Future changes in precipitation intensity over the Arctic projected by a global atmospheric model with a 60-km grid size

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A B S T R A C T

Future changes in precipitation intensity over the Arctic were calculated based on three-member ensemble simulations using a global atmospheric model with a high horizontal resolution (60-km grid) for the period 1872–2099 (228 years). During 1872–2005, the model was forced with observed historical sea surface temperature (SST) data, while during 2006–2099, boundary SST data were estimated using the multi-model ensemble (MME) of the Coupled Model Intercomparison Project, Phase 3 (CMIP3) model, assuming the A1B emission scenario. The annual mean precipitation (PAE), the simple daily precipitation intensity index (SDII), and the maximum 5-day precipitation total (R5d) averaged over the Arctic increased monotonically towards the end of the 21st century. Over the Arctic, the conversion rate from water vapor to precipitation per one degree temperature increase is larger for PAVE than for R5d, which is opposite to the tropics and mid-latitudes. The increases in PAVE, SDII, and R5d can be partly attributed to an increase in water vapor associated with increasing temperatures, and to an increase in the horizontal transport of water vapor from low to high latitudes associated with transient eddies.

1. Introduction

According to the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5; IPCC, 2013), the surface temperature over the Arctic has increased ~1 °C during the past three decades, an amount that is significantly greater than the global mean warming trend, as shown in Fig. 2.22 of the IPCC report (IPCC, 2013). This relatively large warming over the Arctic as compared with the global averaged temperature increase, referred to as polar amplification, is conventionally explained by surface albedo feedback in polar regions due to the recent rapid and drastic reduction in the extent of sea ice. However, other feedback processes may also play an important role in climate change variability over the Arctic, such as lowwave radiation, lapse rate, cloud cover, and atmospheric water vapor content. In particular, Pithan and Mauritsen (2014) showed that different vertical structures of warming at high and low latitudes can accentuate polar amplification. Screen et al. (2012) indicated that warming of the atmosphere over the Arctic is influenced mainly by sea surface temperatures (SSTs) at low latitudes.

Precipitation over land areas of the Arctic has also shown an increasing trend during the last several decades (Pavelsky and Smith, 2006; Rawlins et al., 2010). From a thermodynamic perspective, this increase in precipitation is basically consistent with an increase in water vapor in the troposphere, as indicated by theoretical considerations (Held and Soden, 2006). Several observational studies have highlighted the effect of northward transport of water vapor on climate change in Arctic regions. Oshima and Yamazaki (2004) and Sorteberg and Walsh (2008) provided evidence that water vapor is transported to the Arctic by activity associated with extratropical cyclones.

Kattsov et al. (2007a) reported that simulations of precipitation by the 20th century Climate in Coupled Models (20C3M) over the Arctic by Atmosphere–Ocean General Circulation Models (AOGCMs), as reported in the IPCC Fourth Assessment Report (AR4) (IPCC, 2013), have improved in comparison with previous generation models reported in the IPCC Third Assessment Report (TAR) (IPCC, 2001). Nevertheless, the AOGCMs in IPCC AR4 remain biased with respect to precipitation over Arctic oceans and over major river basins in the Arctic, owing to a horizontal resolution that is insufficient for resolving local orography, errors in large-scale atmospheric circulation, and errors in sea ice distribution. Errors in the sea ice distribution in the Barents Sea might be an especially
important cause of the biases in atmospheric and oceanic circulation over the North Atlantic (Kattsov et al., 2007b).

Section 14.8.2 of the IPCC (2013) summarizes future climate projections over the Arctic based on models of the Coupled Model Intercomparison Project phase 5 (CMIP5) which participated in the IPCC (2013). An ensemble mean of the CMIP5 models shows strong polar amplification over the Arctic at the end of the 21st century, especially in winter. However, the CMIP5 model projections of the precipitation increase over the Arctic are less robust than those of temperature increase. The cause of the precipitation increase over the Arctic can be mostly attributed to precipitation enhancement by intensification of the activity of extratropical cyclones (IPCC, 2013, table 14.3) (Bengtsson et al., 2009; Zappa et al., 2013; Nishii et al., 2014).

The horizontal resolutions of climate models used in climate change studies are generally too low to incorporate the topography of Scandinavia and Greenland, which may result in an underestimation of orographic rainfall over these regions. Bengtsson et al. (2011) conducted a global warming projection using the high-horizontal-resolution (60-km grid size) atmospheric model ECHAM5. They found that the model provides good simulations of the seasonality of precipitation, evaporation over the Arctic, and the distribution of the horizontal transport of water vapor from lower latitudes to the Arctic. In a future warmer climate, meridional water vapor transport to the Arctic is projected to increase.

