Midlatitude Moisture Contribution to Recent Arctic Tropospheric Summertime Variability*

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ABSTRACT

The dynamics of late summer Arctic tropospheric heat content variability is studied using reanalyses. In both trends and interannual variability, much of the August heat content variability in the Arctic mid-troposphere can be explained by the total—sensible plus latent—heat content variability at the midlatitude near surface in July. Climate models suggest that this connection is part of the global warming signal in September–November, but in reanalyses the connection is most strongly present in July–August variability and trends. It is argued that heat content signals are propagated from the midlatitude near surface to the Arctic midtroposphere approximately along climatological moist isentropes. High-frequency data reveal that the propagating signal is primarily driven by a few strong meridional heat flux events each summer season. Composite analysis on these events shows that August meridional heat fluxes into the Arctic midtroposphere are succeeded by positive heat content anomalies in the lower troposphere a few days later. This second connection between the Arctic midtroposphere and the Arctic lower troposphere could be sufficient to explain some of the recent Arctic 850-hPa temperature variability north of 75°N.

1. Introduction

In recent decades, the midtroposphere (defined here as 600–400 hPa) has been warming faster in the Arctic than in the Northern Hemisphere midlatitudes (e.g., Screen and Simmonds 2010); this warming has been accompanied by an increase in Arctic precipitable water (Serreze et al. 2012). Climate model studies in which Arctic warming is driven either by altering surface conditions or as part of the greenhouse warming process have disagreed on the dynamics governing this polar amplification. Some studies suggest that changing heat transport into the Arctic might contribute to the rapid observed Arctic midtroposphere warming (Langen and Alexeev 2007; Graversen and Wang 2009; Screen et al. 2012) while others, primarily greenhouse forcing

simulations (Langen and Alexeev 2007; Graversen and Wang 2009; Screen et al. 2012), emphasize that changes in net atmospheric heat transport into the Arctic as the climate warms are relatively small. This is because warming typically reduces the sensible heat transport and increases the latent heat transport into the Arctic. Stated another way, warming typically increases the fraction of heat brought to the Arctic in the form of latent heat, a point emphasized by several studies (e.g., Alexeev et al. 2005).

To better understand the nature of heat transport in Arctic midtropospheric warming and polar amplification, this study focuses on the dynamical drivers of Arctic midtropospheric heat anomalies in observations. We show that increased heat transport into the Arctic is driven to an important extent by increased precipitable water in the midlatitude planetary boundary layer (PBL). Moist air parcels in the PBL ascend in the warm conveyor belt of synoptic eddies (Eckhardt et al. 2004) and convert their equivalent potential temperature θ_e into potential temperature θ along trajectories roughly aligned with surfaces of constant θ_e . Through this process their latent heat content is converted into sensible heat. This process has been shown by Laliberté et al. (2012) to give rise to the summertime moist overturning

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circulation in the Northern Hemisphere midlatitudes, and Wu and Pauluis (2014) have shown that this process connects midlatitude surface and midtroposphere heat content. The consequent connection between midlatitude stability and moist processes has been identified in observations and as a response to greenhouse gas forcing (Jukes 2000; Frierson 2006).

To confirm that such moist processes are an important contributor to the warming of the Arctic midtroposphere (AMT), we show that the mid- to high-latitude dynamics connect the total—sensible plus latent—heat content at the top of the midlatitude PBL (using 850 hPa as a proxy) with AMT heat content in reanalyzed observations. We build upon a simple conceptual model of AMT warming based on the homogenization of warming along moist (constant θ_e) and dry (constant θ) isentropic surfaces (Laliberté and Kushner 2013, hereafter LK13). Such homogenization leads to increased warming with altitude (for θ and θ_e homogenization) and with drying (for θ_e homogenization). LK13 shows that projected twenty-first-century September–November (SON) AMT warming is greater than what would be expected from the homogenization along dry isentropes and less than expected by the homogenization along moist isentropes for the representative concentration pathway 4.5 (RCP4.5) scenario in most models of phase 5 of the Coupled Model Intercomparison Project (CMIP5). Here, we confirm that this behavior holds for the interannual variability of reanalyzed AMT heat content for late summer but not for SON. We show that in late summer it is associated with variability of meridional total heat fluxes into the AMT. We also demonstrate that the connection between the midlatitudes and the AMT for the interannual variability captures well the secular warming trend in late summer AMT. Finally, we present preliminary evidence for a downward influence of AMT heat content on the Arctic lower troposphere (below 600 hPa).

This paper is organized as follows: In section 2, we describe the thermodynamic framework and detail data and methods used in this study. We present our results in section 3, where we start with a discussion of recent trends in late summer AMT heat content to motivate our analysis and link it to the results of LK13. We then demonstrate that late summer AMT heat content has been strongly correlated with the midlatitude near-surface heat content in recent decades. We follow with an analysis of the AMT moist entropy balance and carry out a composite analysis of August meridional heat influx events, from which we infer a possible connection between the Arctic midtroposphere and Arctic lower troposphere. We summarize and discuss our results in section 4.

2. Theory, data, and methods

a. Second law of thermodynamics and moisture

In this study, we quantify heat content variability of AMT air masses over recent decades. The heat content of an air parcel can be represented by its specific entropy s , which changes along the parcel path only if it experiences a net diabatic heating $\dot{q}$ or if a process with an irreversible entropy production $\dot{s}_{\text{irr}}$ occurs,

$$\frac{ds}{dt} = \frac{\dot{q}}{T} + \dot{s}_{\text{irr}}. \quad (1)$$

The second law of thermodynamics states that the irreversible entropy production $\dot{s}_{\text{irr}}$ is always positive (Ambaum 2010). In numerical models, irreversible entropy production is parameterized through frictional heating $\dot{q}_{\text{fri}}$, turbulent diffusion $\dot{q}_{\text{trb}}$, and gravity wave drag heating $\dot{q}_{\text{gwd}}$. Other unparameterized processes that are typically neglected in numerical models can lead to irreversible entropy production [e.g., the evaporation of water into unsaturated air as discussed in Pauluis and Held (2002)]. The irreversible entropy production from these processes has been evaluated for the Modern-Era Retrospective Analysis for Research and Applications (MERRA) reanalysis and is found to be negligible for July and August in the high-latitude free troposphere at a monthly time scale.

Net diabatic heating $\dot{q}$ includes contributions from net radiative cooling $\dot{q}_{\text{rad}}$ and from moist processes such as $\dot{q}_{\text{mst}}$ like evaporation, condensation, freezing and precipitation. The relative importance of each of these moist processes depends on which formulation for the entropy of moist air is used [see the appendices of Pauluis et al. (2010)]. Using dry entropy $s_d = c_p \ln(\theta_d/T_0)$ will make the contribution from evaporation small and the contributions from condensation, freezing and precipitation large. In this expression, θ_d is the dry potential temperature, $T_0 = 273.15 \text{ K}$ is the triple point of water at 1000 hPa, and $c_p = 1003 \text{ J K}^{-1}$ is the specific heat capacity of dry air. On the other hand, using moist entropy $s_m = c_p \ln(\theta_e/T_0)$, where θ_e is the equivalent potential temperature, will make the contributions from condensation, freezing and precipitation small and the contribution from evaporation large. In the Arctic free troposphere there is little evaporation and with this choice the net contribution to $\dot{q}$ from moist processes $\dot{q}_{\text{mst}}$ is negligible.

Here, we are interested in diagnosing how Eq. (1) is represented over the AMT by the MERRA reanalysis [for an analysis of the vertically integrated polar energy budget, see Cullather and Bosilovich (2012)]. In addition to heatings arising from physical processes, the reanalysis procedure “nudges” the atmospheric model to match the

observational analysis as closely as possible. This nudging is implemented in MERRA through a numerical heating $\dot{q}_{\text{ana}}$ that is set to a constant for every 6-hourly incremental analysis step. A nonzero $\dot{q}_{\text{ana}}$ reflects problems in physical parameterizations in MERRA and/or an imperfect observational analysis.

