



Sensitivity of seasonal circulation response to snow reduction in the Northern Hemisphere and Eurasia and its impact on Eurasian climate

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Abstract

In this study, we analyze the impact of reduced snow cover in the Northern Hemisphere on the atmosphere and if the atmospheric response depends on the model resolution. We use the atmospheric component of the global climate model EC-Earth and perform three experiments: in the first experiment, we reduce the snow cover in the entire Northern Hemisphere by reducing the snow albedo to a constant value of 0.3, in the second experiment, we reduce the snow albedo only over Eurasia, and the third experiment is the control run using normal snow conditions. All experiments are integrated over the period 1980–2015 at standard resolution (~ 80 km) and high resolution (~ 40 km). Experiments comprise 11 and 5 ensemble members at standard resolution and high resolution, respectively. Reducing the snow albedo in the Northern Hemisphere leads to 5–10% snow cover reduction in winter and spring. Significant warm responses are found over northern Eurasia in spring and summer with a warm response reaching 3 °C. Similar but weaker warm temperature responses are found in the middle and upper troposphere (up to 2 °C) and reversed temperature responses in the stratosphere (up to –2 °C), particularly over eastern Eurasia. This is closely associated with westerly jet flow response which is enhanced at high-latitude and weakened at low-latitude in winter and spring over eastern Eurasia. Reduced snow cover leads to warmer surface temperatures that accelerate snow-melting and further lead to different snow-hydrological responses in western and eastern Eurasia and more precipitation occurs over eastern Eurasia (increasing 10–20%), particularly in the Siberian region. When the snow albedo is reduced only in the Eurasian sector, the surface response pattern resembles the results of the Northern Hemisphere experiment. The warm response is slightly weakened about 0.25–0.5 °C over Eurasia and significantly weakened outside of Eurasia. However, the upper air circulation response is much less pronounced over Eurasia. The impact of resolution on the mean surface field response is small yet it is more pronounced on the large-scale circulation response, particularly in spring and winter.

Keywords Snow albedo reduction · High resolution · Snow-hydrological response · Large-scale circulation response

1 Introduction

Snow cover is one of the most important components of the cryosphere and roughly 98% of seasonal snow cover lies in the North Hemisphere (Armstrong and Brodzik 2001). It has long been understood that snow cover variability has strong seasonal distinctions. Snow cover as an important land surface factor plays a crucial role in regulating climate due to its optical properties and impacts on heat fluxes and soil

moisture. Snow cover strongly influences the global energy balance because its strong reflectivity reduces solar radiation absorption, and low thermal diffusivity (e.g. Cohen and Rind 1991; Jones et al. 2001; Gong et al. 2004). Furthermore, snow cover insulates the ground surface and weakens the moisture exchange between the atmosphere and land surface which affects the terrestrial water balance (e.g. Derkson and LeDrew 2000).

During the past few decades, the snow cover in the Northern Hemisphere has substantially decreased due to global warming (e.g. Wu et al. 2020; Mudryk et al. 2020; IPCC 2021), particularly in spring and early summer. Satellite Records from the last five decades show that spring snow cover is disappearing earlier in the year than it did in the past. The area of snow-covered ground is declining most

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rapidly in June (Mudryk et al. 2020; IPCC 2021). Previous studies reveal that snow cover variation triggers the direct local spatial and temporal climate variability (e.g. Fallot et al. 1997; Gong et al. 2004; Brown and Robinson 2011; Peings et al. 2011; Matsumura and Yamazaki 2012; Furtado et al. 2015; Henderson et al. 2018). Changes in the extent of snow cover are expected to give an impact on temperature via snow-albedo feedback, particularly over Eurasia (e.g. Cohen et al. 2007; Hardiman et al. 2008; Allen and Zender 2011; Furtado et al. 2015). Snow cover reduction leads to surface albedo reduction and further enhances the Northern Hemisphere climate sensitivity (Qu and Hall 2007). As one of the critical factors of the Arctic amplification, the snow-albedo feedback implies significant local warm temperatures and further affects the snow-hydrological cycle (e.g. Fletcher et al. 2012; Matsumura and Yamazaki 2012; Thackeray et al. 2021; Abe 2022). The snowmelt-related hydrological processes are associated with late spring and summer climate variability in Eurasia. The hydrological effect is particularly significant in eastern Siberia (e.g. Allen and Zender 2011; Matsumura and Yamazaki 2012; Peng et al. 2013; Chen et al. 2015; Wu et al. 2020). Meanwhile, snow cover also exhibits indirect remote and continental-scale atmospheric circulation responses, and snow melting is associated with the phase of change of Arctic Oscillation (AO) (e.g. Peings et al. 2010, 2012, 2013, 2017; Handorf et al. 2015; Gong et al. 2003, 2004; Cohen et al. 2007, 2013, 2014). Based on observational evidence, a significant snow-AO link has been detected (e.g. Cohen 1999). There is a relationship between autumn Eurasian snow anomalies and the variability of North Atlantic Oscillation (NAO)/AO in wintertime (e.g. Saito and Cohen 2003; Fletcher et al. 2007; Allen and Zender 2011; Cohen et al. 2014). This motivates research into how snow cover reduction may impact high and mid-latitude climates.

In the past few decades, numerous numerical studies have been performed to investigate the local response and the teleconnection between snow cover and large-scale circulation patterns with general circulation models (GCMs). Generally, the majority of models can capture the significant local response (e.g. Peings et al. 2011; Matsumura and Yamazaki 2012; Furtado et al. 2015; Henderson et al. 2018). However, most of the models failed to capture the teleconnection between snow cover and atmospheric circulation, particularly the observed snow-NAO relationship. (Hardiman et al. 2008; Furtado et al. 2015; Handorf et al. 2015; Tyrrell et al. 2018; Peings et al. 2017). This could be due to the strong mid-latitude interannual to decadal variability. The insufficient prescribed snow cover anomalies or the unrealistic representation of stratospheric dynamics in GCMs could also be one of the reasons (Hardiman et al. 2008; Fletcher et al. 2009). Peings et al. (2012) used a nudging methodology to obtain a more realistic representation of the polar

vortex and they found that the correct representation of the stratospheric mean state can help to capture the observed snow-AO teleconnection. Because the relationship between snow cover and NAO plays an important role in regulating Eurasian climate, it is necessary to further investigate the snow-forced local response and atmospheric circulation teleconnection and to see how this will influence the Eurasian climate with a state-of-the-art GCM.

Previous studies mainly concentrated on the impact of snow cover anomalies during certain seasons over some key snow regions, particularly model simulations have so far mainly concentrated on Eurasian or Siberian autumn snow-cover variability and its impact on the Northern Hemisphere atmospheric circulation (e.g. Gong et al. 2003, 2004; Hardiman et al. 2008; Matsumura and Yamazaki 2012; Peings et al. 2011, 2013, 2017; Henderson et al. 2018). However, the North American snow cover may also impact the downstream climate through teleconnection (e.g. Klingaman et al. 2008; Sobolowski et al. 2010; Henderson et al. 2013). Earlier studies investigated the impact of snow cover changes by imposing an artificial change in the snowpack which is inconsistent because snow is added or removed without taking into account heat and water budgets. Satellite observations reveal that there is a dependency between snow cover and albedo (Allen and Zender 2011). Abe (2022) has examined the effect of the snow-albedo feedback using a series of idealized AGCM experiments with constant sea surface temperature and sea ice forcing from 2000 and snow-albedo set to a considerably small constant (0.1). His analysis reveals that the fall and spring surface air temperatures in northern Eurasia become considerably higher and the implied snow cover reduction may also significantly extend the duration of Arctic Oscillation effects when the snow albedo is reduced. We have performed a similar approach as Abe (2022), but in addition, also take into account the impact of the variability of the surface forcing by including ensembles over the entire 1980–2015 period in our experiments. Our study compares two sensitivity experiments, one with snow albedo reduction in the whole Northern Hemisphere and the other with snow albedo reduction only over Eurasia, against a control experiment without imposed changes in the snow albedo. These two experiments will help to investigate the impacts on the Eurasian climate of a snow cover reduction in the whole Northern Hemisphere and Eurasia, respectively.

In addition, the high-resolution simulations are more skillful than the coarse-resolution simulations in terms of the skill of simulating pattern correlations with respect to observations (Sperber et al. 2013). High model resolution is essential for resolving the small-scale processes and topography can be better represented, which has prominent influences on large atmospheric circulation patterns. Previous studies have shown that increasing model resolution yields a more accurate simulation of key features of the physical

climate system in the atmosphere, such as weather regimes (e.g. Fabiano et al. 2020), blocking (e.g. Berckmans et al. 2013; Schiemann et al. 2017), tropical storms (e.g. Boyle and Kelvin 2010; Wehner et al. 2014; Roberts et al. 2020), moisture transports to the continents (e.g. Vanni re et al. 2019) and extreme precipitation (e.g. Bador et al. 2020). High resolution may also improve the slowly varying lower boundary condition (e.g. sea surface temperatures) and can significantly influence the overlying atmosphere (Minobe et al. 2008; Parfitt et al. 2016). Previous studies demonstrate that increasing atmosphere resolution can improve mean climate, however, the increased horizontal resolution does not always improve all processes in GCMs (Fosser et al. 2015; Johnson et al. 2015; Hewitt et al. 2016). Improvements are strongly dependent on the models and the geographical regions. Therefore, it is meaningful and essential to further investigate the climate response to the snow albedo reduction with different resolutions. In order to test the impact of high resolution on the response to snow albedo reduction, we repeated the sensitivity experiments also with the high-resolution configuration EC-Earth3P-HR (Haarsma et al. 2020) with twice the resolution of the standard configuration EC-Earth3P. Details of the model configurations are provided in Sect. 2.1.

The following section describes our detailed experimental design. Sections 3 and 4 describe the results from the Northern Hemisphere snow albedo reduction and the Eurasian snow albedo reduction experiments, respectively. Finally, the main conclusions are presented in Sect. 5.

2 Model description and experimental design

2.1 Model description

The model used in this study is the EC-Earth3P model (Haarsma et al. 2020) which has been jointly developed by the EC-Earth consortium. The atmospheric component of this model is based on the Integrated Forecasting System (IFS) cy36r4 from the European Centre for Medium-Range Weather Forecasts (ECMWF). Variants of the EC-Earth model have been widely used in various studies, e.g. for the Coupled Model Intercomparison Project Phase 5 (CMIP5) (e.g. Hazeleger et al. 2012, 2013) and also in CMIP6 (e.g. D scher et al. 2022). In this study, we run a standard configuration and a high-resolution configuration using atmosphere-only mode at T255 spectral resolution (~ 80 km) and T511 spectral resolution (~ 40 km), respectively. Both configurations have 91 vertical levels and a model top at 0.01 hPa. The snow scheme is a component of the land surface model HTESSEL (Hydrology-TESSSEL, Balsamo et al. 2009), which is an improved version of TESSEL (Tiled

ECMWF Scheme for Surface Exchanges over Land) (van den Hurk et al. 2000) with a hydrology component included. The snow scheme includes an advanced parameterization of snow density, incorporating a liquid water reservoir, and revised formulations for the subgrid snow cover fraction and snow albedo (Dutra et al. 2010). This scheme has been assessed by comparing it with ground-based and remote sensing observations. The scheme can adequately simulate the snow parameters, e.g. snow depth and snow surface albedo (Dutra et al. 2010; Balsamo et al. 2015).

