

### Assessment of the changing role of lower tropospheric temperature advection under arctic amplification using a large ensemble climate simulation dataset

Masatake E. Hori<sup>1</sup> · Masakazu Yoshimori<sup>1</sup>

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### Abstract

The role of temperature advection in the Arctic lower troposphere under changing level of global warming is investigated using a large-ensemble climate simulation dataset. Taking the 30-year climatology of the non-warming simulation (HPB-NAT) as a reference, we examined the difference in temperature advection under changing basic states of the historical experiment (HPB) and 2 K and 4 K warming experiments (HFB-2K and HFB-4K) and decomposed them into terms related to dynamical changes, thermodynamical changes and the eddy term which is a covariance term related to the effect of submonthly transient eddies. Under the HPB experiment, it was found that the total change in advection hangs in a balance between the positive signal located along the sea-ice boundary in the North Atlantic and along the Eurasian continent driven by a stronger dynamical term and a negative signal in the thermodynamical term and eddy term. It is found that with the progression of global warming the dynamical term of advection increases due to changes in the large-scale atmospheric circulation, but the thermodynamical term and eddy term decrease due to weaker temperature gradient and increased sensible heat flux from the newly opened ice-free ocean, respectively. Atmospheric temperature advection terms related to large-scale atmospheric circulation partially cancels one another, and the relative importance of the eddy term diverging locally induced sensible heat from the newly opened ice-free ocean dominates as global warming progresses.

Keywords Arctic amplification · Temperature advection · Sea-ice · Large ensemble model simulation

### 1 Introduction

The near-surface temperature over the Arctic experiences an increased amount of warming compared to the global average under the elevated concentration of greenhouse gases. This Arctic warming phenomenon is known as Arctic amplification (AA). While early studies using global climate models predicted the emergence of AA under increased global warming (Manabe and Wetherald 1975), with strong seasonal and interhemispheric asymmetries in its structure (Manabe and Stouffer 1979), recent observations and model experiments have greatly enhanced our understanding of AA and the associated feedback processes (Taylor et al. 2022; Goosse et al. 2018).

Masatake E. Hori mehori@aori.u-tokyo.ac.jp AA is characterized by a large warming centered in the lower troposphere with an amplitude that is strongest in the fall and winter seasons (Serreze et al. 2009; Screen and Simmonds 2010). Many observational and modeling studies have identified various processes that contribute to AA, some of which are rapid sea ice retreat (Kumar et al. 2010; Lang et al. 2017; Dai et al. 2019), changes in oceanic heat transport (Nummelin and Hezel 2017) and atmospheric heat transport (Henderson et al. 2021; Yoshimori et al. 2017), and cloud radiative effect (Jenkins and Dai 2021).

One of the leading topics in the debate over the cause of AA is the relative importance of locally driven feedback against the remote transport of heat into the Arctic from the mid-latitudes (Pithan and Mauritsen 2014; Yoshimori et al. 2014). The most important locally induced feedback is the ice albedo feedback, in which the reduction in sea ice exposes the surface of the ocean, which in turn increased the upper heat content of the ocean during summer and is released to the atmosphere in the subsequent fall and winter seasons (Hall 2004; Taylor et al. 2013). Otherwise, the sea

<sup>&</sup>lt;sup>1</sup> Atmosphere and Ocean Research Institute, The University of Tokyo, Kashiwa, Chiba, Japan

ice cover insulates the atmosphere from this feedback and generally keeps the lower troposphere from warming (Landrum and Holland 2022). However, the role of large-scale atmospheric heat transport has been disputed and remains as a major source of uncertainty (Henderson et al. 2021). Graversen (2006) used an objective analysis dataset to point out that the atmospheric northward energy transport has increased in recent decades and plays a small but significant part in the Arctic surface air temperature trend. Screen et al. (2012) points out that, while the observed near-surface warming can be mostly explained by changes in sea ice concentration and thickness, remote changes in the sea surface temperature (SST) explain the warming aloft. Large uncertainty arises from the interaction of large-scale transport and accompanying local feedbacks, such as the role of moisture intrusion and the resultant downward longwave radiation from clouds, which acts to enhance AA (Lee et al. 2017).

Shifts in the mean state of the atmosphere due to AA can also act to change the large-scale propagation of waves and atmospheric transport itself. It has been argued that the weakened meridional temperature gradient near the surface reduces the strength of the upper-level jet stream, creating a wavier mean state (Francis and Vavrus 2012, 2015; Vavrus et al. 2017) that leads to a favorable condition for a blocking phenomenon over the Ural Mountains (Gong and Luo 2017; Luo et al. 2018). The increased frequency of blocking strongly influences the advection of warm air into the Arctic, altering not only the mean state of the region but also the location of the sea-ice boundary, thus creating a feedback that favors a weaker meridional temperature gradient (Kim et al. 2021). Other large-scale phenomena, such as Rossby wave propagation from tropical Madden-Julian Oscillation or the El Niño Southern Oscillation reaching the higher latitudes, also play a similar role in remotely changing the mean state of the Arctic (Yoo et al. 2012; Feng et al. 2017). These large-scale phenomena transport heat and moisture to the Arctic and sustain warming in areas with strong interactions between the mid-latitudes. Studies such as Clark et al. (2021) and Dahlke and Maturilli (2017) point out that the dry component of atmospheric circulation, namely the atmospheric temperature advection, also plays a role in AA

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through frequency change in dominant patterns of warming and accompanying anomalous winds.