Since 2002, our global modeling group has been developing a global atmospheric model with a 20-km grid size (20-km model), which has a higher horizontal resolution than conventional climate models. Because the reproducibility of heavy precipitation during the East Asian summer rainy season is improved by use of the 20-km model (Kusunoki et al., 2006), a series of global warming projections using the 20-km model were conducted to study future changes in precipitation intensity over East Asia (Kamiguchi et al., 2006; Kusunoki et al., 2006; Kusunoki and Mizuta, 2008; Kim et al., 2010; Kusunoki et al., 2011; Endo et al., 2012). However, the duration of the target period in these studies was restricted to a few decades at the end of the 21st century, as the 20-km model requires inordinately large computational resources. Consequently, we have been developing a 60-km mesh version (60-km model) of the 20-km model to conduct ensemble simulations that will quantify uncertainties in future climate projections, as the computational time required for a 60-km model is 30 times less than that of the 20-km model.

To investigate long-term climate change over the 228 years from 1872 to 2099, we conducted additional continuous simulations using the 60-km model. This long-term climate projection yielded a continuous decrease in the number of tropical cyclones at global scales (Sugi and Yoshimura, 2012). The reason for this reduction is debated among the tropical cyclone research community, but a weakening of tropical circulation has been proposed as a possible mechanism by Sugi et al. (2002) and Bengtsson et al. (2007). In long-term simulations using the 60-km model, Kusunoki and Mizuta (2013) reported a continuous increase in precipitation intensity over East Asia.

Future changes in precipitation intensity over the Arctic have not yet been extensively investigated, perhaps partly because the amount of precipitation and the population size in the Arctic are much smaller than, for example, in East Asia, and thus the risks of natural disasters associated with heavy rainfall in the Arctic are much less than in other areas. However, the observed polar amplification, the observed increase in precipitation over the Arctic during the last several decades, and the continuous future upward trends in the projected temperature and precipitation in the Arctic (IPCC, 2013) indicate that future changes in precipitation intensity in Arctic regions should receive more attention. Although global warming projections by Bengtsson et al. (2011) used a high-horizontal-resolution model with a 60-km grid size, their target period was restricted to several decades, and the generation of only one simulation result hindered the evaluation of uncertainties in their future climate change projection.

The present study was motivated by the above review of Arctic climate models. The purpose of the present study is to investigate future change in precipitation and its intensity over the Arctic using a global atmospheric model with relatively high horizontal resolution. We further clarify the mechanism of changes in precipitation in terms of horizontal moisture transport and the conversion rate of moisture to precipitation, following the method of Kusunoki and Mizuta (2013).

2. Model and experimental design

This study used the Meteorological Research Institute Atmospheric–General Circulation Model, version 3.2 (MRI-AGCM3.2) developed jointly by the Japan Meteorological Agency (JMA) and the Meteorological Research Institute (MRI) (Mizuta et al., 2012). The model is based on a 60-km horizontal grid spacing and 60 vertical levels with an upper limit at 0.01 hPa (altitude of ~80 km). For cumulus convection, we implemented the so-called Yoshimura scheme (Yoshimura et al., 2014), which is based on the scheme of Tiedtke (1989). The Yoshimura scheme considerably improves the climatology of tropical convection (Mizuta et al., 2012). Our 60-km model is the same as that used in previous studies that investigated changes in tropical cyclones (Murakami et al., 2012; Sugi and Yoshimura, 2012), precipitation over Asia (Endo et al., 2012), and precipitation intensity over East Asia (Kusunoki and Mizuta, 2013).

A time-slice experiment (Bengtsson et al., 1996) was conducted in which the high-resolution AGCM was forced by prescribed external boundary conditions and forcings. For the 134 years from 1872 to 2005, the model was integrated with the monthly means of observed historical SSTs and sea ice concentrations from Hadley Centre Sea Ice and Sea Surface Temperature data set Version 1 (HadISST1) compiled by Rayner et al. (2003). For the 94 years from 2006 to 2099, the boundary SST data were created by superposing: (i) future changes in the multi-model ensemble (MME) of SST projected by the Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset; (ii) the linear trend in the MME of SST projected by the CMIP3 multi-model dataset; and (iii) the detrended observed SST anomalies for 1979–2003. Future changes in the MME of SST were evaluated using the difference between the 20C3M described in the IPCC AR4 (IPCC, 2007) and the future simulation for the IPCC Special Report on Emission Scenario (SRES) A1B emission scenario (IPCC, 2000). In this procedure, the 25-year interannual variations in SST anomalies between 1979 and 2003 were repeatedly added to the CMIP3 SST projection for consecutive future 25-year periods. Future sea-ice concentrations were obtained in a similar fashion. Mizuta et al. (2008) describe the method in more detail. As for greenhouse gases (GHG) such as carbon dioxide and methane, observed historical concentrations were prescribed for 1872 to 2000. After 2001, concentrations based on the A1B emission scenario were prescribed.