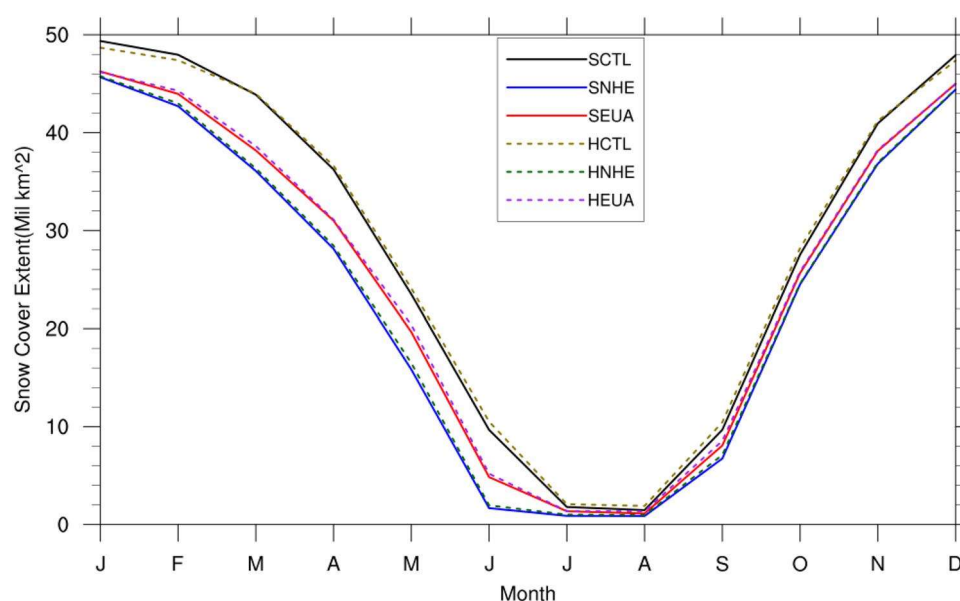
2.2 Experimental design

Previous studies have used GCMs to investigate the possible atmosphere feedback due to anomalous snow cover. To force such an anomalous snow cover scenario, snow conditions (snow cover, snow depth, or snow water equivalent) are usually prescribed and treated as surface boundary conditions (e.g. Gong et al. 2004; Peng et al. 2013; Furtado et al. 2015). Such methodology lacks feedback from the atmosphere to snow cover and moreover, the non-linear interaction between the land surface and atmosphere could be weakened. In addition, snow conditions are not easily prescribed in the model and it would be difficult to conserve water and energy balance and leading to inconsistencies in the model system. Here, we take a similar methodology as Abe (2022) and set the snow albedo in EC-Earth3P to a constant value that subsequently leads to a change in the snow cover in a consistent way. This methodology is better suited to simulate the indirect feedback of snow cover on the atmosphere and allow for investigating the potential climatic response due to freely evolving snow conditions. A similar methodology but using sea ice albedo instead has also been used to study the impact of reduced sea ice cover on climate (Blackport and Kushner 2016; Chripko et al. 2021).

For standard resolution (T255), we have performed one control experiment (SCTL) and two snow albedo reduction experiments (SNHE and SEUA) with the EC-Earth3P model (Haarsma et al. 2020). All experiments were done over the period 1980–2015 and comprised 11 ensemble members. All experiments start from Jan. 1, 1980, and are forced with daily SST and sea ice concentration (SIC) datasets (Kennedy et al. 2017) that have been prepared for the PRIMAVERA project (<https://www.primavera-h2020.eu/>). Each ensemble member starts from a slightly different initial state that is taken from a 20-year long run with constant 1980 forcing. In the control experiment the snow albedo varies with snow age while in the sensitivity experiments, the snow albedo is kept constant. The constant snow albedo value was determined by making a number of shorter experiments. Our test simulations showed that a snow albedo value of 0.3 leads to a snow cover reduction that is closest to the observed snow cover decrease. In the first sensitivity experiment, the snow albedo

Table 1 Summary of sensitivity experiments setup

Experiment	Ensemble size	Analysis period	Model resolution	Remark
SCTL	11	1981–2015	T255 (~ 80 km)	Control run
SNHE	11	1981–2015	T255 (~ 80 km)	Snow albedo set to 0.3 over the whole Northern Hemisphere
SEUA	11	1981–2015	T255 (~ 80 km)	Snow albedo set to 0.3 over Eurasia
HCTL	5	1981–2015	T511 (~ 40 km)	Control run
HNHE	5	1981–2015	T511 (~ 40 km)	Snow albedo set to 0.3 over the whole Northern Hemisphere
HEUA	5	1981–2015	T511 (~ 40 km)	Snow albedo set to 0.3 over Eurasia

Fig. 1 Annual cycle of snow cover extent over the Northern Hemisphere (35–70 °N, 180 °W–180 °E) in the control experiment, the Northern Hemisphere experiment, and Eurasian experiment for standard resolution (SCTL, SNHE, SEUA) and high resolution (HCTL, HNHE, HEUA)

is reduced over the whole Northern Hemisphere (SNHE) (30–90 °N, 180 °W–180 °E). To further investigate the Eurasian climate response to Eurasian snow cover forcing, particularly over Siberia, the snow albedo is only reduced over the Eurasian region (30–90 °N, 0–180 °E) in the second sensitivity experiment (SEUA). The experimental design for the high-resolution sensitivity experiments (T511) is the same as that of the standard resolution experiments except that the ensemble size has been reduced to five members due to expensive computational costs. The detailed experiment setups are summarized in Table 1.

3 Results from the Northern Hemisphere snow albedo reduction experiments

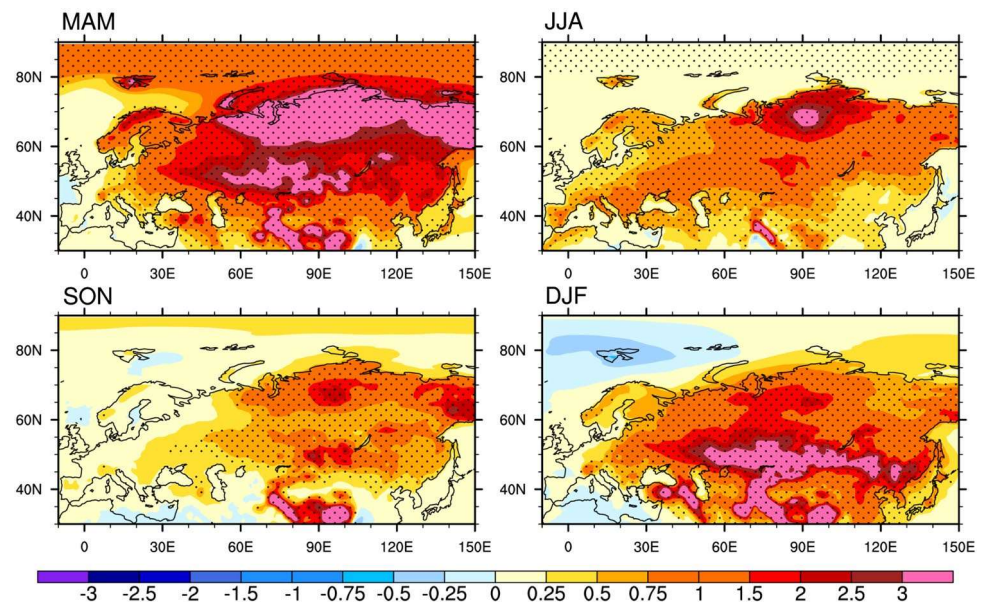
3.1 Surface response over Eurasia

In this section, we first examine the seasonal mean climate response to snow albedo reduction in standard resolution experiments (T255). Figure 1 shows the annual cycle of

snow cover extent over the Northern Hemisphere in the control (SCTL) and sensitivity experiments (SNHE and SEUA). Compared to SCTL, both SNHE and SEUA experiments show systematically reduced snow cover all year round, particularly in winter and spring. The snow cover extent has been reduced by approximately 5–10%. Over the Eurasian region (Supplementary Fig. S1), snow cover has almost disappeared in the summer and the differences between the two sensitivity experiments (SNHE and SEUA) are negligible. This implies that the local snow reduction is mainly due to the quick response to reduced snow albedo. This subsequently leads to a strong surface response in the snow cover reduction region. As seen from Supplementary Fig. S2, the SNHE experiment illustrates strong warm responses over almost the whole Northern Hemisphere in all seasons. More detailed analyses of the Eurasian region are given in the following.

Figure 2 shows the ensemble mean seasonal surface air temperature differences between SNHE and SCTL in spring (MAM), summer (JJA), autumn (SON), and winter (DJF) over Eurasia. Dotted areas are statistically significant at the

Fig. 2 Ensemble Mean surface air temperature differences between the Northern Hemisphere experiment (SNHE) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON, and DJF (Dotted area denotes significant values except 95% confidence level based on a Student's t-test) (unit: °C)

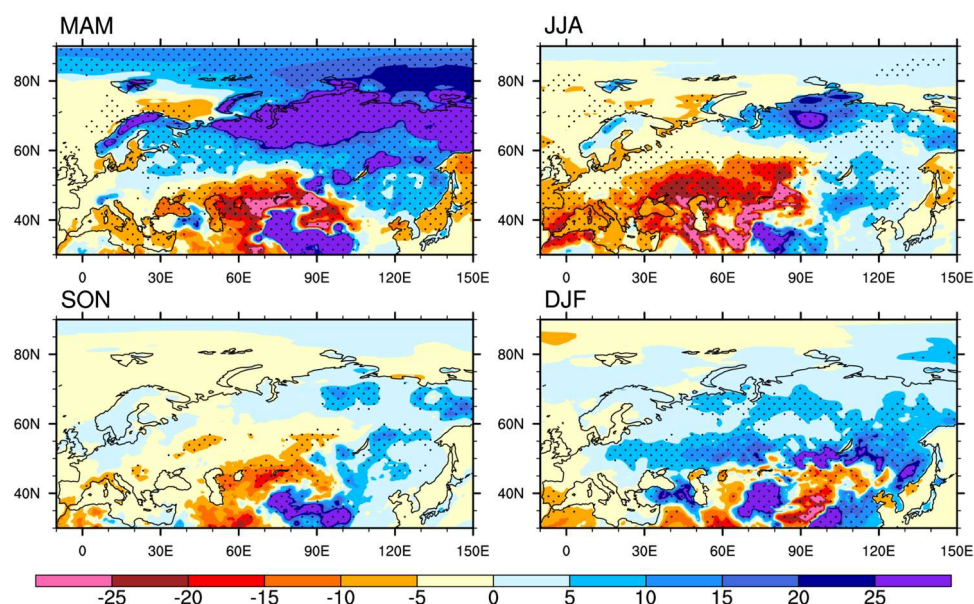


95% confidence level based on a two-sided Student's t-test. Since reduced snow albedo leads to earlier snow melting, this enhanced albedo effect contributes to direct surface heating (Qu and Hall 2007). Snow as an effective insulator is dependent on a number of factors, such as snow depth and snow density. The variation of snow depth and snow density will change snow thermal properties. These changes will further change the snow's thermal energy-transferring capacity and alter the surface temperature. As seen from Fig. 2, the surface air temperature differences between SNHE and SCTL show a strong warm response over the whole Eurasian continent for all seasons. However, the magnitude of responses differs with the season. In spring, the maximum temperature response exceeds 3 °C adjacent to the Arctic coast over the northern part of the continent. Significant strong warm responses are also found over central Eurasia and the Tibetan Plateau. This robust diabatic warming is mainly resulting from the radiative and thermal effects of the reduced snow cover and snow depth. In summer, large warm responses mainly occur over central Siberia along the Arctic coast. As illustrated in Supplementary Fig. S1, snow cover completely disappears over Eurasia during summer in all sensitivity experiments; this warm response is thus an immediate local response to the disappearing snowpack. In autumn, the spatial distribution pattern resembles that of spring which might be due to the fact that autumn and spring are both transitional seasons between winter and summer. The lower snow albedo results in fewer days with snow on the ground and reduced snow cover extent, particularly in high-latitude regions (figures not shown). A smaller snow cover extent reduction in autumn (Fig. 1) has contributed to a much weaker response compared to spring. The magnitude of the warm response is about 1 °C lower compared to the

