleward of 80°N and at 10, 30, 50, 70, and 90 hPa above ground level, where grid points above land are excluded. Using a regular horizontal grid

with uniform vertical spacing in terms of pressure ensures that each trajectory is representative of the same mass ($\approx 1.3 \times 10^{12} \text{ kg}$). From these starting points we compute 10-day kinematic backward trajectories $\mathbf{x}(t)$ using the Lagrangian Analysis Tool (LAGRANTO; Sprenger and Wernli 2015) for all 6-hourly time steps in the study period. Temperature T , θ , θ_c , specific humidity, and surface skin temperature are then interpolated to the trajectory locations. Note that throughout the study, we use for any physical quantity χ traced along a trajectory $\mathbf{x}(t)$ the shorthand notation $\chi(t) = \chi[\mathbf{x}(t), t]$ and $\chi_0 = \chi(t = 0 \text{ h})$ denotes the value at the initialization time of the trajectory. The so-obtained trajectory dataset provides the climatological basis for this study. Selecting trajectories with θ_0^* below the seasonal 5th percentile and in excess of the seasonal 95th percentile of the distribution of θ_0^* for all trajectories, yields the trajectories associated with extreme cold and warm anomalies, respectively.

b. Classification of trajectories

For classifying the trajectories according to their thermodynamic evolution, we consider adiabatic and diabatic temperature changes during the 10 days¹ prior to arrival in the high Arctic, following the procedure by Binder et al. (2017) and Zschenderlein et al. (2019). For χ denoting T or θ , we define $\Delta\chi$ as the maximum absolute difference of χ along the trajectory and its value at the starting time χ_0 , that is,

$$\Delta\chi = \begin{cases} \chi_0 - \chi_{\min}, & \text{if } |\chi_0 - \chi_{\min}| \geq |\chi_0 - \chi_{\max}|, \\ \chi_0 - \chi_{\max}, & \text{else,} \end{cases} \quad (2)$$

where $\chi_{\min}$ and $\chi_{\max}$ denote the minimum and maximum values along the trajectory, respectively. Considering, for example, temperature T , then if $T_{\min}$ deviates more from T_0 than $T_{\max}$, ΔT is positive and the trajectory predominantly experienced warming throughout the previous 10 days, and vice versa if $T_{\max}$ deviates more from T_0 . In a second step we then categorize trajectories according to their location in one of the quadrants in the phase space spanned by ΔT and $\Delta\theta$ as listed in Table 1. As we will see later, this categorization is useful because each category represents a very different thermodynamic evolution of the involved air masses. Throughout the manuscript, we will use the notation $\Delta\theta \pm \Delta T \pm$ for referring to the categories. For example, $\Delta\theta - \Delta T +$

¹ See section 3 of the supplemental material for a sensitivity analysis with regard to the choice of time horizon for the classification of trajectories.

TABLE 1. Definition and key characteristics of trajectory categories.

Category	Definition	Key characteristics	Example of characteristic airstream
$\Delta\theta + \Delta T -$	$\Delta T < 0, \Delta\theta > 0$	Diabatically heated, ascending	Warm conveyor belt
$\Delta\theta + \Delta T +$	$\Delta T > 0, \Delta\theta > 0$	Diabatically heated, little vertical motion	Marine cold air outbreak
$\Delta\theta - \Delta T +$	$\Delta T > 0, \Delta\theta < 0$	Diabatically cooled, subsiding	Subsidence in blocking anticyclone
$\Delta\theta - \Delta T -$	$\Delta T < 0, \Delta\theta < 0$	Diabatically cooled, little vertical motion	Poleward-moving warm air mass

denotes the category comprising air masses with $\Delta\theta < 0$ and $\Delta T > 0$.

An important constraint on the type of airstream represented by each of the categories is obtained from the thermodynamic energy equation [Eq. (1)]. In particular, it implies that trajectories in $\Delta\theta + \Delta T -$ must ascend because the temperature increase due to diabatic heating ($\Delta\theta > 0$) must be overcompensated by adiabatic cooling ($\Delta T < 0$). Hence, the ascending, moist air masses in the warm sector of extratropical cyclones, the so-called warm conveyor belt (e.g., Browning 1990; Madonna et al. 2014), fall into this category. Likewise, trajectories in $\Delta\theta - \Delta T +$ must descend as $\Delta\theta < 0$ and $\Delta T > 0$, which is an important characteristic of subsiding air masses in a blocking anticyclone that are subject to longwave radiative cooling. The $\Delta\theta + \Delta T +$ air masses experience strong diabatic heating and a temperature increase, which, for instance, is typical for air masses in marine cold air outbreaks that are exposed to upward surface sensible heat fluxes [see thermodynamic characterization of marine cold air outbreaks in Papritz and Spengler (2017)]. Finally, a lower-tropospheric, poleward-moving warm air mass subject to longwave radiative cooling, no other substantial diabatic heat sources, and without substantial vertical motion would belong to $\Delta\theta - \Delta T -$.

c. Compositing of flow features

To explore the role of large-scale flow features, we compile composites of 500-hPa geopotential height, cyclone frequency (Wernli and Schwierz 2006), and the frequency of dynamical blocking (Schwierz et al. 2004; Croci-Maspoli et al. 2007) during the 3 days prior to specifically defined events. The identification of these events is based on the 6-hourly time series of the number of warm or cold extreme trajectories started in the high Arctic. Given that all trajectories correspond to the same mass, this measure is proportional to the atmospheric mass associated with lower-tropospheric positive and negative temperature anomalies in the high Arctic. In a first step, all time steps with zero trajectory counts are removed from the time series and the remaining time steps are ranked by decreasing trajectory counts. Events are subsequently identified in an iterative procedure as follows: The highest ranked time step

defines the first event and all time steps from 5 days before until 5 days after the defining time step of the event are removed from the time series. Subsequently, the algorithm proceeds in the same way with the highest ranked time step remaining in the time series. This procedure yields a ranked list of independent events that avoids a double counting in the case that one single event persists for several days. Note that the time interval of ± 5 days corresponds approximately to the decorrelation time of the time series. Tests revealed that the precise choice of the time interval does not have a strong influence on the definition of events and the composites. For the composites we will consider the 100 highest ranked events.

3. Lagrangian climatology of temperature extremes

a. Example episode

To illustrate the Lagrangian approach underlying this study, we first consider the period 15 December 2015–6 January 2016, which comprises the events mentioned in the introduction: one of the most severe outbreaks of cold air masses from the high Arctic over the Nordic seas around 24 December 2015 and an exceptional warm episode in the high Arctic at the turn of the years. Figure 1a shows the time series of the number of trajectories associated with warm and cold extremes in the high Arctic and Figs. 1b–d depict corresponding backward trajectories for three selected times and colored according to their location in one quadrant in $\Delta T - \Delta\theta$ phase space (cf. Table 1).