With these approximations for the Arctic free troposphere in MERRA, the second law of thermodynamics [Eq. (1)] for the dry and moist entropies becomes

$$c_p \frac{d}{dt} \ln(\theta_d/T_0) \approx \frac{\dot{q}_{\text{rad}}}{T} + \frac{\dot{q}_{\text{ana}}}{T} + \frac{\dot{q}_{\text{mst}}}{T} \quad \text{and} \quad (2)$$

$$c_p \frac{d}{dt} \ln(\theta_e/T_0) \approx \frac{\dot{q}_{\text{rad}}}{T} + \frac{\dot{q}_{\text{ana}}}{T}. \quad (3)$$

The sensible heat content of air parcels is represented by θ_d , and it is closely related to the potential temperature θ . The two differ because θ also accounts for the effect of water vapor on the parcel's buoyancy (equivalent to using virtual temperature instead of absolute temperature for its computation). On the other hand, θ_e represents the total—sensible plus latent—heat content of air parcels. These observations are key for the interpretation of our results.

b. Daily data on pressure levels

We obtain temperature T and specific water vapor q_v on the pressure levels of 850, 700, 600, 500, 400, 300, and 250 hPa for two reanalyses. For MERRA (Rienecker et al. 2011), we use the web-based subsetter time-mean feature to obtain the daily-mean variables from instantaneous 3-hourly data [instantaneous basic 3D assimilated fields from Incremental Analysis Update (IAU) corrector (inst3_3d_asm_Cp) dataset variables T and QV]. For the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim; ECMWF 2013), we computed the daily-mean variables from instantaneous 6-hourly data at the analysis time [National Center for Atmospheric Research (NCAR) ds627.0 dataset variables t and q]. These variables are first regridded to an N48 grid using a first-order conservative remapping and low-pass filtered using a 31-day centered running mean to obtain T and q_v . We apply a 31-day running-mean low-pass filter to allow easy comparison with monthly-mean model outputs. These low-pass filtered T and q_v are used to compute the low-pass filtered θ and θ_e . It is important to note that θ_e is a nonlinear function of T and q_v and that applying the low-pass filter before instead of after its calculation will affect our results slightly. Here, we have decided to apply it before to ensure easy comparison with monthly-mean model outputs. We describe in appendix A the procedure used to compute θ and θ_e . We select their 850-hPa value

pressure surface ($\theta|_{850\text{hPa}}$ and $\theta_e|_{850\text{hPa}}$) and compute their zonal mean ($\bar{\theta}$ and $\bar{\theta}_e$). We obtain their mean over the AMT ($\theta|_{\text{AMT}}$ and $\theta_e|_{\text{AMT}}$) by taking their means over the levels of 400, 500, and 600 hPa and taking their meridional area-weighted mean from 75° to 90°N. Similarly, we obtained their mean over the midlatitude near surface ($\theta|_{\text{MNS}}$ and $\theta_e|_{\text{MNS}}$) by computing the meridional area-weighted mean from 45° to 65°N of their 850-hPa level. We denote anomalies from the climatological mean of these quantities with $\delta(\cdot) = (\cdot) - \langle \cdot \rangle$, where the climatological mean is identified as $\langle \cdot \rangle = (34 \text{ yr})^{-1} \int_{1979}^{2012} (\cdot) dt$.

Finally, we obtain the equivalent potential temperature propagated along moist isentropes $\delta\theta_e|_{\text{prop}}$ using the same method used in LK13. We review this method in appendix B. Results denoted by REANALYSES were obtained by computing the average of all these quantities for MERRA and ERA-Interim.

c. High-frequency MERRA data for budget calculations

We obtain high-frequency and high-resolution data from MERRA to compute the different terms of the second law in Eqs. (2) and (3). We used model-level 3-hourly zonal mass fluxes $u\Delta p$ (time averaged), meridional mass fluxes $v\Delta p$ (time averaged), vertical pressure velocity ω (time averaged) and layer edge pressure p_e (time staggered and instantaneous) on a C grid at $0.67^\circ \times 0.5^\circ$ longitude–latitude [MERRA accumulated 3-hourly chemistry-related transport fields on 3D model layers (MAT3NVCHM), MERRA accumulated 3-hourly chemistry-related transport fields at 3D model layer edges (MAT3NECHM), and MERRA instantaneous 3-hourly chemistry-related fields for off-line transport at 3D model layer edges (MAI3NECHM) datasets]. At every horizontal grid point and at every 3-hourly time step, we obtain a modified vertical pressure velocity $\tilde{\omega}$ by integrating the mass continuity equation from the top of the atmosphere down to the surface. This $\tilde{\omega}$ is very close to ω and it preserves mass at the grid scale, ensuring that the analysis is consistent with mass conservation. To compute thermodynamic quantities, we use model-level 3-hourly T and q_v on a $1.25^\circ \times 1.0^\circ$ longitude–latitude grid [MERRA time-averaged 3-hourly chemistry-related fields on 3D model layers (MAT3FVCHM) dataset]. On the same horizontal grid, we obtain pressure-level 3-hourly temperature tendencies from analysis $\dot{T}_{\text{ana}}$, radiation $\dot{T}_{\text{rad}}$, friction $\dot{T}_{\text{fri}}$, gravity wave drag $\dot{T}_{\text{gwd}}$, turbulence $\dot{T}_{\text{trb}}$, and moist physics $\dot{T}_{\text{mst}}$ [MERRA time-averaged 3-hourly temperature tendencies on 3D pressure levels (MAT3CPTDT) dataset]. The specific water vapor and the temperature are regridded to the $0.67^\circ \times 0.5^\circ$ C-grid center using a distance remapping. The temperature

tendencies are also regridded to the C grid but using a bilinear remapping and are linearly interpolated from pressure levels onto the model-level vertical grid without extrapolation.

In section 3c we use these data to obtain the moist entropy budget for the months of July and August in the

MERRA reanalysis. To carry out this analysis, we average Eq. (3) over the AMT (75°–90°N between model layer edges at 600 and 375 hPa) and integrate it in time to obtain the August $\theta_e|_{\text{AMT}}$ in terms of the July $\theta_e|_{\text{AMT}}$, the radiative forcing, the reanalysis nudging, and the heat convergence,

$$c_p \{ [\overline{\ln(\theta_e|_{\text{AMT}}/T_0)\Delta p}]_{\text{AMT}} \}_{\text{August}} \approx c_p \{ [\overline{\ln(\theta_e|_{\text{AMT}}/T_0)\Delta p}]_{\text{AMT}} \}_{\text{July}} + \text{RADIATION} + \text{NUDGING} + \text{BOTTOM} + \text{TOP} + \text{MERIDIONAL}, \quad (4)$$

where RADIATION represents the radiative cooling; NUDGING represents sources and sinks of heat associated with the analysis term; and BOTTOM, TOP, and MERIDIONAL represent the fluxes of moist entropy out of the bottom, top, and lateral boundary at 75°N, respectively. In this expression, $[\cdot]_{\text{AMT}}$ represents an average over the AMT divided by the mass of the AMT and $\{\cdot\}_{\text{August}}$ and $\{\cdot\}_{\text{July}}$ represent August and July monthly means.

The RADIATION and NUDGING terms are given by

$$\text{RADIATION} = \tau [\{\overline{\Delta p \dot{q}_{\text{rad}}/T}\}]_{\text{AMT}} \quad \text{and}$$

$$\text{NUDGING} = \tau [\{\overline{\Delta p \dot{q}_{\text{ana}}/T}\}]_{\text{AMT}},$$

where $\{\cdot\}$ is a double 31-day centered running mean and $\tau = 31$ days. The BOTTOM, TOP, and MERIDIONAL terms are related to the convergence of entropy,

$$\text{BOTTOM} = \tau c_p [\{\overline{\ln(\theta_e/T_0)\tilde{\omega}^*}\}]_{\text{AMT}}^{375\text{hPa}},$$

$$\text{TOP} = -\tau c_p [\{\overline{\ln(\theta_e/T_0)\tilde{\omega}^*}\}]_{\text{AMT}}^{600\text{hPa}}, \quad \text{and}$$

$$\text{MERIDIONAL} = -\tau c_p [\{\overline{\ln(\theta_e/T_0)v\Delta p}\}]_{\text{AMT}}^{75\text{N}},$$

where $[\cdot]_{\text{AMT}}^{375\text{hPa}}$, $[\cdot]_{\text{AMT}}^{600\text{hPa}}$, and $[\cdot]_{\text{AMT}}^{75\text{N}}$ are integrals on the layer edge at 375 hPa, on the layer edge at 600 hPa, and on the 75°N boundary, respectively, and all three are divided by the mass of the AMT. We have made use of the continuity equation to expand the material derivative into its conservation form to obtain the right-hand side. We describe in appendix C how these different terms were obtained.