spring response. The lower solar radiation in the high latitude region reduces the effects of the snow albedo modification compared to the middle-latitude region in winter and the largest warm response extends from the center of Eurasia to the Tibetan Plateau. The spatial distribution pattern of the warm response is similar to the results of Abe (2022) in the Northern Hemisphere. This warm response due to the snow albedo effect is also consistent with previous studies (e.g. Qu and Hall 2007; Lorant et al. 2014; Fletcher et al. 2015; Thackeray et al. 2021). However, the magnitude of warm response in Abe (2022) is almost twice stronger as in our study in all seasons. Since the magnitude of the snow-albedo effect is determined by several factors, snow cover variation could only explain part of the surface warming (Qu and Hall 2007; Fernandes et al. 2009; Fletcher et al. 2012, 2015). Part of surface warming might be due to different model configurations (e.g. model resolution, SST, and sea ice forcing).

Reduced amounts of snow also exert a strong control on the hydrological cycle. The induced changes in the runoff distribution from snow melting will change the soil moisture and exhibit a strong land-atmosphere coupling between evaporation and precipitation. Considering the lagged impact of soil moisture due to its climate memory, this could have particularly robust effects in summer (Matsumura and Yamazaki 2012). Figure 3 illustrates the P-E (precipitation minus evaporation) response between SNHE and SCTL. The spatial distribution of the relative change of P-E in spring is characterized by widespread wetter conditions over eastern Siberia and the Tibetan Plateau and dryer conditions over central Eurasia. Over eastern Siberia and the Tibetan Plateau, the P-E responses are coincident with the surface air temperature responses (Fig. 2). The more melting snow due to warmer temperature partly contributes to more

Fig. 3 Same as Fig. 2 but for relative changes of mean P-E (precipitation minus evaporation) (unit: %)



surface evaporation into the atmosphere. This indicates that the increased precipitation over these regions is largely due to local evaporation (figures not shown). This is consistent with previous studies (Kurita et al. 2004; Matsumura and Yamazaki 2012). This also coincides with more cloud cover over P-E increasing regions (Supplementary Fig. S3). The spatial pattern of total cloud cover in spring (Supplementary Fig. S3) has a similar spatial distribution as precipitation (figures not shown) and P-E (Fig. 3) and also shows a close resemblance to surface air temperature responses (Fig. 2). This confirms the dominant role of surface air temperature in the snow melting process. However, in spring only minor patches of snow cover are found south of 50 °N over central Eurasia. The warmer air temperature dries the soil which enhances the soil moisture deficit and results in less precipitation. In summer, lower snow albedo leads to snow-free conditions over Eurasia except for the Tibet region. Over eastern Siberia, the P-E response is much weaker compared to spring. This could be explained that decreased snow cover in spring could result in less snowmelt and correspondingly lower soil moisture in summer (Kurita et al. 2004; Matsumura and Yamazaki 2012). Matsumura et al. (2010) pointed out that the spring snow cover variation over Eurasia is closely related to the summertime atmospheric and hydrological circulation. From western Europe to central Eurasia, striking continental-wide negative P-E responses are seen over the mid-latitude region (south of 50 °N). Similar to spring, strong surface heating leads to drier soil moisture over this region (figures not shown). In autumn, P-E decreases noticeably over regions south and east of the Caspian Sea which are desert or semi-desert regions. In direct contrast, P-E is increasing over the Tibet region. Similar spatial distribution is seen in winter, however, more significant

P-E responses are seen over Central Europe, East Siberia, and East Asia. Compared to Abe (2022), more pronounced snow-hydrological effects are found in all seasons from our study. Because surface air temperature variability is sensitive to soil moisture conditions after snowmelt (Matsumura and Yamazaki 2012), a stronger warm response in Abe (2022) could lead to much drier soil and appear to contribute less pronounced P-E variability.

It is well established that surface temperatures are closely related to the surface energy balance. The altering of surface energy components will contribute to the surface air temperature warming through different processes. In our experiments, the dominant effect of reduced snow albedo causes strong radiative warming at the underlying surface. This will directly alter the turbulent fluxes (sensible and latent heat fluxes), which are dominant terms in the surface energy budget. Figures 4 and 5 illustrate the ensemble mean differences of sensible and latent heat fluxes between SNHE and SCTL (positive downward) experiments. The sensible heat flux response in spring (Fig. 4) is considerably large and the stronger response (the magnitude of response is more than -15 Wm^{-2}) coincides with the larger warm temperature response (warmer than 3 °C) (Fig. 2). These regions are mainly localized to regions where snow cover has been reduced which are associated with increased short wave radiation absorption and emitted thermal radiation partially balanced the enhanced surface heating (Supplementary Figs. S4 and S5). These response patterns are similar to Abe (2022), except that the magnitude of response is much weaker in our study. This is mainly due to the stronger warm response in Abe (2022). In northern Eurasia, latent heat flux changes (Fig. 5) due to the increased surface

Fig. 4 Same as Fig. 2 but for surface sensible heat fluxes (unit: W m^{-2} , positive downward)

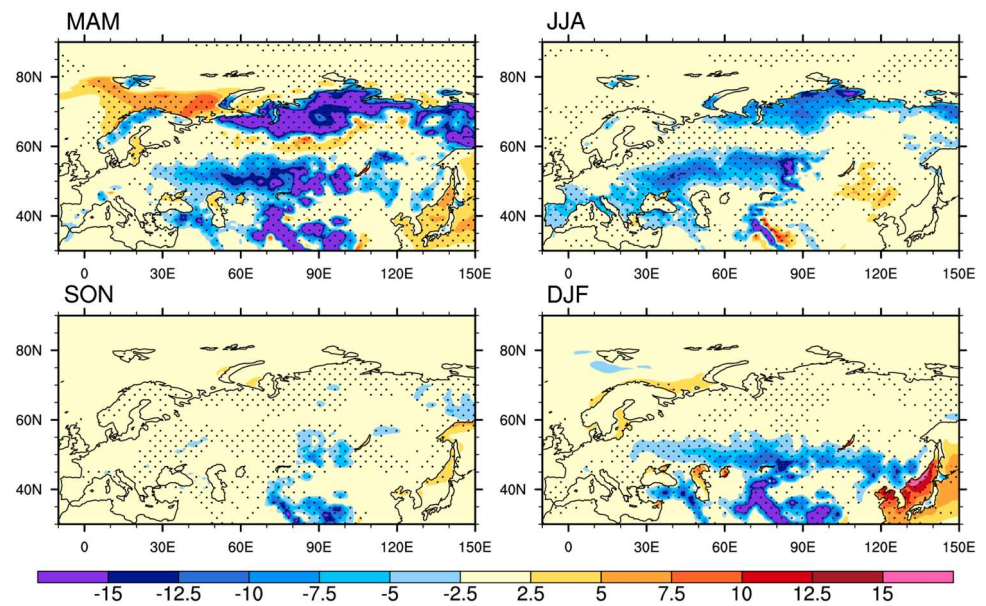
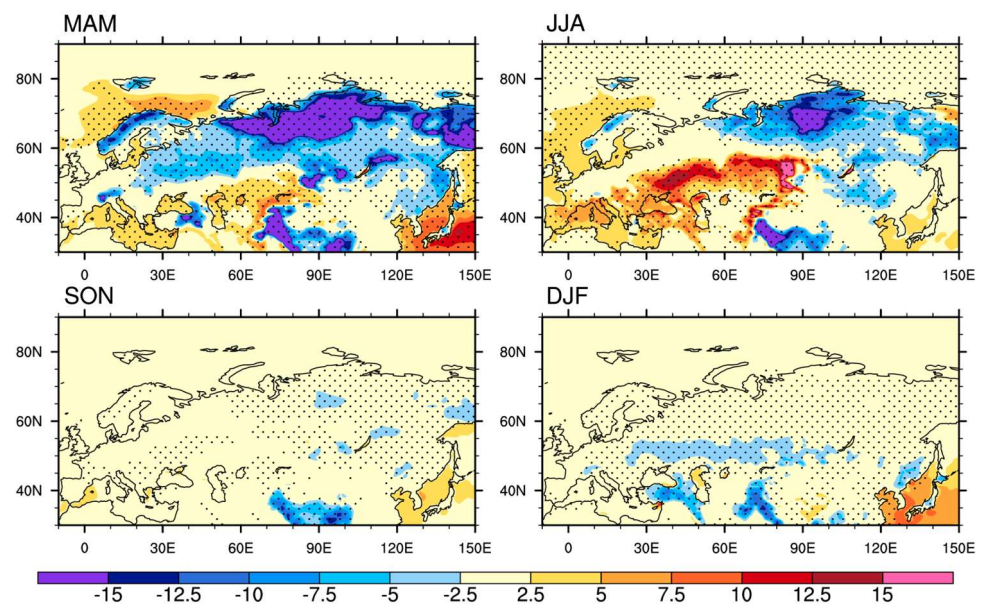


Fig. 5 Same as Fig. 2 but for surface latent heat fluxes (unit: W m^{-2} , positive downward)



evaporation and local precipitation (Fig. 3) contribute about equally to the heat loss into the air as sensible heat flux changes (Fig. 4) which are due to increased surface air temperature. Vice versa, the changes in latent and sensible heat fluxes show opposing responses over southern Eurasia (south of 50°N) excluding the Tibetan Plateau. In spring the reduced snow cover over these regions in combination with decreased precipitation leads to drier soils which implies that the warm surface response over this region (south of 50°N) is mainly via the sensible heat flux, partly due to a lack of latent heat flux. The enhanced net short-wave energy fluxes are also a contributing factor (Supplementary Fig. S4). In summer, both sensible and latent

heat fluxes (Figs. 4 and 5) show similar spatial patterns as in spring, respectively. The magnitude of the difference in the sensible heat flux received by the atmosphere is much smaller compared to spring over the whole of Eurasia. This is mainly because of the weaker warm temperature response (Fig. 2), which might be partly influenced by dry soil. Ye and Lau (2019) found that snowmelt-produced soil moisture anomalies do not significantly impact the late spring-summer local climate anomalies in Northern Eurasia due to the delayed remote responses of atmospheric circulation and climate to the melting of Eurasian snow. The snow cover almost completely disappears in summer, the largest outgoing latent heat flux occurs mainly in the

