



The Impact of CO₂-Driven Climate Change on the Arctic Atmospheric Energy Budget in CMIP6 Climate Model Simulations

OLIVIA LINKE 

JOHANNES QUAAS 

**Author affiliations can be found in the back matter of this article*

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ABSTRACT

The Arctic amplification is driven by several intertwined causes that are embedded in an overall changing energy balance of the atmosphere and ocean. We investigate the impact of quadrupled CO₂ concentrations on the Arctic atmospheric energy budget in CMIP6 models. The decomposition of the energy budget accounts for the atmospheric radiation budgets, the sensible and latent heat flux at the surface, and the convergence of atmospheric energy transport. The CO₂ response of these components is found to strongly depend on the Arctic season and underlying surface type. While the widespread Arctic radiative-advective equilibrium remains intact during boreal summer, profound changes are restricted to the winter season: Strongly increasing surface heat fluxes over areas of retreating sea ice are largely counteracted by dropping positive heat fluxes over open Arctic ocean. For retreating sea ice, the increase in the surface fluxes is stronger for a subset of climate models with weaker Arctic amplification. For these regions, we propose an intermediate transformation of the local radiative-advective equilibrium to a radiative-convective equilibrium. The wintertime changes in the components of the atmospheric energy budget strongly relate to alterations at the surface, concerning the modification of sea ice extent, surface temperature and stability. We find robust linear correlations for the mediating effect during winter. The energy transport convergence is derived as residual in our energetic framework as main mechanism to ensure the local energy budget. On a large scale, we find an overall decreasing transport convergence to balance the surplus energy from the surface which outruns the intensification of the Arctic radiation deficit in a warmer climate.

CORRESPONDING AUTHOR:

Olivia Linke

Leipzig Institute for
Meteorology, Universität
Leipzig, Leipzig, Germany

olivia.linke@uni-leipzig.de

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1. INTRODUCTION

The globally rising near-surface air temperature is driven by increasing anthropogenic greenhouse gas concentrations, with CO₂ playing the major role (IPCC, 2018). The temperature change is strongest in the Arctic, with a rate 2–3 times faster than the global average, a phenomenon which is commonly referred to as Arctic amplification (AA; Serreze and Barry, 2011; Wendisch et al., 2017). The recent warming trend in the northern polar regions is mostly confined to the surface, and particularly evident during boreal winter (Graversen et al., 2008; Bintanja and Van der Linden, 2013). The Arctic amplification is part of a highly interactive system and results from various operating causes and feedbacks, prominently involving the severe sea ice decline during past decades (Carmack et al., 2015).

Previous studies indicated a strong relation of the Arctic warming with the atmospheric energy budget (AEB), concerning the alteration of radiative and turbulent heat fluxes, and the transport of heat and moisture into the Arctic (e.g., Semmler et al., 2012; Mayer et al., 2016). The large-scale Arctic AEB is driven by a net annual radiation deficit at the top-of-the-atmosphere (TOA) which is mostly compensated by atmospheric and ocean heat transport, with the atmosphere playing the major role (Serreze and Barry, 2014; Mayer et al., 2019). Thereby, on a long-term annual basis, the large-scale AEB can be described as a radiative-advective equilibrium. The components of the Arctic AEB exhibit a pronounced seasonality which is mediated by the annual cycle of incoming solar radiation. In addition, the budget components show strong regional disparities across the Arctic, which is linked to the impact of variable surface types on the surface energy budget (Lauer et al., 2020).

This study provides a decomposition of the Arctic AEB and its response to radiative forcing and feedbacks through quadrupling atmospheric CO₂ concentrations. We seek to understand how local surface types and seasonality mediate the alteration of individual energy fluxes and their composition in context of the Arctic radiative-advective equilibrium. Our motivation is encapsulated in an improved understanding of how the alteration of the AEB in a warmer climate shapes high-latitude feedbacks driving the Arctic amplification (e.g., Feldl et al., 2020). The idea is that looking at the topic from an energetic framework will help to better understand AA in the future. In this study however, we are not explicitly investigating AA and climate feedbacks in the Arctic.

We use state-of-the-art climate models from the 6th phase of the Coupled Model Intercomparison Project (CMIP6 Eyring et al., 2016). CO₂-driven changes of the AEB components are derived by comparing idealised model experiments within CMIP6, namely the pre-industrial control (piControl) run and the climate change experiment with increasing CO₂ levels at a rate of 1% per year (1pctCO2).

Details about the CMIP6 experiments and the analysis approach are provided within Section 2. In a first step, the annual cycle of the large-scale Arctic AEB is derived from one single model in the control run (Section 3.1). We present the following results as multi-model means using 30-year averages and are limited to the region north of 70°N. Section 3.2 quantifies the impact of increasing CO₂ levels on individual AEB components by comparing both reference and warming scenario during boreal summer, winter, and as annual-average changes, respectively. We further analyse the spatial pattern of changes by deriving statistical relationships between the diagnostics through spatial correlations across climate models (Section 3.4). Principally, we seek to constrain the impact of reducing and retreating sea ice on the energetic fluxes by further impacting surface temperature and atmospheric stability.

2. DATA AND METHODS

Data from CMIP6 are used to investigate the impact of increasing CO₂ levels on the Arctic AEB and the interplay of individual components therein. To quantify changes in the AEB components, pairs of model simulations are used to account for the model responses under idealised CO₂ forcing. The pre-industrial control run provides the baseline of the climate system in near-equilibrium, with a forcing corresponding to pre-industrial conditions at around 1850. The 1pctCO2 experiment branches off the piControl run and simulates the climate response to a global annual-mean CO₂ concentration that is gradually increasing at a rate of 1% per year starting at pre-industrial levels prescribed by piControl.

In total, 12 CMIP6 models ([Table 1](#)) provide all required diagnostics for this study. All results are presented as climatological means of the last 30 years of a 150-year simulation when CO₂ levels have approximately quadrupled in the 1pctCO2 experiment with respect to piControl. This strong forcing scenario is chosen such that a large and unambiguous signal due to CO₂-induced warming is expected. We don't aim to derive a quantitatively realistic scenario here, but a clear signal and large signal-to-noise ratio allows us to draw conclusions for potential real-world warming scenarios.

