Enhanced Arctic warming amplification revealed in a low-emission scenario

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The Arctic region has warmed faster than the global mean in past decades. Future climate change projections also suggest this Arctic warming amplification will continue. Here, using 50-member historical and future scenario simulations by a single climate model, we find that Arctic warming amplification is stronger in a low-emission scenario, compared to a high-emission scenario, after the mid-2040s. This is because in the low-emission scenario, sea ice continues to exist beyond 2040 and the ice-albedo feedback therefore maintains Arctic warming, unlike other latitudes. By contrast, in the high-emission scenario summer sea ice melts away by about 2050. Multi-model analyses show that the strength of Arctic amplification in the low-emission scenario is highly correlated with the amount of sea-ice reduction, whereas this relationship weakens in the high-emission scenario. Our results indicate that climate change mitigation may have a side effect because Arctic warming persists even if the global warming is stabilized.

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In future climate projections, as estimated by global climate models (GCMs) with increasing greenhouse gases, surface and lower-tropospheric temperatures over the Arctic region increase at a faster rate than the global average. This phenomenon is known as Arctic warming amplification (AA)^{1,2}, which is a robust feature in GCMs^{3,4} and has also been identified in observational data for past decades^{5,6}. AA is only a recent emergence⁷, and the Arctic has warmed approximately four times as fast as the globe⁸. Many studies also have reported that AA affects the atmospheric circulation not only over the regional scale^{9,10} but also over the hemispheric scale^{11,12} and is linked to mid-latitude extreme weather in both summer¹³ and winter^{14–16}. In contrast, recent studies suggest that the link between AA and mid-latitude has become weaker^{17,18}. Therefore, a deeper physical understanding of AA is important for increasing the robustness of climate change projections at various scales^{19–21}.

The mechanism for AA has been explained well in the literature: temperature feedbacks (Planck response and lapse rate feedback)^{22–26}, longwave feedback associated with polar clouds and water vapor^{3,27,28}, changes in the poleward heat transport by the atmosphere and ocean^{29–31}, and the ice-albedo feedback^{32–35}, together warming the Arctic region preferentially in response to the increase in greenhouse gases. The lapse rate feedback^{22–24}, which is intensified by sea ice loss and ice-albedo feedback^{36,37}, and the ice-albedo feedback^{24,32–35} are frequently cited as significant contributors to AA, whereas the relative contributions of the above processes to AA are still debatable^{20,21}.

Future projections by the Coupled Model Intercomparison Project Phase 6 (CMIP6) GCMs have shown that the Arctic sea ice in September will robustly decrease, regardless of the emission scenarios following shared socioeconomic pathways (SSPs)³⁸. A previous study suggests that a large sea-ice loss in the high-emission scenario causes a large AA³⁹. However, differences in the future change in AA between different emission scenarios have not yet been investigated. Given that the ice-albedo feedback is an important driver of AA, the difference is likely to emerge depending on the sea-ice change in the scenarios, which is a hypothesis of the present study.

Here, using historical and two SSP (SSP1-2.6 and SSP5-8.5) simulations, having 50 members each, by a single climate model MIROC6 (ref. 40), along with observational data^{41–44} (see “Methods”), we show that AA in the future climate is enhanced in the low-emission scenario (SSP1-2.6) compared with the high-emission scenario (SSP5-8.5). This counter-intuitive result can be explained by the fact that sea ice remains only in SSP1-2.6, in which the ice-albedo feedback works in late summer, and, subsequently, the heat release from the ocean to the atmosphere in late autumn intensifies AA. The different magnitude of AA between the two scenarios was first identified in MIROC6 and is further supported by CMIP6 multimodel simulations. Notably, the present study compares the Arctic warming scaled by global warming, which is not absolute warming.

Results

Arctic amplification in the MIROC6 large-ensemble. MIROC6 is one of CMIP6 GCMs, and the historical simulation reproduces the observed increase in the surface air temperature (SAT) since 1980 averaged both over the globe and the Arctic (northward of 65°N) (Supplementary Fig. 1a, b). The Arctic-mean SAT increases more than the global-mean SAT (Fig. 1a) and their ratio is defined as the Arctic amplification index (AAI, see “Methods”). For the past 25 years (1990–2014), the observed AAI is 3.9 K K^{-1} (ref. 41), which lies in the ensemble spread of the MIROC6 historical simulation ($3.3 \pm 0.8 \text{ K K}^{-1}$, see “Methods”).