Siberian region. Over southern Eurasia, a drier continental climate leads to stronger downward latent heat fluxes. The reduced total cloud cover also plays an important role (Supplementary Fig. S3). In autumn, although there are still strong warm atmospheric responses (Fig. 2), both sensible and latent fluxes show very small variation across the whole of Eurasia excluding the Tibetan Plateau. As snow cover is mainly reduced in the high latitude or elevation regions during autumn, temperature warming over these regions is due to enhanced solar radiation. During winter, only sensible heat fluxes show a significant response over southern Eurasia (Fig. 4). However, reduced snow albedo exerts not only seasonal changes in snow cover but also in snow depth. At high latitudes the differences in snow cover between sensitivity and control experiments are small, yet we still find an impact on the temperature and heat fluxes. The lower snow albedo in this region does not lead to less snow cover but instead a reduction in the snow depth and subsequently a weaker contribution to local warming (Baker et al. 1992; Fallot et al. 1997). However, over southern Eurasia, lower snow albedo leads to snow-free regions and a stronger local warming response. Both solar radiation and long-wave radiation fluxes illustrate similar response signals (Supplementary Figs. S4 and S5).

In order to further examine the impact of horizontal resolution and facilitate the comparison between T255 and T511 experiments, the variables for the comparison are further regridded to $0.5^\circ \times 0.5^\circ$ using bilinear interpolation. Since the number of ensemble members differs for T255 and T511 simulations, the uncertainty due to ensemble size is not considered. To compare the changes in the modeled response due to different resolutions, we first calculate the response for each resolution as the difference between the sensitivity experiment and the control experiment and then compute the difference between the high-resolution response (T511) and standard resolution response (T255). In general, the differences in the responses to low snow albedo between high and low resolutions are marginal for most surface fields (e.g. surface air temperature, precipitation, P-E, heat fluxes). The impact of increased resolution on surface air temperature response is small in all seasons and usually smaller than 0.5°C , although there are some regions where there are notable response differences, except for SON. Colder temperature response differences along the Arctic coast during MAM and warmer temperature response differences in eastern Eurasia during JJA and DJF at high resolution are observed for the Northern Hemisphere snow albedo experiment (Supplementary Fig. S6, Left column). These large response differences may arise due to the differences in the surface mean climate in T255 and T511 experiments. Notable changes in surface temperature are observed between SCTL and HCTL simulations (figures

not shown). These evident differences in the mean climate state could lead to large response differences between the resolutions.

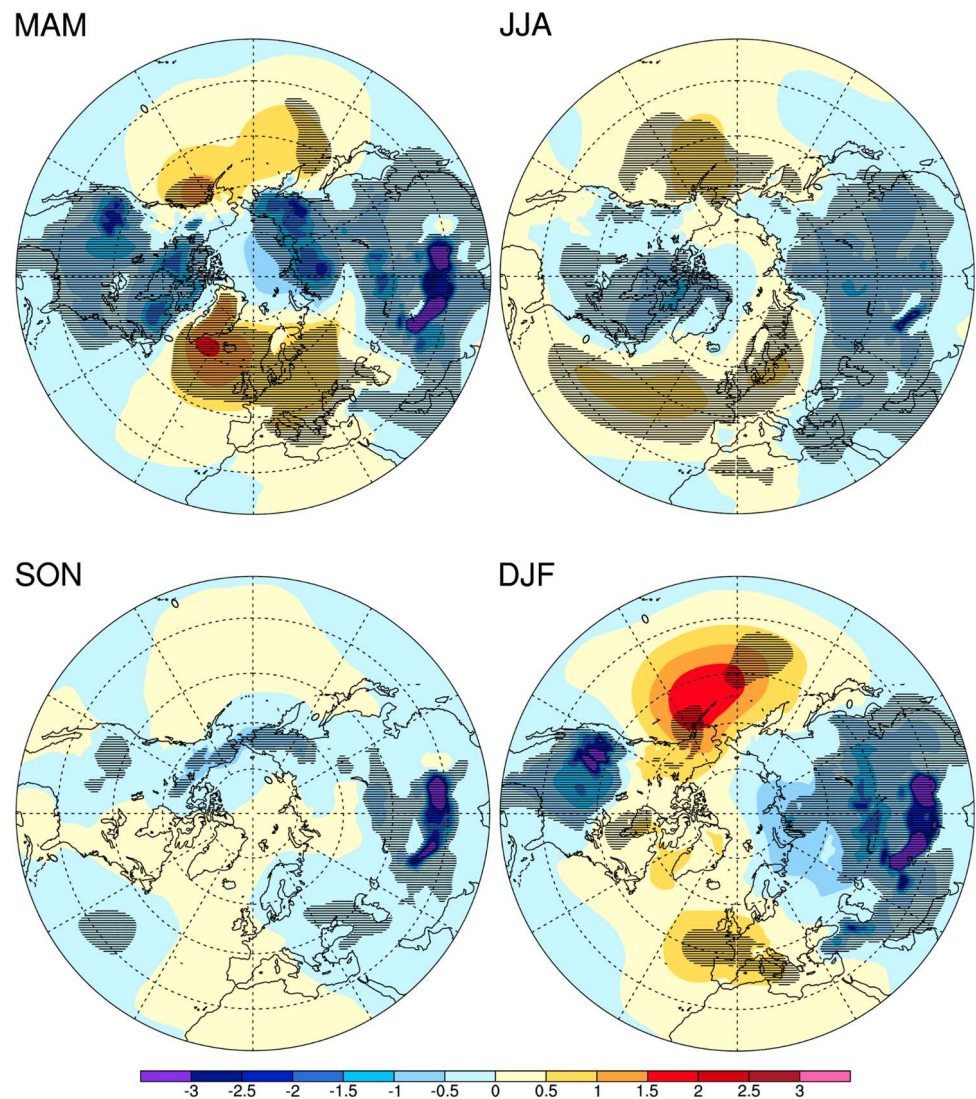
3.2 Atmospheric circulation response

The above analyses reveal that there is a clear surface response to a snow cover reduction in all seasons. Less or absent snow cover over land enhances diabatic heating at the surface and in the lower troposphere. This might modify the stationary wave pattern and transfer less energy from the troposphere to the stratosphere, resulting in a strong polar vortex (e.g. Cohen et al. 2007; Furtado et al. 2015). This snow-atmosphere coupling can regulate both the local and the remote climate. In this section, we will investigate the influence of snow cover reduction on atmospheric circulation.

Figure 6 shows the seasonal difference in the mean sea level pressure (SLP) between SNHE and SCTL. The reduced snow cover over the land region warms the overlying atmosphere and generates a negative SLP response. This is particularly evident over East Eurasia in winter and spring. A strong positive SLP response is seen over the North Atlantic Ocean, the North Pacific Ocean, and western Eurasia, which could partly be due to the sharp land-sea thermal contrast from the anomalous warming over land. This indicates that the Icelandic low and Aleutian low is weakened. And this circulation response will further modulate the magnitude of NAO and influence the Eurasian climate. A similar but much stronger response is also found in Abe (2022). In the other seasons, although the notable surface warming response occurs over the whole of Eurasia, the SLP anomalies show similar spatial patterns, but the magnitude weakens considerably.

To further understand how the tropospheric circulation may respond to the snow cover reduction, we first analyze the 500 hPa geopotential height (GPH) response. The simulated stationary eddies patterns are shown in Fig. 7. This stationary component is expressed as the deviation from the zonal mean of 500 hPa GPH field. The strong zonal asymmetry circulation patterns in the Northern Hemisphere are primarily determined by the complex topography, e.g. the Rockies and Himalayas, and land-sea thermal contrast (Chang 2009). The variations of stationary wave patterns play an important role in regulating heat and moisture poleward transportation and exhibit different warm/cold conditions over continents. It is apparent that the stationary eddy response has strong seasonal variation. In winter, the spatial stationary wave pattern shows ridges over northwestern North America and Western Europe and troughs over northeastern North America and eastern Asia in the SCTL experiment (solid and dashed black contour in Fig. 7). The ridge and trough are particularly strong over Eurasia. The

Fig. 6 Ensemble Mean sea level pressure differences between the Northern Hemisphere experiment (SNHE) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON and DJF (Hatched area denotes significant values except 95% confidence level based on a Student's t-test) (unit: hPa)