We define four surface types, representing Arctic land areas, open ocean, sea ice, and sea ice retreat (SIR). Grid cells are classified as sea ice covered if the sea ice concentration (SIC) exceeds a threshold of 15% (e.g., Serreze et al., 2007). Grid cells with SIR are those that are identified as ice-covered ocean in the piControl run, but open ocean in the 1pctCO2 simulation.

The computation of surface-type averages applies model data at original spatial resolution. For the representation of spatial patterns, the data is regridded to a 64 × 128 Gaussian grid that equates to a horizontal resolution of approximately 250 km.

MODEL ACRONYM	RESOLUTION LAT × LON	REFERENCE
CNRM-CM6-1-HR	360 × 720	Voldoire et al. (2019)
CNRM-ESM2-1	128 × 256	Séférian et al. (2019)
EC-Earth3-Veg	256 × 512	Dööscher et al. (2021)
GISS-E2-1-G	90 × 144	Miller et al. (2021)
INM-CM4-8	120 × 180	Volodin et al. (2018)
IPSL-CM6A-LR	143 × 144	Boucher et al. (2020)
MIROC-ES2L	64 × 128	Hajima et al. (2020)
MPI-ESM-1-2-HAM	96 × 192	Tegen et al. (2019)
MPI-ESM1-2-HR	192 × 384	Müller et al. (2018)
MPI-ESM1-2-LR	96 × 192	Mauritsen et al. (2019)
MRI-ESM2-0	160 × 320	Yukimoto et al. (2019)
UKESM1-0-LL	144 × 192	Sellar et al. (2019)

Table 1 12 CMIP6 models that provide all diagnostics necessary for this study.

2.1. ENERGY BUDGET FRAMEWORK

The tendency in energy storage within an atmospheric column can be estimated as

$$\frac{\partial E_a}{\partial t} = R_a + Q_H - \nabla \cdot \mathbf{F}_a, \quad (1)$$

where the respective terms on the r.h.s. are the net atmospheric radiation budget R_a , the net turbulent heat flux at the surface Q_H , and the convergence of atmospheric energy transport $-\nabla \cdot \mathbf{F}_a$. We use Eq. 1 to describe the energy budget of any atmospheric column extending from the surface to the TOA. Each atmospheric column is linked to the underlying ocean column or land surface through the surface energy budget.

The radiation budget R_a is derived as the sum of the net downward radiative flux at the TOA and the upward radiative flux at the surface. R_a is further broken down into the contributions from atmospheric short-wave (SW_a) and long-wave (LW_a) components. The net turbulent heat flux at the surface is composed of sensible and latent heating, Q_h and Q_e , respectively. Positive values account for a net transfer of energy into the atmospheric column, i.e., positive vertical fluxes at the surface are directed upward from the ground, and downward through the TOA, respectively.

The horizontal convergence of energy transport can be computed as

$$-\nabla \cdot \mathbf{F}_a = -\nabla \cdot \int_0^{p_{srf}} (c_p T + gz + Lq + k) \mathbf{V} \frac{dp}{g}, \quad (2)$$

accounting for the integrated value from pressure $p = 0$ at the TOA to the surface pressure p_{srf} . The horizontal wind vector is denoted as $\mathbf{V}$. The total atmospheric energy is composed of all terms within the parentheses on the r.h.s. which are the internal, potential, latent, and kinetic

energy, respectively. The internal energy is the product of the specific heat of the atmosphere at constant pressure ($c_p = 1005 \text{ J K}^{-1} \text{ kg}^{-1}$) and the temperature T . The potential energy is the gravitational acceleration ($g = 9.81 \text{ ms}^{-2}$) multiplied by the height z . The latent energy is the product of the latent heat of vaporisation ($L = 2.5 \cdot 10^6 \text{ J kg}^{-1}$) and the specific humidity q (constants according to WMO, 1966). The kinetic energy is derived as $k = \frac{1}{2}(u^2 + v^2)$, with u and v , the horizontal wind components in zonal and meridional direction, respectively. The sum of internal and potential energy accounts for the dry static energy (DSE) which typically dominates the transport convergence within the Arctic (Serreze and Barry, 2014; Armour et al., 2019).

The tendency in atmospheric energy storage has been presented as part of the energy budget in Eq. 1, but it can be directly computed as

$$\frac{\partial E_a}{\partial t} = \frac{\partial}{\partial t} \int_0^{p_{srf}} (c_p T + \Phi_{srf} + Lq + k) \frac{dp}{g}, \quad (3)$$

where the term Φ_{srf} gives the surface geopotential which is not a function of the pressure coordinate (see e.g., Trenberth, 1997).

3. THE ARCTIC ATMOSPHERIC ENERGY BUDGET

3.1. BUDGET CLOSURE

A detailed energy balance analysis of the Arctic atmosphere is derived from the output of the IPSL-CM6A-LR model (Boucher et al., 2020). IPSL-CM6A-LR was the only model to provide all diagnostics at the necessary high temporal resolution to apply Eq. 2 and 3, respectively (see Appendix for more details). **Figure 1** gives the annual cycle of the pan-Arctic AEB from the control run, averaged over a 30-year period. **Figure 1** is supported by the respective seasonal quantities including uncertainty ranges in **Table 2**.

The energy storage tendency is close to zero on a long-term annual basis and shows a moderate seasonal variability throughout the year: During spring, the atmosphere gains energy when the incoming solar radiation increases at its strongest rate, and loses energy in fall when the solar radiation input declines.

During wintertime, the incoming short-wave radiation is close to zero and the long-wave radiation deficit is mostly compensated by advective heating. The turbulent heat fluxes are at their annual minimum in winter due to a widespread near-surface temperature inversion which limits the depth of vertical flux exchange (Serreze and Barry, 2014). During summertime, both the absorbed short-wave radiation and the long-wave radiative loss reach a seasonal maximum. The transport convergence is smaller due to weaker meridional temperature gradients between the Arctic and lower latitudes. The turbulent heat fluxes at the surface are mainly positive throughout