We used 3-dimensional natural and anthropogenic aerosol distributions calculated by the MRI-Earth System Model (Yukimoto et al., 2011) based on historical and A1B scenario aerosol emission data. Aerosols from volcanic eruptions were included only for the Mt. Pinatubo eruption of 1991. The 3-dimensional distributions of stratospheric ozone calculated by the MRI-Chemical Transport Model (CTM) (Shibata et al., 2005) based on historical and A1B scenario aerosol emission data were prescribed. To evaluate uncertainties originating from the internal variability of the model atmosphere, three-member ensemble simulations were performed.
for three different atmospheric initial conditions. The experiment design was identical to that adopted by Kusunoki and Mizuta (2013). We mainly analyzed daily precipitation data archived for the entire 228-year period from 1872 to 2099.

3. Present-day climate simulations

3.1. Verification data

To verify simulated precipitation results, we used the One-Degree Daily (1dd) data from the Global Precipitation Climatology Project (GPCP) V1.1 compiled by Huffman et al. (2001), which has a horizontal resolution of 1° in both longitude and latitude. We selected the GPCP 1dd data for verification purposes, as these daily data are required for calculating precipitation extreme events, although the period is limited to 1997 onwards. We terminated the GPCP data at 2005, as the model was forced with observed SSTs from 1872 to 2005.

We also used monthly data of the GPCP v2.2 compiled by Adler et al. (2003) for 1979–2005 (27 years), which has a horizontal resolution of 2.5° in both longitude and latitude. The monthly GPCP v2.2 data cover a longer period than the GPCP 1dd data, but the horizontal resolution is lower. We note that observed precipitation data over the Arctic include large uncertainties, mainly due to the low precipitation amounts over the Arctic and the absence of observation stations over the Arctic Ocean. Other possible causes of uncertainty may be related to the under-catchment of small precipitation by rain gauges, as indicated in section 6.1.1 of Serreze and Barry (2005) and by Kattsov et al. (2007a).

To verify simulated large-scale circulation, we used the Japanese 55-year Reanalysis [JRA-55] (Kobayashi et al., 2015) for 1986–2005 (20 years), which has a horizontal resolution of 1.25° in both longitude and latitude.

3.2. Indices of precipitation intensity

We adopted several of the 10 indicators of precipitation intensity proposed by Frich et al. (2002). The Simple Daily Precipitation Intensity Index (SDII; units, mm/day) is defined as the total annual precipitation divided by the number of rainy days (precipitation ≥ 1 mm/day); if no rainy days occurred at a grid point, a missing flag was applied to that point. The SDII is widely used in modeling studies (e.g., Dai, 2006; Global Climate Projection I, chapter 10 in IPCC, 2007). We also used the maximum 5-day precipitation total (R5d; units, mm) and calculated the annual mean precipitation (PAE; units, mm/day) as a basic measure of model performance.

3.3. Precipitation climatology

The Arctic is formally defined as the region north of the Arctic Circle (66.55°N) (Serreze and Barry, 2005). However, the target region as defined in previous studies relating to Arctic climate has varied depending on the focus and scope of the research. In this paper, we define the Arctic as the region north of 67.5°N, according to the definition adopted in the Atlas of Global and Regional Climate Projections, Annex I, figure A1.8–11, of IPCC (2013).

Fig. 1 compares the climatology of the GPCP 1dd (1997–2005) with the simulated climatology (1986–2005) of the present-day climate. The observed PAVE from the GPCP (Fig. 1a) is large over the Greenland Sea, the Norwegian Sea, Siberia at around 90°E, and the Chukchi Sea, but is small around the North Pole and over northern Greenland. In general, precipitation over the Arctic is much higher in summer than in winter (figure not shown). However, the spatial distribution of precipitation maxima over the Atlantic sector is related to the increase of cyclonic activity along the North Atlantic during winter (Serreze and Barry, 2005, Section 6.1.2), as well as high SSTs in the Atlantic sector which activate convection over the warm ocean during cold-air outbreaks from the Arctic. Similar to PAVE, the observed SDIs from the GPCP (Fig. 1b) are large over the Greenland Sea, the Norwegian Sea, Siberia at around 90°E, and the Chukchi Sea. The spatial distribution of R5d (Fig. 1c) is similar to that of the SDII (Fig. 1b).

The simulated PAVE (Fig. 1d) agrees well with the observed precipitation maximum over the Greenland Sea and the Norwegian Sea. The regional average annual precipitation according to the model (Fig. 1d) is 1.1 mm, which is larger than the observed value of 0.88 mm (Fig. 1g); this positive bias of 22% is partly due to model overestimates of precipitation in the North Pole region (Fig. 1g), where observations show a local minimum (Fig. 1a). The spatial correlation coefficient between observations (Fig. 1a) and simulations (Fig. 1d) is as high as 0.84. Kattsov et al. (2007a, Fig. 2) reported that AOGCMs in IPCC AR4 tend to overestimate the annual precipitation over Alaska and the western Arctic, whereas they tend to underestimate annual precipitation over the eastern Arctic and in the Norwegian–Barents Sea region. Excessive precipitation over Alaska and the western Arctic in our results (Fig. 1g) is consistent with the results of Kattsov et al. (2007a), but ours is 0.007 mm/day over the eastern Arctic and the Norwegian–Barents Sea region (Fig. 1g) is opposite to the results of Kattsov et al. (2007a). These differences in the bias distribution can be attributed to differences in used models and in the time periods of the analyses. We used an atmospheric model forced by observed SSTs and sea ice distributions, while Kattsov et al. (2007a) used AOGCMs, which incorporate errors in estimates of SSTs and sea ice distributions.