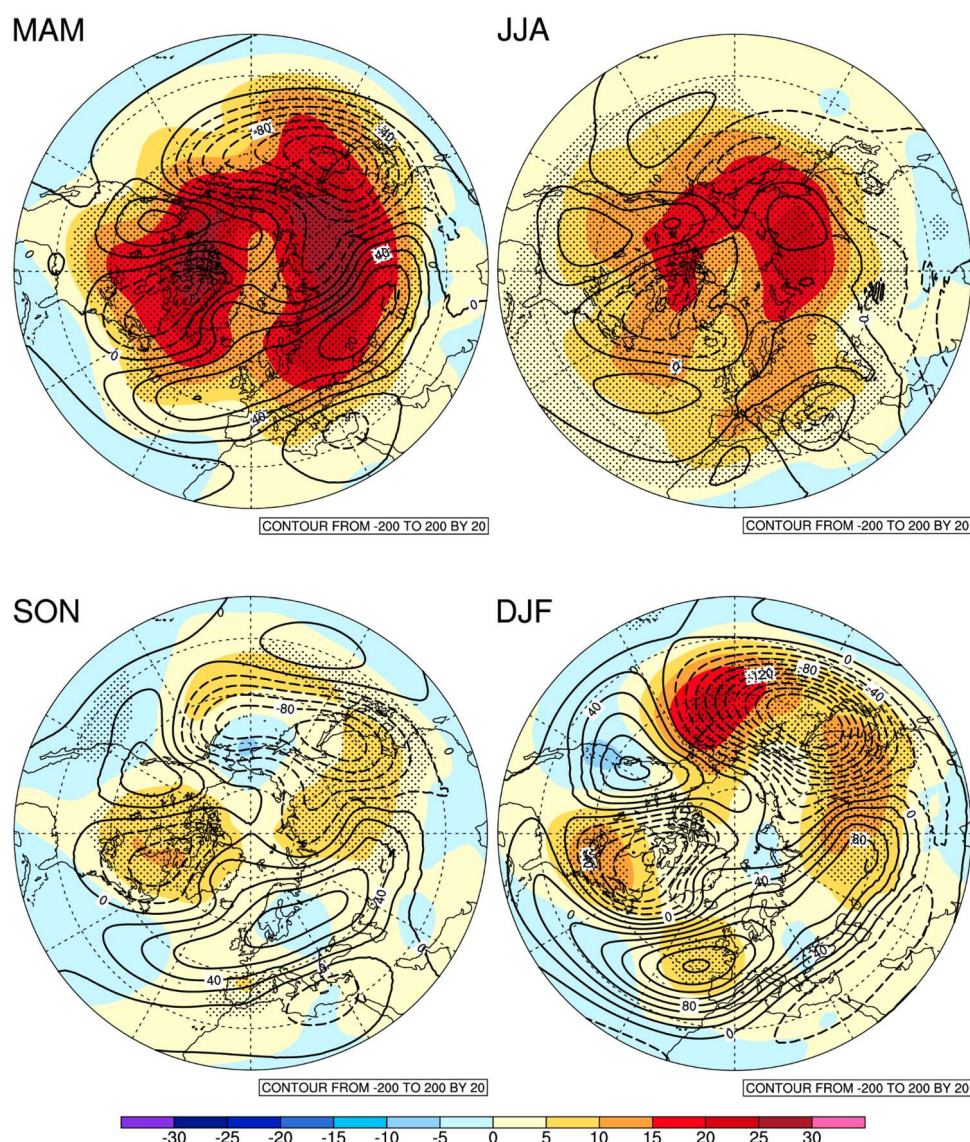


positive stationary eddy response over Eurasia indicates that the ridge over western Eurasia is slightly enhanced, but the trough over East Eurasia is slightly displaced and weakened. The strongest positive response is located in the North Pacific Ocean sector. The ridge and trough over northern America are also slightly southward displaced and the amplitude of the stationary eddy response is relatively smaller compared to the Pacific Ocean sector. The positive geopotential height response over high latitude regions is more robust in summer and particularly in spring. This is consistent with the pronounced warm surface temperature response over high latitude regions (Fig. 2), which indicates that the magnitude of surface warming could modulate the vertical energy transfer of the atmosphere. In autumn, weaker surface warming yields smaller stationary eddy responses. The positive eddy response pattern over eastern Eurasia in all seasons exhibits a clear intensification of stationary eddy activity and this indicates that more moisture is

transferred to eastern Eurasia, particularly over the Siberian region. This is consistent with a stronger P-E increase over Eastern Eurasia (Fig. 3).

From the above analyses, both surface and circulation responses illustrate distinctly different responses between the western and eastern parts of Eurasia. Snow cover reduction plays an important role in regulating large-scale circulation via radiative and thermodynamic effects, particularly over the Siberian region. The large-scale circulation response could further influence the vertical energy exchange and leads to different temperature response in the upper air. Therefore, we first assess the vertical temperature response over eastern Eurasia (averaged between 60 and 150 °E) between SNHE and SCTL (Fig. 8). Strong warm responses in the troposphere and cold responses in the stratosphere are apparent in all seasons except in autumn. In the lower troposphere, the warm response extends from the mid-latitude until the northern polar

Fig. 7 Ensemble mean stationary eddy differences of the 500 hPa geopotential height between the Northern Hemisphere experiment (SNHE) and control experiment (SCTL) for standard resolution during MAM, JJA, SON and DJF (shaded area) (solid and dashed black lines denote the zonal mean eddy from control experiment) (Dotted area denotes significant values except 95% confidence level based on a Student's t-test) (unit: gpm)



region. This vertical response pattern implies that systematically reduced snow cover and a weakened Siberian high lead to a decreased energy transfer from the troposphere to the stratosphere (Cohen et al. 2007). This vertical temperature response pattern resembles Abe (2022), except the magnitude of the warm response is much weaker in our results. This might be due to much stronger surface warming in Abe (2022) and indicates that the upper air is very sensitive to local surface warming, particularly in the lower troposphere. Different model configurations could also partly influence the magnitude of warm response. A model with a well-resolved stratosphere exhibits a faster and weaker response to snow forcing (Fletcher et al. 2009). Over western Eurasia, the vertical temperature response pattern resembles the eastern Eurasian pattern but its magnitude is about 0.5–1 °C colder in all seasons (Supplementary Fig. S7).

The cold temperature response in the upper troposphere and the stratosphere tends to strengthen the polar vortex (Cohen et al. 2007). This will enhance the westerly flow near the polar region. However, the warm response will weaken the westerly flow in the troposphere and lower stratosphere over the mid-latitude region. The strong positive zonal-wind response extends from the surface to the stratosphere at high latitudes while at mid-latitudes the negative zonal-wind responses are primarily in the troposphere. This vertical pattern is more pronounced in winter and spring (Fig. 9). In summer and autumn, the westerly flow only slightly weakens in the troposphere over the mid-latitudes (Fig. 9). This stronger zonal wind over high latitude is associated with the strengthened polar vortex which propagates down into the troposphere and generates zonal-wind anomalies at mid-latitudes with opposite signs and a northward displacement of the jet stream (Cohen

Fig. 8 Ensemble Mean vertical temperature profile differences (averaged between 60 and 150 °E) between the Northern Hemisphere experiment (SNHE) and control experiment (SCTL) for standard resolution during MAM, JJA, SON and DJF (shaded area) (solid and dashed black lines denote the zonal mean temperature profile from control experiment) (unit: °C)

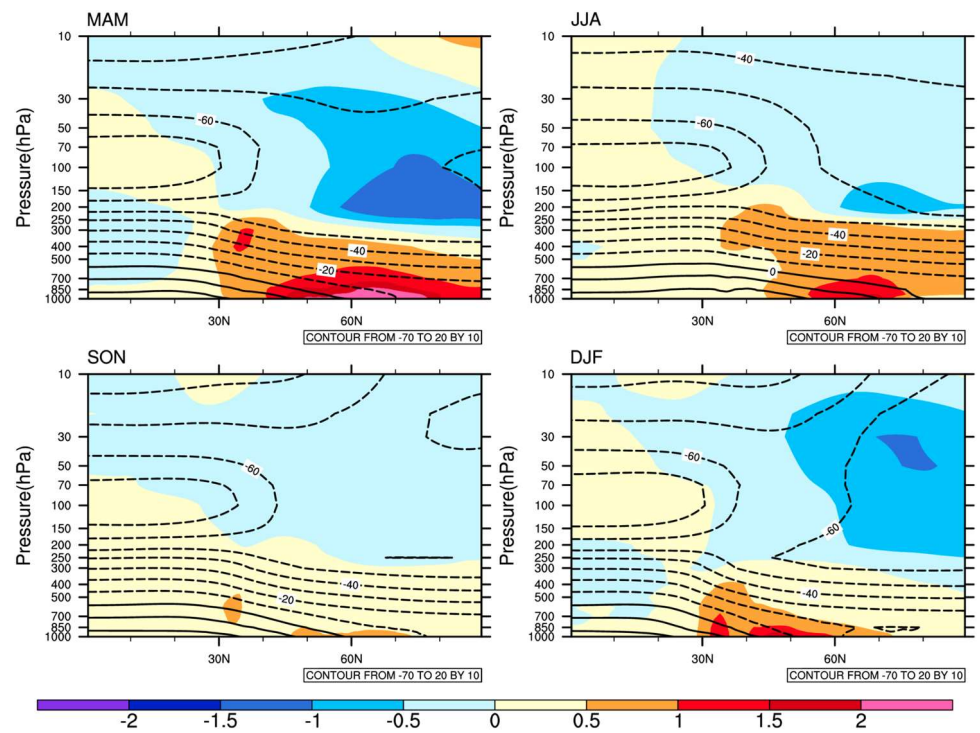
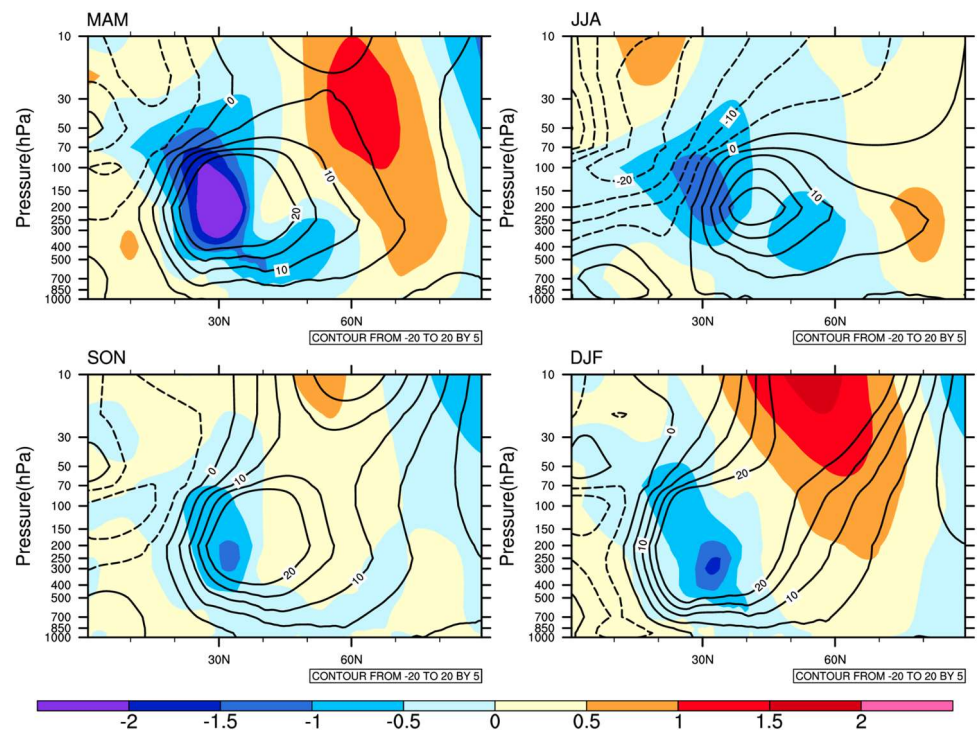


Fig. 9 Same as Fig. 8 but for zonal U wind component (unit:m/s)

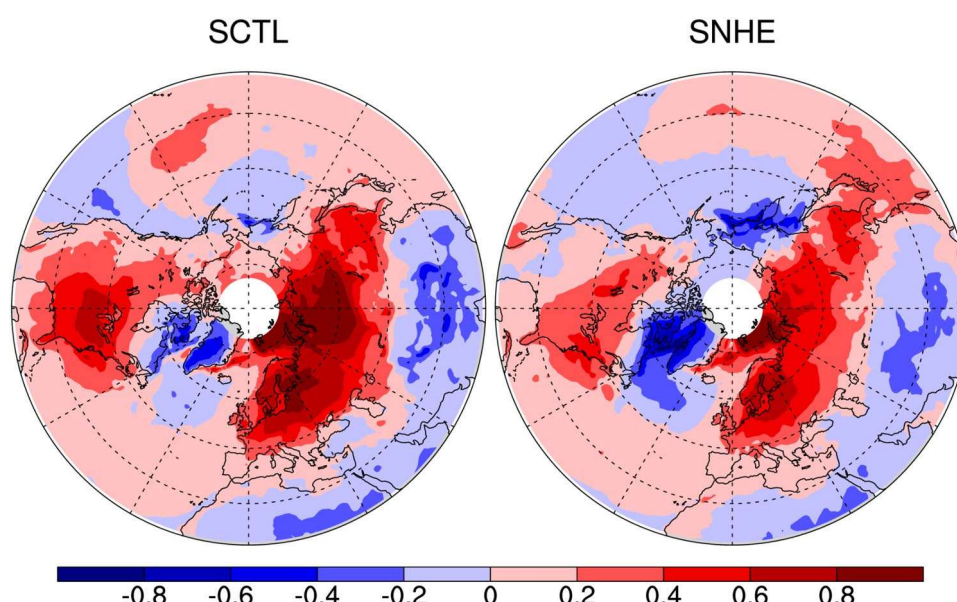


et al. 2007). Over western Eurasia, the zonal profile is similar to that of eastern Eurasia. However, the strengthened high-latitude jet is primarily confined to the stratosphere in winter and spring. The mid-latitude jet only slightly weakens over the troposphere except in summer, when

the low-latitude jet has strongly weakened (Supplementary Fig. S8).