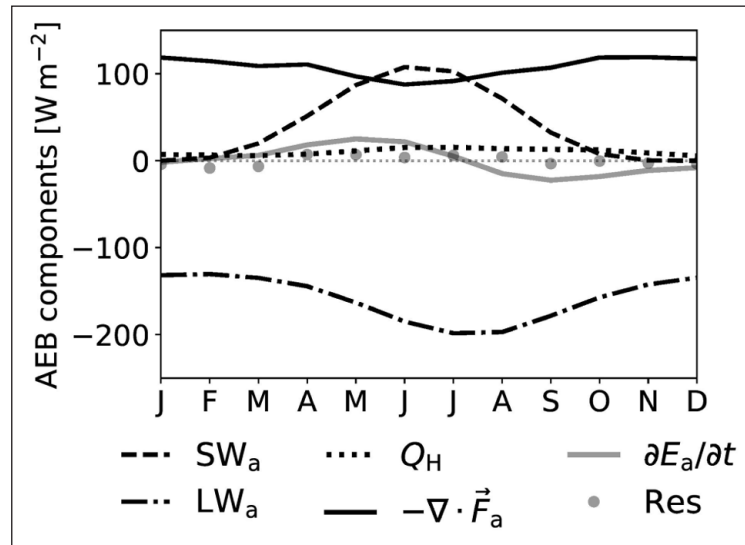


Figure 1 Annual cycle of the pan-Arctic AEB. All values are derived as 30-year climatological averages from piControl of the IPSL-CM6A-LR model. AEB components account for the atmospheric energy storage $\partial E_a / \partial t$, the atmospheric radiation budget (SW_a and LW_a contributions), the net surface turbulent heat flux ($Q_H = Q_h + Q_e$), and the convergence of atmospheric energy transport $-\nabla \cdot \vec{F}_a$. The residual is derived by summing up all budget components and subtracting the energy storage tendency.

CHANGES IN ENERGY FLUXES AND STORAGE TENDENCY [$W\ m^{-2}$]							
SEASON	SW_a	LW_a	Q_h	Q_e	$-\nabla \cdot \vec{F}_a$	$\partial E_a / \partial t$	RESIDUAL
MAM	52.8 ± 0.1	-147.4 ± 0.4	0.9 ± 0.2	7.3 ± 0.2	105.5 ± 1.2	16.5 ± 0.7	2.6 ± 1.0
JJA	94.0 ± 0.1	-193.5 ± 0.4	4.1 ± 0.1	10.7 ± 0.1	93.5 ± 0.8	4.1 ± 0.3	4.7 ± 0.7
SON	13.5 ± 0.0	-159.3 ± 0.4	1.5 ± 0.2	9.9 ± 0.2	114.9 ± 1.1	-17.3 ± 0.5	-2.2 ± 0.8
DJF	1.1 ± 0.0	-132.3 ± 0.6	-0.8 ± 0.3	7.2 ± 0.2	116.9 ± 2.1	-2.4 ± 1.0	-5.4 ± 1.2
Annual mean	40.4 ± 0.0	-158.1 ± 0.3	1.4 ± 0.1	8.8 ± 0.1	107.7 ± 0.7	0.2 ± 0.1	-0.1 ± 0.5

Table 2 Components of the pan-Arctic AEB from the IPSL-CM6A-LR model averaged over spring (MAM), summer (JJA), fall (SON), and winter (DJF), derived from 30-year averages 1970–1999 of piControl. The corresponding annual cycle is presented in Figure 1. Uncertainty intervals are calculated as the inter-annual standard error.

the year, but are of secondary importance in the large-scale AEB analysis.

The directly computed energy transport convergence as in Eq. 2 takes values between $94\ W\ m^{-2}$ in summer and $117\ W\ m^{-2}$ in winter. The annual-average value is $108\ W\ m^{-2}$, which is in good agreement with previous estimates of lateral energy transport into the Arctic (e.g., Serreze et al., 2007; Porter et al., 2011; Semmler et al., 2012).

The AEB residual is very small on a long-term annual basis which is an encouraging result for the applicability of Eq. 1. It should be noted that the residual shows a moderate seasonal signal with positive values during spring and summer and negative values during fall and winter. The seasonal residual can be related to the shortcomings of Eq. 1 which does not account for factors like the conversion between liquid water and precipitating ice (Mayer et al., 2019).

In the following we consider only summer, winter, and annual averages where the tendency in atmospheric energy storage is below 2% relative to the full budget and thus negligible. This substantially simplifies Eq. 1

and allows us to derive the transport convergence as residual of Eq. 1. Thereby, we link the energy transport convergence solely to the atmospheric radiation budget and the net surface turbulent heat flux. The simplified energetic framework allows us to assess the AEB from monthly-mean diagnostics available from 12 models and additionally avoids the significant lack of closure when using 6-hourly model data to estimate the transport term (for details see Appendix).

We further use this approach when assessing the AEB components in the $4 \times CO_2$ climate, although the gradual increase of CO_2 mediates a global TOA imbalance of $3.6\ W\ m^{-2}$ with respect to the control climate. However, due to the small heat capacity of the atmosphere, virtually no energy is stored in the atmosphere itself. Consequently, also in the warmer climate, the AEB is closed.

3.2. SEASONAL AND SURFACE-TYPE CONTROL ON THE ARCTIC ATMOSPHERIC ENERGY BUDGET

The AEB components from the control and perturbed run as well as their corresponding differences (1pctCO2

minus piControl) are presented in [Figure 2](#). Seasonal values are derived for boreal summer and winter as spatial averages over the predefined surface types. An additional discrimination gives one half of the model ensemble with weaker AA, and the other half with stronger AA, respectively. We define the AA factor as the difference in Arctic atmospheric surface temperatures between perturbed and control climate, divided by the corresponding global-mean difference.

CO₂-driven changes in the AEB are further analysed in terms of their geographical distributions as annual-mean values ([Figure 3](#)). The LW radiation budget is decomposed into contributions from the surface and TOA. [Figure 3](#) additionally accounts for annual-mean changes in SIC and atmospheric stability. The latter is defined as the difference between the atmospheric surface temperature and the maximum temperature within the atmospheric layer ranging from the surface to the 500-hPa pressure level. Note that the discrimination of individual SW

radiation fluxes at the surface and TOA was left out in [Figure 3](#). This matter will be discussed in the following.

During summertime, the AEB components in the control climate ([Figure 2a](#)) are very similar across different surface types, and the inter-model variability is small. The AEB is mainly regulated by the absorption of short-wave radiation (95 W m⁻²), the long-wave cooling (−190 W m⁻²), and the mediating convergence of energy transport (80 W m⁻²) within the Arctic. The turbulent heat fluxes contribute little to the energy balance in summer and are mainly directed from the surface towards the atmosphere. Only the sensible heat flux is slightly negative over the ice covered ocean (including SIR in piControl), where ice melt constrains the surface skin temperature at the freezing point of sea water, maintaining a surface-based temperature inversion (Serreze and Barry, 2014).