d. Composite analysis of entropy flux events into the AMT

In section 3c, we composite on strong poleward total heat flux events that are identified as follows: For every 6-day window in July and August of each year, we use 3-hourly data to identify the maximum in $[\overline{\ln(\theta_e/T_0)v\Delta p}]_{\text{AMT}}^{75\text{N}}$. There will be one for every 6-day

window in July and August, but some of these maxima might be identified more than once. For each one of the identified maxima, we compute

$$\text{MERIDIONAL}^{\text{first}} = -\tau c_p [\{\overline{\ln(\theta_e/T_0)v^{\text{first}}\Delta p}\}]_{\text{AMT}}^{75\text{N}},$$

where v^{first} is set to v for a 6-day window centered on the maximum and set to 0 otherwise. The fraction

$$\frac{\text{MERIDIONAL}^{\text{first}}}{\text{MERIDIONAL}}$$

represents how much of that year's meridional entropy flux into the AMT is associated with the maximum. We selected the event that maximizes this fraction to obtain the first maximum. We repeated this procedure with v^{first} replacing v to obtain the second maximum and so on for the subsequent maxima. We discuss in section 3c how different time windows might affect our results.

3. Results

a. Recent Northern Hemisphere temperature trends

We begin by describing how the sensible heat content of Northern Hemisphere air masses has evolved during August in recent decades. The focus on the month of August is motivated by the lag-correlation analysis presented in section 3b which shows that the isentropic homogenization process described in section 1 is clearest in August.

In the top panels of Fig. 1, we quantify this evolution by plotting trends of $\bar{\theta}$ for MERRA and ERA-Interim using color shading. It is important to exercise caution when interpreting these trends. The sparsity of data in the Arctic and changing observational networks over the reanalysis period might make some of these trends an artifact of reanalysis products. For this study, we use these trends to describe the structure of the recent Northern Hemisphere warming; most of the dynamical analysis will use detrended data. Using black contours,

FIG. 1. (top) August 1979–2012 climatology of zonal-mean potential temperature $\bar{\theta}$ (black contours) and August year-to-year trends of zonal-mean $\bar{\theta}$ for (left) MERRA and (right) ERA-Interim (color shaded). Regions that are not shaded have trends that are statistically significant at 95% using the t test. The black box indicates the AMT (75°–90°N, 600–400 hPa) and the white box indicates the MNS (45°–65°N, 850 hPa). (bottom) As in (top), but with the equivalent (moist) potential temperature $\bar{\theta}_e$ instead of the potential temperature $\bar{\theta}$.

we indicate surfaces of constant $\langle \bar{\theta} \rangle$. We use a black box to identify the AMT (defined here as the region 75°–90°N between 400 and 600 hPa) and a white box to identify the midlatitude near surface (MNS; defined here as the region 45°–65°N at 850 hPa). Trends that are shaded in gray are not statistically significant at 95% using the t test.

Trends range from 0.3 K decade^{−1} in the midlatitudes to 0.6 K decade^{−1} at 850 hPa in the Arctic and their patterns are similar between the two reanalyses. At each pressure level, trends are larger in the Arctic than anywhere else in the Northern Hemisphere for both reanalyses, illustrating the Arctic amplification of warming. In the Arctic, both reanalyses exhibit a relative maximum in trends at 500 hPa. Here, we will show that this relative maximum can be explained by a transport of total heat content from the midlatitudes. In LK13, this was argued based on the warming profile of $\bar{\theta}_e$, a quantity that represents more closely the total heat content of air masses.

Motivated by LK13, we thus quantify the evolution of the total heat content by plotting $\bar{\theta}_e$ trends using color shading for the month of August in the bottom panels of Fig. 1. Black contours indicate surfaces of constant $\langle \bar{\theta}_e \rangle$. The boxes have the same meaning as in the top panels; again, trends that are not statistically significant at 95%

using the t test are shaded in gray. As in LK13, $\bar{\theta}_e$ trends are vastly different from $\bar{\theta}$ trends. Although $\bar{\theta}$ trends are similar in the two reanalyses, $\bar{\theta}_e$ trends are weaker in ERA-Interim than in MERRA. Since both reanalyses have a similar $\langle \bar{\theta}_e \rangle$ (black contours), it suggests the recent evolution of their August water vapor fields differ in the Northern Hemisphere.

In this picture, trends of $\bar{\theta}_e$ in the AMT (identified as $\theta_{e|AMT}$) have a magnitude similar to trends of $\bar{\theta}_e$ at the MNS (identified as $\theta_{e|MNS}$). Moreover, as in the analysis of LK13 for climate models, color contours have a weak but negative gradient along the surfaces of constant $\langle \bar{\theta}_e \rangle$, resulting in $\theta_{e|MNS}$ trends slightly but substantially larger than $\theta_{e|AMT}$ trends for both reanalyses.

In Fig. 2 we use trends of $\theta_{|MNS}$, $\theta_{|AMT}$, $\theta_{e|AMT}$, and $\theta_{e|MNS}$ for August to quantify this weak gradient. We show results for MERRA, ERA-Interim, and the average of the two, which is identified as REANALYSES. The horizontal line inside the box indicates the least squares trend. From the t test, the box indicates the inner quartiles and the whiskers indicate the one-sided 95% confidence interval. From this comparison, it is clear that trends of $\theta_{|MNS}$ are smaller than trends of $\theta_{|AMT}$, which are similar to trends in $\theta_{e|AMT}$, and these are in turn smaller than trends of $\theta_{e|MNS}$. This relationship is somewhat more robust in MERRA but it is not statistically

FIG. 2. For each reanalysis, from left to right, trends of 31-day low-pass filtered zonal-mean $\bar{\theta}$ at the MNS (identified by $\theta|_{\text{MNS}}$) in green, $\bar{\theta}$ in the AMT (identified by $\theta|_{\text{AMT}}$) in black, θ_e in the AMT (identified by $\theta_e|_{\text{AMT}}$) in black, and θ_e at the MNS (identified by $\theta_e|_{\text{MNS}}$) in red. Using one-sided t tests, boxes indicate inner quartiles and whiskers indicate the inner 95%.

strong in either reanalyses because of the large uncertainty in AMT trends. In LK13 it was argued that a dynamical connection through moist processes in synoptic systems would lead to such a relationship between these four quantities. Such a relationship would then mean that the increase in total heat content in the midlatitudes as measured by $\theta_e|_{\text{MNS}}$ is sufficient to explain the recent warming in the AMT during August.

This analysis suggests that there might be a dynamical connection between trends in MNS and the AMT through θ_e homogenization during August. However, because of uncertainty surrounding these trends and because the hypothesized mechanism involves moisture transport by synoptic eddies, these connections should be observed in interannual variability of detrended data. Therefore, in the next section we investigate more closely whether the MNS is indeed linked to the AMT by performing a lag-correlation analysis on the detrended time series.

b. Influence of midlatitude near-surface heat content on the Arctic midtroposphere

In the left panel of Fig. 3, we show the August lag-correlation analysis of detrended $\delta\theta_e|_{850\text{hPa}}$ as predictor with detrended $\delta\theta_e|_{\text{AMT}}$ as response. In this figure, we observe a patch of high correlation of $\delta\theta_e|_{\text{AMT}}$ with $\delta\theta_e|_{850\text{hPa}}$ between 45° and 65°N and lags between 0 and -40 days. In Fig. 1 we can see that the region between 45° and 65°N , the region we have identified as the MNS, corresponds to where surfaces of constant $\langle\bar{\theta}_e\rangle$ from the AMT intersect the 850-hPa surface. This suggests that the MNS has a causal influence on the AMT. The lag of maximum correlation goes from approximately -30 days at 60°N to no lag at 45°N . The absence of lag at 45°N could either describe a strong instantaneous connection with the AMT or be the result of two causal influences that cancel each other: one from 45°N to the AMT with

negative lag and one from the AMT to 45°N with positive lag (consistent with the typical trajectories described by the circulation in moist isentropic coordinates; Laliberté et al. 2012).