To further extend the understanding of how atmospheric heat transport from remote regions and local air-sea interactions play a crucial role in the evolution of AA, we use a large ensemble model dataset with multiple experiments to quantify the role of atmospheric temperature advection in the lower troposphere and its relation to the local change in the mean state of the atmosphere. We aimed to address the following questions in this study. First, how does the horizontal temperature advection change under changing degree of global warming? Second, given that the Arctic atmosphere warms faster than the global average, is the change in horizontal temperature advection influenced more by the hemispheric change in temperature gradient or more locally driven by effects such as the dynamical change in the atmosphere? Finally, how does the change in sea ice affect the overall balance of such changes?

The data and methods used in this study are described in Sect. 2. The results of the atmospheric advection analysis are presented in Sect. 3, followed by discussion in Sect. 4 and conclusions in Sect. 5.

### 2 Data and methods

#### 2.1 Large ensemble model dataset

The dataset used in this study is the Database for Policy Decision-Making for Future Climate Change (d4PDF), which is a large ensemble climate simulation constructed using the Meteorological Research Institute's Atmospheric Global Circulation Model version 3.2 (MRI-AGCM3.2) (Mizuta et al. 2017; Fujita et al. 2019). The model has a horizontal resolution of 60 km and 64 vertical levels with the top at 0.01 hPa. Four sets of experiments were used in this study: a historical climate simulation (HPB), a non-warming simulation (HPB-NAT), and 2 K and 4 K future climate simulations (HFB-2K and HFB-4K, respectively). The experiment design is summarized in Table 1.

The historical simulation uses the observed monthly mean SSTs and sea-ice concentrations based on the

 Table 1
 Experiment design for the d4PDF dataset

Number of ensembles	SST and sea ice	GHG
100	Observed value for 1951–2010	Observed
100	Detrended to the average of 1900–1919	Value at 1850
54 (9 $\times$ 6 CMIP5 models)	Detrended observed SST + warming pat- tern between 1900–1919 and 2031–2050	Value at 2040 of RCP8.5
90 (15×6 CMIP5 models)	Detrended observed SST + warming pat- tern between 1900–1919 and 2080–2099	Value at 2090 of RCP8.5
	Number of ensembles 100 100 54 (9×6 CMIP5 models) 90 (15×6 CMIP5 models)	Number of ensemblesSST and sea ice100Observed value for 1951–2010100Detrended to the average of 1900–191954 (9×6 CMIP5 models)Detrended observed SST + warming pattern between 1900–1919 and 2031–205090 (15×6 CMIP5 models)Detrended observed SST + warming pattern between 1900–1919 and 2080–2099

Centennial Observation-Based Estimates of SST version 2 and climatological monthly sea-ice thickness from Bourke and Garrett (1987) as the lower boundary conditions and was run from 1951 to 2010. Observed value for globalmean concentrations of greenhouse gases are given for each year. Also, three-dimensional distributions of ozone from the MRI Chemistry-Climate Model (MRI-CCM; Deushi and Shibata 2011) and aerosols from the MRI Coupled Atmosphere–Ocean General Circulation Model, version 3 (MRI-CGCM3; Yukimoto et al. 2012) are used for forcing. Small SST perturbations on the order of analytical errors were added to construct 100 ensemble experiments. The non-warming simulation, which also consisted of 100 members, assumed that global warming has not taken place since the preindustrial period. Greenhouse gases were set to the estimated level of 1850, and aerosols such as sulfate, black carbon, and organic carbon were set to preindustrial values. The long-term trend in SST was removed to form a SST baseline averaged between 1900 and 1919, which did not have significant warming compared to the preindustrial level. Future simulations were constructed using the observational SST with the long-term trend removed, after which the climatological SST warming patterns for six Coupled Model Intercomparison Project Phase 5 (CMIP5) models were added. The 6 models are CCSM4 (Community Climate System Model version 4), GFDL (Geophysical Fluid Dynamics Laboratory), HadGEM2-AO (Hadley Centre Global Environmental Model version 2 AO), MIROC5 (Model for Interdisciplinary Research on Climate version 5), MPI-ESM-MR (Max Planck Institute for Meteorology Earth System Model MR), and MRI-CGCM3 (Meteorological Research Institute Coupled Atmosphere-Ocean General Circulation Model version 3). For each model, the difference in average SST for the historical experiment and Representative Concentration Pathway 8.5 (RCP8.5) experiments was taken between 1991-2010 and 2031-2050 (2080-2099) and used for the SST warming pattern of HFB-2K (HFB-4K). Each warming pattern is multiplied by a scaling factor to give a global-mean surface air temperature warming of 2 K (4 K) for HFB-2K (HFB-4K). Also, the greenhouse gas concentration was set to the level of 2040 (2090) under RCP8.5. For each model, a 60-year 9 (15) member ensemble experiment was conducted for HFB-2K (HFB-4K) with a total of 54 (90) ensembles.

It is important to note that even though the observational data used as the basis of the SST boundary condition has the inhomogeneity and uncertainty inherit to observational data, it nevertheless still incorporates the effect of air-sea interaction within the SST. The same can be said for the warming pattern of the 6 coupled models added to the historical data where each SST warming pattern can be seen to be the result of air-sea interaction. The atmospheric circulation calculated by the AGCM reflect the air-sea interaction through the boundary conditions. Therefore, disregarding the difference between atmospheric models of the 6 coupled models used for acquiring the SST warming pattern and the MRI-AGCM3.2, the temperature advection presented in our analysis is physically consistent with that of the fully coupled model, and the interpretation of results is not restricted to that of the AGCM. However, due to experimental design, the AGCM only simulates the atmospheric condition which is consistent with the given boundary condition, and an attribution of causality between the boundary condition and the atmosphere is not possible.