During winter ([Figure 2b](#)), the net short-wave radiation is at its annual minimum, and the long-wave radiation loss decreases by about one third, compared to summer,

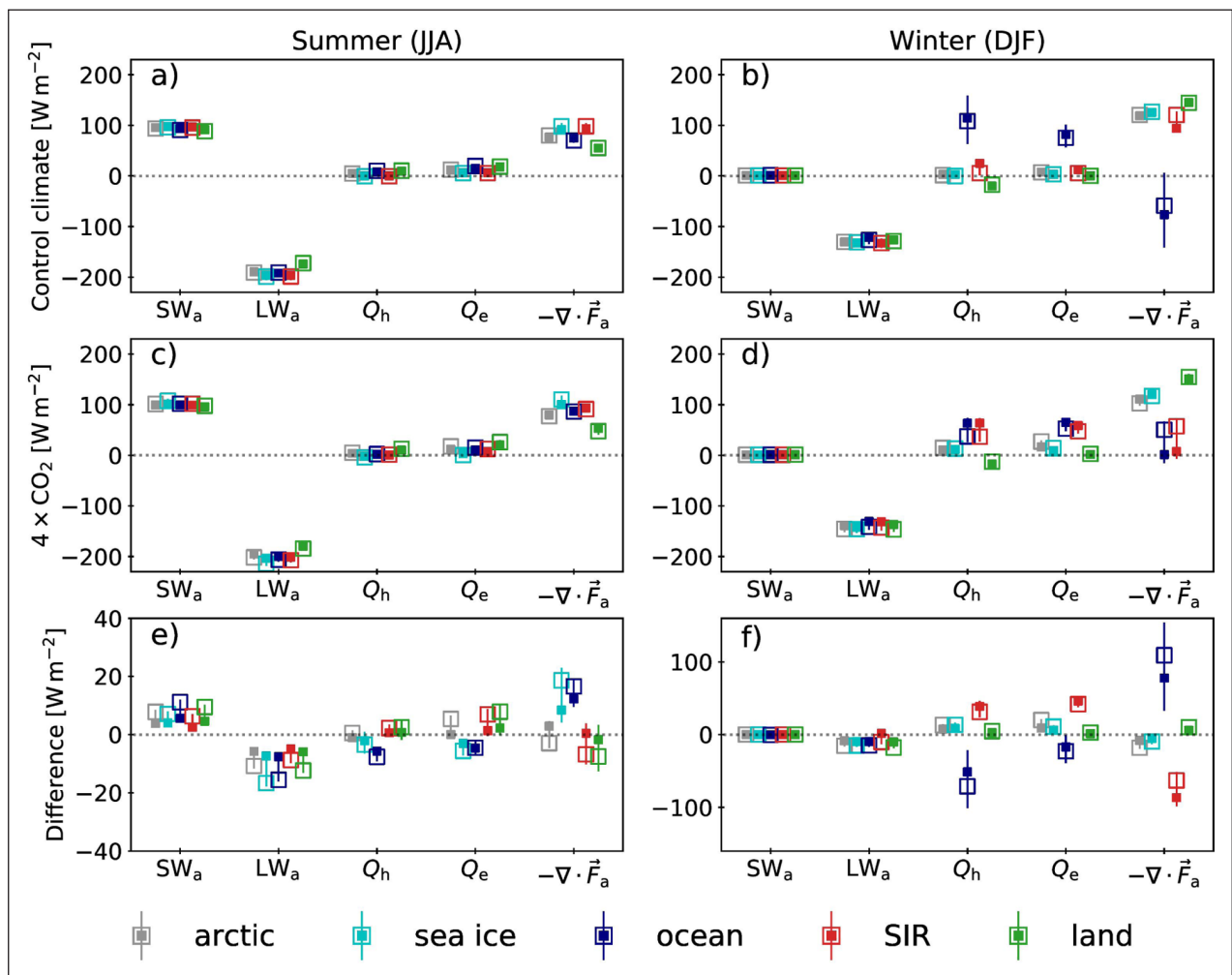


Figure 2 Seasonal components of the Arctic AEB from CMIP6 simulations, with left panels representing summertime, and right panels showing wintertime, respectively. Panels **a** and **b**: control model run (piControl), **c** and **d**: quadrupled-CO₂ climate from 1pctCO₂, **e** and **f**: the corresponding differences (1pctCO₂ minus piControl). The tendency in atmospheric energy storage is negligible, i.e., the AEB components add up to zero. All components are derived as climatological (30-year) averages, and are further categorised into four surface types. From the ensemble of 12 climate models, larger squares indicate the average of six models with higher AA, smaller squares give the model average of six models with weaker AA, respectively. Error bars indicate the inter-model standard deviation. Note that the scale of changing fluxes in panel e and f is 1:4 for summer vs. winter.