During the first half of the period, anomalously cold conditions prevailed in the high Arctic. Associated cold trajectories belonged predominantly to $\Delta\theta - \Delta T -$ (Fig. 1a), which represents air masses that experienced diabatic cooling and a temperature decrease. As revealed by Fig. 1b, many of them originated at high latitudes over Siberia. The subsequent period was then characterized by the joint occurrence of cold and warm extremes in the high Arctic following the export of a substantial amount of cold air into the Nordic seas [cf. Papritz and Sodemann (2018) for a synoptic overview]. While north of Fram Strait and in the western Arctic still

FIG. 1. Trajectories associated with cold and warm extremes in the high Arctic during the period from 0000 UTC 15 Dec 2015 to 0000 UTC 6 Jan 2016. (a) Time series of the number of backward trajectories in each category starting in the high Arctic at a given time. First 6 days of backward trajectories started at selected times: (b) 1200 UTC 23 Dec 2015, (c) 0000 UTC 27 Dec 2015, and (d) 0000 UTC 1 Jan 2016. Colors indicate the category, and gray dots show the trajectory starting points in the high Arctic.

extremely cold air masses prevailed, a plume of extremely warm air masses entered the high Arctic farther east via the Barents and Kara Seas (Fig. 1c). After 29 December 2015 and with the onset of the exceptional warm episode (Moore 2016; Cullather et al. 2016; Binder et al. 2017), the trajectory counts were solely dominated by extremely warm air masses. As opposed to the thermodynamic evolution of cold air masses, these warm air masses were the result of the combination of three distinct airstreams with different pathways (Fig. 1d) and thermodynamic evolution (see also Binder et al. 2017).

These comprise originally Arctic air masses that were diabatically heated and showed a net temperature increase ($\Delta\theta + \Delta T +$), diabatically cooled air masses with rising temperature (i.e., subsiding; $\Delta\theta - \Delta T +$), and diabatically cooled air masses with decreasing temperature and a subtropical origin ($\Delta\theta - \Delta T -$). The most important contribution comes from $\Delta\theta + \Delta T +$ and $\Delta\theta - \Delta T +$, whereas $\Delta\theta - \Delta T -$ clearly contributes less.

The findings from this example episode hint at three important characteristics regarding the categorization of the air masses associated with lower-tropospheric warm

FIG. 2. The ΔT – $\Delta\theta$ phase space diagrams showing trajectory counts for wintertime (a) cold and (b) warm extremes. Note the logarithmic color scale.

and cold extremes: First, the processes that can lead to the formation of an anomalously cold air mass are limited, requiring diabatic cooling and the absence of strong adiabatic warming due to descent. Hence, they are largely restricted to $\Delta\theta$ – ΔT –. Second, and in contrast to cold extremes, warm extremes can be to a large extent the result of the interplay of air masses with different thermodynamic characteristics, comprising air masses from $\Delta\theta$ + ΔT +, $\Delta\theta$ – ΔT +, and $\Delta\theta$ – ΔT –. Furthermore, it is apparent that air masses from several categories often co-occur. Third, because we are considering here lower-tropospheric air masses, ascending air masses cannot be an important contributor. Hence, air masses that are diabatically heated and at the same time experience a temperature increase ($\Delta\theta$ + ΔT –) contribute to neither warm nor cold extremes in a substantial way.

b. Climatological contributions of categories to cold and warm extremes

In Fig. 2 we consider climatological trajectory counts in ΔT – $\Delta\theta$ phase space diagrams for wintertime extreme cold and warm anomalies. The diagrams corroborate the conclusions drawn from the example episode, namely, that a broad range of combinations of adiabatic and diabatic temperature changes can lead to warm extremes ($\Delta\theta$ + ΔT +, $\Delta\theta$ – ΔT +, $\Delta\theta$ – ΔT –; Fig. 2b), whereas cold extremes are mostly associated with $\Delta\theta$ – ΔT – only (Fig. 2a). Contributions from $\Delta\theta$ + ΔT – are, as expected, the least important. Furthermore, the diagrams reveal remarkably high amplitudes of temperature changes reaching up to changes of 60 K within 10 days.

Quantifying the fractional contributions of each category, we find that more than 80% of the trajectories

associated with cold extremes in winter (60% in summer) belong to $\Delta\theta$ – ΔT – (Figs. 3a,c). In addition, about 10% of the trajectories in winter (25% in summer) belong to $\Delta\theta$ – ΔT +, which are also cooled diabatically but at the same time are subject to adiabatic warming due to subsidence that exceeds the temperature decrease associated with the diabatic cooling. Categories $\Delta\theta$ + ΔT + and $\Delta\theta$ – ΔT – are the most important ones for wintertime warm extremes contributing nearly 40% each and an additional about 20% of the trajectories belong to $\Delta\theta$ – ΔT – (Fig. 3b). In summer, in contrast, nearly 70% of the air masses belong to $\Delta\theta$ – ΔT – of subsiding trajectories with the remaining ones split between $\Delta\theta$ – ΔT – (about 20%) and $\Delta\theta$ + ΔT – (about 10%; Fig. 3d). Note that the relative contributions of the four categories remain fairly constant throughout the study period (cf. Fig. S2). A noteworthy, albeit weak variation is evident as a summertime increase of the contribution of subsiding air masses ($\Delta\theta$ – ΔT +) at the expense of $\Delta\theta$ – ΔT – in the last 13 years of the study period. In the following, we will only consider categories contributing at least 5% of all trajectories associated with cold and warm extremes, respectively. For cold extremes, these are $\Delta\theta$ – ΔT – and $\Delta\theta$ – ΔT –, whereas for warm extremes additionally also $\Delta\theta$ + ΔT – is considered.

Six-hourly trajectory counts for categories associated with warm extremes are moderately correlated (r between 0.3 and 0.6) with rather small differences between summer and winter. Hence, also from a climatological perspective, several categories, and, thus, air masses with a very different thermodynamic evolution, contribute jointly to individual warm extremes. Similarly, also $\Delta\theta$ – ΔT – and $\Delta\theta$ – ΔT – associated with cold extremes show a notable correlation ($r \approx 0.5$).

FIG. 3. Fraction of trajectories in each category for (a),(b) wintertime and (c),(d) summertime extreme (left) cold and (right) warm anomalies. Categories with a fraction of less than 5% (dashed line) are not considered in this study.

Interestingly, warm and cold extreme categories are only weakly anticorrelated ($r < -0.2$), indicating that the co-occurrence of warm and cold extremes in the high Arctic—as during the example episode (e.g., Fig. 1c)—may not be an exceptional case.

c. Thermodynamic characterization of air masses

Insight into the key features of the thermodynamic evolution, and in particular the role of vertical motion, is obtained from the fingerprints of each category in θ - T diagrams. Figures 4a and 4b depict for each category of warm and cold extreme trajectories the evolution of median θ and T during winter and summer. Adiabatic flow is represented by horizontal displacements with temperature changes directly related to vertical motion, that is, pressure changes. For diabatic flow, there exist two special cases, namely, isothermal diabatic flow in parallel to the vertical axis, which is characterized by a compensation of diabatic temperature changes by

opposing adiabatic temperature changes due to vertical motion, and isobaric diabatic flow along sloping lines of constant pressure. To relate the thermodynamic evolution to the origin of the air masses and the transport pathways, we will, as an additional diagnostics, also consider the spatial distribution of trajectories 10 and 3 days before the occurrence of the extreme temperature anomaly (Fig. 5), as well as maximum subsidence and maximum poleward motion of the air masses (Fig. 6).