The AAI for the future projection seems similar between different emission scenarios (slopes in Fig. 1a), unlike the global- or

Arctic-mean SAT changes that are highly scenario-dependent (Supplementary Fig. 1a, b). However, a close look at the two 50-member ensembles shows that the ensemble mean of AAI averaged over 2015–2100 is larger in SSP1-2.6 ($3.4 \pm 0.2 \text{ K K}^{-1}$) than in SSP5-8.5 ($3.0 \pm 0.1 \text{ K K}^{-1}$). This implies that AA does not scale with the global warming level, unlike the argument in the Sixth Assessment Report of the Intergovernmental Panel on Climate Change^{45,46}. We have only three simulations for the other two scenarios of SSP2-4.5 and SSP3-7.0, which show AAIs of 3.3 ± 0.2 and $3.3 \pm 0.1 \text{ K K}^{-1}$, respectively. The difference in AAI between the high- and low-emission scenarios is detectable owing to the large-ensemble size (Fig. 1b). In MIROC6, the AAIs in SSP1-2.6 and SSP5-8.5 can be clearly separated with 10 or more members. Therefore, the enhanced Arctic amplification in a low-emission scenario may not be identified by a single-member projection.

The rate of change in the global- and Arctic-mean SAT, having a non-constant slope (Fig. 1a), suggests that the difference in AA among emission scenarios varies as climate change evolves. To identify the time of emergence, we compared the time series of the AAI between the two scenarios (Fig. 2). They are very close in the early period of this century, but the difference becomes statistically significant after the mid-2040s. The AAI increases from 3.0 K K^{-1} in 2017 to 3.6 K K^{-1} in 2098 in SSP1-2.6, whereas the AAI increases to 3.1 K K^{-1} until 2050 s and gradually decreases to 3.0 K K^{-1} in SSP5-8.5 (Fig. 2a). It is challenging to diagnose the impact of internal variability on AA^{47,48}. Yet, the difference (0.6 K K^{-1}) in the ensemble-mean AAI between the two scenarios is sufficiently larger than the respective ensemble spread that represents internal variability (Fig. 2a). The September sea-ice area (SIA) scaled by the global- and annual-mean SAT change is nearly unchanged and is similar between SSP1-2.6 and SSP5-8.5 by the 2040 s, but afterwards, a significant difference emerges (Fig. 2b), which is similar to the AAI. Hence, the emergent time of the AAI difference between the two scenarios appears linked to that of the scaled SIA difference. The results do not depend on the window length chosen for the running mean (Supplementary Fig. 2).

Here we investigate the impact of the SAT change at each latitude on the AAI because the global-mean SAT change includes temperature changes in the Arctic region. In both scenarios, temperature changes are larger (smaller) in the north (south) of 16°N than the global mean (Supplementary Fig. 3a, b). In SSP1-2.6, the region with strong warming extends southward after the 2050s. In the southern mid- to high-latitudes, the region with a weak warming slightly expands by the end of the twenty-first century. When compared the SAT changes averaged over the global (90°S–90°N), Arctic (65°–90°N) and non-Arctic (90°S–65°N) regions in SSP1-2.6, the global- and non-Arctic-mean SAT changes weaken after the 2070s, whereas the Arctic-mean SAT change remains almost constant (Supplementary Fig. 3c). As a result, the AAI increases in time. In SSP5-8.5, the Arctic-mean SAT change increases at a smaller rate than the global- and non-Arctic means after the 2080s (Supplementary Fig. 3d), and thus the AAI decreases.

Mechanism of the enhanced AA in SSP1-2.6. Why is the Arctic warming maintained even when the global warming is stabilized in the low-emission scenario? To elucidate the mechanism of the enhanced AA scaled by the global-mean SAT change in SSP1-2.6 after the middle of the twenty-first century, we further investigated changes in the SAT and energy budgets over the Arctic region during two periods: 2020–2039 and 2080–2099. To compare the climate response between the two scenario experiments, the change in the quantity of interest is normalized by the annual global-mean SAT change for each scenario, as in Fig. 2b.