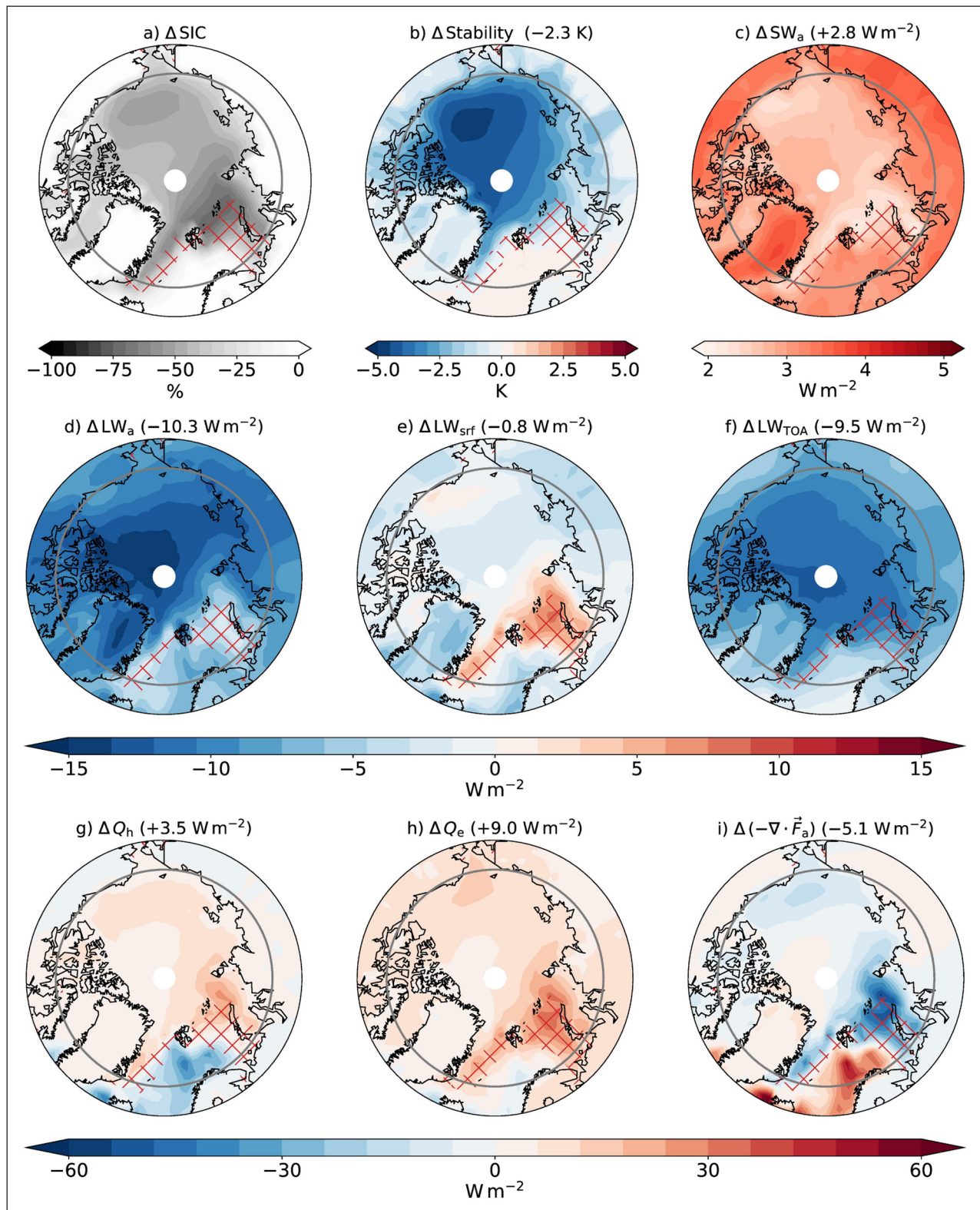


Figure 3 CO_2 -driven annual differences in the AEB components between the quadrupled- CO_2 and control climate from CMIP6 simulations. Changes are further shown for SIC (panel **a**) and stability (panel **b**). The LW_0 term is additionally broken down into net contributions from the surface (panel **e**), and the TOA (panel **f**). All components are calculated as 30-year climatological mean, and multi-model averages from 12 CMIP6 models. The 70°N latitude is marked in gray. Pan-Arctic averages are given on top of each panel. Each map shows the distribution of maximum occurrence of the SIR surface type marked by red hatching.

as a consequence of dropping atmospheric temperatures. The Arctic is in approximate balance between about -130 W m^{-2} long-wave radiative cooling and the same amount of advective heating. The exception is over ocean, where the large amount of heat stored in the ocean mixed-

layer is released by turbulent heat fluxes and balanced by long-wave radiative and advective cooling. Small positive heat fluxes over sea ice are driven by conduction from the ocean to the air-ice interface, further supported by ice growth at the ice bottom, where heat release by

fusion maintains an upward conductive gradient through the ice layer as shown by e.g., Serreze and Barry (2014); Carmack et al. (2015). The atmosphere above Arctic land areas experiences rather strong negative sensible heating, associated with snowpack warming through the downward temperature gradient aligned with a surface-based inversion.

Changes induced by the 1pctCO2 experiment correspond to CO₂ quadrupling in the reference period. Both the net absorbed short-wave radiation during summer and the annual long-wave radiative cooling intensify by additional warming (Figure 2e, and f). Both summer and winter changes in SW₀ and LW₀ contribute to the annual response in Figure 3c, and d, respectively. The intensification in long-wave radiation cooling by overall -10.3 W m^{-2} outruns the increase in short-wave radiation absorption by $+2.8 \text{ W m}^{-2}$, thereby increasing the overall Arctic radiation deficit.

Looking at the SW fluxes at the surface and TOA separately (not shown in the Figure), the surface flux decreases over sea ice, SIR and land due to additional snow and ice melt. This decrease of SW radiation at the surface transfers to a decreasing outgoing flux at the TOA. However, the decrease in outgoing SW radiation at the TOA is slightly stronger, i.e., also the atmosphere accumulates SW radiative energy in the warmer climate. We relate this to an increased atmospheric SW radiation absorptivity by surplus moistening. In fact, an analysis across individual models indicates that models with larger increase in water vapor concentration exhibit a stronger SW₀ increase. The local dampening of the atmospheric SW₀ gain over regions of strong SIC reduction (compare Figure 3a and c) can be linked to the surface-albedo feedback. The local suppression of LW₀ decrease is also tied to retreating sea ice, where the long-wave radiation input from the surface increases more than the corresponding downward surface flux (Figure 3d and e).

Alterations of the turbulent heat fluxes show a strong dependency on the surface type (Figure 2e and f).

Retreating sea ice is associated with increasing energy input from the surface during both summer and winter. For open ocean, turbulent heat fluxes are mainly driven by the ocean-atmosphere temperature gradient, which is weakened in the $4 \times \text{CO}_2$ climate (from 11 K to 4 K in winter, and 1 K to 0 K in summer). Both effects carry more weight during winter where the temperature difference between the warmer ocean and the overlying colder atmosphere is large. This results in a contrast of substantially increased, and decreased surface heating over SIR regions, and the ice free ocean, respectively (Figure 3g and h).

The arising dipole pattern in the annual-mean anomalous turbulent heat fluxes strongly shapes the mediating transport convergence change (Figure 3i): Over SIR, additional heat input from the ocean invokes the divergence of surplus energy from the atmospheric column. Over open ocean, the transport convergence increases, thereby switching sign in winter (from divergence to convergence) to balance the severe heat loss at the surface. Considering the entire Arctic without sub-division, the transport term is the main mechanism to ensure the large-scale energy balance through decreasing values (-5.1 W m^{-2}) as the atmosphere is inefficient at radiating away surplus energy from the surface.