Further inspection of the advantages of the high resolution shows that the impact of resolution on the large-scale circulation response over the North Hemisphere is more

Fig. 10 Regression of DJF surface temperature anomaly onto the NAO index for the control experiment (SCTL, Left) and the Northern Hemisphere experiment (SNHE, Right) for standard resolution (T255) (unit: °C)



robust compared to the temperature response. In the North Hemisphere snow albedo experiment (Supplementary Fig. S9, left column), the response differences of mean SLP are more pronounced in winter (DJF), and the Siberian high and Aleutian low over the Pacific Ocean sector is slightly weaker at high resolution. This could arise due to a slightly enhanced Eurasia warming response in the high resolution (Supplementary Fig. S6, left column). Vice versa, in MAM, high resolution leads to strong Azores high over the Atlantic Ocean and strong Siberia high due to the enhanced cold temperature response in high resolution (Supplementary Fig. S6, left column).

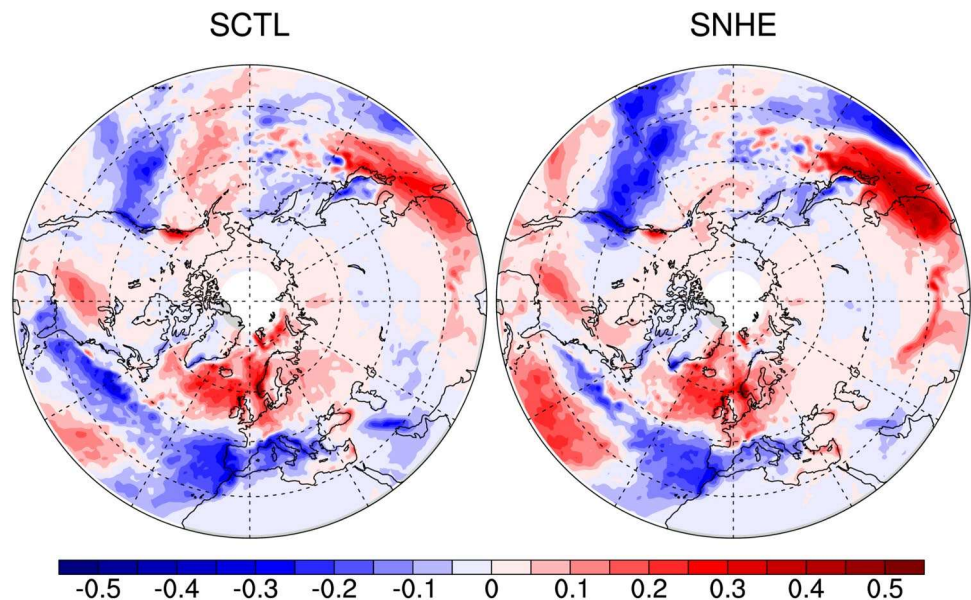
3.3 The link between atmospheric circulation and the Eurasian climate

Over Eurasia, the phase of the North Atlantic Oscillation (NAO) plays a crucial role in the links between the climate of the Arctic and the mid-latitudes. The interannual variability of the Eurasian climate is influenced by NAO through atmospheric teleconnection. Some observational studies found a causal relationship between snow and the winter NAO or Arctic Oscillation (AO). Lacking snow over Siberia in autumn could favor the positive phase of the NAO (e.g. Cohen and Entekhabi 1999; Bojariu and Gimeno 2003). However, this relationship is not conclusive. The strong mid-latitude interannual to decadal variability and the time period of observation data could strongly influence this relationship. Peings et al. (2013) found that the snow–(N) AO connection was episodic and thus not reproducible throughout the twentieth century based on reanalysis data. Some GCM simulations also failed to capture the significant response to the Siberian snow anomalies during autumn (e.g.

Peings et al. 2017), and the snow–(N)AO teleconnection is also absent and being attributed to weak vertically propagating waves (e.g. Furtado et al. 2015). The discrepancies among model sensitivity studies imply that the model may have deficiencies in response to surface forcing associated with snow forcing (e.g. Hardiman et al. 2008; Henderson et al. 2018). Henderson et al. (2018) point out that it could be that only certain configurations of the atmosphere allow for a large-scale effect of snow anomalies.

Based on our sensitivity analysis, the most robust responses to snow cover anomalies occur in winter and spring. Handorf et al. (2015) also identify that changes in Eurasian snow cover may modulate the mid-latitude atmospheric circulation throughout winter and spring. To further explore the possible physical mechanisms for this teleconnection and investigate how snow reduction will impact the large circulation, we have analyzed the relationship between the NAO index and surface temperature and precipitation during winter. The model simulated NAO index is defined as the first leading mode of an Empirical Orthogonal Function (EOF) analysis of monthly sea-level pressure anomalies between 20–80 °N, 90 °W–40 °E. Figure 10 illustrates the regression coefficients of winter surface temperature anomalies onto the standardized winter NAO index. In the SCTL experiment, the spatial pattern is characterized by a strong positive correlation in the high-latitude regions and a negative correlation in the midlatitude. This spatial pattern is broadly similar to that found in previous observational studies (e.g. Cohen and Entekhabi 1999; Bojariu and Gimeno 2003) and modeling experiments studies (e.g. Abe 2022). Compared to SCTL, the SNHE experiment produces a similarly warm-cold pattern over Eurasia, but the magnitude is slightly weaker. Previous studies identify

Fig. 11 Same as Fig. 10 but for precipitation (unit: mm/day)



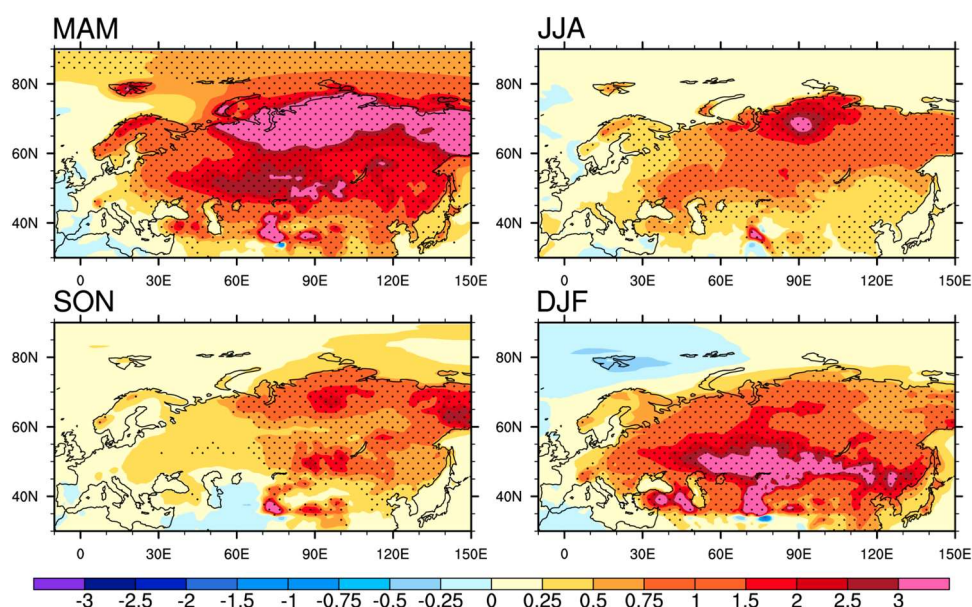
that the anomalous snow cover in autumn could modulate the variability of following winter (N)AO and this could further modulate the surface air temperature variation (e.g. Gong et al. 2003; Saito and Cohen 2003; Smith et al. 2011; Handorf et al. 2015; Gastineau et al. 2017). Although the snow cover has decreased by about 5–10% during autumn and winter (Fig. 1) and the surface air temperature has substantially increased in the SNHE experiment (Fig. 2), there is no pronounced increase of positive phase NAO events in the SNHE experiment compared to SCTL experiment (figure not shown). Peings et al. (2013) find that the snow-AO relationship is not stationary. The snow cover variation is not only influencing NAO's phase but also the strength and interannual variability pattern. This indicates that (N)AO only partly regulates the surface air temperature variability. This is consistent with Gong et al. (2017) who found that the AO-related surface air temperature anomalies are generally weak over the Eurasian continent. In spring, although there is significant snow cover reduction (Fig. 1) and much larger warm responses (Fig. 2), the regression coefficient between surface temperature and spring NAO index shows no evident response over Eurasia (Supplementary Fig.S10). The above analyses suggest that anomalous snow cover which is persistent year-round in our experiments yields limited impact on the (N)AO pattern and only partly explains the large warm responses in winter and spring.

Figure 11 shows the regression coefficients of winter precipitation anomalies onto the standardized winter NAO index. It is found that the largest impact of the NAO index on precipitation is over the ocean instead of over the land. Compared to the control run (SCTL), the negative correlation is slightly reduced in the North Atlantic Ocean and slightly enhanced in the eastern North Pacific Ocean.

There is no apparent response over the Siberian region. This is consistent with Fig. 6 where there is no clear strong circulation response over the Eurasian continent even though snow cover has been substantially reduced. This further confirms that the NAO pattern is only slightly influenced by snow cover reduction, even though snow albedo reduction leads to strong interannual changes in snow cover over Eurasia in the SNHE experiment (Fig. 1).

As noted in Sect. 3.2, snow cover reduction in high resolution leads to a more significant impact on the large-scale circulation. This altered large-scale circulation pattern response coincides with a change in the correlation between the NAO index and surface air temperature. The regression coefficients of winter surface temperature anomalies onto the standardized winter NAO index in high-resolution simulations (HCTL, HNHE) are depicted in Supplementary Fig. S11. Compared to HCTL, the regression coefficients have been substantially increased in HNHE due to stronger warm responses over Eurasia. This relationship is reversed in the standard resolution experiment (Fig. 10), in which a warmer response in SNHE has led to a weaker correlation between the NAO index and surface air temperature compared to SCTL (Fig. 10). This indicates that the causality between snow reduction and the NAO circulation pattern is still ambiguous because different model resolutions may lead to different large circulation responses.