Fig. 1 Relationship between global- and Arctic-mean SAT changes from 1990 to 2100. a Scatter plots of the Arctic-mean (northward of 65°N) SAT change (K) against the global-mean SAT change (K) from observation (HadCRUT5 (ref. 41); black) and the MIROC6 large-ensemble (historical; gray, SSP1-2.6; blue, SSP5-8.5; red, each plot indicates the 50-member ensemble mean), relative to 1980–2009 mean. Background shading and lines represent AAI (K K⁻¹). **b** Box-and-whisker plots of AAI averaged over 2015–2100 in SSP1-2.6 (blue) and SSP5-8.5 (red) as a function of ensemble size. The vertical dashed line indicates a threshold beyond which the two ensembles are statistically separated.

Spatial distributions of the annual-mean SAT change over the Arctic region show a strong warming in the Barents-Kara Seas (Supplementary Fig. 4). The difference between the two scenarios is substantial in 2080–2099, indicating a greater relative warming around the Barents-Kara Seas in SSP1-2.6 compared to SSP5-8.5. The AA is also known to show a pronounced seasonality^{49–51}. The zonal-mean SAT warming north of 65°N lags behind the upper-ocean warming by two months, suggesting heat transport from the ocean to the atmosphere in late autumn (Supplementary Figs. 5, 6; see Supplementary Note 1).

Using surface energy budgets averaged over the Arctic, we identify major contributors to the AA difference between the two scenarios in late summer (July–September; JAS) and late autumn (October–December; OND) (Fig. 3). Naturally, changes in the Arctic temperature, sea ice, and surface energy budgets are larger with SSP5-8.5 than with SSP1-2.6 when they are not scaled by the global-mean SAT change (Supplementary Fig. 7). Since changes per 1 K increase in the global-mean SAT matter for AA, however, energy budgets in Fig. 3 are shown after scaling with the global-mean SAT in each scenario.

In 2020–2039, changes in surface energy budgets are quite similar between SSP1-2.6 and SSP5-8.5 (blue and red bars in Fig. 3a, b); net excess energy absorbed by the ocean increases the upper-ocean temperature (positive ΔT_o) in late summer, which is then released to the atmosphere via surface heat fluxes (positive ΔLH and ΔSH) in late autumn. Consistent with the continuous reduction of sea ice (negative ΔSIA), the ice-albedo feedback parameter (λ_a) estimated over the Arctic (Supplementary Fig. 8; see “Methods”) is positive in both scenarios in late summer resulting in more shortwave radiation absorption by the ocean (negative ΔSR) but near-zero in late autumn when insolation is small. These changes fit the typical mechanism of AA^{39,51}.

In 2080–2099, however, the surface energy budget changes are significantly different between the two scenarios (Fig. 3c, d and Supplementary Fig. 9). In SSP1-2.6, the changes are similar to those in early period, with some amplification. In SSP5-8.5, sea ice in late summer has almost disappeared and λ_a is very small, leading to the net energy being much smaller than that in SSP1-2.6. The small energy release to the atmosphere in the following season (positive ΔLH) results in weak AA and slight sea-ice formation.

Fig. 2 Time evolution of changes in AAI and sea-ice area (SIA) under two emission scenarios in the MIROC6 large ensemble. **a**, **b** Time series of (a) AAI (K K^{-1}) and (b) September SIA change scaled by the global-mean SAT change ($10^6 \text{ km}^2 \text{ K}^{-1}$) from 2015 to 2100 (SSP1-2.6; blue, SSP5-8.5; red). All time series are smoothed with a 5-year running mean. The thick line denotes the ensemble mean, and the shading represents one standard deviation. Gray shading indicates that differences between SSP1-2.6 and SSP5-8.5 are significant at the 95% confidence level. In (b), long- and short-dashed lines indicate that SIA in September is zero, [defined as the value of zero minus the 1980–2009 mean (scaled by the global-mean SAT change)], for SSP1-2.6 and SSP5-8.5, respectively.

The above results indicate that the presence/absence of the ice-albedo feedback in late summer is a primary contributor to differentiate the level of AA between the two scenarios in the long-term future. Meanwhile, the contribution of oceanic heat transport to the upper-ocean temperature in late summer is not larger than surface heat flux (Supplementary Fig. 10; see Supplementary Note 2).