Over sea ice, both Q_{H} and Q_{E} decrease in summer, and increase during winter, with the latter dominating the annual response. This intensification of the overall negative summer Q_{H} and positive winter Q_{H} is linked to the increased seasonal cycle of summer ice melt and winter ice production: To illustrate that, Figure 4 gives the annual cycle of the average SIC over the entire sea ice extent, for both the control and $4 \times \text{CO}_2$ climate. The annual cycle is derived from the model-averaged SIC, using the 15% threshold to identify persisting sea ice cover in both simulations. In short, this concerns the area over which all fluxes derived over sea ice are averaged in Figure 2. Figure 4 shows that the warming naturally corresponds to an overall reduction in SIC throughout

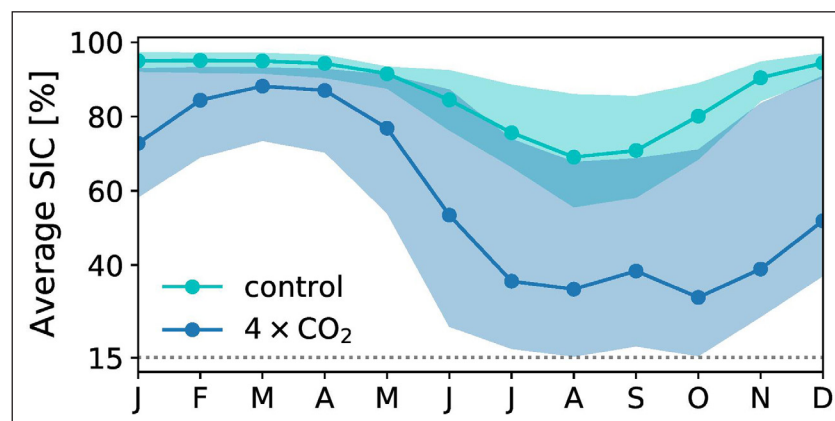


Figure 4 Annual cycle of pan-Arctic SIC averaged over the ice-covered ocean. The annual cycle is given for both the model control simulation and quadrupled-CO₂ conditions from CMIP6. The curves show the multi-model average and the shaded areas include data from all 12 climate models.

the year, but the annual cycle gives a clear intensification from a maximum SIC during March/April to the minimum at the end of melt season. The steeper slope of SIC increase (decrease) during the DJF-period (JJA-period), corresponds to stronger latent heat release by fusion (latent heat absorption by ice melt), which mediates the response of anomalous turbulent heat fluxes over the sea ice covered ocean.

From the perspective of changing sea ice extent it should be noted that at the end of melt season, for more than half of the models (including the entire subset of strong-AA models) sea ice almost entirely vanishes in the $4 \times \text{CO}_2$ climate. In the model-average SIC distribution on the other hand, sea ice persists even under quadrupled- CO_2 conditions. However, the corresponding model-mean SIC value is biased towards higher values in [Figure 4](#), as not all models contribute to the derivation of SIC averaged over the sea ice extent.

Due to the strong discrepancy of annual sea ice extent across models we want to put another emphasis on the fractional contribution of sea ice and SIR surface types to the overall Arctic AEB change: The discrimination of different AA scenarios in [Figure 2](#) shows that the energy flux changes averaged over the entire Arctic region (grey squares) are close to the corresponding values over sea ice for weaker AA models, respectively. Thereby, the fractional contribution of sea ice dominates the overall change of the AEB components. However, looking at the subset of models with stronger AA, the pan-Arctic flux changes are closer to the SIR signal, respectively. Thereby, the SIR signal increasingly contributes to the pan-Arctic AEB changes in a warmer climate, determined by the fractional contribution to the domain.

3.3. HOW DOES THE ENERGY EQUILIBRIUM DEPEND ON THE STRENGTH OF ARCTIC AMPLIFICATION?

The discrimination between weak and strong AA indicates that the AEB changes in [Figure 2e](#) and [f](#) are less intense (or zero) in the weaker-AA ensemble (small squares). The exception is found over SIR during wintertime, where the increase in turbulent heat fluxes is controlled by the vertical temperature gradient at the ocean-atmosphere interface. For models with weaker AA, the temperature gradient in the $4 \times \text{CO}_2$ climate is slightly larger (4K) than for strong AA (3K). This aids a stronger increase of both the turbulent heat fluxes and the terrestrial heat input from the surface (which even overcompensates the cooling at the TOA).

From the perspective of energy equilibrium, the CO_2 forcing does not substantially affect the widespread radiative-advective balance during summer. Even during the winter period, where the overall Arctic warming (+17K) is about three times larger than during summer (+6K), changes at the surface are small over sea ice and land, which leaves the local equilibrium quasi unaffected.

Substantial changes in the state of equilibrium are restricted to the dipole region over open Arctic ocean and SIR during winter: Considering the $4 \times \text{CO}_2$ equilibrium ([Figure 2d](#)), models with lower AA indicate that the net turbulent heat flux (128 W m^{-2}) and the long-wave radiative cooling (-131 W m^{-2}) are in approximate equilibrium, leaving the columnar transport convergence close to zero. Over ocean (including SIR in 1pct CO_2), models with low AA show thus in winter a radiative-convective equilibrium. For models with stronger AA, the Q_H increase over SIR (decrease over open ocean) is lower (higher). Both implies that in the $4 \times \text{CO}_2$ climate, Q_H is smaller (90 W m^{-2}) compared to weaker AA, so that the radiation deficit (-141 W m^{-2}) is additionally balanced by advective heating (51 W m^{-2}). This suggests that during winter for an increasing degree of AA, the intermediate radiative-convective equilibrium re-transfers towards radiative-advective equilibrium as is the case during the summer season.

3.4. SPATIAL CORRELATIONS

Motivated by the surface-type control on individual AEB components, we examine spatial relationships across all models. Linear correlation coefficients are derived from spatial variability of the annual-mean diagnostics over the Arctic-ocean domain, with sea ice conditions in the model control run. Correlations are computed between changes of the AEB components and ambient factors, namely SIC, atmospheric stability (S), the surface-near atmospheric temperature (T_{os}), water vapor path (WVP), and total cloud cover. The latter total cloud cover includes both large-scale and convective clouds. The LW radiation budget is broken down into the fluxes at the surfaces and TOA. Given the highly correlated individual SW flux distributions at the surface and TOA we leave out the discrimination in the discussion of spatial relationships. The transport convergence is not included in the investigation of statistical relationships given the depending role as residual of the energy budget.