The use of d4PDF allows for the quantification of internal variability of the atmosphere using multiple ensembles. Also, the future change in atmospheric temperature advection can be quantified relative to the historical HPB experiment using the non-warming HPB-NAT experiment as a baseline for comparison. While the dataset was created using a single model (MRI-AGCM3.2), the d4PDF experiment framework has the advantage of addressing the multi-model uncertainty in future projections by the use of multiple SST/ sea-ice boundary conditions for the HFB-2K and HFB-4K experiments.

### 2.2 Daily climatology and horizontal temperature advection

Out of the 60 years of each experiment, we used the latter 30 years in daily resolution for our analysis. In the case of HPB this corresponds to the 1981–2010 climatology which is compared to the climatology of the same year under the preindustrial experiment (HPB-NAT). The climatology was calculated by removing the extra leap day during this period and retaining up to the third-order annual harmonics. Then, the February 29 leap day was re-inserted by taking the average of the climatological value of February 28 and March 1. Daily anomalies were obtained by removing the corresponding daily climatology from the daily data. Temperature advection was calculated using this daily dataset, after which the seasonal mean for December-January-February (DJF) and ensemble means over all members were taken. This study focused on the cold season of DJF, when the hemispheric meridional temperature gradient is largest, and thus the role of horizontal advection becomes strongest.

# 2.3 Decomposition of advection into contributing terms

We address the difference in horizontal temperature advection between each experiment, which can be written as the left-hand term of Eq. (1), where V and T represent the wind vector and air temperature, respectively;  $\Delta$  represents the difference between experiments, which is defined as the difference against the non-warming experiment HPB-NAT;  $\nabla_H$  represents the horizontal gradient; and the overbars represent the seasonal average over the DJF period.

Following the convention of Wang et al. (2019), we broke down the difference in horizontal temperature advection into two terms corresponding to the climatological mean field and the anomaly by inserting  $T = \tilde{T} + T'$ ,  $V = \tilde{V} + V'$ , where  $\tilde{T}, \tilde{V}$  represents the daily climatology of the 30-year period.

$$-\Delta\left(\overline{V}\cdot\nabla_{H}T\right) \approx -\Delta\left(\overline{\tilde{V}}\cdot\nabla_{H}\tilde{T}\right) - \Delta\left(\overline{V'}\cdot\nabla_{H}T'\right), \tag{1}$$
$$= -\left(\overline{\Delta\tilde{V}}\cdot\nabla_{H}\tilde{T}\right) - \overline{\tilde{V}}\cdot\Delta(\nabla_{H}\tilde{T}) - \overline{\Delta\tilde{V}}\cdot\Delta(\nabla_{H}\tilde{T}) - \overline{\Delta(V'}\cdot\nabla_{H}T')} \tag{2}$$

Since  $\overline{T'}$  and  $\overline{V'}$  are nearly equal 0 with a small residual due to Fourier smoothing, the difference in temperature advection can be written as Eq. (1). We call the first term the climatological term because it depends only on the mean state and represents the advection of the climatological temperature gradient by the climatological wind. The second term is called the eddy term, and it represents the advection of an anomalous temperature gradient by anomalous wind due to transient eddies. The approximation comes from a residual from the averages and represents the sub-seasonal and interannual variabilities. This is also documented in Wang et al. (2019) and is small compared to the leading terms.

Expanding the first term yields Eq. (2), in which the climatological term is broken down into three contribution components. The first term represents the temperature advection due to changes in climatological wind and is called the dynamical term. Similarly, the second term represents the change in temperature advection due to changes culated in this study includes the contribution of both the sub-monthly timescale and the inter-annual timescale. However, the contribution of inter-annual variability is negligible compared to the sub-monthly timescale (figure not shown). Therefore, the eddy term can be regarded as the contribution of transient eddies and not to be confused with stationary eddies of the atmosphere.

in the temperature gradient and is referred to as the thermo-

dynamical term. The third term is called the second-order

term because it represents the advection of a change in the

temperature gradient due to a change in the wind field. Each

term was calculated at a daily timescale, after which it was

One detail to be noted is that the way climatology is cal-

averaged over the season and over multiple ensembles.

### **3 Results**

## 3.1 Features of lower tropospheric temperature change

To investigate the representation of AA within the d4PDF dataset, the linear-trend coefficient of the 2-m temperature for the 52-year period of 1959–2010 for the HPB experiment is compared to the corresponding years of ERA5 (Hersbach et al. 2017; Bell et al. 2021) and JRA55 produced by the Japan Meteorological Agency (JMA; Kobayashi et al. 2015;



**Fig. 1** Linear trend in DJF average 2-m temperature for 1959–2010 using the **a** 100-member ensemble of the HPB experiment of d4PDF, **b** ERA5 dataset, and **c** JRA55 dataset. Dotted areas signify regions

where the difference is statistically significant at a 95% level using a two-tailed Student's t test

Harada et al. 2016) in Fig. 1. Both ERA5 and JRA55 show a strong warming in the North Atlantic along the Fram Strait and Barents-Kara Sea. The ensemble mean trend in HPB also shows this feature, albeit with a more zonal structure and stronger warming in the Barents-Kara Sea and central Arctic. A weak negative trend can be seen in HPB along the Canadian archipelago. This is due to a cooler representation of temperature during the 1990s in the region and marks a stark contrast against the observation. Still, the trend coefficient becomes smaller in the region for both ERA5 and JRA55, resulting in an east-west gradient of warming over the Arctic. The focus of the following results is centered on the difference observed along the North Pacific and along the Eurasian continent and North Pacific. The influence of this negative bias in the d4PDF dataset is discussed further in Sect. 4.