The difference in the heat release to the atmosphere in late autumn is not only explained by the latent and sensible heat fluxes but also by the surface longwave heating. Geographically, these changes occur mainly over the Barents-Kara Seas where the SAT difference is the largest between SSP1-2.6 and SSP5-8.5 (Supplementary Fig. 9). Over a wide area of the Arctic, cloud cover and specific humidity increase in late autumn in SSP1-2.6, together enhancing the downward longwave radiation (Supplementary Fig. 11). The greater AAI in the low-emission scenario could also be supported by differences in atmospheric stability (Supplementary Fig. 12; see Supplementary Note 3).

Based on the surface energy budget, the mechanism of the enhanced AA, which is scaled by the global-mean SAT change, in the low-emission scenario is schematically shown in Fig. 4. In SSP1-2.6 with a slower increase in the greenhouse gases forcing, sea ice remains in late summer and the surface warming exerts ice-albedo feedback leading to more insolation that warms the upper ocean. In late autumn, the accumulated heat is released from the ocean to the atmosphere by the surface heat fluxes and longwave radiation, acting to warm the surface. In SSP5-8.5, in contrast, the Arctic Ocean is almost ice-free in late summer, resulting in very weak ice-albedo feedback. The SAT increases in late autumn by increasing latent heat flux, which supports the AA, but the increase is small.

In the present study, we focused on the difference in AA at the fixed time intervals between the two scenarios. However, the question of whether the AA is explained in terms of the global

warming level remains^{45,46}. Therefore, we also compared the difference in AA between the two scenarios at the same magnitude of the global-mean SAT and/or sea-ice loss (Supplementary Figs. 13–15; see Supplementary Note 4).

Robustness of the enhanced AA in CMIP6 climate models.

Given that the future decline of SIA depends not only on the emission scenario but also on the GCM³⁸, the robustness of our results must be verified using multimodel ensembles. To this end, we performed analyses of historical and two SSP simulations by 28 CMIP6 climate models (Supplementary Table 1).

Four out of the 28 models produced an initial-condition ensemble with 10 or more members (Supplementary Table 2), so that we first compared the AAI in 2020–2039 and 2080–2099 obtained from these CMIP6 models with the MIROC6 large ensemble. As in MIROC6, there is no discernible scenario dependence of the AAI in 2020–2039 (Supplementary Fig. 16a) when the late summer sea ice is present in both scenarios (Supplementary Fig. 17). In 2080–2099, all but one model (ACCESS-ESM1-5 (ref. ⁵²)) show the enhanced AAI in SSP1-2.6. This result apparently supports our finding, but the absolute value of AAI varies considerably across models. Indeed, the multimodel ensemble mean using one realization from 28 GCMs shows the insignificant difference between scenarios, even in 2080–2099 (Supplementary Fig. 16b). Differences in the model structure and emission scenario are the major cause of uncertainty for the September SIA projections in CMIP6 models, except for the near-term future⁵³. This seems to explain the large-ensemble spread of AAI. It should also be noted that even when the Arctic warming is normalized by the global-mean SAT change (i.e., AAI), scenario differences still make a sizable contribution to uncertainty in model projections, which is comparable in magnitude to the contribution of multimodel spread within a high-emission scenario due to the model structural uncertainty.

Fig. 3 Changes in the Arctic surface energy budgets in SSP1-2.6 and SSP5-8.5 in MIROC6. **a, b** Changes in JAS and OND for 2020–2039, respectively, in (from left to right) SIA ($10^6 \text{ km}^2 \text{ K}^{-1}$), ocean potential temperature for the upper 200 m (T_o , $^\circ\text{C K}^{-1}$), surface net energy flux (denoted as NET, $\text{W m}^{-2} \text{ K}^{-1}$), and individual components of NET (net shortwave radiation, SR, net longwave radiation, LR, sensible heat flux, SH, and latent heat flux, LH). The ice-albedo feedback parameter for the Arctic region (λ_a , $\text{W m}^{-2} \text{ K}^{-1}$, see “Methods”) is shown on the right. All values are scaled by the global- and annual-mean SAT change (K), and positive values of energy fluxes represent fluxes upward from the ocean to the atmosphere. Error bars indicate the 50-member ensemble spread of the 20-year means. Open bars of SIA signify that there is still sea ice remaining, which is calculated assuming the no-ice condition (see caption of Fig. 2 for the definition). **c, d** As in (a, b) but for 2080–2099.