The simulated SDII (Fig. 1e) reproduces the intense precipitation over the Greenland Sea, but underestimates precipitation over other regions (Fig. 1h). The regional average SDI from the model (Fig. 1e) is 3.3 mm/day, which is 8% smaller than the observed value of 3.6 mm/day (Fig. 1h). The underestimate of the SDII (see definition in section 3.2) over the Arctic originates from the overestimate of the number of rainy days, indicating that the model tends to predict too many weak rainfall events (figure not shown).

The spatial correlation coefficient between observations (Fig. 1b) and our simulation (Fig. 1e) is as low as 0.68. The model (Fig. 1f) reproduces the observed R5d distribution (Fig. 1c) reasonably well, but overestimates it around Svalbard (Fig. 1i). The regionally averaged R5d from the model (Fig. 1f) shows a positive bias of 17% (Fig. 1i). The spatial correlation coefficient between observations (Fig. 1c) and the simulation (Fig. 1f) is 0.69, which is similar to that of the SDII (Fig. 1b and e).

Model simulations always generate uncertain results, especially at middle and high latitudes, on account of internal atmospheric variability (Deser et al., 2010). As we have conducted three-member ensemble simulations using different initial atmospheric conditions, we can estimate the magnitude of the uncertainty (spread) originating from atmospheric non-linear internal dynamics. Model spreads, as measured by the standard deviation of three-member ensemble simulation results, are shown by the hatched regions in Fig. 1d–f. Large areas of spread are commonly found over the North Atlantic, high latitudes in Eurasia, and Alaska; these areas generally correspond to high and intense precipitation regions, mainly caused by the activity of extratropical cyclones.

In contrast to the relatively large spreads at each grid point, the spread of regional average precipitation is very small. In the case of PAVE, the standard deviation (SD) of three runs is 0.007 mm/day, which is 0.7% of the regional average value of 1.1 mm/day. In the case of the SDII, the SD is 0.006 mm/day, which is 0.2% of the regional average value of 3.3 mm/day. In the case of R5d, the SD is 0.23 mm, which is 0.7% of the regional average value of 34 mm.
Consequently, we have confirmed that the model has the ability to reproduce the observed annual precipitation and precipitation intensity of present-day climate, although the model biases of precipitation intensity depend on the selection of skill measure of precipitation intensity.

### 3.4. Horizontal transport of water vapor

Fig. 2 compares observed and simulated values of the vertically integrated annual mean water vapor flux and net precipitation (precipitation minus evaporation). Observations (Fig. 2a) show large northeastward moisture transport over the Atlantic sector, as well as northward moisture transport over the Pacific sector through the Bering Strait. This northward moisture transport is also consistent with results of an observational study by Bengtsson et al. (2011). The water vapor flux associated with stationary atmospheric circulation makes a large contribution to the northward fluxes over the Atlantic and Pacific sectors (figure not shown). Peixoto and Oort (1992, section 12.3.2.2) and Jakobson and Vihma (2010) both indicated that this moisture transport is generally accompanied by transient cyclones, such as subpolar lows.

Moisture transported from mid-latitudes circulates anticlockwise around the North Pole (Fig. 2a). The model reproduces this observed flow of moisture well (Fig. 2c). In Fig. 2a, we also show regions of moisture flux convergence and divergence (with shading), which can be interpreted as net precipitation (Fig. 2b). The simulated moisture flux convergence and divergence (Fig. 2b) generally resembles the observed distribution (Fig. 2a). However,
northeastward moisture flux over the Norwegian Sea and the Barents Sea is overestimated by the simulation (Fig. 2e). The noisy pattern of bias in the flux convergence and divergence (Fig. 2e) might partly be the result of the noisy pattern of observational data (Fig. 2a).

Simulated net precipitation (Fig. 2d) shows a distribution...
similar to observations (Fig. 2b), but the model fails to simulate negative net precipitation over the North Atlantic (Fig. 2d). The model tends to underestimate net precipitation over Greenland and Baffin Bay, but overestimates net precipitation over the North Atlantic (Fig. 2f), due in part to the underestimation of evaporation over this area, especially in winter (figure not shown).

4. Long-term variation in precipitation

Fig. 3 shows time-series of simulated precipitation indices averaged over the Arctic. The target area is the same as in Figs. 1 and 2. In the case of PAVE (Fig. 3a), precipitation increases slightly from the 1870s through the 1980s, but later increases almost monotonically. Simulated precipitation overestimates the observed GPCP 1d1v11 data (red line) which is consistent with positive bias in Fig. 1g. Model also overestimates the observed GPCP 2.5-degree data (orange line). The amplitude of the year-to-year variability in the 21st century seems to be larger than that in the 20th century, but the change is not statistically significant based on F-tests using variances. An increase in the annual precipitation is consistent with the projection shown in annex I (Atlas of Global and Regional Climate Projections, figures AI.10 and 11) of IPCC (2013), although the emission scenario and target season of our study are different to those in annex I.