1) WINTER (DJF)

Figure 4a shows the θ - T diagram for winter. Categories $\Delta\theta-\Delta T+$ and $\Delta\theta-\Delta T-$ associated with cold extremes are characterized by diabatic cooling at a rate of $1\text{--}1.5\text{ K day}^{-1}$, which is typical for uninhibited radiative cooling in the Arctic (e.g., Cavallo and Hakim 2013; Papritz and Spengler 2017). These air masses largely originate from within the Arctic poleward of 60°N

FIG. 4. The θ - T diagrams for extreme cold ($\Delta\theta - \Delta T +$, $\Delta\theta - \Delta T -$) and warm ($\Delta\theta + \Delta T +$, $\Delta\theta - \Delta T +$, $\Delta\theta - \Delta T -$) anomalies showing the evolution from $t = -240$ h to $t = 0$ h of medians of θ and T for (a) winter and (b) summer. Dots show values in 24-hourly intervals with $t = 0$ h indicated by black circles. Isobars are shown by gray dashed lines in intervals of 50 hPa.

(Figs. 5a,b). Maximum poleward motion and subsidence further indicate that throughout the entire 10-day period they mainly remain in the Arctic lower troposphere, that is, at less than 20° latitude and 200 hPa from their final location (Figs. 6a,b). Remaining located over sea ice or land (cf. densities for $t = -72$ days in Figs. 5a and 5b), they are sheltered from major diabatic heat sources such as the release of latent heat or surface sensible heat fluxes, which allows for a maximum temperature decrease due to radiative cooling. Differences between the two categories of trajectories exist in terms of their precise origin in the interior of the Arctic and the amount of subsidence they experience in the Arctic lower troposphere. As evident from the high trajectory density over Siberia (Fig. 5a), trajectories in $\Delta\theta - \Delta T +$ predominantly originate from the climatologically coldest regions of the Arctic over northern Siberia and to a lesser extent also the Canadian Arctic (cf. Fig. S1a), where they tend to remain until nearly 3 days prior to arrival in the high Arctic. Furthermore, they subside by about 150 hPa in the median (Fig. 6a), which leads to a nearly isothermal evolution with adiabatic warming only slightly overcompensating the T decrease associated with radiative cooling. In contrast, air masses in $\Delta\theta - \Delta T -$ are initially more widely spread across the Arctic with highest trajectory densities in the high Arctic (Fig. 5b) and less subsidence compared to air masses in $\Delta\theta - \Delta T +$ and the climatology (Fig. 6b).

The fingerprints in θ - T space of the three relevant categories associated with warm extremes, in contrast, strongly differ from each other. The $\Delta\theta + \Delta T +$ trajectories

are subject to diabatic (radiative) cooling at a rate of slightly more than 1 K day^{-1} throughout the first 6–7 days (Fig. 4a). The bulk of these air masses subsides by less than 200 hPa (Fig. 6a), yet adiabatic warming is sufficiently strong to compensate for the radiatively induced decrease of T such that the evolution is quasi-isothermal (Fig. 4a). Consequently, the initial thermodynamic evolution of these air masses bears similarities to that of $\Delta\theta - \Delta T +$ trajectories associated with cold extremes albeit with higher initial θ and T values. Interestingly, this analogy is further reflected in a similar origin of an important fraction of air masses in the Siberian Arctic (54%) and a smaller portion in the Canadian Arctic, north of the Labrador Sea (17%; Fig. 5c).² While maximum subsidence and maximum poleward motion are close to climatology (Fig. 6a), the distinctive feature of these air masses is the subsequent period of rapid diabatic heating that lasts for 48–72 h (Fig. 4a) and coincides with a shift of high trajectory densities toward open ocean with maxima at $t = -72$ h in the Barents Sea and in a band extending from the Labrador Sea around the southern tip of Greenland into the Nordic seas (Fig. 5c). The polar, lower-tropospheric origin of the air masses and their fingerprint in θ - T space, characterized by continuous radiative cooling

² Trajectories are counted in box 60° – 150°E , 60° – 80°N for Siberia and in box 120°W – 55°E , 60° – 80°N for the Canadian Arctic, and fractions are given with respect to the total count of wintertime $\Delta\theta + \Delta T +$ trajectories.

FIG. 5. Air mass origin maps showing the probability of finding a trajectory of a specific category at a certain location 10 days ($t = -240$ h; shading) and 3 days [$t = -72$ h; dashed; contours at 1, 4, and 16% (10^5 km^2) $^{-1}$] prior to the occurrence of the extreme temperature anomaly for (a)–(e) wintertime and (f)–(k) summertime extreme cold anomalies in (a), (b), (f), and (g) and warm anomalies in (c)–(e) and (h)–(k). Note the logarithmic color scale.

FIG. 6. (a),(c) Maximum subsidence (pressure at $t = 0$ h minus minimum pressure) and (b),(d) maximum poleward motion (latitude at $t = 0$ h minus minimum latitude) of air masses contributing to extreme cold ($\Delta\theta-\Delta T+$, $\Delta\theta-\Delta T-$) and warm ($\Delta\theta+\Delta T+$, $\Delta\theta-\Delta T+$, $\Delta\theta-\Delta T-$) anomalies during (top) winter (DJF) and (bottom) summer (JJA). Additionally shown is the respective climatology comprising all air masses that reach the lower troposphere of the high Arctic in the given season. Whiskers indicate the 10th–90th percentile range.

with moderate subsidence followed by a phase of rapid diabatic heating, are akin to the properties of cold air outbreak air masses (Papritz and Spengler 2017). As we will see below, many of the extremely warm air masses in this category are indeed related to cold air outbreaks.

The thermodynamic evolution of trajectories in $\Delta\theta-\Delta T+$ is dominated by pronounced subsidence by more than 300 hPa in the median (Fig. 6a) and associated adiabatic warming throughout the entire 10 days (Fig. 4a). Despite decreasing θ due to radiative cooling,

this results in a substantial net increase of T . The bulk of these air masses has its origin south of the high Arctic, farther equatorward than the climatological air mass (Fig. 6b), with density maxima over the Nordic seas and the Barents Sea throughout the 10-day period (Fig. 5d). The thermodynamic evolution dominated by subsidence indicates a close linkage of the occurrence of such air masses with atmospheric blocking (e.g., Pfahl and Wernli 2012).