Next, we investigated the representation of AA, including the experiments on the future climate. Figure 2 shows the difference in 2-m temperature between the ensemble mean DJF climatology of HPB-NAT and other experiments. Note that the range of the shading differs between each experiment to highlight the regional features.  $\Delta$ HPB shows significant warming in most areas, with a more pronounced signal over the Eurasian continent and over the Arctic Sea. The largest warming exists over the Barents-Kara Sea, where the sea ice is retreating. A weak negative signal in the Beaufort Sea is due to a difference in the seasonal evolution during late February, where the temperature in HPB-NAT becomes warmer but HPB remains cold. This may be due to a model bias, but it is mostly statistically insignificant. The contrasting feature of a strong warming over the Eurasian continent and a weaker warming over 90°-180° W is consistent with observational findings (Xiao et al. 2020; Hu et al. 2021).  $\Delta$ HFB-2K shows a similar feature with increased warming over the Barents-Kara Sea and central Arctic Sea. Warming also becomes stronger along the coast of the Eurasian continent and the Bering Strait. The rate of warming in the coastal region increases in  $\Delta$ HFB-4K, which coincides with the decrease in sea ice in this region. Compared to the HPB-NAT experiment, the difference in the global average 2-m temperature is + 0.72 °C in HPB, + 2.10 °C in HFB-2K, and + 4.30 °C in HFB-4K. The Arctic average defined as the area average north of 60° N experiences a + 1.20 °C warming for HPB, + 4.39 °C warming for HFB-2K, and + 10.81 °C warming for HFB-4K.

Because we used a large ensemble dataset with multiple boundary conditions, it was important to verify that the temperatures over the analysis domain are comparable to each other, and that taking an ensemble mean is a reasonable representation of the experiment. To elucidate this point, we took the 2-m temperature averaged over the region of  $60^{\circ}$ -90° N for all ensembles within each experiment and



Fig.3 Probability density distribution of DJF averaged-area average temperature for  $60-90^\circ$  N produced with a Gaussian kernel density estimation



**Fig. 2** Difference in DJF mean 2-m temperature for **a** HPB, **b** HFB-2K, and **c** HFB-4K against the baseline climatology of HPB-NAT. Dotted areas signify regions where the difference is statistically significant at a 95% level using a two-tailed Student's *t* test

presented them as a probability density using the Gaussian kernel density estimation. Here, the optimal bandwidth was estimated by Scott's rule of thumb using the SciPy toolkit (Virtanen et al. 2020). It should be noted that the seasonal mean for the 30-year period was used over all 100 ensembles for HPB and HPB-NAT, 54 ensembles for HFB-2K, and 90 ensembles for HFB-4K. The result is shown in Fig. 3.

Overall, the probability density has a single distinctive peak for each experiment, which validates the use of ensemble means as a representation. The median value for the historical experiment HPB is - 19.43 °C with a standard deviation of 0.93 °C, whereas the non-warming experiment has a median value of - 20.60 °C and a standard deviation of 0.80 °C. The slightly larger variance of HPB was expected because the historical experiment is not detrended and includes the long-term change in temperature. The probability density function for the warming experiment HFB-2K (HFB-4K) has a medium value of - 16.13 °C (-9.83 °C) and a standard deviation of 0.83 °C (0.68 °C). The distribution of HFB-2K is slightly skewed with a mode of - 16.06 °C. This is because one of the models has a smaller SST warming pattern compared to the rest of the models for HFB-2K. However, the difference in the ensemble mean is not significant, and the results are qualitatively the same even with this model removed. Therefore, we used all 54 ensembles in HFB-2K to obtain our results.

### 3.2 Changes in temperature advection in progression of global warming

We now turn our attention towards the feature of the temperature advection in the lower troposphere. Figure 4 shows the ensemble mean horizontal temperature advection for the 925-hPa level for HPB-NAT averaged over the DJF season. We focused on 925 hPa, the second-lowest pressure level, because the lowest level in the dataset (1000 hPa) is mostly undefined below the surface pressure level, and the 925-hPa level captures the effect of turbulent heat flux from the ocean more accurately. The following results were also checked for the third-lowest 850-hPa level, and the basic features and differences remained.

Figure 4a shows that the climatology of the overall advection is positive over the Arctic Sea and the Eurasian continent, while negative advection is seen over the North Atlantic and Bering Strait. Locations with positive advection correspond to where the local temperature is warmed by horizontal advection. In contrast, regions with negative advection correspond to locations where incoming sensible heat from the ocean is dissipated by the atmosphere via the eddy term, creating a net negative advection.

In regions over the ice-free ocean, the atmosphere is cold relative to the warm sea surface and the sensible heat flux is positive (figure not shown). The situation is reversed over the sea ice and continents, where the atmosphere is being cooled against the cold surface. From an atmospheric point of view, the atmospheric heat transport over the icefree ocean is divergent, dissipating the incoming sensible heat flux from the ocean through advection due to transient eddies which turns the eddy term negative. The opposite can be said over the sea ice and continents where the atmosphere loses heat against the cold surface whereby the atmospheric heat transport due to transient eddies becomes convergent and the eddy term becomes positive.