In the Arctic, $\delta\theta_e|_{\text{AMT}}$ is strongly correlated to $\delta\theta_e|_{850\text{hPa}}$ north of 75°N . This correlation appears skewed toward positive lags, indicating that the AMT might be leading the 850-hPa signal in the Arctic. We will revisit this feature later in section 3e.

In the right panel of Fig. 3, we show the zonal decomposition of the lag correlation from the top-left panel at -30 -day lag. We obtained this decomposition by computing the covariance of detrended $\delta\theta_e|_{850\text{hPa}}$ with detrended $\delta\theta_e|_{\text{AMT}}$, normalized in such a way that its zonal mean equals its value in the left panel. We have added the 45° and 65°N latitudes in white to indicate the MNS. The correlation patterns for MERRA and ERA-Interim (not shown) are very similar to the patterns for REANALYSES.

In this plot, we see strong positive correlations between 45° and 65°N over North America, western Russia, and the Pacific. Positive albeit weaker correlations are also found everywhere in the Atlantic. It is somewhat counterintuitive that the strongest correlations are found over land. To better understand this correlation pattern, we compared this spatial pattern to the pattern of correlation of $\delta\theta_e|_{850\text{hPa}}$ with the zonal-mean $\delta\bar{\theta}_e|_{850\text{hPa}}$ (not shown) and found that the two patterns were similar. This similarity means that years with high zonal-mean $\delta\bar{\theta}_e|_{850\text{hPa}}$ are usually associated with a high $\delta\theta_e|_{850\text{hPa}}$ over these land regions. These land regions are therefore more likely to be represented in the connection between the MNS and the AMT. In addition, there is considerable active convective instability over these land regions during August, which might amplify their connection with the AMT.

In the top-left panel of Fig. 4, we investigate the causal influence of the MNS on the AMT by performing a lag-correlation analysis of detrended $\delta\theta|_{\text{MNS}}$ as the predictor with detrended $\delta\theta_e|_{\text{AMT}}$ as the response. This plot confirms that correlations between 0.4 and 0.6 occur mostly from mid-July to mid-September. We indicate using a black curve the lag of maximum correlation. In August, the maximum correlation occurs when the $\delta\theta|_{\text{MNS}}$ signal precedes the $\delta\theta_e|_{\text{AMT}}$ signal by about 20 days. In the middle-left panel, we show the regression coefficient λ for the relationship $\lambda\delta\theta|_{\text{MNS}} = \delta\theta_e|_{\text{AMT}}$ with $\delta\theta|_{\text{MNS}}$ preceding $\delta\theta_e|_{\text{AMT}}$ by the number of days corresponding to the maximum correlation. The dark shading indicates the inner quartiles and the light shading indicates the one-sided 95% confidence interval from the t test. Note that over the first half of August the regression coefficient λ is larger than 1.0.

FIG. 3. (left) Lag correlation of detrended $\delta\theta_e|_{\text{AMT}}$ response during August with detrended zonal-mean $\delta\theta_e|_{850\text{hPa}}$ predictor with lag. Strong correlations at negative lag means that the 850-hPa signal precedes the AMT signal. (right) The zonal structure of the correlations shown in (left) along the -30 -day lag. Performing a zonal mean of the correlations shown in (right) is equal to the correlations in (left) along the -30 -day lag line. The white parallels indicate 45° , 65° , and 80°N . The lag correlations were obtained for every day of August and only the average of these days is shown. Results for REANALYSES are shown.

In the bottom-left panel, we show the time series $\lambda\delta\theta|_{\text{MNS}}$ on 15 July and $\delta\theta_e|_{\text{AMT}}$ on 15 August; in this panel, the anomaly time series have not been detrended but the coefficients λ are derived from the detrended time series. Even though these time series were not detrended and include the warming signal, we observe that $\lambda = 1.83$ captures well both the interannual variability and the secular trend. Thus, while midsummer $\theta|_{\text{MNS}}$ is a good statistical predictor of late summer $\theta_e|_{\text{AMT}}$ (high correlations in the top-left panel), midsummer $\theta|_{\text{MNS}}$ variability is about half the magnitude required to quantitatively explain late summer $\theta_e|_{\text{AMT}}$ variability [scale factor λ is $O(2)$].

In the right panels of Fig. 4, we perform the same analysis as in the left panels of Fig. 4, using $\delta\theta_e|_{\text{MNS}}$ instead of $\delta\theta_e|_{\text{MNS}}$ as the predictor. With $\delta\theta_e|_{\text{MNS}}$, correlations of more than 0.5 are found from mid-July to September with correlations of more than 0.6 during the first half of August, which is higher than the maximum correlation found in the top-left panel of Fig. 4. This means that July $\delta\theta_e|_{\text{MNS}}$ explains about 40% of the detrended variance of August $\delta\theta_e|_{\text{AMT}}$. The maximum correlation, indicated by a black curve, occurs when the MNS signal precedes the AMT signal by between 20 and 40 days. In the middle panel, we see that the linear regression coefficient $\lambda \lesssim 1.0$, lies mostly between 0.5 and 1.4 during August, and has a much narrower spread than in the left panel. In the bottom-right panel, we

show the time series $\lambda\delta\theta_e|_{\text{MNS}}$ on 15 July and $\delta\theta_e|_{\text{AMT}}$ on 15 August. We observe that $\lambda = 0.97$ captures well both the interannual variability and the secular trend. Thus, taking into account sensible and latent heat content improves the statistical connection between midsummer $\theta|_{\text{MNS}}$ and late summer $\theta_e|_{\text{AMT}}$ (higher correlations in the top-right panel) and provides a quantitative correspondence between the two quantities [scale factor λ is $O(1)$]. Moreover, because $\lambda = 0.97$ captures the secular trend well, it suggests that this quantitative correspondence might explain a large part of the trend in August $\theta_e|_{\text{AMT}}$ over the last decades.

As mentioned earlier, the MNS corresponds to the region where surfaces of constant $\langle\theta_e\rangle$ from the AMT intersect the 850-hPa level. We would therefore expect that the moist propagation model of LK13 would paint a picture similar to the one obtained in Fig. 4. We verify that this is indeed the case in Fig. 5, where we carry out the same analysis as in the two top-right panels of Fig. 4 but replace the predictor $\delta\theta_e|_{\text{MNS}}$ by the equivalent potential temperature propagated along climatological moist isentropes $\delta\theta_e|_{\text{prop}}$ (see section 2b). The results are similar to the top-right panels of Fig. 4, except that correlations at positive lags disappear. This similarity suggests that during August the influence of the MNS on the AMT identified in Figs. 3 and 4 can be modeled as the homogenization of $\delta\theta_e$ along $\langle\theta_e\rangle$ surfaces.

FIG. 4. (top left) Lag correlation of detrended $\delta\theta_e|_{\text{AMT}}$ response on the x axis with detrended $\delta\theta|_{\text{MNS}}$ predictor. Strong correlations at negative lag mean that the MNS signal precedes the AMT signal. (middle left) Solid black curve indicates the linear regression coefficient λ at the negative lag with maximum correlation [solid black curve in (top left)]. Using one-sided t tests, dark gray shading indicates inner quartiles and light gray shading indicates the inner 95%. (bottom left) The $\delta\theta_e|_{\text{AMT}}$ on 15 Aug in black and the product of λ on 15 Aug with $\delta\theta|_{\text{MNS}}$ on 15 Jul in green. The two time series were not detrended in (bottom left). (right) As in (left), but with $\delta\theta_e|_{\text{MNS}}$ instead of $\delta\theta|_{\text{MNS}}$ as the predictor. Results for REANALYSES are shown.

The LK13 analysis of projected warming for SON in CMIP5 simulations suggests that is only after the 2020–30 period that the AMT warming response emerges from interannual variability: after that period, it becomes possible to statistically determine to which extent the AMT warming response is consistent with the homogenization of $\delta\theta_e$ along $\langle\theta_e\rangle$ surfaces. However, in LK13 the midlatitude response homogenized along isentropes and the AMT response during SON are not correlated in the twentieth-century historical integrations (as exhibited for two models in the bottom-center and bottom-right panels of Fig. 2 in LK13). The reasons for this are being investigated, but the differences in seasonality and signal strength suggest that details of the midlatitude–AMT connection might be distinct in the CMIP5 simulations and the reanalyses.