As for the SDII (Fig. 3b), precipitation increases slightly from the 1870s through the 1980s, but later increases almost monotonically, in a manner similar to PAVE. The simulated precipitation intensity is underestimated when compared with observations, which is consistent with Fig. 1b and e. In the case of R5d (Fig. 3c), the long-term trend resembles the trends in PAVE and the SDII. The simulated precipitation intensity is overestimated as compared with observations, which is consistent with Fig. 1c and f. In summary, both annual precipitation and precipitation intensity increase monotonically towards the end of the 21st century.

5. Geographic distribution of future changes in precipitation

The geographic distribution of future changes in precipitation is illustrated in Fig. 4 for the periods 2046–2065 and 2080–2099. In the case of PAVE, for 2046–2065 (Fig. 4a) the precipitation increases over nearly all regions except for the oceans to the east of Greenland. For the period 2080–2099 (Fig. 4d) the regions of statistically significant precipitation increase are much larger than for 2046–2065 (Fig. 4a). The geographic distribution of changes in annual precipitation for the period 2080–2099 resembles the projections of the 21 multi-model ensemble of CMIP3 models, as shown in Fig. S11.28 of IPCC (2013). Moreover, the precipitation increase over land in Eurasia and North America is consistent with the results of Kattsov et al. (2007a). A general increase in precipitation over the Arctic at the end of 21st century is also consistent with results of Walsh et al. (2013).

As for the SDII, between 2046 and 2065 (Fig. 4b) the precipitation increases over all regions, but the region of statistically significant increase is smaller than for PAVE (Fig. 4a). For the period 2080–2099 (Fig. 4e) the precipitation intensity increases over all regions, with statistically significant regions being much larger than for 2046–2065 (Fig. 4b). In the case of R5d (Fig. 4c), precipitation intensity generally increases over almost all regions, but the statistically significant regions are small when compared with PAVE (Fig. 4a) and SDII (Fig. 4b). For the period 2080–2099 (Fig. 4f) the precipitation intensity increases over the entire region. In terms of the SDII and R5d, precipitation intensity generally increases over the Arctic for 2046–2086 and 2080–2099. The degree of increase and the area of statistically significant regions are much larger for 2080–2099 than for 2046–2065.

6. Frequency distribution of intense precipitation

The probability distribution function (PDF) of intense precipitation is illustrated in Fig. 5 as a function of daily precipitation amount. For each year and at each grid point, we counted the number of days that fall in the range of precipitation defined by each bin (bin interval, 1 mm/day). We then normalized the number of days for each bin by the total number of days in the target year (365 for non-leap years; 366 for leap years). Finally, we averaged all PDFs in the target years to derive the observed and simulated climatology at each grid point. Regional averages were calculated on the basis of the grid-point-derived climatology.

In Fig. 5a, the observed frequency of precipitation over oceans (long black dashes) is higher than that over land (short black dashes), indicating that the recycling of water is much more efficient over oceans than over land. On the other hand, the simulated frequency of precipitation over land (short blue dashes) is higher than that over oceans (long blue dashes). It is possible that the contrast in the hydrologic cycle between ocean and land is not properly reproduced by the model. Moreover, the model (blue lines) tends to overestimate the frequency of intense precipitation, especially precipitation of over 60 mm/day.

In Fig. 5b, the increase in precipitation frequency during 2080–2099 (red) is larger than that during 2046–2065 (green). In the present-day (2086–2005) climate, as well as in the future periods 2046–2065 and 2080–2099, the frequency of precipitation over land (short dashed line) is higher than that over oceans (long dashed line). The larger increase in frequency during 2080–2099 compared with 2046–2065 can also be confirmed by the ratio of future change relative to the present-day climate (Fig. 5c). Most of the changes in Fig. 5c are statistically significant at the 95% confidence level.

7. Mechanism of precipitation change

The mechanisms driving the increase in PAVE, SDII, and R5d (indicated in Figs. 3 and 4) can be roughly divided into two categories: thermodynamic environmental changes and dynamic circulation changes.

7.1. Thermodynamic effect

In principle, increases in the precipitation amount and precipitation intensity can be ascribed to the increased availability of water vapor associated with an increase in the temperature of the atmosphere (IPCC, 2007, 2013). The annual surface air temperature will rise by approximately 2°C–6°C over the Arctic during 2080–2099 (Fig. 6a), causing an increase in annual precipitation (Fig. 6b). Fig. 6c shows the rate of annual precipitation change per 1°C increase in surface air temperature; this quantity can be regarded as a conversion rate from water vapor to precipitation, which is referred to as the 'hydrological sensitivity' in section 9.3.4 of the IPCC TAR (IPCC, 2001). The hatched regions in Fig. 6c show the rate of precipitation increase (7.5%/°C) that is theoretically expected from the Clausius–Clapeyron (C–C) relationship (Vecchi and Soden, 2007). The hydrological sensitivity of PAVE (Fig. 6c) does not reach the level of the C–C relationship in most areas. However, the hydrological sensitivity exceeds the level of the C–C relationship over Greenland, partly because of orographic effects on rainfall that might be caused by Greenland’s elevation.