Before discussing statistical relationships it should be noted that the diagnostics show a strong collinearity, particularly over regions where sea ice retreats: Changes in the energy fluxes are correlated with each other, and with the degree of SIC decrease, T_{os} increase, and stability loss, with individual $r > 0.7$. Further collinearity exists for changes in SIC, S , and T_{os} : Retreating sea ice strongly increases the surface temperature as it exposes the warmer ocean surface. This breaks the surface-based temperature inversion and thus decreases atmospheric stability. In addition, the decline in SIC and S shows a positive correlation with increasing cloud cover over SIR. The multicollinearity over SIR regions makes the interpretation of statistical relationships challenging.

To facilitate the interpretation, the network plot in [Figure 5](#) shows all robust correlations ($r > 0.6$) that are statistically significant, but over persisting sea ice (excluding SIR), where the ambient factors show less

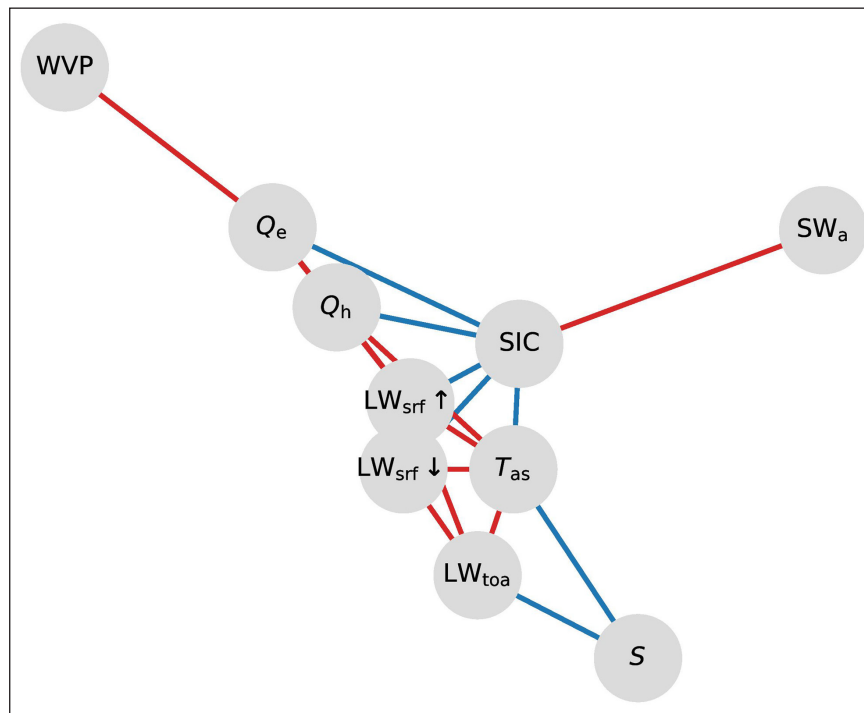


Figure 5 Correlation network plot including all Pearson correlations of annual-mean diagnostics that are significant at the 99% level and above the r threshold of 0.6. The distance of the nodes accounts for the degree of correlation, with smaller distances indicating a stronger correlation. Red (blue) colouring indicates positive (negative) correlations. Spatial correlations are restricted to the sea-ice covered ocean domain, where sea ice does not retreat.

collinearity. The network presents changes in the SIC as the central local impact on the spatial variability of changing energy fluxes. This is evident from the highest number of connections to the other variables. Four individual linear systems can be described, which are linked to one another, and directly or indirectly related to the reduction in sea ice:

- 1) $SIC-SW_a$: Over sea ice, areas with stronger SIC reduction correspond to more dampening of the local short-wave absorption of the atmosphere. The robust positive correlation appears during the sunlit season and links the spatial variability in SW_a alterations to the SIC-mediated surface albedo change (the correlation is also robust over land corresponding to snow melt).
- 2) $SIC-T_{as}-LW_{srf}$: Changes in the LW_{srf} budget are mostly linked to the surface temperature, which relates to the SIC decline: The warmer surface temperature through SIC decrease links to stronger terrestrial heating from the surface, thereby also increasing the corresponding downward flux. These coupled processes are strongly linear throughout the year, with individual correlation coefficients $r > 0.9$. The connection of this mechanism to the SIC degrades in summer, when the ocean impact on the terrestrial heating is outpaced by the warmer atmosphere.
- 3) $T_{as}-(LW_{srf})$ -stability- LW_{toa} : Changes in the LW cooling to space are positively correlated with mechanism 2), mostly via the surface temperature, but also link to the stability parameter: Increasing T_{as} links to

more LW cooling at the TOA. This process is stronger over regions with larger sea ice reductions. The extra outgoing LW radiation at the TOA originates mostly from the surface warming and is aided by decreasing atmospheric stability. The negative stability- LW_{toa} correlation implies that decreasing atmospheric stability leads to more efficient LW cooling at the TOA, i.e., the surplus warming at the surface is more efficiently removed from the surface.

- 4) $Q_h-Q_e-SIC(-WVP)$: Regions with stronger decrease in SIC are linearly related to regions with more Q_h increase with the same argument as for the LW_{srf} budget. Regions of stronger Q_h are further associated with stronger surface temperature increase as it acts through the modification of vertical temperature gradients. Changes in the Q_e are positively correlated with alterations of the WVP by the modification of local evaporation rates.

The network plot suggests that cloud changes have no strong effect on both SW and LW components over sea ice as the local increase in cloud cover is small (overall +0.9%). However, over SIR (not shown here), the downward LW_{srf} flux increase is stronger where cloudiness increases more ($r = 0.7$). Over SIR, the change in cloud cover is additionally linked to the alteration of stability loss over SIR ($r = -0.6$). Thereby, we suggest a cloud signal in the downward LW_{srf} flux change due to the cloud Greenhouse effect, but no cloud signal in the altered SW radiation components at the surface and TOA.