The correlations at positive lags in Fig. 4 mean that August AMT heat content is a relatively good predictor of MNS heat content in September. If the dynamics tend to homogenize heat content anomalies along surfaces of constant $\langle\theta_e\rangle$ as suggested by Fig. 5, then it would be expected that it could act both ways, as suggested earlier: August AMT heat content anomalies could cause heat content anomalies at the MNS in September. Note

that September $\delta\theta_e|_{\text{prop}}$ represents near-surface heat content anomalies that are equatorward of the MNS, which might explain why the correlations at positive lags disappear in Fig. 5.

FIG. 5. As in Fig. 4 (top right), (middle right), but with the predictor set to $\delta\theta_e|_{\text{prop}}$ and the propagated $\delta\theta_e$ as in LK13. Results for REANALYSES are shown.

We have thus described a connection between the MNS and AMT in late summer where the total heat content at the MNS appears to be dynamically driving the heat content in the AMT. It remains, however, unclear which atmospheric phenomena mediate this connection. Our analysis shows that a 31-day running mean is enough to ensure that $\delta\theta_e|_{\text{AMT}}$ is sufficiently homogenized so that we obtain a strong signal from the MNS to the AMT. This would suggest that the phenomena at play 1) reliably occur every year during July and August and 2) have a strong moist component. Next, we show that this connection is associated with meridional total heat transport into the AMT and is associated with mostly discrete heat flux events that occur reliably over July and August.

c. High-latitude entropy balance

To demonstrate that total heat transport is likely to be an important contributor to AMT variability, we look at the entropy balance Eq. (3) over the AMT and integrate it in time to obtain the August $\theta_e|_{\text{AMT}}$ in terms of the July $\theta_e|_{\text{AMT}}$, the radiative forcing, the analysis nudging, and the heat convergence. We use Eq. (4) to compute this balance:

$$\{\theta_e|_{\text{AMT}}\}_{\text{August}} \approx \{\theta_e|_{\text{AMT}}\}_{\text{July}} + \text{RADIATION} + \text{NUDGING} + \text{CONVERGENCE}, \quad (5)$$

where CONVERGENCE is given by

$$\text{CONVERGENCE} = \text{BOTTOM} + \text{TOP} + \text{MERIDIONAL}, \quad (6)$$

the sum of the contributions from the different directions. We show in Fig. 6 a schematic of this balance. We define STORAGE by the change in equivalent potential temperature in the AMT between the month of July and August,¹

$$\text{STORAGE} = \{\theta_e|_{\text{AMT}}\}_{\text{July}} - \{\theta_e|_{\text{AMT}}\}_{\text{August}}.$$

STORAGE represents how much the AMT cools between July and August.

In the top panel of Fig. 7 we show the climatology of these different terms and we see that the STORAGE term is small and comparable to the NUDGING. The

FIG. 6. Schematic diagram explaining the different contributors setting the August equivalent potential temperature in the AMT.

RADIATION cooling is thus balanced almost exclusively by the CONVERGENCE. More precisely, the MERIDIONAL heat fluxes into the AMT are sufficient to balance about 75% of the RADIATION and are thus the dominant term in the CONVERGENCE of moist entropy. This observation suggests that there is a mostly meridional turnover of air masses in the AMT. This would mean that August AMT temperatures are primarily maintained by the large-scale circulation that continuously replaces AMT air masses by new, warmer, and moister ones from lower latitudes.

The fact that the main climatological balance is between MERIDIONAL heat fluxes and RADIATION does not guarantee that this balance controls the AMT variability. To assess this, we first plot in the middle panel of Fig. 7 the time series of $\{\theta_e|_{\text{AMT}}\}_{\text{August}}$ computed directly in black and reconstructed from the right-hand side of Eq. (5) in gray. Small errors were introduced from the regridding and other approximations made to obtain Eq. (3), resulting in a correlation between the two time series of about 0.8. In the bottom panel of Fig. 7, we then detrend each term on the right-hand side of Eq. (5) and compute its covariance with the detrended $\{\theta_e|_{\text{AMT}}\}_{\text{August}}$. We normalize this covariance by the product of the standard deviation of detrended $\{\theta_e|_{\text{AMT}}\}_{\text{August}}$ and the standard deviation of the

¹ The storage was obtained by multiplying $c_p \delta \{ [\ln(\theta_e|_{\text{AMT}}/T_0) \Delta p]_{\text{AMT}} \}_{\text{July}} - c_p \delta \{ [\ln(\theta_e|_{\text{AMT}}/T_0) \Delta p]_{\text{AMT}} \}_{\text{August}}$ with the linear conversion factor α in Eq. (C1).

FIG. 7. (top) The climatology for each term of the entropy balance in July–August. (middle) The August interannual anomalies of $\{\delta\theta_e|_{AMT}\}_{August}$ computed in two different ways: 1) directly (black) and 2) from the right-hand side of Eq. (5) (gray). (bottom) The detrended covariances of $\{\theta_e|_{AMT}\}_{August}$ with the different terms in the right-hand side of Eq. (5). Covariances are normalized by the product of the detrended standard deviations of $\{\theta_e|_{AMT}\}_{August}$ and of the right-hand side of Eq. (5). Note that year 2010 was excluded from this analysis.

detrended right-hand side of Eq. (5). We see that $\{\theta_e|_{AMT}\}_{August}$ is positively correlated with $\{\theta_e|_{AMT}\}_{July}$ and with the MERIDIONAL component. The latter is the main contributor to the interannual variability. The vertical convergence BOTTOM + TOP and the NUDGING are both negatively correlated with $\{\theta_e|_{AMT}\}_{August}$. The NUDGING contributes to the correlation with $\{\theta_e|_{AMT}\}_{August}$ by as much as $1/5$ of the contribution from MERIDIONAL. The NUDGING thus plays here an outside role. This is also true for trends associated with these different terms (shown in Table 1). The trends of all quantities but BOTTOM + TOP are significant. In particular, we see significant trends in MERIDIONAL, RADIATION, and NUDGING. The trends from these three quantities almost sum up to zero. Unfortunately, we cannot identify which component of the entropy balance [Eq. (5)] the NUDGING trend is compensating for. While our analysis would suggest that the August AMT heat content variability is controlled by the variability in MERIDIONAL heat fluxes into the AMT, the NUDGING adds uncertainty to this inference, especially if one would like to understand

TABLE 1. Trends in the different terms of Eq. (5) in kelvins per decade. The uncertainty corresponds to the 95th percentile of the t test.

	Trend (K decade ⁻¹)
MERIDIONAL	1.99 ± 1.19
BOTTOM + TOP	-0.18 ± 1.13
$\{\theta_e _{AMT}\}_{July}$	0.40 ± 0.27
RADIATION	-0.95 ± 0.19
NUDGING	-1.09 ± 0.61
$\{\theta_e _{AMT}\}_{August}$	0.51 ± 0.38

what controls the trend found in the reanalysis. This nevertheless shows that the dominant mechanism driving the August AMT heat content variability is also an important contributor to the secular trend.

d. Composites of influx events

A large fraction of the MERIDIONAL component in the entropy balance [Eq. (5)] is the result of several more or less discrete influx events. To demonstrate this, we show in Fig. 8 a Hovmöller diagram of normalized meridional heat flux anomalies into the AMT for 1981 over July and August (see section 2c for the method used to compute these anomalies). There is nothing particular about 1981, except that it represents well the intermittent nature of meridional heat fluxes, with events easily identified around 20 July, 10 August, and 25 August. We describe in section 2d a method to identify and order these events based on how much they contribute to the MERIDIONAL component. We apply this method every year to find the five events with the largest fractional contribution to the MERIDIONAL component, and we list in Table 2 their climatological fractional contribution. The event with the largest fractional contribution accounts for a quarter of the MERIDIONAL component and the fractional contribution from the three largest is 0.57. Messori and Czaja (2013) obtained similar fractional contributions for the heat flux at 850 hPa using a different definition of extreme events (their Table 2). To quantify the time scale of these events and their effect on the Arctic tropospheric column heat content, we have performed a composite analysis of these events. For every year, we have chosen the event with the largest fractional contribution (listed in Table 3) and composited events for the entire reanalysis record, except for the year 2010 because of missing data.

In the top panel of Fig. 9 we show the composite of normalized meridional heat fluxes anomalies into the AMT from these events. The result is an anomaly propagating poleward over a time scale of approximately 5 days. This is comparable to the time scale of warm events into the Arctic found by Graversen et al. (2008).