Yet, we can identify a systematic relationship between the AAI increase and the sea-ice decrease among 28 CMIP6 models. The highest correlation is found between the OND-mean AAI change and the JAS-mean SIA change scaled by the global-mean SAT change (Supplementary Fig. 18), consistent with the arguments obtained from the MIROC6 large ensembles (Figs. 3, 4).

For 1990–2014, the AAI is significantly correlated with the SIA change in late summer ($r = -0.84$, Fig. 5a), which is in good agreement with a previous study³⁹. The significant negative correlation remains in 2020–2039 ($r = -0.88$) and 2080–2099 ($r = -0.79$) under SSP1-2.6 (Fig. 5b, c), but is weaker under SSP5-8.5 in 2080–2099 (Fig. 5e). When 28 models are divided into two groups with reference to the AAI for 1990–2014 (high-AAI models in orange and low-AAI models in purple), their difference can still be seen in the future projection, except for 2080–2099 for SSP5-8.5, when the late summer SIA is nearly ice-free in all models (Supplementary Fig. 17). Fourteen high-AAI models better reproduce the observed changes in AAI and the SIA decrease except for one observation⁴¹. MIROC6 large ensemble also captures the observed AAI and SIA changes within the spread. Using the relationship between the OND-mean AAI change and the JAS-mean SIA change scaled by global-mean SAT change, we can constrain the future AAI, which is likely larger than the multimodel mean in SSP1-2.6 (Fig. 5c, e).

To examine the impact of internal variability on the systematic relationship between the AAI increase and the SIA decrease, we additionally analyzed another large ensemble by CanESM5

(ref. 54). Significant negative correlations are found in both MIROC6 and CanESM5 regardless of the scenarios and periods (Supplementary Fig. 19). However, when each member of two models are divided into high- and low-AAI groups with reference for 1990–2014, their difference is unclear in future projections. This result indicates that the emergent constraint is largely influenced by internal variability.

Discussion

Our results imply that the Arctic region will warm preferably relative to the rest of the globe in the low-emission scenario, even though the global- and annual-mean SAT increase since the pre-industrial era is limited below 2°C (Supplementary Fig. 1). In MIROC6, a weakening of westerly winds at high latitudes is stronger in the low-emission scenario than the high-emission scenario to balance the weakened meridional temperature gradient due to the enhanced AA extending to the lower troposphere below 800 hPa (Supplementary Figs. 20, 21). The weakening of the low-level polar vortex may cause a cold air outbreak from the Arctic more frequently, as has already been suggested to occur in response to the sea-ice decrease over the past decades^{14–16}. However, the CMIP5 models do not show robust evidence that future AA will cause changes in the mid-latitude atmospheric circulation⁴. Therefore, it would be premature to conclude the robust connection between enhanced AA and mid-latitude circulation response identified in MIROC6 SSP1-2.6.

Fig. 4 Mechanisms for the enhanced AA in the low-emission scenario. **a, b**, Schematic diagram illustrating mechanisms for AA, which is scaled by the global-mean SAT change, under SSP1-2.6 and SSP5-8.5, respectively, in 2080–2099. In the low-emission scenario of SSP1-2.6, sea ice still exists, and the ocean surface warming results in the sea-ice retreat causing increased shortwave absorption at the open water and further warming (ice-albedo feedback) in late summer (left). In late autumn (right), the accumulated heat in the surface ocean is released to the atmosphere, enhancing the surface and atmospheric warming in the Arctic. In the high-emission scenario of SSP5-8.5, ice-albedo feedback is nearly absent due to the no-ice condition in late summer (left). In late autumn (right), small upward latent heat flux results in weak atmospheric warming and slight sea-ice formation.