Fig. 12 Ensemble Mean surface air temperature differences between the Eurasian experiment (SEUA) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON and DJF (Dotted area denotes significant values except 95% confidence level based on a Student's t-test) (unit: °C)



4 Results from Eurasia snow albedo reduction experiments

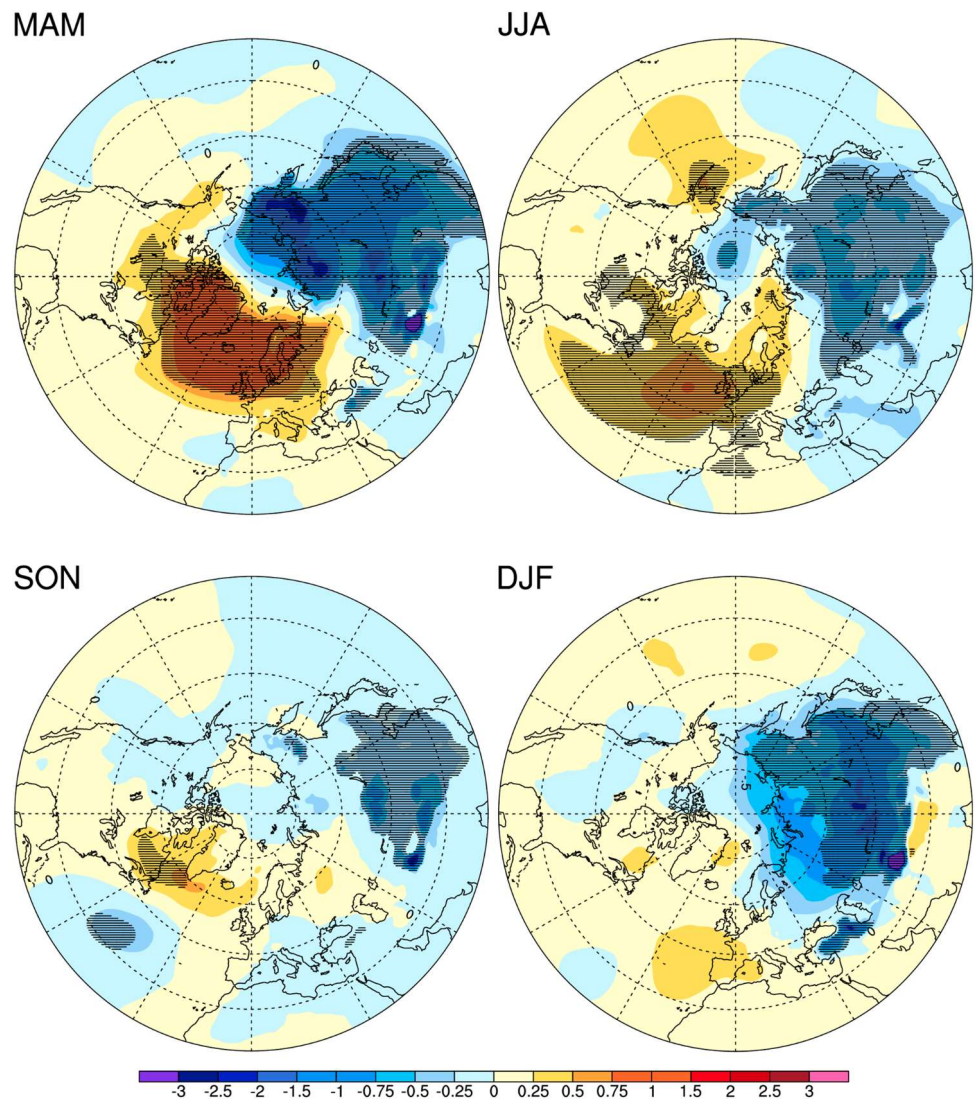
The above analysis of the Northern Hemisphere experiment showed that the changes in snow cover are closely correlated with surface temperature. Lower snow cover is coincident with strong and large-scale surface warming. Similar but weaker temperature responses are noted in the middle and upper troposphere while the temperature response signal in the stratosphere is reversed. Strong surface warming over high latitudes tends to weaken the poleward temperature gradient at mid-high latitudes and affect the mid-latitude jet and alter stationary eddy activities associated with changing surface forcing. However, the Northern Hemisphere experiments so far do not allow us to differentiate between these local effects and potential remote effects from snow reductions over North America, even though we have found that local processes are of large importance for Eurasian warming. How will the atmosphere respond when snow albedo is only reduced over the Eurasian region? To further elucidate the causality between the Eurasia snow cover reduction and the Eurasian climate, the Eurasian snow albedo reduction experiments (SEUA and HEUA) are analyzed.

In general, the SEUA experiment produces a similar response pattern for surface air temperature over the Northern Hemisphere as in the SNHE experiment (Supplementary Fig. S2), except that it only exerts a strong warm response over Eurasia with a minor warm response in the rest region of the Northern Hemisphere (Supplementary Fig. S12). Compared to the SNHE experiment (Fig. 2), the snow cover extents over Eurasia are almost unchanged in the SEUA experiment (Supplementary Fig. S1) and the

magnitude of response is slightly weaker over the Eurasian region and the warm response has been reduced by about 0.25–0.5 °C (Fig. 12). This indicates that the largest impact on the surface air temperature is mainly coming from a local snow cover reduction. Further inspection of high resolution experiments (HEUA) illustrates that the impacts of high resolution on snow cover (Supplementary Fig. S1) and surface air temperature response differences (Supplementary Fig. S6, right column) are minor and only marginally field significant.

Similar responses are also found in other surface parameters, e.g. Precipitation and heat fluxes (figures not shown). The response in the SEUA experiments is mainly confined to the Eurasian region with much weaker responses outside Eurasia. Despite slightly weaker surface warm responses, there still is a strong impact on the circulation variations, particularly in the upper air. Compared to the SNHE experiment (Fig. 6), the spatial SLP response patterns in the SEUA experiments have been altered in all seasons, especially in winter and spring (Fig. 13). In winter, strong negative SLP response is confined over Eurasia. In spring, except for the negative SLP response over Eurasia, a strong positive SLP response is seen over the North Atlantic Ocean which implies a weakening of the northeastern Siberian high and the Icelandic low. The SLP responses over North America and the North Pacific Ocean are minor in all seasons. This will alter the zonal pressure gradient between land and ocean and may further affect atmospheric circulation patterns. Similar to the North Hemisphere experiments, snow cover reduction in the Eurasian experiments also leads to a significant response in large-scale circulation in high resolution experiments (HEUA). Due to the stronger land-sea thermal contrast over Eurasia (Supplementary Fig. S12),

Fig. 13 Ensemble Mean sea level pressure differences between the Eurasian experiment (SEUA) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON and DJF (Hatched area denotes significant values except 95% confidence level based on a Student's t-test) (unit: hPa)

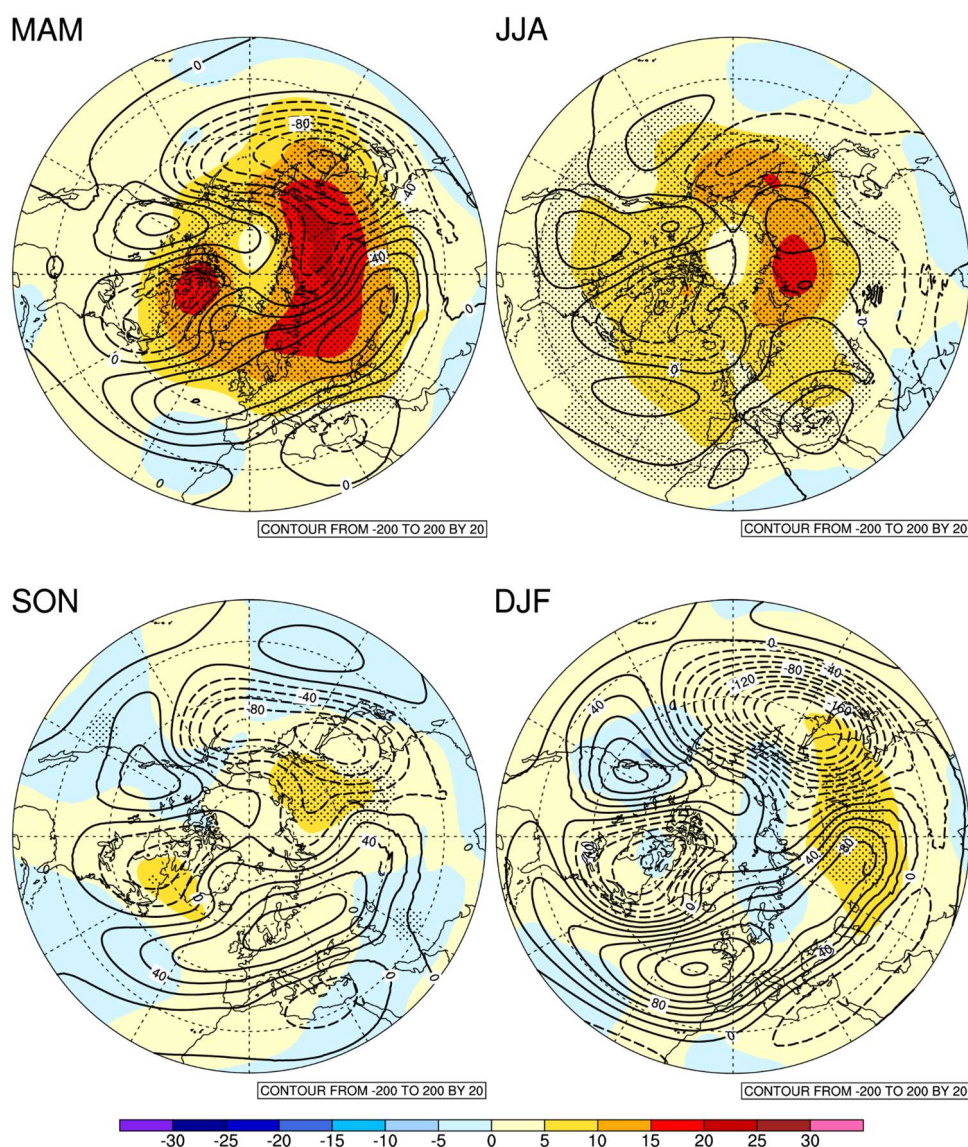


the Eurasia snow albedo experiments produce a more robust large-scale circulation response in high resolution in MAM and DJF (Supplementary Fig. S9, right column).