Fig. 7 shows precipitation (PAVE, SDII, and R5d) changes averaged over the Arctic as a function of the surface air temperature change and the temperature change averaged over the same region. For example, the increase in PAVE during the near-future decade of
2020–2029 ranges from 6.2% to 10.6% for individual simulations (blue 'x' symbols), with a value of 8.3% for the ensemble average (blue circles). These changes are associated with a surface air temperature increase of ~1.9°C, resulting in a hydrological sensitivity rate of 3.2%/°C to 5.5%/°C for individual simulations and 4.4%/°C for the ensemble average. During the last decade of the 21st century (2090–2099), changes in temperature and PAVE are much larger than during the period 2020–2029, but the hydrological sensitivity rate of the ensemble average is 4.6%/°C, which is similar to that for 2020–2029. The hydrological sensitivity rate of PAVE, calculated using only ensemble average values estimated by linear regression, is 4.8%/°C. This is consistent with findings from previous global warming studies, that the conversion from water vapor to precipitation is less effective than that expected from the C–C relationship (7.5%/°C) between water vapor and temperature (Held and Soden, 2006; Vecchi and Soden, 2007). In the case of the tropics, this low sensitivity might be attributed to the weakening of atmospheric circulation (Held and Soden, 2006; Vecchi and Soden, 2007).

Fig. 3. Time-series of precipitation indices averaged over the Arctic region (67.5°N–90°N; see Fig. 1) from 1872 to 2099 (228 years). Red line shows observations from GPCP 1d dv11 data for 1997–2005 (9 years). Orange line shows observations from GPCP 2.5 data for 1979–2005 (27 years). Thick line shows the three-member ensemble average (MAVE). Maximum and minimum ranges from the individual simulations are shaded. Green lines show three 20-year target periods (1986–2005, 2046–2065, and 2080–2099) for analyses in Figs 1 and 4. (a) PAVE (mm/day). (b) SDII (mm/day). (c) R5d (mm).
2007); however, it is not clear that this explanation can be applied to high-latitude areas.

In the case of the SDII and R5d, the hydrological sensitivity rate estimated by a procedure similar to that described above is 5.6%/°C and 4.2%/°C, respectively. The hydrological sensitivity rates measured by indices of heavy precipitation are similar to those estimated by annual precipitation. Moreover, in the East Asia region, the hydrological sensitivity of heavy precipitation is greater than that of moderate precipitation (Kusunoki and Mizuta, 2013), whereas the reverse is true over the Arctic (i.e., the hydrological sensitivity of heavy precipitation is less than that of moderate precipitation).

We further investigated the dependence of the hydrological sensitivity rate on regional variations. Fig. 8 compares the hydrological sensitivity rate for the future period 2080–2099 with the present-day (1986–2005) climatology for different latitudinal zones. In the case of PAVE (blue), the hydrological sensitivity over the Arctic and Antarctica is larger than that of the global average.
Fig. 5. Probability distribution function (PDF) of the daily precipitation intensity (mm/day) over the Arctic region (67.5°N–90°N). Solid lines denote statistics for all areas of the Arctic. Long dashed lines denote statistics over oceans. Short dashed lines denote statistics over land. (a) Observations and model simulation for the present-day. Unit is 10^{-6}. Black line showing observations from the GPCP 1ddv1.1; data for 1997–2005 (9 years). Blue lines show the simulated climatology (1986–2005) using the three-member ensemble average. (b) Model simulations for the present-day and the future. Unit is 10^{-6}. Blue lines show the simulated present-day climatology (1986–2005). Green lines show the near-future climatology (2046–2065). Red lines show the future climatology (2080–2099). (c) Future change ratio in the frequency relative to the present climatology. Closed circles indicate a confidence level exceeding 95%, based on Student’s t-test.
Larger hydrological sensitivity values over the Arctic, as compared with the global average, are consistent with the results of Bintanja and Selton (2014). This enhancement of hydrological sensitivity over the Arctic might be partly attributed to a decrease in sea ice over the Arctic Ocean, which results in enhanced evaporation of water from the ocean (Bintanja and Selton, 2014). In the global case, the hydrological sensitivity of intense precipitation (i.e., the SDII and R5d) is higher than that of the annual precipitation (PAVE); this relationship also holds for mid-latitudes and the tropics. In contrast, the hydrological sensitivity of intense precipitation (R5d) over the Arctic and Antarctica is lower than that of PAVE, which suggests that the mechanism of intense rainfall over high-latitude regions is somewhat different to that in low-latitude regions. As compared with the tropics and mid-latitudes, lower temperatures in high latitudes lead to a lower water-vapor-holding capacity, which might inhibit the effective conversion of water vapor to intense precipitation.