FIG. 8. Normalized anomalies of 3-hourly $[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}$ over the months of July and August 1981.

Using different time windows (5–15-day windows were analyzed) has a limited qualitative influence on this picture and always yields a similar time scale of approximately 5 days. The 6-day window was chosen to capture this time scale and maximize anomalies in the middle and bottom panels.

In the middle and bottom panels of Fig. 9 we show the composite of normalized total heat content and latent heat anomalies, respectively. We observe that influx events are followed by a warm and moist AMT anomaly that appears to extend to the lower troposphere 2–3 days later. This composite analysis suggests that MERIDIONAL heat fluxes into the AMT directly influence the lower troposphere.

e. Is Arctic midtropospheric variability driving variability in the Arctic lower troposphere?

In light of these results, we revisit the high correlations between detrended $\delta\theta_e|_{\text{AMT}}$ and detrended $\delta\theta_e|_{850\text{hPa}}$ north of 75°N that were noted in Fig. 3. In Fig. 10, we carry out the same analysis as for Fig. 4 but using detrended $\delta\theta_e|_{\text{AMT}}$ as the predictor and, in turn, detrended $\delta\theta|_{850\text{hPa}}$ and detrended $\delta\theta_e|_{850\text{hPa}}$ as the responses. In the top panels, for both predictors, we observe high correlations skewed toward the AMT signal preceding the 850-hPa signal. In particular, the maximum correlation indicated by the black curve is almost everywhere

TABLE 2. Climatological fraction contribution to MERIDIONAL heat fluxes by the five events that contribute the most from each year. See section 2d for a description of the method used to identify those events.

Max	Fraction	Cumulative fraction
1st	0.24	0.24
2nd	0.19	0.43
3rd	0.14	0.57
4th	0.10	0.67
5th	0.08	0.75

negative, in line with our results from the composite analysis. These results suggest that the dynamics of total heat influx events might connect heat content changes in the AMT and in the lower troposphere.

In the middle panels, we note that the linear regression coefficient for the dry detrended $\delta\theta|_{850\text{hPa}}$ response is less than 1.0 and that the linear regression coefficient for the moist detrended $\delta\theta_e|_{850\text{hPa}}$ response is greater than or equal to 1.0. These results suggest that if the dynamical connection between the AMT and 850 hPa is due to air brought down adiabatically then it would be sufficient to explain the sensible heat content but not necessarily enough to explain the total heat content at 850 hPa. This is not surprising since these downward motions would likely have an impact across the PBL and be associated with enhanced latent heat fluxes at the surface, as hinted by the composite analysis in the bottom panel of Fig. 9. These enhanced surface latent heat fluxes would increase the latent heat content within the PBL, implying that the net anomaly in total heat content at 850 hPa would be larger than the one expected from purely adiabatic motions descending from the AMT. In the bottom-left panel we show time series of $0.68\delta\theta|_{\text{AMT}}$ in green and $\delta\theta|_{850\text{hPa}}$ in black while in the bottom-right panel we show time series of $1.08\delta\theta_e|_{\text{AMT}}$ in red and $\delta\theta_e|_{850\text{hPa}}$ in black. In both cases, the linear regression coefficient captures well both the interannual variability and the secular trend. This would suggest that in August there is a quantitative correspondence between $\delta\theta_e|_{\text{AMT}}$ and $\delta\theta_e|_{850\text{hPa}}$ and that this correspondence captures a large part of the secular trend.

4. Summary and discussion

We have shown that recent trends in August midlatitude total—sensible plus latent—heat content at 850 hPa are comparable to recent trends in August Arctic midtropospheric (AMT) heat content. Based on this observation, we have described what appears to be a causal connection between the midlatitude total heat

TABLE 3. Dates and times of the composited events and their fractional contribution to that year's MERIDIONAL heat fluxes.

Date	Time (UTC)	Fraction	Date	Time (UTC)	Fraction
13 Aug 1979	1030	0.196	—	—	—
3 Aug 1980	0430	0.425	23 Jul 1981	1330	0.346
21 Jul 1982	2230	0.120	29 Jul 1983	0430	0.270
10 Aug 1984	1030	0.321	26 Jul 1985	0730	0.228
7 Aug 1986	1030	0.273	26 Jul 1987	1030	0.331
23 Jul 1988	0730	0.262	3 Aug 1989	0430	0.213
1 Aug 1990	0430	0.317	30 Jul 1991	1030	0.251
10 Aug 1992	1030	0.223	9 Aug 1993	0730	0.195
12 Aug 1994	0430	0.217	19 Jul 1995	1330	0.202
25 Jul 1996	0130	0.245	1 Aug 1997	2230	0.274
2 Aug 1998	1030	0.188	7 Aug 1999	1330	0.366
18 Jul 2000	1030	0.186	7 Aug 2001	1630	0.183
30 Jul 2002	0130	0.331	9 Aug 2003	1030	0.221
25 Jul 2004	1030	0.211	9 Aug 2005	0430	0.217
26 Jul 2006	1330	0.155	4 Aug 2007	1330	0.256
29 Jul 2008	0130	0.212	29 Jul 2009	0130	0.170
—	—	—	7 Aug 2011	2230	0.271
12 Aug 2012	0130	0.172	—	—	—

content at 850 hPa in July and AMT heat content in August using detrended data. The analysis shows that the midlatitude total heat content variability at 850 hPa in July explains a large part of the AMT heat content variability in August over recent decades (it explains about 40% of the detrended variance). This relationship thus suggests that the variability of AMT heat content could be driven by the transport of air masses from the midlatitude near surface. Moreover, we have confirmed that during August this transport can be modeled as a homogenization along climatological moist isentropes as hypothesized by LK13 for SON warming over the twenty-first century. We have however found that during SON interannual variability in the AMT is not linked to variability at the MNS in reanalyzed observations. Consequently, our results show that reanalyzed observations cannot easily help us understand why LK13 found that warming in the AMT can be linked to warming in midlatitudes in twenty-first-century simulations with greenhouse gas forcing.

We have confirmed that meridional heat fluxes are the main contributor to the heat content variability in the AMT. Our results show that AMT heat content variability is controlled by discrete meridional heat flux events in which signals propagate from the MNS to the AMT over the course of 20–40 days. Fleshing out the dynamics of these events, including determining the reason for the 20–40-day time scale, why the July–August period is favored for the events, and the characteristics of the eddies involved in this transport, remains to be done. The analysis of Shaw and Pauluis (2012) however

FIG. 9. Composites of events identified in Table 3. (top) Meridional heat flux anomalies, (middle) total heat content anomalies, and (bottom) latent heat content anomalies. All anomalies were normalized by their standard deviation over the months of July and August. Note that the year 2010 was excluded from this analysis.

indicates that eastward-propagating transient waves at the planetary scale (with wavenumber less than or equal to 3) are responsible for most of the latent heat transport into the Arctic during June–August (JJA). Their analysis suggests that between 40° and 60°N during JJA transient waves propagate eastward rather slowly, which could explain why the typical time scale we have observed here is longer than that of typical weather patterns.

We have highlighted that a trend in nudging within the MERRA reanalysis makes it difficult to directly attribute the recent AMT warming to an increase in meridional heat fluxes. Furthermore, we have shown that strong meridional heat fluxes events into the AMT are followed 2–3 days later by positive anomalies in total and latent heat contents though the tropospheric column and the surface. Our interpretation is that the events that influence AMT heat content tend to almost simultaneously influence the Arctic lower troposphere. We have

FIG. 10. (left) As in Fig. 4 (left), but with detrended $\delta\theta_e|_{\text{AMT}}$ as predictor and detrended zonal-mean $\delta\theta|_{850\text{hPa}}$ over $75^{\circ}\text{--}90^{\circ}\text{N}$ as the response. (right) As in (left), but with $\delta\theta_e|_{850\text{hPa}}$ over $75^{\circ}\text{--}90^{\circ}\text{N}$ as the response. Strong correlation at negative lag means that the AMT signal precedes the Arctic 850-hPa signal. Note that the time series in (bottom) were not detrended. Results for REANALYSES are shown.

demonstrated that the connection between the AMT and the Arctic lower troposphere is such that variability of total heat content in the Arctic lower troposphere cannot be fully explained by variability of heat content in the AMT but that changes in the lower tropospheric energy budget could be necessary to explain the variability of total heat content in the Arctic lower troposphere.