4. DISCUSSION AND CONCLUSIONS

We have presented an AEB framework in its simplified form where the tendency in energy storage is solely mediated by the atmospheric radiation budget, the surface turbulent heat fluxes, and the energy transport convergence. We show the closure of the large-scale energy balance and find that secondary terms like the conversion of liquid water and ice are negligible.

CO₂-driven changes of the AEB components show a profound seasonality and dependence on the underlying surface type, consistent with previous work (e.g., Lauer et al., 2020; Boeke et al., 2021). Strong changes are restricted to the winter season and govern the annual response, most notably the dipole of anomalous turbulent heat fluxes over open Arctic ocean and retreating sea ice. The sea-ice → ocean transformation releases additional energy from newly opened ocean, but once ice free, the flux trend changes sign and counteracts the excessive warming at the surface.

The SIR-ocean dipole strongly shapes the atmospheric response in terms of anomalous transport convergence, which we present as the main mechanism to ensure the local energy balance, while being mediated by large-scale diffusion (Armour et al., 2019). Our overall atmospheric energy convergence decreases with increasing CO₂, but we suggest a tipping point for an increasingly melting Arctic, where the effect of SIR on the anomalous transport convergence is outrun by the open ocean signal. That implies an overall increasing energy transport convergence mediated by more open ocean areas.

In our energetic approach, we do not explicitly relate the transport to the large-scale circulation, but we find similarities in other studies that are largely based on reanalyses: Mayer et al. (2016) indicate a comparable North-South dipole for the directly calculated transport trends, and negative trends concentrated along the ice edge, consistent with our findings. The seasonally contradicting signal of slightly positive changes in summer, and strong decrease during winter corresponds to Semmler et al. (2012), who show no major changes in the large-scale circulation in summer, but weakened circulation cells in winter aligned with reduced advection into the Arctic.

Regarding the performance of models to reproduce individual diagnostics we want to point out that this has not been assessed further by including e.g., observational datasets or reanalyses. However, the main conclusions presented are qualitatively consistent across models. We do not explicitly calculate the changing AEB fluxes under real-world conditions, but the emphasis is on the sign of changes in a warming world forced by CO₂. Different quantities across models allow us to discriminate for different AA scenarios and to derive a potential transformation of the radiative-advective equilibrium over melting sea ice: The sea-ice → ocean transformation leads to a deviation from the local radiative-advective equilibrium through excessive heat

release from the ocean, which balances the radiation deficit in a state of radiative-convective equilibrium. With advanced warming, the surface heat fluxes decrease again, so that advective heating regains influence on the overall balance.

We further derive spatial relationships over sea ice to link the changing energy fluxes to anomalous sea ice concentration, surface temperature and atmospheric stability. The sea ice decline is of central importance by aiding turbulent and terrestrial fluxes from the surface and warming the lower boundary layer. In addition, increasingly warm surface temperatures weaken atmospheric stability, both of which contributing to additional LW cooling at the TOA. We find robust diagnostic relationships for these processes, with the intensity being largely constrained by the degree of SIC reduction. In summer, the atmosphere warms and outpaces the impact of SIC-mediated ocean-atmospheric flux alterations. The linearity is limited to an imprint of the surface-albedo feedback, which locally mutes the overall increase of SW₀ by surface absorption. In sum, we show that the presented changes in the Arctic AEB are tied to the regional and seasonal control of SIC-mediated surface fluxes and temperature alterations. We find no significant relationship between alterations in the AEB components and clouds, as changes in the Arctic cloud cover are quantitatively small in the model average. However, clouds are notoriously difficult to simulate in global climate models, especially Arctic clouds in the boundary layer. Thereby, more work is needed to study the impact of clouds and the role in AA.

While the transport convergence is treated as compensatory term to meet the energetic demands, we argue that the interplay of excessive surface heating and transport convergence mediates the vertical shape of the Arctic atmospheric warming: The vertically non-uniform warming is considered to be a positive contribution to the AA as Arctic warming tends to maximise near the surface, but gets muted at higher altitudes. This eventually dampens the ability of the atmosphere to radiate away the surplus energy to space. The coupled view on anomalous surface heating and transport implies an intensification of this process over retreating sea ice, and conversely, a negative feedback contribution over open ocean. Within the context of an intermediate radiative-convective equilibrium over areas of retreating sea ice, we further suggest the local feedback response to be less affected by remote influences, namely the anomalous transport. However, more work is needed to understand how local Arctic feedbacks respond to enhanced greenhouse gas concentrations within an energy budget perspective.

APPENDIX

The tendency in atmospheric energy storage is derived via Eq. 3 by approximating the time derivative with

central finite differences from extrapolated 6-hourly fields of T , z , q , u , and v . The direct computation of the energy transport convergence follows Eq. 2 and equally applies 6-hourly fields. However, the extrapolation of instantaneous fields to 6 hours in CMIP6 has led to spurious results in the transport estimation, with systematic errors in the order of 22% (MAM), 59% (JJA), 22% (SON), 9% (DJF), and 22% in the annual mean. We avoid this problem by using the instantaneous model output of the mass weighted vertical integral of the DSE transport which constitutes the major part of the total energy transport in the control model run. Unfortunately, the instantaneous output was not routinely provided within CMIP6, and was only facilitated by the IPSL model group. The contribution of both latent and kinetic energy to the total energy transport was calculated from 6-hourly values as in Eq. 2, and added to the instantaneous DSE transport output.

The transport divergence is computed in spherical coordinates with R_e the Earth's radius, so that

$$-\nabla \cdot \mathbf{F}_a = \frac{1}{R_e \cos \phi} \left(\frac{\partial F_{a,\lambda}}{\partial \lambda} + \frac{\partial (F_{a,\phi} \cos \phi)}{\partial \phi} \right), \quad (4)$$

with λ and ϕ , the longitude and latitude coordinate, respectively. The partial derivatives are approximated by central finite differences along each coordinate. All vertical integrals are calculated with full vertical resolution and carried out using the trapezoidal rule from the lowest pressure level to the highest.

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COMPETING INTERESTS

The authors have no competing interests to declare.

AUTHOR AFFILIATIONS

Olivia Linke  orcid.org/0000-0002-5286-2185

Leipzig Institute for Meteorology, Universität Leipzig, Leipzig, Germany

Johannes Quaas  orcid.org/0000-0001-7057-194X

Leipzig Institute for Meteorology, Universität Leipzig, Leipzig, Germany

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