7.2. Dynamic effects

Thermodynamic effects alone cannot be used as the basis for future changes in precipitation amount and intensity, as dynamic effects also play an important role in mid-latitudes through horizontal advection processes (Meehl et al., 2005), and in the tropics through vertical motions (Emori and Brown, 2005). Here, we highlight the contribution of the combined effects of dynamic circulation and water vapor transport on changes in precipitation amount and intensity, as projected in our simulations.

Fig. 9a–c illustrates future changes in the vertically integrated annual mean water vapor flux and its convergence for 2080–2099 relative to the present-day (1986–2005) climatology and its convergence, and in annual precipitation (Fig. 9d). The total water vapor flux is calculated for 6-hourly data of specific humidity and wind (Fig. 9a). We decomposed the total flux into two parts: the stationary flux, which is calculated by the monthly average of
specific humidity and wind for each month and year (Fig. 9b), and
the transient flux, which is the contribution of transient eddies,
defined as the deviation of 6-hourly specific humidity and wind
from the monthly mean for each month and year (Fig. 9c).
In practice, the transient flux is calculated by subtracting the sta-
tionary flux from the total flux for each month and year. The
transient water vapor flux associated with transient eddies plays an
important role in the hydrologic cycle and budget in the Arctic
(Peixoto and Oort, 1992; Groves and Francis, 2002; Oshima and
Yamazaki, 2004).

As for the total flux (Fig. 9a), the northeastward water vapor flux
increases in the 0°–90°E and 90°W–180°W sectors, suggesting that
increasingly large amounts of water vapor are transported from
lower latitudes. Changes in the anticlockwise water vapor flux
are evident around the North Pole. Changes in the convergence of
the water vapor flux (shaded) are generally positive (blue) over the
Arctic, the North Atlantic, Greenland, and northern Eurasia (Fig. 9).
These changes may partly contribute to the increase in precipita-
tion and precipitation intensity over these regions (Fig. 4d–f). For
other regions, a correspondence between the spatial distribution
of moisture flux convergence and divergence patterns (Fig. 9a) and
the spatial distributions of annual precipitation changes (Fig. 9d)
is not evident. This is partly because the distribution of evaporation
also changes in the future.

The spatial pattern and magnitude of the stationary flux (Fig. 9b,
arrows) are similar to those of the total flux (Fig. 9a, arrows). The
convergence of the stationary flux (Fig. 9b, shading) is generally
positive over the target region. In contrast, the transient flux
(Fig. 9c, arrows) is north-directed almost everywhere in the Arctic,
although the magnitude of the flux is less than that of the sta-
tionary flux (Fig. 9a, arrows). The convergence of the transient flux
(Fig. 9c, shading) is large and negative in several regions, which
correlates to the negative convergence of the total flux in some
regions (Fig. 9a, shading). Changes in the annual precipitation can
be partly interpreted to result from changes in the stationary flux.

The calculations for Fig. 9 considered both the contributions of
changes in water vapor (a thermodynamic effect) and wind (a dy-
namic effect). To separate the contributions of these effects on
changes in the water vapor flux, we conducted the additional an-
alyses shown in Fig. 10. Considering that most of the water vapor is
concentrated in lower levels of the troposphere, we selected the
850-hPa level as a target level. Fig. 10a shows future changes in
the total water vapor flux. Fig. 10b shows future change in the sta-
tionary flux. The spatial pattern of the stationary flux (Fig. 10b) is
similar to that of the total flux (Fig. 10a), indicating that transient
flux makes a small contribution to the total flux.

The present-day monthly mean wind and the future monthly
mean water vapor were used to calculate future changes in the
water vapor stationary flux (Fig. 10c), which can be regarded as the
contribution from the water vapor change. On the other hand,
Fig. 10d shows the application of present-day monthly mean water
vapor and future monthly mean wind to calculate future changes in
the water vapor stationary flux, which can be regarded as the
contribution due to wind change. The patterns in Fig. 10c generally
resemble those in Fig. 10b, suggesting the dominance of the water
vapor change effect; this is consistent with the results of previous
modeling studies by Cassano et al. (2007), Skific et al. (2009), and
Bintanja and Selton (2014). However, the anticlockwise water vapor
flux change over the North Atlantic (Fig. 10b) is only represented in
Fig. 10d; this means that the water vapor flux change over the North
Atlantic can be attributed to the change in the wind field.