In the North Atlantic, the boundary between positive and negative total advection closely follows the sea-ice



**Fig. 4** Ensemble mean DJF climatology of the **a** total advection, **b** climatological term, and **c** eddy term for HPB-NAT. The area-averaged value over the Arctic region north of  $60^{\circ}$  N is also shown in each panel

boundary. Even though a large advection is associated with the climatological southwesterly in the region, the in-situ advection becomes negative due to the cancellation by the eddy term. In the Bering Strait, the sea-ice boundary is located near 60° N, and the negative total advection in this region corresponds not only to the open sea but the negative temperature advection due to the climatological northerly wind. The climatological term shows that the atmosphere is indeed warmed over the North Atlantic and the western half of the Eurasian continent due to the strong climatological southwesterly coming from the mid-latitudes. The Pacific side of the Arctic Sea and the Bering Strait shows a negative signal where locally formed cold air is advected southwards by the climatological wind. While the overall pattern of total advection is governed by the eddy term, the climatological term also has an effect of 0.3 K/day over the 60°-90° N region, amounting to almost half of the total advection over the Arctic.

To show the effect of global warming and the accompanying AA on temperature advection, we show the difference in total advection, the climatological term, and the eddy term for the HPB, HFB-2K, and HFB-4K experiments. We used the non-warming experiment HPB-NAT as the baseline climatology to quantify the contributing factors with Eq. (1). In Fig. 5a, the difference in total advection for the HPB experiment shows a distinct positive/negative dipole in the North Atlantic through the Barents-Kara Sea. This corresponds to regions where the sea-ice concentration has receded in recent decades and the location of the negative signal agrees with the newly opened ocean. Also, a large positive difference in advection can be seen near the Bering Strait and in the Beaufort Sea, which correspond to an increase in the



Fig. 5 Difference in  $\mathbf{a}$ ,  $\mathbf{d}$ ,  $\mathbf{g}$  total advection,  $\mathbf{b}$ ,  $\mathbf{e}$ ,  $\mathbf{h}$  climatological term, and  $\mathbf{c}$ ,  $\mathbf{f}$ ,  $\mathbf{i}$  eddy term for the HPB, HFB-2K, and HFB-4K experiments using the non-warming HPB-NAT experiment as the baseline climatology. Indices at the top right show the area-averaged value

over  $60^{\circ}$ – $90^{\circ}$  N. Dotted areas signify regions where the difference is statistically significant at a 95% level using a two-tailed Student's *t* test

climatological term. While the signal in the Bering Strait corresponds to a weaker northerly in the region (discussed later), the signal over the Beaufort Sea should be interpreted with caution because of the negative bias in the climatology, as seen in Fig. 2a.

In Fig. 5b, a large area of warming can be seen in the climatological term extending from the North Atlantic into the Arctic Sea. A strong negative signal exists in the western side of Greenland in the Canadian archipelago near 60° W. This corresponds to the strengthening of the climatological northerly advecting cold air into the mid-latitudes. The eddy term (Fig. 5c) shows a negative signal over the North Atlantic Sea and Bering Strait, which agrees with the reduced sea ice in the region. Here, the upward sensible heat flux from the sea surface increases where sea ice retreats (discussed later in Fig. 9). Simultaneously, the position of the maximum temperature gradient also moves along with the shift of the ice edge thereby changing the atmospheric heat transport by transient eddies which acts to eliminate the temperature gradient. Heat transport decreases or increases where the temperature gradient becomes smaller or larger, respectively. Therefore, horizontal heat transport where sea ice has retreated becomes divergent due to transient eddies which appears as a negative signal in the eddy term.

While regional features dominate in the advection field, it is still useful to compare the area average value to understand how the temperature advection contributes to AA. The area-averaged advection increased under HPB by +0.051 K/ day with the contribution of the climatological term being larger than that of the eddy term at +0.073 K/day.

The role of atmospheric advection changes drastically for the experiments on future climate. For the HFB-2K experiment (Fig. 5d), the difference in total advection over the Arctic becomes - 0.024 K/day, with a negative difference in advection extending from the North Atlantic through the Bering Strait along the coast of the Eurasian continent. The climatological term (Fig. 5e) still plays a role in warming the atmosphere, but the location of the positive signal is shifted northward into the Arctic Sea, and a negative signal emerges over the Russian plain and north of the Scandinavian Peninsula. The difference in total advection over the Bering Strait now becomes negative, with the eddy term overcoming the positive advection of the climatological term. The overall feature of the eddy term (Fig. 5f) resembles that of the total advection and now plays the dominant role at - 0.080 K/ day averaged over the Arctic. HFB-4K shows these features strengthening, with stronger negative advection in the Barents-Kara Sea and Bering Strait, and along the Eurasian coast (Fig. 5g). The positive signal in the climatological term (Fig. 5h) becomes weaker and shifts further to the north, while the negative signal becomes stronger. The areaaveraged difference in total advection becomes - 0.18 K/ day with the climatological term having a diminished role of 0.02 K/day and the difference in eddy term dominating at - 0.17 K/day (Fig. 5g-i).

From an atmospheric temperature advection perspective, the results show that its contribution to Arctic warming is largely inhomogeneous and is localized near the sea-ice boundary both in the Atlantic and Pacific regions, albeit with a difference in latitudes. The potential of atmospheric temperature advection warming the Arctic weakens as global (or Arctic) warming becomes stronger. While the atmospheric circulation pattern may show a dominant mode, such as the Arctic Oscillation (AO), the ability of the accompanying advection to warm the region is actually weaker when locally induced eddy activity takes over.