This work describes a quantitative connection between midlatitude moisture and the August AMT at the interannual time scale that also captures well the observed secular trend in August AMT temperatures. This work therefore highlights how increasing midlatitude moisture associated with greenhouse gas forcing might influence the August AMT warming. It is generally agreed that AMT warming is driven by processes outside the Arctic, as argued by Screen et al. (2012) and as confirmed by attribution studies (e.g., Taylor et al. 2013). This work would thus confirm these previous results and would suggest that the increase in AMT heat content originates from an increase in midlatitude near-surface latent heat content that is propagated along climatological moist isentropes. There is also strong modeling evidence that the Arctic amplification of surface warming can be attributed to local feedbacks (Kay et al. 2012), suggesting that Arctic midtropospheric warming and

Arctic lower tropospheric warming would be driven by different processes. This work however demonstrates that the Arctic lower and middle troposphere are highly correlated from July to October and that their recent warmings are well represented by this correlation. We have shown that strong meridional influx events into the AMT could enable this coupling, but our results cannot entirely rule out the possibility that other processes like increased Arctic convection could contribute to this coupling.

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APPENDIX A

Computing Moist Thermodynamic Quantities

For each pressure level 1000, 925, 850, 700, 600, 500, 400, 300, and 250 hPa, the `pyteos_air` package from the python

version of Thermodynamic Equation of Seawater 2010 (TEOS-10; Feistel et al. 2010) was used to compute the saturation specific humidity $q_{\text{sat}}(p, T)$ on a regular T grid; the potential temperature $\theta(p, \mathcal{H}, T)$ on a regular relative humidity $\mathcal{H}$ and T grid; and the moist entropy $s_m(p, \mathcal{H}, T)$ on the same grid. We tabulate these three quantities on these regular grids in the supplementary material.

To compute the potential temperature θ , the dry potential temperature θ_d and the equivalent potential temperature θ_e we use p and T to find $q_{\text{sat}}(p, T)$ by bilinear interpolation on the regular p and T grid. We then use q_v to obtain the relative humidity $\mathcal{H}(p, q_v, T) = [(1 - q_v)^{-1} - 1][(1 - q_{\text{sat}})^{-1} - 1]^{-1}$. Using this relative humidity, we find $\theta(p, \mathcal{H}, T)$ and $s_m(p, \mathcal{H}, T)$ by trilinear interpolation on the regular $\mathcal{H}$ grid, T grid, and p grid. By setting $\mathcal{H}$ to 0%, we obtain the dry entropy $s_d = s_m(p, 0\%, T)$, allowing us to obtain the dry potential temperature as $\theta_d = T_0 \exp(s_d/c_p)$. Similarly, we obtain the equivalent potential temperature using the expression $\theta_e = T_0 \exp(s_m/c_p)$. The resulting θ_e takes into account the latent heat of sublimation and vaporization and deviates only slightly from the definition of Emanuel (1994) and the one used in LK13.

APPENDIX B

Propagated Equivalent Potential Temperature

For each day, the propagated equivalent potential temperature $\delta\theta_e|_{\text{prop}}$ is set to a uniform value along $\langle\theta_e\rangle$ surfaces. It is set to the $\bar{\theta}_e$ value where the $\langle\theta_e\rangle$ surface intersects the 850-hPa pressure surface. This is accomplished through the following procedure: We first find $\phi_{850}(\phi, p)$, the latitude at which the $\langle\theta_e\rangle$ surface passing through (ϕ, p) intersects the 850-hPa pressure surface. It is found by solving the equation $\langle\theta_e\rangle(\phi, p) = \langle\bar{\theta}_e\rangle(\phi_{850}, 850 \text{ hPa})$ for each (ϕ, p) , by linearly interpolating $\langle\bar{\theta}_e\rangle(\phi, 850 \text{ hPa})$ along ϕ . Then, for every day t and at each point in the latitude and pressure plane (ϕ, p) we set $\theta_e^{\text{prop}}(t, \phi, p) = \theta_e[t, \phi_{850}(\phi, p), 850 \text{ hPa}]$. Finally, we define anomalies as $\delta\theta_e|_{\text{prop}} = \theta_e^{\text{prop}} - \langle\bar{\theta}_e\rangle$.

APPENDIX C

Computing the Different Terms in Moist Entropy Budget

The model layer thickness Δp is calculated as the difference between the pressures at layer edges. Using the method described in appendix A, the dry and moist entropies $\ln(\theta_d/T_0)$ and $\ln(\theta_e/T_0)$ are computed at the C-grid center with the midlevel center pressure as the

average between the top and bottom layer edge values, allowing us to compute their mass-weighted zonal-mean $\overline{\ln(\theta_d/T_0)\Delta p}$ and $\overline{\ln(\theta_e/T_0)\Delta p}$. The moist and dry entropies are regridded to the C-grid edge using a bilinear remapping and are linearly interpolated onto the model-level layer edge. The zonal-mean fluxes of dry entropies in the meridional $\overline{\ln(\theta_d/T_0)v\Delta p}$ and in the vertical $\overline{\ln(\theta_d/T_0)\bar{\omega}}$ directions are then obtained. The zonal-mean 3-hourly discrete time derivative in entropy contents $\Delta_{3\text{hr}}\overline{\ln(\theta_e/T_0)\Delta p}$ and $\Delta_{3\text{hr}}\overline{\ln(\theta_e/T_0)\Delta p}$ are also computed. The zonal-mean entropy forcing from analysis $\Delta p\dot{q}_{\text{ana}}/T$, radiation $\Delta p\dot{q}_{\text{rad}}/T$, friction $\Delta p\dot{q}_{\text{fri}}/T$, gravity wave drag $\Delta p\dot{q}_{\text{gwd}}/T$, turbulence $\Delta p\dot{q}_{\text{trb}}/T$, and moist physics $\Delta p\dot{q}_{\text{mst}}/T$ are obtained using the corresponding regridded temperature tendencies.

To present our results, we convert the units of RADIATION, NUDGING, BOTTOM, TOP, and MERIDIONAL from joules per kelvin per second to kelvin per second by multiplying them by the following conversion factor:

$$\alpha = \frac{T_0}{c_p} \exp(\langle\{\overline{\ln(\theta_e|_{\text{AMT}}/T_0)\Delta p}\}_{\text{AMT}}\rangle_{\text{August}}) = \frac{305.98 \text{ K}}{1003 \text{ J K}^{-1}}. \quad (\text{C1})$$

We use the same conversion factor to obtain equivalent potential temperature anomalies from moist entropy anomalies,

$$\delta\{\theta_e|_{\text{AMT}}\}_{\text{July}} = \alpha c_p \delta\{\overline{\ln(\theta_e|_{\text{AMT}}/T_0)\Delta p}\}_{\text{AMT}}\}_{\text{July}} \quad \text{and}$$

$$\delta\{\theta_e|_{\text{AMT}}\}_{\text{August}} = \alpha c_p \delta\{\overline{\ln(\theta_e|_{\text{AMT}}/T_0)\Delta p}\}_{\text{AMT}}\}_{\text{August}}.$$

The zonal-mean meridional fluxes of dry and moist entropies ($\overline{\ln(\theta_d/T_0)v\Delta p}$ and $\overline{\ln(\theta_e/T_0)v\Delta p}$) are obtained and then used to compute the zonal-mean vertical fluxes of moist entropies $\overline{\ln(\theta_e/T_0)\bar{\omega}^*}$ as a residual from the moist physics entropy forcing. This is accomplished by subtracting Eq. (2) from Eq. (3) and integrating the following residual equation from the model layer edge at 375 hPa down to the layer edge at 600 hPa for every latitude:

$$\Delta_p [\overline{\ln(\theta_e/T_0)\bar{\omega}^*} - \overline{\ln(\theta_d/T_0)\bar{\omega}}] = \frac{1}{c_p} \overline{\Delta p\dot{q}_{\text{mst}}/T} - \Delta_{3\text{hr}} \overline{\ln(\theta_e/\theta_d)\Delta p} - \Delta_\phi \overline{\ln(\theta_e/\theta_d)v\Delta p}, \quad (\text{C2})$$

where Δ_p and Δ_ϕ represent cell-centered vertical and meridional discrete derivatives on the C grid, respectively. This correction captures the effect of convective processes that are not resolved at the 3-hourly time scale. This

correction is usually small in the climatology but important to obtain the right interannual variability.