To quantify the amount of water vapor transported from lower
latitudes to the Arctic, we analyzed the longitudinal profile of
change in the meridional component of the vertically integrated
annual mean water vapor flux at 67.5°N (Fig. 11). The observed total
flux (Fig. 11a, black solid line) shows a large northward component
over the Atlantic sector, and a large southward component at
90°W. The longitudinal average in the observed total flux is pos-
itive (3.36 kg/m²s). The longitudinal profile of the observed sta-
tionary flux (Fig. 11a, black long-dashed line) is similar to that of the
total flux, but the southward component of the observed stationary
flux is larger than that of the total flux. This larger southward flux
leads to the smaller longitudinal average of the observed stationary
flux, which, at 1.28, is 15.3% of the longitudinal average of the total flux. The observed transient flux (Fig. 11a, black short-dashed line) is northward at all locations, although the magnitude is smaller than the magnitudes of the total flux and the stationary flux. The longitudinal average of the observed transient flux is positive (7.08 kg/m/s), is larger than that of the stationary flux, and is 84.7% of the longitudinal average of the total flux. Profiles of the simulated fluxes (Fig. 11a, in blue) are similar those of the observed

![Diagram](image_url)

Fig. 9. (a) Future change in the vertically integrated annual mean water vapor flux (arrows; Kg/m/s) for 2080–2099 relative to the present-day climatology (1986–2005) and its convergence (shading; mm/day), based on 6-hourly specific humidity and wind. The unit of convergence is converted to mm/day assuming a density of liquid water of 1 g/cm³. Red arrows denote regions where the confidence level exceeds 95%, based on Hotelling’s T² statistics (Storch and Zwiers, 1999; Wilks, 2011), obtained using the three-member ensemble average. (b) Same as (a) but for the stationary flux, based on monthly averages of specific humidity and wind. (c) Same as (a) but for the transient eddy flux, based on deviations of 6-hourly wind and specific humidity, from monthly averages. (d) Changes in the annual mean precipitation (same as Fig. 4d but in units of mm/day).
Fig. 11b compares simulated fluxes for present-day climate (blue) with those of future climate (red). The northward total and stationary fluxes increase over the Atlantic sector and Alaska, and the southward total flux and stationary fluxes increase at around 90°W. Increasing northward transport over the Atlantic sector and Alaska is attributed in part to the stronger cyclonic activity in the North Atlantic and North Pacific regions, respectively. By analyzing the global warming projections using the same 60-km model as that used in the present study, Mizuta et al. (2011) found that the increase in the frequency of intense cyclones is observed on the polar and downstream sides of Atlantic and Pacific storm tracks in winter. The longitudinal average of the change in the total flux is 1.24 kg/m/s, indicating that a substantial amount of water vapor is transported from lower latitudes to the Arctic. The longitudinal average of the change in the transient flux of 1.09 kg/m/s is 88% of the change in the total flux. Therefore, the future change in water vapor transport from lower latitudes to the Arctic is caused mainly by transient eddies.

Fig. 12 shows the seasonality of the longitudinally averaged change in the meridional component of the vertically integrated annual mean water vapor flux at 67.5°N for 2080–2099, relative to present-day (1986–2005) climatology. Changes in the total flux (gray) are positive for all seasons. The contribution of the transient flux (orange) to the positive change in the total flux (gray) is much larger than that of the stationary flux (purple) for all seasons. The changes in the total flux (gray) and transient flux (orange) in summer are the largest of all seasons. Time-slice simulations by Kug et al. (2010) demonstrated that a poleward shift of storm tracks in the Northern Hemisphere could enhance polar warming and moistening. They indicated that the largest contribution of transient eddies to precipitation over the Arctic occurs in summer (Kug et al., 2010, Fig. 4c). Our results are consistent with theirs, as the largest northward transport of water vapor in summer (Fig. 12) can be attributed at least in part to the enhanced activity of transient eddies in summer.

8. Conclusions

Our results are summarized as follows.
1. The model reproduces observed annual precipitation, precipitation intensity, and horizontal transport of water vapor of the present-day climate (1986–2005) over the Arctic (67.5°–90°N) reasonably well.

2. Annual precipitation and precipitation intensity averaged over the Arctic both increase monotonically towards the end of the 21st century.

3. Annual precipitation and precipitation intensity increase in most regions over the Arctic for the period 2080–2099.

4. The statistically significant area of annual precipitation and precipitation intensity increase are larger for the period 2080–2099 than for 2046–2065.

5. The increases in annual precipitation and precipitation intensity can be attributed in part to an increase in water vapor associated...
with warming. The conversion rate from water vapor to PAVE, SDII, and R5d per 1°C increase in the surface air temperature (hydrological sensitivity) is estimated at 4.8%/°C, 5.6%/°C, and 4.2%/°C, respectively.

6. The hydrological sensitivity of intense precipitation (R5d) over the Arctic and Antarctica is lower than that of PAVE, while the opposite is true in the tropics and mid-latitudes where the hydrological sensitivity of intense precipitation (SDII, R5d) is higher than that of PAVE.

7. The increase in the horizontal transport of water vapor from lower latitudes to the Arctic is responsible for increases in annual precipitation and intense precipitation over the Arctic, mainly as a result of the influence of transient eddies.

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Annex


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