Finally, $\Delta\theta-\Delta T-$ trajectories are confined to the lower troposphere (Fig. 6a), experiencing considerably

less subsidence (Fig. 4a). Instead, they originate over western Eurasia (Fig. 5e) from temperate latitudes and they move between 30° and 40° latitude poleward (Fig. 6b). A further, albeit weaker branch of trajectories originates in the Pacific and enters the high Arctic near Bering Strait (cf. trajectory densities for $t = -72$ h in Fig. 5e). The $\Delta\theta - \Delta T -$ air masses are characterized by high initial θ and T values (Fig. 4a), which ensures that, despite radiative cooling on their journey north (Fig. 4a), the air masses are still sufficiently warm to induce an extreme warm anomaly when they arrive in the high Arctic. The southerly origin and warm initial temperature suggest a close relationship with warm extremes in the high Arctic induced by an enhanced poleward heat flux.

2) SUMMER (JJA)

In summer, the fingerprints of the individual categories in θ - T space are qualitatively similar to those in winter (Fig. 4b). Temperature anomalies, however, are smaller in magnitude along with overall slightly smaller changes in both θ and T . This applies most notably to the categories of diabatically heated ($\Delta\theta + \Delta T +$) and subsiding ($\Delta\theta - \Delta T +$) trajectories associated with extreme warm anomalies. In summer they are subject to weaker diabatic heating and subsidence-induced adiabatic warming, respectively, than in winter. Furthermore, all air masses, with the exception of $\Delta\theta - \Delta T -$ air masses associated with extreme warm anomalies, originate from the Arctic and remain confined to within about 15° latitude from the pole in the median (Figs. 5f-i and 6d). This shift toward a more high-latitude origin of air masses is in line with climatological transport studies into the Arctic troposphere (Stohl et al. 2006; Orbe et al. 2015). During summer, the sea ice covered portion of the Arctic Ocean is the climatologically coldest part of the Arctic (Fig. S1b). Thus, air masses associated with extreme cold anomalies originate from there and predominantly remain in this area (Figs. 5f,g and 6d). Interestingly, also the poleward motion of air masses leading to extreme warm anomalies is clearly weaker in summer than in winter, yet it exceeds the climatological poleward motion (Fig. 6d). While the origin of the diabatically heated ($\Delta\theta + \Delta T +$) and to a lesser extent also the subsiding ($\Delta\theta - \Delta T +$) air masses is fairly close to the pole and by that similar to that of air masses associated with summertime extreme cold anomalies, $\Delta\theta + \Delta T +$ air masses move southward throughout the 10-day period before returning back into the high Arctic (Figs. 5h,i and 6d). The $\Delta\theta - \Delta T -$ air masses, finally, are initially spread out widely over the subarctic and with notable contributions from the North Pacific. The maximum poleward motion is between 10° and 40°

latitude (Figs. 5k and 6d), and, therefore, reveals a remarkably wide spread.

d. Physical nature of diabatic heating processes

The category of diabatically heated air masses ($\Delta\theta + \Delta T +$) with Arctic origin contributes about 40% of the air masses associated with wintertime extreme warm anomalies. This suggests that in addition to the poleward transport of warm and humid air masses from southerly latitudes, other more local processes also play an important role in the formation of warm extremes in the high Arctic. Indeed, the importance of originally polar air masses that are diabatically heated for warm extremes in the high Arctic has been noted before. Binder et al. (2017) found that 37% of the air masses contributing to the previously discussed extreme warm event at the turn of 2015/16 were of Arctic origin and experienced strong diabatic heating. They identified this airstream as an outbreak of polar cold air masses over the Labrador Sea where the air masses were exposed to intense upward surface heat fluxes that lead to a rapid diabatic warming. Subsequently, the air masses were advected around the tip of Greenland and injected into the high Arctic near Fram Strait (see also Fig. 1d).

To shed light on the physical nature of the diabatic heating in warm extreme $\Delta\theta + \Delta T +$ air masses, we first consider in Fig. 7a the spatial distribution of the air masses at the time of the largest 24-hourly θ increase. Three preferential regions are apparent, namely, the Labrador Sea, the Greenland and Iceland Seas, and the Barents Sea. These regions are well-known hot spots of marine cold air outbreaks (Kolstad et al. 2009; Fletcher et al. 2016; Papritz and Spengler 2017), as confirmed here by the high winter mean frequency of the air-sea potential temperature difference with respect to 900 hPa exceeding 4 K ($\theta_{\text{SST}} - \theta_{900} > 4$ K; for details about cold air outbreak identification, see, e.g., Papritz et al. 2015). Furthermore, the air masses are located near the sea surface at a median pressure of 980 hPa (interquartile range: 44 hPa). They attain a maximum air-sea potential temperature difference of about 8 K in the median, which is unmatched by any other category (Fig. 7c) and indicates strong cold air outbreaks, which are generally associated with intense upward fluxes of sensible and latent heat, as well as convective overturning (Papritz and Spengler 2017). These maximum values of $\theta_{\text{SST}} - \theta_{900}$ are reached in the same regions as the largest 24-hourly θ increase, though slightly closer to the sea ice edge and land-sea boundaries where the air masses are first exposed to open ocean (Fig. 7b). Based on these findings, we conclude that the diabatic heating in $\Delta\theta + \Delta T +$ warm air masses is to a large extent the result of

FIG. 7. (a),(b) Probability maps for $\Delta\theta + \Delta T +$ trajectories associated with extreme warm anomalies at the time of (a) maximum 24-h mean diabatic heating rate and (b) maximum air-sea potential temperature difference ($\theta_{\text{SST}} - \theta$). Red contours in (a) and (b) indicate the wintertime climatological frequency of cold air outbreaks as identified from the air-sea potential temperature difference with respect to the 900-hPa level exceeding 4 K ($\theta_{\text{SST}} - \theta_{900} > 4 \text{ K}$; contours from 20% in intervals of 10%). Details about the identification of cold air outbreaks can be found in Papritz and Spengler (2017). (c) Boxplot of maximum air-sea potential temperature difference ($\theta_{\text{SST}} - \theta$) along trajectories of the different categories. Whiskers indicate the 10th–90th percentile range.

the upward heat fluxes during cold air outbreaks and the associated release of latent heat in convective clouds.

e. Moisture intrusions associated with warm extreme air masses

Given the strong impact of water vapor on the radiation budget, the question how much water vapor the lower-tropospheric extreme air masses import into the high Arctic is of paramount importance. Unsurprisingly, the air masses associated with cold extremes carry considerably less moisture than the climatological lower-tropospheric air masses in the high Arctic (Figs. 8a,c). In winter, warm air masses, in contrast, have about 3–4 times higher specific humidity than climatological air masses with interestingly only small differences between the categories (Fig. 8a). Specifically, cold air outbreak air masses ($\Delta\theta + \Delta T +$) tend to inject the same amount of moisture as air masses originating from lower latitudes ($\Delta\theta - \Delta T -$). In addition, the originally very dry air masses of $\Delta\theta - \Delta T +$ rapidly pick up moisture by mixing with the surrounding air and via ocean evaporation as they subside into the planetary boundary layer. This process is likely favored by the fact that the subsidence preferentially occurs over the Barents and Kara Seas from where the air masses move over the ice-free Barents Sea and the Nordic seas (cf. Fig. 5d), which allows them to equilibrate their moisture deficit. Even though the poleward-moving air masses ($\Delta\theta - \Delta T -$) do not have higher specific humidity at $t = 0 \text{ h}$, they have larger maximum moisture content before reaching the

high Arctic than the other categories (Fig. 8b). An important part of the moisture is, however, lost during the poleward journey of the air masses (Fig. 8a), most likely due to precipitation. The associated latent heat release compensates part of the cooling effect of longwave radiation and contributes to the final warm anomaly.