Finally, we compute normalized anomalies of 3-hourly $[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}$ as follows: At each 3-hourly time step over the months of July and August, we obtained the climatological anomaly by subtracting the multiyear-mean meridional entropy fluxes,

$$\delta[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}} = [\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}} - \langle [\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}} \rangle.$$

This removes the intraseasonal cycle. To remove any trends, we then remove the mean climatological anomaly in July and August for every year,

$$\delta[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}|_{\text{detrend}} = \delta[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}} - 0.5(\delta\{[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}\}_{\text{August}} + \delta\{[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}\}_{\text{July}}).$$

Finally, for each year we normalize $\delta[\ln(\theta_e/T_0)v\Delta p]_{\text{AMT}}^{75\text{N}}|_{\text{detrend}}$ using its standard deviation over the months of July and August of that year. We perform the same procedure to obtain the normalized anomalies of moist entropy $\ln(\theta_e/T_0)\Delta p$ and specific humidity $q_v\Delta p$ averaged from 75° to 90°N on the model level.

REFERENCES

- Alexeev, V., P. Langen, and J. Bates, 2005: Polar amplification of surface warming on an aquaplanet in “ghost forcing” experiments without sea ice feedbacks. *Climate Dyn.*, **24** (7–8), 655–666, doi:[10.1007/s00382-005-0018-3](https://doi.org/10.1007/s00382-005-0018-3).
- Ambaum, M. H. P., 2010: The first and second laws. *Thermal Physics of the Atmosphere*, John Wiley & Sons, 17–37.
- Cullather, R. I., and M. G. Bosilovich, 2012: The energy budget of the polar atmosphere in MERRA. *J. Climate*, **25**, 5–24, doi:[10.1175/2011JCLI4138.1](https://doi.org/10.1175/2011JCLI4138.1).
- Eckhardt, S., A. Stohl, H. Wernli, P. James, C. Forster, and N. Spichtinger, 2004: A 15-year climatology of warm conveyor belts. *J. Climate*, **17**, 218–237, doi:[10.1175/1520-0442\(2004\)017<0218:AYCOWC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<0218:AYCOWC>2.0.CO;2).
- Emanuel, K., 1994: *Atmospheric Convection*. Oxford University Press, 592 pp.
- European Centre for Medium-Range Weather Forecasts, cited 2013: ERA-Interim project. CISL Research Data Archive. [Available online <http://rda.ucar.edu/datasets/ds627.0/>.]
- Feistel, R., and Coauthors, 2010: Numerical implementation and oceanographic application of the thermodynamic potentials of liquid water, water vapour, ice, seawater and humid air—Part 1: Background and equations. *Ocean Sci.*, **6**, 633–677, doi:[10.5194/os-6-633-2010](https://doi.org/10.5194/os-6-633-2010).
- Frierson, D. M. W., 2006: Robust increases in midlatitude static stability in simulations of global warming. *Geophys. Res. Lett.*, **33**, L24816, doi:[10.1029/2006GL027504](https://doi.org/10.1029/2006GL027504).
- Graversen, R. G., and M. Wang, 2009: Polar amplification in a coupled climate model with locked albedo. *Climate Dyn.*, **33**, 629–643, doi:[10.1007/s00382-009-0535-6](https://doi.org/10.1007/s00382-009-0535-6).
- , T. Mauritsen, M. Tjernström, E. Källén, and G. Svensson, 2008: Vertical structure of recent Arctic warming. *Nature*, **451**, 53–56, doi:[10.1038/nature06502](https://doi.org/10.1038/nature06502).
- Juckes, M. N., 2000: The static stability of the midlatitude troposphere: The relevance of moisture. *J. Atmos. Sci.*, **57**, 3050–3057, doi:[10.1175/1520-0469\(2000\)057<3050:TSSOTM>2.0.CO;2](https://doi.org/10.1175/1520-0469(2000)057<3050:TSSOTM>2.0.CO;2).
- Kay, J. E., M. M. Holland, C. M. Bitz, E. Blanchard-Wrigglesworth, A. Gettelman, A. Conley, and D. Bailey, 2012: The influence of local feedbacks and northward heat transport on the equilibrium Arctic climate response to increased greenhouse gas forcing. *J. Climate*, **25**, 5433–5450, doi:[10.1175/JCLI-D-11-00622.1](https://doi.org/10.1175/JCLI-D-11-00622.1).
- Laliberté, F., and P. J. Kushner, 2013: Isentropic constraints by midlatitude surface warming on the Arctic midtroposphere. *Geophys. Res. Lett.*, **40**, 606–611, doi:[10.1029/2012GL054306](https://doi.org/10.1029/2012GL054306).
- , T. Shaw, and O. Pauluis, 2012: Moist recirculation and water vapor transport on dry isentropes. *J. Atmos. Sci.*, **69**, 875–890, doi:[10.1175/JAS-D-11-0124.1](https://doi.org/10.1175/JAS-D-11-0124.1).
- Langen, P., and V. Alexeev, 2007: Polar amplification as a preferred response in an idealized aquaplanet GCM. *Climate Dyn.*, **29** (2–3), 305–317, doi:[10.1007/s00382-006-0221-x](https://doi.org/10.1007/s00382-006-0221-x).
- Messori, G., and A. Czaja, 2013: On the sporadic nature of meridional heat transport by transient eddies. *Quart. J. Roy. Meteor. Soc.*, **139**, 999–1008, doi:[10.1002/qj.2011](https://doi.org/10.1002/qj.2011).
- Pauluis, O., and I. M. Held, 2002: Entropy budget of an atmosphere in radiative–convective equilibrium. Part II: Latent heat transport and moist processes. *J. Atmos. Sci.*, **59**, 140–149, doi:[10.1175/1520-0469\(2002\)059<0140:EBOAAI>2.0.CO;2](https://doi.org/10.1175/1520-0469(2002)059<0140:EBOAAI>2.0.CO;2).
- , A. Czaja, and R. Korty, 2010: The global atmospheric circulation in moist isentropic coordinates. *J. Climate*, **23**, 3077–3093, doi:[10.1175/2009JCLI2789.1](https://doi.org/10.1175/2009JCLI2789.1).
- Rienecker, M. M., and Coauthors, 2011: MERRA: NASA’s Modern-Era Retrospective Analysis for Research and Applications. *J. Climate*, **24**, 3624–3648, doi:[10.1175/JCLI-D-11-00015.1](https://doi.org/10.1175/JCLI-D-11-00015.1).
- Screen, J. A., and I. Simmonds, 2010: The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature*, **464**, 1334–1337, doi:[10.1038/nature09051](https://doi.org/10.1038/nature09051).
- , C. Deser, and I. Simmonds, 2012: Local and remote controls on observed Arctic warming. *Geophys. Res. Lett.*, **39**, L10709, doi:[10.1029/2012GL051598](https://doi.org/10.1029/2012GL051598).
- Serreze, M. C., A. P. Barrett, and J. Stroeve, 2012: Recent changes in tropospheric water vapor over the Arctic as assessed from radiosondes and atmospheric reanalyses. *J. Geophys. Res.*, **117**, D10104, doi:[10.1029/2011JD017421](https://doi.org/10.1029/2011JD017421).
- Shaw, T. A., and O. Pauluis, 2012: Tropical and subtropical meridional latent heat transports by disturbances to the zonal mean and their role in the general circulation. *J. Atmos. Sci.*, **69**, 1872–1889, doi:[10.1175/JAS-D-11-0236.1](https://doi.org/10.1175/JAS-D-11-0236.1).
- Taylor, P. C., M. Cai, A. Hu, J. Meehl, W. Washington, and G. J. Zhang, 2013: A decomposition of feedback contributions to polar warming amplification. *J. Climate*, **26**, 7023–7043, doi:[10.1175/JCLI-D-12-00696.1](https://doi.org/10.1175/JCLI-D-12-00696.1).
- Wu, Y., and O. Pauluis, 2014: Midlatitude tropopause and low-level moisture. *J. Atmos. Sci.*, **71**, 1187–1200, doi:[10.1175/JAS-D-13-0154.1](https://doi.org/10.1175/JAS-D-13-0154.1).