### 3.3 Contributing terms in the temperature advection

To elucidate the contributing factors in the difference in temperature advection, the climatological term was further separated into the dynamical, thermodynamical, and second-order terms using Eq. (2). Figure 6a–c shows that the difference in climatological term for HPB is mainly governed by the dynamical term.

In the North Atlantic, this corresponds to the climatological temperature gradient advected by the increase in southwesterly wind in the region (discussed later in Fig. 7) resembling the positive phase of the AO. Similarly, a southerly anomaly in the Bering Strait (also discussed later in Fig. 7) is responsible for the positive dynamical term in the region. A weak negative signal in the thermodynamical term in the Atlantic Sea and the western half of the Eurasian continent is related to the weakening of the temperature gradient. The second-order term has a smaller value in the order of magnitude and is considered negligible. Under the future simulation, the contribution of the dynamical term increases further (Fig. 6d, g), with increased advection along the North Atlantic as well as the Bering Strait. However, weakening of the temperature gradient in HFB-2K and HFB-4K leads to a strong negative advection in the thermodynamical term (Fig. 6e, h), canceling the effect of the dynamical term. A statistically significant warming in the thermodynamical term is visible in the central Arctic as well as the Bering Strait. This is due to an increase in the temperature gradient, which follows the movement of the ice edge northward into the central Arctic. These multiple features of the dynamical term and thermodynamical term cancel or enhance each other to shift the positive signal in the climatological term northward (as seen in Fig. 5e, h).

Figure 7 shows the ensemble mean difference in sea level pressure (SLP) and 925-hPa wind for each experiment. The SLP shows a positive anomaly in the mid-latitudes over the North Atlantic and North Pacific, as well as a negative anomaly over the Arctic in all experiments resembling the



**Fig. 6** Difference in **a**, **d**, **g** dynamical term, **b**, **e**, **h** thermodynamical term, and **c**, **f**, **i** second-order term for the HPB, HFB-2K, and HFB-4K experiments based on the formulation of Wang et al. (2019). Dot-

ted areas signify regions where the difference is statistically significant at a 95% level using a two-tailed Student's *t* test

positive phase of the Arctic Oscillation (AO). Such a feature is well known under the global warming experiments, such as CMIP6, albeit with strong inter-model variability (IPCC 2021). An increase in the meridional component of wind can be seen over the North Atlantic and the Bering Strait. A strong southerly anomaly becomes more evident in HFB-4K along the Scandinavian Peninsula and the Eurasian coastline of 30-70° E, and a northerly anomaly in East Siberia around 150-180° E also becomes stronger. While the overall feature of the SLP resembles the AO, a cyclonic anomaly appears as a regional feature over the Bering Strait for HFB-2K and HFB-4K (Fig. 7b, c). This corresponds to a model response where the Aleutian low is weakened and/or shifted northward under global warming and is consistent with past studies (Hori and Ueda 2006). The positive signal in the dynamical term observed over the North Atlantic (Fig. 6a, d, g) corresponds to the change in SLP and wind where a more AO-like response strengthens the southwesterly in the North Atlantic. Lower pressure in the Bering strait which corresponds to the northern shift of the Aleutian low contributes to the change in the dynamical term (Fig. 6a, d, g), but the induced change in local temperature gradient cancels this effect (Fig. 6b, e, h) and the combined effect of advection becomes weaker in the Bering Strait as global warming becomes stronger (Fig. 5b, e, h). It should be stressed that while the atmospheric circulation pattern may show a dominant mode, such as the AO, the ability of the accompanying advection to warm the region is actually weaker when locally induced eddy activity takes over.

To clarify the balance of the dynamical and thermodynamical terms and their contribution to the climatological term in the North Atlantic, a latitude section of the



**Fig. 7** Ensemble mean difference in DJF average climatological SLP (shadings) and 925-hPa wind (vectors) between **a** HPB, **b** HFB-2K, and **c** HFB-4K. For clarity, wind vectors are shown only for differences exceeding 0.5 m/s

difference between HPB-NAT averaged over 30° W-90° E for each experiment is shown in Fig. 8. It should be noted that the shading represents the standard error, which shows the accuracy of the ensemble mean in each experiment. While the difference in the climatological term for HPB has a peak at 77.5 °N, the location of the peak is shifted northward to 80.0 °N for HFB-2K and 82.5 °N for HFB-4K. This is because the change in temperature gradient in the high Arctic creates a stronger peak in the northward shift of the thermodynamical term with increased Arctic warming. The peak of the dynamical term is located at 75 °N and does not change between experiments. This corresponds to the stronger meridional wind accompanying the AO-like response in HFB-2K and HFB-4K. The negative signal of the thermodynamical term south of 75-80 °N becomes stronger for the future climate, which signifies a weakened temperature gradient contributing to less advection in the region.

The difference in atmospheric advection is closely tied to the underlying sea-ice concentration and sensible heat flux from the open sea. Figure 9 shows the difference in sensible heat flux defined as upward positive and sea-ice concentration for each experiment. For HPB, an increase of sensible heat flux along the Greenland coast and Barents-Kara Sea near 75° N corresponds to the decrease in sea ice in the same region. In the future climate experiments, the increase in sensible heat is pronounced over the Barents-Kara Sea and over the Barents Strait for HFB-2K and HFB-4K, which is in good agreement with the reduction of sea ice. In the open sea region of the North Atlantic, especially south to the seaice boundary, a strong negative difference in sensible heat is visible. This is because the warming of the atmosphere under the future experiment suppresses the heat flux coming out of the already open sea, thus creating a negative difference compared to that of HPB-NAT.