Because the MIROC6 large ensembles are available only for SSP1-2.6 and SSP5-8.5, we have not examined climate changes in other emission scenarios. The difference in AAI between the two scenarios may not be caused by the difference in CO₂ forcing alone but affected by changes in other forcing agents such as aerosols, methane, and CFCs/HFCs^{21,55}. For instance, emissions trajectories⁴⁵ show that CO₂ forcing reduces under SSP1-2.6 since the early twenty-first century but continues to increase until the late twenty-first century under SSP5-8.5. On the other hand, the anthropogenic aerosol emissions decrease in both scenarios, likely acting to warm the globe. Although the CO₂ forcing is dominant for global warming in all SSP scenarios (Figure SPM.4 in ref. ⁴⁵), an assessment of the degree to which other radiative forcings' impact on AA will remain for future works.

Methods

Observational data. Observed global- and Arctic-mean (northward of 65°N) SAT were derived from Met Office Hadley Centre/Climatic Research Unit global surface temperature anomalies version 5 (HadCRUT5)⁴¹, and the observed SIA data were derived from the COBE-SST2 (ref. ⁴²) data set, NASA Team algorithm⁴³, and Bootstrap algorithm⁴⁴. Changes in all quantities were defined as an anomaly relative to the 1980–2009 mean.

MIROC6 large ensemble. The climate model used in the present study is the Model for Interdisciplinary Research on Climate version 6 (MIROC6)⁴⁰. Here, we briefly describe model configuration (see ref. ⁴⁰ for the detailed structure and evaluation). There are two major updates from the previous version MIROC5 (ref. ⁵⁶): a finer atmospheric vertical resolution with a higher model top and the incorporation of a shallow convective parameterization. The horizontal resolution of the atmospheric component is a T85 spectral truncation (150 km), and there are 81 vertical levels up to 0.004 hPa. The horizontal resolution of the oceanic component is a nominal 1° to the south of 63°N and approximately 60 km over the central Arctic Ocean. There are 62 vertical levels, the lowermost level of which is located at a depth of 6,300 m. The sea-ice component has the same framework and

parameters⁵⁷ as MIROC5 (ref. ⁵⁶). Using MIROC6 (ref. ⁴⁰), we performed 50 members of SSP1-2.6 and SSP5-8.5 experiments under the CMIP6-ScenarioMIP⁵⁸ protocol following historical experiments with different initial conditions.

CMIP6 models. We used 28 climate model simulations listed in Supplementary Table 1 from the CMIP6 archive. For one ensemble member of each model, annual global- and Arctic-mean SATs and monthly SIAs were calculated using variable names *tas* and *siconc* from historical, SSP1-2.6, and SSP5-8.5 experiments. The SIA was calculated from the remapped *siconc* on a 1° × 1° area using the Climate Data Operators tool (<https://code.mpimet.mpg.de/projects/cdo>). Changes in the SAT and SIA were defined as deviations from the 1980–2009 mean. Along with MIROC6 (ref. ⁴⁰), four climate models with a large-ensemble member (Supplementary Table 2) were used in Supplementary Figs. 16, 17, and 19.

AAI and SIA. The Arctic amplification index (AAI) is defined as the ratio of the Arctic-mean to the global-mean SAT changes relative to the 1980–2009 mean, where the global- and Arctic-mean SATs are defined as the area-weighted average. The Arctic sea-ice area (SIA) is defined as the sum of the grid cell area multiplied by the sea-ice concentration (>0%) over the Northern Hemisphere. In this study, the error bars for the AAI represent plus and minus one standard deviation. For the past 25 years (1990–2014), for example, we have first calculated the 25-year mean of AAI for each member of the 50-ensemble, and then the 50-member ensemble mean (3.3 K K^{−1}) and spread (0.8 K K^{−1}, one standard deviation). The time series of the AAI and SIA change were smoothed with a 5-year window (Fig. 2) and with 3-, 11-, 15-, 21-, and 31-year windows (Supplementary Fig. 2). The AAI can also be defined as the ratio of linear trends between the Arctic and the globe. However, in SSP1-2.6, the trends are almost zero in the late twenty-first century. In the present study, therefore, we did not adopt this definition.