Moisture content is considerably larger in summer in line with the Clausius–Clapeyron relationship, yet the relative differences in the distributions are similar to winter (Figs. 8c,d). A noteworthy change is that differences between warm air masses and climatology are smaller in summer than in winter, which may be owed to the more local origin of the air masses. In addition, $\Delta\theta - \Delta T -$ air masses carry somewhat more moisture than the other categories, which gives them a more distinctive role in shaping humidity anomalies in the high Arctic.

f. Relative importance of anomalous transport and diabatic processes for the formation of extreme temperature anomalies

We have seen that diabatic processes play a key role in the formation of lower-tropospheric temperature extremes in the high Arctic. However, transport, for example, subsidence, can also contribute to the formation of temperature anomalies. The relative importance of diabatic and transport processes can be quantified in the potential temperature framework by considering the deviation of potential temperature from its local climatology θ_c . Specifically, the potential temperature

FIG. 8. (a),(c) Specific humidity at $t = 0$ h and (b),(d) the maximum specific humidity along the trajectories for the climatology and the relevant categories contributing to cold and warm extremes for (top) DJF and (bottom) JJA. Whiskers indicate the 10th–90th percentile range.

anomaly $\theta^* = \theta - \theta_c$ of an air mass at time $t = 0$ h is the result of $\theta^*(t = -240$ h) [initial anomalies (INI)] and changes of θ and θ_c along the air parcel trajectory, that is,

$$\begin{aligned} \theta^*(t = 0 \text{ h}) &= \underbrace{\theta^*(t = -240 \text{ h})}_{\text{INI}} \\ &+ \underbrace{\theta(t = 0 \text{ h}) - \theta(t = -240 \text{ h})}_{\text{DIAB}} \\ &+ \underbrace{\theta_c(t = -240 \text{ h}) - \theta_c(t = 0 \text{ h})}_{\text{TRANS}}, \end{aligned} \quad (3)$$

where DIAB denotes diabatic processes and TRANS represents transport from a region with different climatology. Thus, a positive θ^* can emerge due to diabatic

heating and the transport from a climatologically warmer into a colder region. Analogously, a negative θ^* can emerge due to diabatic cooling and transport from a climatologically colder into a warmer region. In the following, it is our goal to pinpoint systematic differences in INI, TRANS, or DIAB compared to *typical* Arctic air masses that account to the formation of warm and cold anomalies. For that purpose, we will consider departures of these terms for each category from the medians of the climatological distributions. They reveal, for instance, whether air masses associated with extreme cold anomalies experience stronger radiative cooling than the median air mass.

Figure 9 depicts the climatological contributions of the individual terms in Eq. (3) for all trajectories from the high Arctic, including air masses that are not

FIG. 9. Climatological distributions of final (ANOM $t = 0$ h) and initial potential temperature anomalies (ANOM $t = -240$ h), transport (TRANS), and diabatic processes (DIAB) for (a) DJF and (b) JJA. Whiskers indicate the 10th–90th percentile range. Note the smaller temperature range for summer compared to winter.

necessarily associated with a strong potential temperature anomaly. While INI is relatively small, diabatic (radiative) cooling (DIAB) of in the median 15 K in winter (Fig. 9a) and 8 K in summer (Fig. 9b) is balanced by transport (TRANS) from climatologically warmer regions toward the high Arctic. Furthermore, consistent with the relatively lower spread of initial and final anomalies compared to TRANS and DIAB, the latter two terms are anticorrelated with $r = -0.65$ (DJF) and $r = -0.79$ (JJA). This balance between cooling and transport is the default mode for air masses arriving in the high Arctic lower troposphere.

Normalizing departures from climatology by the magnitude of the final potential temperature anomaly [$\theta^*(t = 0 \text{ h})$] provides a measure for the *relative contributions* of deviations of INI, DIAB, and TRANS as depicted in Fig. 10. The horizontal and vertical axes indicate the relative contributions of transport and diabatic processes, respectively, and quadrants highlight whether diabatic processes or transport dominate. If transport and diabatic processes fully account for the final anomaly, the values align on the thick gray lines with the lower line representing negative and the upper line positive anomalies. Departures from these lines

along the diagonal (black line) toward the center indicate a positive contribution from the initial anomaly, and departures away from the center indicate a negative contribution.

Wintertime extreme cold anomalies form predominantly due to anomalous transport and, for $\Delta\theta - \Delta T+$, a pronounced initial anomaly also contributes (Fig. 10a), while enhanced radiative cooling with respect to climatology occurs for only about 25% of the trajectories. For $\Delta\theta - \Delta T+$ air masses, the initial cold anomaly typically accounts for about 50% of the final anomaly. Furthermore, their median radiative cooling is slightly weaker than for climatological air masses, which may be related to the higher initial altitude of $\Delta\theta - \Delta T+$ air masses where water vapor is less abundant. These findings indicate that in winter, extremely cold $\Delta\theta - \Delta T+$ air masses have an exceptionally long residence time in the high Arctic. For $\Delta\theta - \Delta T-$ air masses, in contrast, the initial anomaly is less important and is, in fact, positive for more than 25% of the air masses.

For extreme cold anomalies in summer, there is a notable widening of the distributions in terms of the relative importance of diabatic cooling and transport, as well as a shift toward stronger contributions from

FIG. 10. Contributions of transport and diabatic processes normalized by the magnitude of the potential temperature anomaly at $t = 0$ h for (a) DJF and (b) JJA. The thick (thin) colored lines indicate the 25th–75th (10th–90th) percentile range for each of the categories. Note that before computing the contributions for each term the median from the seasonal climatology (Fig. 9) has been removed such that the terms represent departures from the “median trajectory.” Isolines of the sum of the transport and diabatic terms are indicated by gray lines (solid and dashed) from -1.5 to 1.5 in intervals of 0.5 . For trajectories located on the thick gray lines (values -1 and 1), transport and diabatic processes fully account for the final positive (upper line) or negative (lower line) anomaly. Displacements along the diagonal (black line) toward (away from) the center reveal positive (negative) contributions of the initial anomaly. The quadrants indicate four different regimes dominated either by diabatic processes (white shading) or transport (gray shading).

enhanced diabatic cooling (Fig. 10b). This is consistent with the fact that the coldest portions of the Arctic in summer are located in the high Arctic where sea ice remains (Fig. S1b), which implies that transport is less efficient than in winter when the coldest regions are outside the high Arctic (Fig. S1a). In addition, the warmer temperatures in summer go along with larger variations of the atmospheric water content, which makes the efficiency of radiative cooling more variable and, thus, allows air masses to experience substantially more cooling than the average air mass.