**Fig.8** Latitude cross section of the difference in the **a** climatological term, **b** dynamical term, and **c** thermodynamical term for each experiment averaged over  $30^{\circ}$  W– $90^{\circ}$ E. The second-order term is small and not shown. Lines show the ensemble mean, and the shading signifies the standard error

### **4** Discussion

The role of atmospheric temperature advection in the lower troposphere of the Arctic during winter is argued to be twofold. One is the large-scale advection of heat through the climatological term and the other is the dissipation of locally induced heat through the eddy term. As shown in Fig. 4, while the climatological temperature advection averaged over the whole Arctic is nearly split between the two terms, the strong regionality of the eddy term dominates the signal of the total advection.

Taking the difference between the basic state of HPB-NAT, the changes in temperature advection in the North Atlantic show a distinct dipole pattern along the sea-ice boundary of the Barents-Kara Sea. The anomaly pattern in Fig. 5a acts to shift the negative advection center in Fig. 4a northwards. Note that the climatological value of total advection in the North Atlantic is negative (Fig. 4a), and this difference amounts to less dissipation of heat due to transient eddies. This is due to the warmer atmosphere dampening the outgoing sensible heat flux from the ocean. The positive advection of the climatological term also shifts northward due to the balance between the contribution of the dynamical term driven by large-scale atmosphere pattern having an AO signature with the northern shift of the peak in the thermodynamical term. The climatological component of the advection competes with the eddy term, with the eddy term taking over once sea ice in the region retreats and sensible heat from the ocean becomes dominant. Over the region where sea ice has receded and newly opened sea exists, the difference in sensible heat flux is positive (Fig. 9a-c) and the corresponding total advection is negative due to the eddy term.

While the climatological term acts to warm the atmosphere in a large-scale pattern, the effect of locally induced heat from the ocean dissipating through the eddy term dominates. In contrast, regions still having sea ice experience a neutral or positive difference in total advection due to the climatological term and eddy term combined, meaning that the remote feedback of atmospheric temperature advection is more dominant in such regions.

The role of the dynamical and thermodynamical terms changes drastically under global warming. Stronger southerly winds over the ocean lead to stronger dynamical advection over the North Atlantic under global warming, which is consistent with the finding of Clark et al. (2021). However, the weakening of the temperature gradient, which is mainly in the meridional direction but also in the zonal direction along the Eurasian coastline, acts to cancel this effect through the thermodynamical term. The total picture of the climatological term is fluid, with the resultant warm advection shifting northward with stronger Arctic warming and negative advection becoming more evident in the lower latitudes of the Arctic.

Clark et al. (2021) found that the change in atmospheric temperature advection is governed by the increase in frequency for some self-organizing map patterns. Because our study separates the contribution eddy component from that of the climatological mean changes, a direct comparison between the two studies cannot be made. While there



Fig. 9 Difference in a-c sensible heat flux and d-f sea-ice concentration between each experiment and the baseline climatology of HPB-NAT

remains a possibility that the climatological mean change is caused by the frequency change in the sub-seasonal eddy variabilities, further investigation is needed to verify this point.

It should be noted that while the positive AO-like response driving the dynamical term is a robust feature under global warming throughout the CMIP5/CMIP6 models outputs (Shindell et al. 2001; Hori et al. 2007; IPCC 2021), there are various theories about the cause of such response, including the role of sea ice retreat in the Arctic or the influence of the stratosphere, among others. The cause of positive AO-like response within d4PDF dataset is considered to be the same but the precise mechanism is not understood. Whether the cause of a positive AO-like response lies in the SST boundary condition in the polar regions or in the low latitudes is an important question addressed in previous studies (Hoerling et al. 2001; Yamaguchi and Noda 2006; Landrum and Holland 2022), but it is beyond the scope of this study.

The area-averaged total atmospheric heat transport over the Arctic undergoes drastic change. Each component of the atmospheric advection, as well as the global and Arctic change in temperature, is summarized in Fig. 10. Here, we use the standard deviation for each term to emphasize the internal variability of the atmospheric temperature advection arising from the perturbation in the SST boundary condition. The significance of difference from the basic state of the HPB-NAT experiment is signified by the circles above each term. Under the HPB experiment, total advection, led by the dynamical term, is positive and the effect of the thermodynamical and eddy terms canceling this positive signal is still weak. Once global warming and the resultant AA intensifies, the dynamical term still increases linearly, but the canceling effect of the thermodynamical and eddy terms also increases. The total atmospheric temperature advection becomes negative, which means that the local process of airsea interaction dominates via the eddy term.

The advantage of using an ensemble dataset with a large number of members can be seen in Fig. 10a–c, where the change in the thermodynamical term falls within the internal variability of the HPB experiment and its effect is uncertain under the current objective reanalysis. Furthermore, future changes are robust against the standard deviation of all ensemble members.

It should be noted that while the change in total advection is positive for HPB and negative for future experiments of HFB-2K and HFB-4K, the precise timing of when the total advection turns negative within the model experiments or in the real world remains unknown. Further studies need to be undertaken with multiple emission scenarios to understand the time evolution of the balance among competing terms of atmospheric heat transport.