Ice-albedo feedback parameter. We estimated the ice-albedo feedback parameter to investigate the contribution of the ice-albedo feedback to the difference in AA between scenarios. This parameter is defined as $\lambda_a = (\partial R / \partial I)(dI / dT)$ (W m^{−2} K^{−1}), where R (W m^{−2}) is the net shortwave radiative flux change at the top of the atmosphere (negative downward), I (m²) is the SIA change in the Northern Hemisphere, and T (K) is the Arctic-mean SAT change. The differential coefficients $\partial R / \partial I$ and dI / dT are calculated using a 5-year window, and finally, the ice-albedo feedback parameter is obtained by multiplying $\partial R / \partial I$ and dI / dT for each scenario

Fig. 5 Relationship between sea-ice reduction and AAI in the CMIP6 multimodel ensemble. **a–e** Scatter plots of the OND-mean AAI ($K K^{-1}$) against the JAS-mean SIA change scaled with the global-mean SAT change ($10^6 km^2 K^{-1}$) for **(a)** 1990–2014, **(b, d)** 2020–2039, and **(c, e)** 2080–2099. Results from **(b, c)** SSP1-2.6 and **(c, e)** SSP5-8.5 simulations are shown separately. The black closed circle, square and triangle in **(a)** denote observations^{41–44}, in which the JAS-mean SIA changes are derived from COBE-SST2⁴², NASA Bootstrap⁴⁴, and NASA Team⁴³, respectively. The X symbol in **(a–e)** denotes MIROC6 run 1. Orange and purple circles show CMIP6 models, grouped with respect to the value of AAI for 1990–2014. The two box-and-whisker plots in **(a)** indicate the range of the MIROC6 large ensemble for AAI and $\Delta SIA/\Delta SAT$, respectively. Ensemble means and spreads for each group of CMIP6 models are indicated by a plus sign in **(b–e)**. The correlation coefficient between the AAI and the sea-ice change using all CMIP6 models is shown in each panel.

(Supplementary Fig. 8). Note that we changed the sign of R when calculating $\partial R/\partial I$ to represent the ice-albedo feedback parameter as a positive value. In both scenarios, λ_a shows a positive value up to $3.0 W m^{-2} K^{-1}$ only when sea ice exists and insolation reaches the surface. In SSP5-8.5, JAS-mean λ_a decreases toward zero as sea ice declines, which is in contrast to λ_a remaining positive in SSP1-2.6 by 2100.

Statistics. The statistical significance of the difference between SSP1-2.6 and SSP5-8.5 and the correlation coefficient were evaluated using a two-tailed Student's t test. The dependence of the AAI on the ensemble size (Fig. 1b) was investigated based on a bootstrap method. First, the ensemble means of global- and Arctic-mean SAT changes were calculated for ensemble sizes of 3, 5, 10, 15, 20, 25, 30, 35, 40, and 45, which were randomly selected from 50 members and repeated 100 times. Then, for each ensemble sizes, 100 different AAI for the entire period (2015–2100) were calculated, and finally, the maximum, minimum, mean, and 25 and 75 percentiles were calculated.

Data availability

The HadCRUT5 (ref. ⁴¹) data set is available from Met Office Hadley Centre website (<https://www.metoffice.gov.uk/hadobs/hadcrut5/data/current/download.html>). The COBE-SST2 (ref. ⁴²) data set is available from the NOAA Physical Sciences Laboratory website (<https://psl.noaa.gov/data/gridded/data.cobe2.html>). Sea-ice area data based on the NASA Team algorithm⁴³ and Bootstrap algorithm⁴⁴ are available from the National Snow & Ice Data Center (NSIDC) (<https://nsidc.org/data/g02202>). The CMIP6 model output analyzed in the present study is available from the Earth System Grid Federation (<https://esgf-node.llnl.gov/search/cmip6/>). The MIROC6 data (historical⁵⁹, SSP1-2.6 (ref. ⁶⁰), SSP2-4.5 (ref. ⁶¹), SSP3-7.0 (ref. ⁶²), and SSP5-8.5 (ref. ⁶³)) are available from the Earth System Grid Federation.

Code availability

Codes used for creating main figures are available from the corresponding author on reasonable request.

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Author contributions

J.O. and M.W. designed the research. H.T., Y.K., M.A., and J.O. conducted the numerical experiments. J.O. performed the analysis advised by M.W., Y.K., and M.A. and wrote the paper. All authors discussed the results and commented on the manuscript.

Competing interests

The authors declare no competing interests.

Additional information

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