The contribution from initial anomalies to wintertime warm extremes can be of either sign and is generally relatively smaller than that of the combined effect of diabatic processes and transport, as evident from the distributions’ approximate centering on the thick gray line and the narrow spread perpendicular to it (Fig. 10a). For the most part, the three categories separate into diabatically dominated ($\Delta\theta+\Delta T+$) and transport-dominated ($\Delta\theta-\Delta T+$, $\Delta\theta-\Delta T-$) regimes. The diabatic category ($\Delta\theta+\Delta T+$) has only weakly negative contributions from anomalous transport, whereas 25% of the other two categories ($\Delta\theta-\Delta T+$, $\Delta\theta-\Delta T-$) comprise air masses with anomalous diabatic

heating exceeding the contribution from anomalous transport. In summer, the most notable difference (Fig. 10b) is a shift of the category of poleward-moving air masses ($\Delta\theta-\Delta T-$) toward more positive contributions from anomalous diabatic processes at the expense of anomalous transport. Potential reasons for the enhanced contributions from the diabatic term are likely the stronger dampening of longwave cooling due to cloud cover, incident shortwave radiation, and latent heat release—the latter, however, limited by the lack of ascent. By and large, $\Delta\theta-\Delta T+$ and $\Delta\theta-\Delta T-$ nevertheless remain in the transport-dominated regime.

When during the 10-day evolution of the air masses do the temperature anomalies typically emerge? To address this question, we consider in Fig. 11 the evolution of the medians of θ and θ_c traced along the trajectories and depict them in the same fashion as the evolution of θ and T in the previously discussed θ – T diagrams (Fig. 4). Note that θ_c represents the local climatology interpolated to the current trajectory position. In this diagram, deviations from the diagonal to the upper left and lower right indicate positive and negative θ^* , respectively. Displacements along the vertical axis denote

FIG. 11. The $\theta - \theta_c$ diagrams for extreme cold ($\Delta\theta - \Delta T+$, $\Delta\theta - \Delta T-$) and warm ($\Delta\theta + \Delta T+$, $\Delta\theta - \Delta T+$, $\Delta\theta - \Delta T-$) anomalies showing the evolution from $t = -240$ h to $t = 0$ h of medians of θ and θ_c for (a) winter and (b) summer. Dots show values in 24-hourly intervals with $t = 0$ h indicated by black circles. Isolines of θ^* are shown by gray dashed lines in intervals of 5 K.

diabatic temperature changes, whereas displacements along the horizontal axis from right to left represent transport into a climatologically colder region and likewise into a climatologically warmer region for displacements from left to right.

Cold extreme air masses are associated with a slow but gradual increase of θ^* as they are cooled radiatively (Fig. 11a). As seen before, during winter these air masses largely originate from the climatologically coldest parts of the Arctic. Accordingly, transport toward their final destination further amplifies the anomaly during the last 2–3 days, a feature that is absent in summer as the high Arctic is already the climatologically coldest region (Fig. 11b).

For the transport-dominated categories associated with wintertime warm extremes ($\Delta\theta - \Delta T+$, $\Delta\theta - \Delta T-$), the median air masses remain close to climatology with $\theta^* < 2.5$ K until approximately 2 days ($\Delta\theta - \Delta T+$) and 4–5 days ($\Delta\theta - \Delta T-$), respectively, before reaching the peak value (Fig. 11a). Thus, the emergence of an extreme warm anomaly due to vertical transport ($\Delta\theta - \Delta T+$) is more rapid than due to horizontal poleward transport ($\Delta\theta - \Delta T-$). Initially, diabatically heated air masses ($\Delta\theta + \Delta T+$) cool at approximately the same rate as they subside into a climatologically colder region (cf. Fig. 4a). As noted before, their initial evolution is, thus, similar as for cold extremes. It is only 3 days before arriving in the high Arctic that they first move into a climatologically warmer region, while at the same time diabatic heating leads to a rapid increase of θ . During

the final 1–2 days, θ^* rises rapidly due to intense diabatic heating and finally also the transport of the air masses back into the climatologically colder high Arctic. The fingerprints of the air masses in $\theta - \theta_c$ space in summer are qualitatively similar to winter (Fig. 11b), albeit with a slightly slower emergence of the anomalies.

4. Linkage to dynamical flow features

In this section we seek to identify large-scale flow anomalies that lead to the formation of near-surface temperature extremes in the high Arctic. For that purpose, we compile composites of 500-hPa geopotential height, frequency of cyclones, and dynamical blockings for the three days preceding the 100 most intense extreme cold and warm events in the high Arctic as defined in the methods (section 2). Given that several trajectory categories can contribute simultaneously to temperature extremes (see section 3b), we do not define separate events for each of the categories but instead we consider the total counts of warm and cold extreme trajectories. To test the robustness of the composites, we perform a Monte Carlo resampling of the events with repetitions and 1000 iterations, which yields distributions of the sample mean anomalies. If the magnitude of the sample mean anomalies exceeds the 25th–75th and 10th–90th percentile ranges of these distributions, we consider composite anomalies—defined with respect to the corresponding seasonal mean—as robust and highly robust, respectively.

FIG. 12. Composite of 500-hPa geopotential height anomalies for the three days prior to the top 100 (a),(c) cold and (b),(d) warm events for (top) winter and (bottom) summer. Dark blue contours indicate the climatological values of 500-hPa geopotential height in intervals of 100 m. Light blue hatched and cross-hatched areas indicate regions where the amplitude of the anomaly exceeds the 25th–75th and 10th–90th percentile range, respectively, according to a Monte Carlo resampling of events with repetition.

a. 500-hPa geopotential height

Wintertime cold and warm extremes are preceded by highly robust dipoles of 500-hPa geopotential height anomalies (Figs. 12a,b). In the case of cold extremes, a negative anomaly is located over the high Arctic with its center slightly shifted toward Siberia and a positive,

weaker anomaly is found over northern Canada. This configuration indicates a slight shift of the center of the tropospheric polar vortex toward the pole and Siberia and an overall strengthening of the cyclonic midtropospheric flow that tends to shelter the high Arctic from meridional exchange processes with lower latitudes. In the case of warm extremes, the dipole is reversed in sign

and the centers of the anomalies are somewhat shifted compared to the centers associated with cold extremes. Specifically, the negative anomaly is centered over Greenland and the positive anomaly over the Barents and Kara Seas. As a result, the midtropospheric flow prior to warm extremes is conducive to enhanced poleward transport of air masses over the Nordic seas. Overall, the 500-hPa geopotential height anomalies closely resemble those identified by Messori et al. (2018) associated with surface warm and cold extremes. An important difference is, however, that for cold events they did not find a positive anomaly over Canada and Greenland.