Fig. 10 Area-averaged difference in a-c advection terms and d-f 2-m temperature. For the advection terms, the average is taken over the Arctic at 60°–90° N. For the 2-m temperature, the annual mean global average value is shown along with the winter-averaged Arctic

mean. Error bars denote the standard deviations for all ensembles and the filled circles above some terms show that the difference is statistically significant at a 95% level using a two-tailed Student's *t* test

Another point to consider is the use of the AGCM in the d4PDF dataset, which may change the strength of the eddy term. It can be speculated that the high frequency air-sea interaction near the sea-ice boundary may dampen the effect of the eddy term when the ocean is warmed from the atmosphere and the total amount of sensible heat flux decreases. A sensitivity experiment using a coupled GCM with the air-sea interaction turned off may shed light on this issue. It should also be noted that while the boundary condition for HFB-2K and HFB-4K is derived from 6 CMIP5 models, the d4PDF dataset uses a single-model (MRI-AGCM3.2) for calculation. Therefore, a consideration of model bias should be made using multiple models.

It should be emphasized that this study does not explain the causality of Arctic Amplification per se but clarifies the role of atmospheric temperature advection under Arctic Amplification, how the feedback of temperature advection contributes to the maintenance of the Arctic Amplification through the balance of each term. Because the current study focused on the future change of horizontal atmospheric advection and did not quantify the role of vertical advection, diabatic heating and deposition of turbulent heat fluxes, which partially cancel each other, this study cannot accurately account for the total warming shown in Fig. 2. Furthermore, the effect of moisture advection and the resultant downward longwave radiation from water vapor and clouds is also an important factor that should be studied separately.

While the use of a large-scale gridded dataset for Arcticwide atmospheric temperature advection is inevitable, the effect of sub-grid-scale turbulent heat or a near-surface inversion layer, which models cannot represent accurately, should not be ignored. These factors may alter the near-surface stability and the mixed layer of the ocean, which may play a role in the aggregated effect of the transient eddies (Inoue and Hori 2011). Use of high-resolution objective analysis, such as the Arctic System Model version 2, may bring further insights but is constrained by short-term data availability and large uncertainty.

### 5 Conclusions

We analyzed multiple experiments with the large ensemble model dataset d4PDF to investigate the role of atmospheric temperature advection in the lower troposphere under a changed basic climate state and a strong AA.

The non-warming experiment HPB-NAT was taken as the baseline climatology. The historical experiment HPB used observed SST and sea-ice values. The future experiments HFB-2K and HFB-4K used the boundary condition of six CMIP5 models under a snapshot of the RCP8.5 scenario. HPB, HFB-2K, and HFB-4K all exhibited a robust response of AA throughout multiple ensembles. AA within the dataset is characterized by strong warming in the Barents-Kara Sea, the central Arctic, and the coastline of the Eurasian continent consistent with the change in sea ice.

Atmospheric temperature advection is characterized by a strong warming in the central Arctic and the Eurasian continent with a net cooling in the North Atlantic and the Bering Strait associated with the local dissipation of heat from the ocean. The climatological term signifying the advection of mean temperature gradient by the mean wind shows an overall positive value from the North Atlantic through western Eurasia. The eddy term associated with the local dissipation of heat through transient eddies shows a net cooling in the North Atlantic and dictates the overall structure of the total advection. Differences under global warming were investigated by taking the advection of the non-warming HPB-NAT experiment as the basic state and breaking down the difference in the climatological term into a dynamical term, where the mean temperature gradient is advected by changes in the wind field, and a thermodynamical term, where the change in temperature gradient is advected by the mean wind.

Under the HPB experiment, it was found that the total change in advection is governed by the competing effect between the atmospheric temperature advection terms with the stronger dynamical term along the sea-ice boundary in the North Atlantic and along the Eurasian continent partially canceled by the negative signal of the thermodynamical term and eddy term.

With the progression of global warming the dynamical term of advection increases due to changes in the largescale atmospheric circulation, but the thermodynamical term and eddy term decreases. While the decrease in the thermodynamical term is due to a weaker temperature gradient between the lower latitudes and the high Arctic, the decrease in eddy term is related to the increased sensible heat flux from the newly opened ice-free ocean. This change in balance between each term of the atmospheric heat transport enhances with the effect of transient eddies diverging the locally induced sensible heat from the icefree ocean dominating as global warming progresses. While the present result is limited to the use of a single AGCM, our results show that the balance of each term in the horizontal temperature advection under AA are heavily regulated by the underlying sea-ice boundary, and while the dynamical effect of advection becomes stronger under global warming, its contributing effect may be suppressed by the effect of transient eddies diverging the locally induced heat from the ocean. Future studies are needed to further clarify the effect of moisture advection and its feedbacks to fully understand the importance of atmospheric advection under AA.

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Author contributions All authors contributed to the study conception and design, and interpretation of the results. Material preparation, data collection, analysis and graphical representation were performed by MEH. The first draft of the manuscript was written by MEH and commented by MY. All authors read and approved the final manuscript.

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Availability of data and material The d4PDF dataset used in this study is maintained by DIAS and can be downloaded from http://d4pdf. diasjp.net/. Details of the dataset are also available at https://www.miroc-gcm.jp/d4PDF/. The monthly ERA5 data used in this study were downloaded from https://cds.climate.copernicus.eu/cdsapp/. The daily JRA55 data used in this study were downloaded from DIAS (http:// search.diasjp.net/ja/dataset/JRA55).

#### Declarations

**Conflict of interest** The authors have no competing interests to declare that are relevant to the content of this article.

Ethical approval and consent to participate Not applicable.

Consent for publication Not applicable.

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