The different magnitude of the center of the AO pattern in the North Pacific Ocean and the North Atlantic Ocean could further modulate the remote atmospheric teleconnection (Gong et al. 2017). A stationary eddy response is less robust and apparent in the SEUA experiment (Fig. 14). In spring and summer, the 500 hPa stationary wave response resembles the spatial patterns as in the SNHE experiment (Fig. 7), but the magnitude is much weaker and mainly enhanced over Eurasia. The stationary eddy response is ignorable in autumn and winter. Both Fig. 15 and Supplementary Fig. S13 indicate that the direct diabatic heating due to Eurasian snow reduction leads to a much weaker meridional temperature gradient and westerly jet flow response. This implies that a less warm surface response in the SEUA experiment compared to the SNHE experiment could lead to

a less pronounced temperature meridional gradient response, however, this immediate meridional gradient response is sufficient to weaken upward temperature fluxes, hence leading to more energy absorption in the stratosphere and causing a smaller negative response in the troposphere (Fig. 15). Compared to the SNHE experiment, this is associated with a weakened westerly jet in the SEUA experiment (Supplementary Fig. S13) which could be partly due to a slightly weakened polar vortex (e.g. Gong et al. 2003; Fletcher et al. 2007; Cohen et al. 2014; Peings and Magnusdottir 2016). The impact of reduced snow cover is also more pronounced in the troposphere and stratosphere with high resolution, particularly over eastern Siberia in the HEUA experiment (figures not shown). The strong response in the upper troposphere has altered the meridional temperature gradient and the westerly jet flow change could further favor the formation of strong slow-moving or stationary anticyclones at high latitudes.

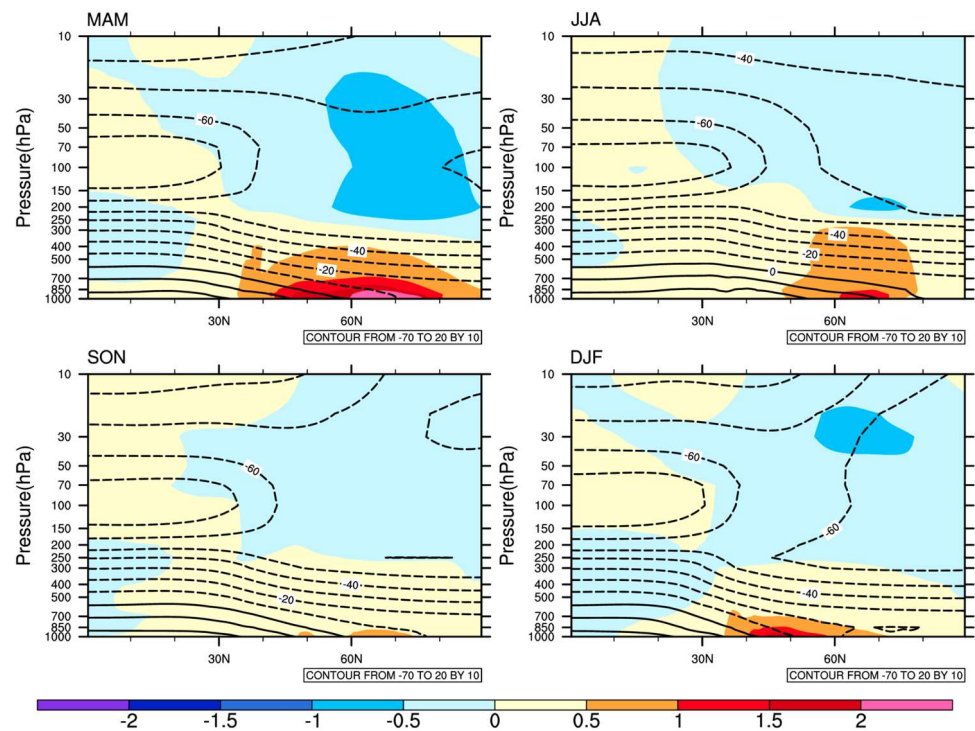
Fig. 14 Ensemble mean stationary eddies differences of the 500 hPa geopotential height between the Eurasian experiment (SEUA) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON and DJF (shaded area) (solid and dashed black lines denote the zonal mean eddy from control experiment) (Dotted area denotes significant values except 95% confidence level based on a Student's t-test) (unit: gpm)



Overall, snow reduction over Eurasia leads to a similar surface warming response as the SNHE and HNHE experiments, except that the magnitude is slightly weakened. This identifies that surface warming is mainly due to local snow cover reduction. However, the upper air circulation response is much less pronounced. This suggests that the large circulation response in the upper air over Eurasia is also partly influenced by snow reduction over North America. Klingaman et al. (2008) found that a regional North American snow cover anomaly in an area of high inter- and intra-annual snow cover variability has a strong teleconnection to European winter climate and the link between anomalous Great Plains snow cover and winter-time western Eurasian surface air temperatures facilitated by a snow-induced shift in the NAO toward positive phase. Compared to the SNHE experiment, there is no evident

surface warm response over North America in the SEUA experiment (Supplementary Fig. S12). Similar patterns are also found in high resolution experiments (figures not shown). Sobolowski et al. (2010) found that anomalous snow forcing that persists through the entire snow season is necessary to yield the atmospheric teleconnection pathway and remote transient-state response in Eurasia during winter-spring. Therefore, a snow-forced atmospheric teleconnection pathway that stretches from North America to eastern Siberia via transient eddy activity could contribute to a more robust circulation response over Eurasia (Sobolowski et al. 2010). This could partly explain the weaker upper air circulation response in the SEUA and HEUA experiments compared to SNHE and HNHE experiments, respectively.

Fig. 15 Ensemble Mean vertical temperature profile differences (averaged between 60 and 150 °E) between the Eurasian experiment (SEUA) and control experiment (SCTL) for standard resolution (T255) during MAM, JJA, SON and DJF (shaded area) (solid and dashed black lines denote the zonal mean temperature profile from control experiment) (unit: °C)



5 Summary and discussion

Previous studies have explored the snow-atmosphere coupling using prescribed surface snow forcing or nudging methods (e.g. Gong et al. 2003; Peings et al. 2012; Furtado et al. 2015). The surface snow-atmosphere interaction is absent from such methodologies. In this study, we adopt a similar methodology as Abe (2022) and previous sea ice albedo studies (Blackport and Kushner 2016; Chripko et al. 2021). We use the snow albedo to modify the snow cover in the EC-Earth3P model which is forced with daily SST and sea ice concentration (SIC) datasets. The snow albedo is set to a low constant value of 0.3 to enforce a reduction of the snow cover. The caveat of this method is that a constant low albedo is not realistic for real-world conditions. Two sensitivity experiments are performed in which the snow albedo is reduced in the Northern Hemisphere and over Eurasia, respectively, to understand how the lower snow albedo and subsequent snow cover reduction impact the Eurasian climate. To further investigate the impact of resolution, additional high-resolution experiments are performed. The major findings are summarized as follows:

- (1) In the Northern Hemisphere experiments, the reduced snow albedo results in robust thermodynamic and dynamic responses, spanning from the surface to the stratosphere. The induced snow cover reduction is closely correlated with surface temperature. Lower snow cover is coincident with strong and large-scale

surface warming in all seasons, particularly in winter and spring. Similar but weaker warm temperature responses are also observed in the middle and upper troposphere and reversed temperature responses in the stratosphere in all seasons except in autumn. The weakened poleward temperature gradient at mid-high latitude is associated with the enhanced high-latitude westerly jet flow and weakened mid-latitude westerly flow. Compared to Abe (2022), the response patterns are similar in both studies. However, the magnitude of response is about 50% weaker in our study. This is partly due to different snow albedo in the sensitivity experiment. The different model versions and surface forcing also play an important role in regulating the snow-albedo feedback.

Reduced snow cover is also characterized by increased surface temperature, which is accompanied by enhanced turbulent and radiative fluxes. This leads to different precipitation and P-E responses in western and eastern Eurasia and more precipitation occurs over eastern Eurasia, particularly in the Siberian region. The snow-NAO regression analysis reveals that the substantially reduced snow cover does not exhibit a higher correlation between the NAO index and surface temperature in our experiments. Although observational Siberian snow-NAO studies and a few numerical experiments have successfully captured this teleconnection (e.g. Cohen and Entekhabi 1999; Saito et al. 2001; Gong et al. 2003; Cohen et al. 2007, 2014;

Peings et al. 2013; Han and Sun 2018), the majority of models failed to represent such a relationship (e.g. Hardiman et al. 2008; Furtado et al. 2015; Handorf et al. 2015; Tyrrell et al. 2018; Peings et al. 2017). Hardiman et al. (2008) pointed out that the reason might be due to insufficient snow forcing over Siberia or the snow-driven local atmospheric anomaly being too confined longitudinally. This could prevent sufficient wave activity in the stratosphere to weaken the polar vortex. In our study, the SLP response shows a negative SLP response over Eurasia and this may be associated with weakened energy transferring from the troposphere to the stratosphere and has led to a negative temperature response in the stratosphere. Fletcher et al. (2009) found that a well-resolved stratosphere leads to a faster and weaker response to snow forcing compared to the response in the same model with a low-resolved stratosphere. EC-Earth3P model has 38 vertical levels above 100 hPa, which are many levels compared to most other climate models, and this could improve the representation of the lower-stratosphere circulation. This can also partly explain the weak snow-NAO relationship in our snow reduction experiments.

- (2) Compared to the Northern Hemisphere snow albedo reduction experiment, snow albedo reduction over the Eurasian region leads to slightly weaker surface responses over Eurasia in all seasons. As seen from Supplementary Fig. S12, the impact of strong anomalous snow reduction is generally limited to the vicinity of Eurasia and the response signal is minor outside Eurasia. The strong surface warming over Eurasia also leads to a strong circulation response upstream of the region where snow cover has been forced in spring and summer. However, the upper air circulation response is less pronounced and weakened from the lower stratosphere to the lower troposphere, suggesting that snow reduction over North America may also play an important role in regulating the large circulation response in the upper air over Eurasia.
- (3) The impact of resolution on the surface field responses (e.g. surface air temperature, precipitation) is small yet it is more pronounced on the large-scale circulation response, particularly in MAM and DJF. The different responses for large-scale circulation in the North Hemisphere and Eurasia snow albedo reduction experiments indicate that the causality relationship between snow reduction and the NAO circulation pattern is still unclear. The large circulation response differences may arise due to the differences in the surface mean climate state in T255 and T511 experiments. Notable changes in surface temperature are observed between SCTL and HCTL simulations. The evident differences in the sur-

face mean climate state could lead to large circulation response differences between the resolutions.

As we know, the Eurasian climate is strongly regulated by lower-frequency variability of atmospheric circulation in the North Atlantic region, due to the remote snow-atmosphere relationship, the Eurasian climate is also influenced by the snow variability in North America (Klingaman et al. 2008; Ge and Gong 2009; Sobolowski et al. 2010). A regional North American snow cover anomaly could lead to a stronger teleconnection to the European climate. Sobolowski et al. (2010) suggest that North American snow anomalies can initiate and maintain a physically plausible atmospheric teleconnection with consequences for Eurasian climate, via enhanced North Atlantic storm-track activity. This partly explained why the climate response is weaker in our Eurasian snow reduction experiments and confirmed that snow reduction in North America can influence the Eurasian climate through teleconnection. To further investigate planetary wave generation and propagation and better understand the mechanism behind the impact of snow reduction in North America, a more detailed analysis of the Rossby wave dynamics simulated by the model is required, which is beyond the scope of the current study.

Supplementary Information The online version contains supplementary material available at <https://doi.org/10.1007/s00382-023-06867-8>.

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Data availability Data is available by contacting the corresponding author.

Declarations

Conflict of interest The authors declare that they have no conflict of interest.

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