In summer, in contrast, the dominant anomalies show a highly robust monopole structure with the centers in both cases located over and confined to the high Arctic (Figs. 12c,d). As in winter, summertime cold extremes are preceded by a strengthening of the midtropospheric cyclonic circulation over the pole that shelters the high Arctic from airmass exchanges with lower latitudes. Furthermore, the strongly positive midtropospheric geopotential height anomaly prior to warm extremes suggests an important role for anticyclones and blockings in the high Arctic—an aspect that will be confirmed below.

b. Cyclone and blocking frequencies

Here, we link the previously discussed anomalies of 500-hPa geopotential height to modulations in the occurrence frequencies of mid- and high-latitude cyclones and dynamical blocking. These dynamical flow features play an important role in lower-tropospheric airmass transports and the formation of near-surface temperature extremes (Pfahl 2014). In addition, blockings are important drivers of large-scale subsidence (e.g., Pfahl and Wernli 2012). Figures 13 and 14 depict anomalies of cyclone and blocking frequencies, respectively, as well as the respective climatologies.

Let us first consider frequency anomalies associated with wintertime cold extremes. The strengthening of the tropospheric polar vortex is directly reflected in a higher-than-normal cyclone frequency in the Barents, Kara, and Laptev Seas (Fig. 13a), which provides favorable conditions for the advection of cold air masses from Siberia into the high Arctic. Furthermore, the occurrence of blocking is suppressed over most of the high Arctic (Fig. 14a). In addition to these highly robust patterns in the high Arctic, the storm track over the North Atlantic is shifted toward the southwestern parts of the basin with strongly enhanced cyclone frequencies near Newfoundland and a suppression of cyclone activity between Greenland and Iceland and in the western Nordic seas (Fig. 13a). Such a flow configuration is

associated with a reduced lower-tropospheric poleward transport of midlatitude, maritime air masses across the North Atlantic and the Nordic seas. Finally, the previously identified positive geopotential height anomaly over northern Canada (Fig. 12a) matches with an area of enhanced blocking frequency that extends from northern Canada toward the North Atlantic (Fig. 14a). This is, however, not a highly robust feature, which indicates that it occurs only with some extreme cold events and is not a prerequisite for their occurrence. We hypothesize that this also the reason why no positive 500-hPa geopotential height anomaly was found by Messori et al. (2018). Yet, the presence of a blocking anticyclone over northern Canada favors the transport of cold continental air masses into the high Arctic and can, thus, be of key importance for some of the cold events.

In the case of wintertime warm extremes, the North Atlantic storm track shows opposite anomalies compared to cold extremes. Specifically, the storm track is shifted poleward with reduced cyclone frequency in a band that stretches from Newfoundland across the North Atlantic into the eastern Nordic seas and the Barents Sea (Fig. 13b). At the same time, a strongly enhanced cyclone frequency is found near Iceland and in the western Nordic seas, providing favorable conditions for the poleward transport of maritime, warm air masses ($\Delta\theta - \Delta T^-$) and of diabatically warmed cold air outbreak air masses from the Labrador Sea ($\Delta\theta + \Delta T^+$). A feature of particular interest is the enhanced cyclone frequency north of Greenland. Cyclogenesis in that region in phase with the poleward flow of air masses over the Nordic seas has been recognized by Messori et al. (2018) as an important prerequisite for the injection of air masses from the North Atlantic all the way into the high Arctic. In addition, these positive storm-track anomalies are favorably located for injecting air masses with low potential vorticity into the Arctic upper troposphere, contributing to ridge amplification and eventually promoting the formation of blocking anticyclones poleward and downstream (see below; Pfahl et al. 2015; Wernli and Papritz 2018). In fact, blocking frequency is strongly enhanced in the Barents and Kara Seas—a region where the climatological frequency is very low—while it is reduced over Greenland (Fig. 14b). The positive blocking frequency anomaly in the Barents and Kara Seas coincides with the region where most of the subsiding air masses reside before they move into the high Arctic (cf. Fig. 5d). Moreover, blocks, that are located sufficiently deep in the high Arctic, can access cold air masses over Siberia and advect them on their southern flank across Novaya Zemlya over the Barents Sea, where they are warmed

FIG. 13. As in Fig. 12, but for the frequency of cyclones. Blue contours indicate the climatological frequency of cyclones in intervals of 8%.

and subsequently advected back into the high Arctic by the anticyclonic flow.

Summertime frequency anomaly patterns are generally fuzzier, and anomalies are less intense and less robust (Figs. 13c,d and 14c,d). Consistent with the earlier finding that the origin of air masses contributing to cold and warm extremes in summer is more strongly confined to high latitudes than in winter, highly robust frequency anomalies are almost exclusively located in the high

Arctic. Specifically, the strengthened tropospheric polar vortex prior to cold extremes is reflected in a notable reduction of the frequency of blocking in the high Arctic (Fig. 14c), whereas cyclone frequency anomalies show a less clear picture with only a small region with enhanced frequency near the pole (Fig. 13c). For warm extremes, however, highly robust anomalies in the high Arctic are more pronounced. A strongly reduced frequency of cyclones is found over the pole (Fig. 13d), while blocking

FIG. 14. As in Fig. 13, but for dynamical blocking. Blue contours indicate the climatological frequency of dynamical blocks in intervals of 2%.

frequency shows a large-scale positive anomaly extending over all of the high Arctic (Fig. 14d), which coincides with the region of origin of subsiding air masses (Fig. 5i). In addition, a belt comprising several robust regions of enhanced cyclone frequencies surrounds the high Arctic (Fig. 13d). This may be relevant for warm events with important contributions from diabatically heated air masses that move off the ice, are warmed, and move back into the high Arctic ($\Delta\theta + \Delta T+$),

as well as the import air masses from the surrounding continents ($\Delta\theta - \Delta T-$).

5. Conclusions

Using 10-day kinematic backward trajectories, we investigated the thermodynamic characteristics and evolution of air masses that lead to the formation of extreme warm and cold anomalies (top 5% of all air

masses) in the high Arctic ($\geq 80^\circ\text{N}$) lower troposphere (lowermost 100 hPa) during winter (DJF) and summer (JJA). Based on a categorization of air masses according to their thermodynamic evolution in $\Delta\theta$ – ΔT phase space, we quantified the relative importance of transport (e.g., subsidence and poleward transport) and diabatic processes (e.g., surface sensible heat fluxes). Composites for the 100 most intense extreme warm and cold events provided further insight into the linkage of air mass transport and evolution to dynamical, large-scale flow features, namely, mid- and high-latitude cyclones and blockings.

a. Synthesis

Synthesizing the results from the various analyses, our principal findings regarding the formation of *extreme warm anomalies* are as follows:

- 1) For about 60% and 90% of the air masses in winter and in summer, respectively, the formation of warm anomalies is dominated by vertical and to a lesser extent also horizontal air mass transport from regions with a higher climatological potential temperature. For the remaining air masses, the potential temperature anomaly emerges predominantly due to enhanced diabatic heating.
- 2) Subsidence-induced adiabatic warming is by far the most important process for air masses with transport governing the formation of the warm anomaly. This is particularly true in summer, where 70% of all warm extreme air masses are due to subsidence. Most of these air masses originate in the Arctic midtroposphere, and the subsidence is predominantly driven by anomalous blocking over the Barents, Kara, and Laptev Seas in winter and in the high Arctic in summer.
- 3) In both seasons, lower-tropospheric poleward transport of warm air masses from lower latitudes contributes only about 20%. Such transport is favored in winter by a poleward shift of the North Atlantic storm track as reflected in enhanced cyclone frequency near Iceland and in the western parts of the Nordic seas, as well as north of Greenland concomitant with a reduction near Newfoundland and south of Greenland. The origin of these air masses in summer is largely confined to the Arctic, the Nordic seas and the North Pacific, in part linked to a shift of cyclone activity from the high Arctic toward the marginal seas.
- 4) Air masses predominantly affected by diabatic heating are a wintertime phenomenon contributing 40% of all air masses in DJF and 10% in JJA. They are related to marine cold air outbreaks in the Barents and Nordic seas, as well as the Labrador Sea that form as the result of the advection of radiatively

cooled polar air masses over the relatively warm ocean surface, where they are heated by surface sensible heat fluxes.

As a contrasting case to warm extremes, we have identified the following particularities associated with *extreme cold anomalies*:

- 1) Cold extremes do not result from enhanced instantaneous (radiative) cooling rates of Arctic air masses but form due to uninterrupted cooling over a prolonged period resulting from an enhanced residence time of air masses in the Arctic.
- 2) They are preceded by a strengthening of the tropospheric polar vortex, a reduction of high-latitude blocking, and in winter also a southward shift of the North Atlantic storm track with a strong reduction of cyclone frequencies near Iceland and in the western Nordic seas—a configuration opposed to that associated with warm extremes. These characteristics of the large-scale flow shelter the high Arctic from meridional air mass exchanges. During winter, in addition, enhanced frequencies of cyclones in the Kara and Laptev Seas and of blocking over Canada favor the transport of cold air masses from the climatologically coldest regions over Siberia and Canada into the high Arctic.

b. Concluding remarks

Our findings emphasize the predominantly rather high-latitude origin of air masses and the relevance of atmospheric dynamical and thermodynamic processes in the Arctic for the formation of lower-tropospheric extreme warm anomalies. These processes include, in particular, subsidence-induced adiabatic compression, as well as diabatic warming associated with air–sea heat exchanges. In spite of the rapid warming of the Arctic, no systematic changes in the relative importance of these processes have occurred in the study period (see Fig. S2). An exception is a notable but relatively small increase in the contribution of subsiding air masses to summertime warm extremes after 2005, which is likely related to the increase of the frequency of anticyclonic circulation patterns and blocking in the high Arctic (Ding et al. 2017; Wernli and Papritz 2018).

In fact, anomalous blocking poleward of the main climatological blocking regions is found to be an important driver of subsidence and of the long-range transport of polar air masses over the ocean, where they are heated by surface sensible heat fluxes. While these findings confirm the results of Binder et al. (2017) from a climatological perspective, the importance of subsidence is also in agreement with what is known

about the formation of near-surface temperature anomalies during midlatitude heat waves (Pfahl and Wernli 2012; Bieli et al. 2015; Quinting and Reeder 2017; Zschenderlein et al. 2018, 2019). Yet, our findings complement the dominating view that warm extremes in the high Arctic are essentially the result of the injection of already warm air masses from midlatitudes. As quantified in this study, such injections constitute only around 20% of all extremely warm air masses in the lowermost 100 hPa. Nevertheless, injections of warm and humid air masses are likely of a higher importance for temperature extremes at higher altitude as well as for cloud formation in the Arctic, which directly affects the surface energy balance via their radiative impact (D.-S. R. Park et al. 2015; H.-S. Park et al. 2015; Woods and Caballero 2016). This view is supported by the observation that intense poleward moisture transports do not peak near the surface but further aloft (Laliberté and Kushner 2014; Woods and Caballero 2016; Dufour et al. 2016; Naakka et al. 2019)—a consequence of the isentropic slope at the edge of the polar dome that requires poleward-moving air masses to ascend (Bozem et al. 2019). In addition, injections of air masses with low potential vorticity from lower latitudes into the Arctic upper troposphere play an important role for the amplification of upper-tropospheric ridges and the formation of blocks, which in turn are important drivers of subsidence (Pfahl et al. 2015; Ding et al. 2017; Wernli and Papritz 2018).

Synoptic flow anomalies preceding warm and cold extremes reveal similar spatial structures with opposite signs over the North Atlantic and in the Arctic. In terms of 500-hPa geopotential height anomalies, they are characterized in winter by a dipole in the Atlantic–European sector poleward of 60°N and in summer by a monopole that is largely confined to the high Arctic. Specifically, one pole of the winter pattern is located near Iceland, going along with modulations in cyclone frequency there and in the Nordic seas, and modulations of opposite sign farther southward, thus, projecting onto the North Atlantic Oscillation. The second and opposite pole is located poleward in the eastern Arctic and is related to blocking in the Barents, Kara, and Laptev Seas. This is in line with previous studies noting the importance of blocking preceding warm extremes in the Arctic (Woods et al. 2013; Liu and Barnes 2015; H.-S. Park et al. 2015). Yet, as we find here, the blocking occurrence is farther poleward than the canonical Ural blocking, in a region where climatological blocking frequencies are generally low (Pfahl et al. 2015). The notion that it is the anomalous nature of blocking that matters for warm extremes is further fueled by strongly enhanced blocking frequencies in the high Arctic prior to summertime warm extremes. Furthermore, the absence

of blocking and, thus, the more cyclonic circulation in the Arctic provides favorable conditions for the sustained cooling of air masses and the formation of an extremely cold air mass. Given the emphasis on blocking, we suggest that future work should focus on the dynamical mechanisms that govern the formation and evolution of anomalous blocking in the Barents, Kara, and Laptev Seas, as well as in the high Arctic itself. In this context, a promising avenue is to investigate if and how anomalous blocking is linked to storm-track anomalies, for example, via the injection of air masses with low potential vorticity into the Arctic upper troposphere (see also Wernli and Papritz 2018).

Because of the radiative forcing associated with water vapor and clouds in the Arctic, another important aspect concerns the transport of moisture into the high Arctic during warm extremes. While specific humidity of air masses associated with warm extremes is consistently above climatological values, we found that humidity values are very similar, irrespective of airmass origin and the dominating mechanisms that shape the warm anomalies, that is, poleward transport of already warm air masses, subsidence, or diabatic heating. Concomitant with the rather high-latitude origin of many of these air masses, this underlines that in addition to the classical injection of humid air masses from lower latitudes (e.g., Woods and Caballero 2016), the uptake of moisture locally in the Arctic also plays an important role (see also Dufour et al. 2016), which in recent years has increased substantially (Boisvert et al. 2015). An important goal of future research is, therefore, to quantify the contributions of local and remote moisture sources during warm temperature extremes in the high Arctic and to explore how moisture transport and uptake are linked to the thermodynamic evolution of air masses.

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