ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 35, SEPTEMBER 2018, 1129–1136

• **Original Paper** •

Changes in the Proportion of Precipitation Occurring as Rain in Northern Canada during Spring–Summer from 1979–2015

Wei HAN[1,2], Cunde XIAO[*3], Tingfeng DOU[4], and Minghu DING[2]

[1]*Nanjing University of Information Science and Technology, Nanjing* 210044, *China*
[2]*Institute of Polar Meteorology, Chinese Academy of Meteorological Sciences, Beijing* 100081, *China*
[3]*State Key Laboratory of Earth Surface and Resource Ecology, Beijing Normal University, Beijing* 100875, *China*
[4]*College of Resources and Environment, University of Chinese Academy of Sciences, Beijing* 100049, *China*

(Received 15 September 2017; revised 24 February 2018; accepted 24 February 2018)

ABSTRACT

Changes in the form of precipitation have a considerable impact on the Arctic cryosphere and ecological system by influencing the energy balance and surface runoff. In this study, station observations and ERA-Interim data were used to analyze changes in the rainfall to precipitation ratio (RPR) in northern Canada during the spring–summer season (March–July) from 1979–2015. Our results indicate that ERA-Interim describes the spring–summer variations and trends in temperature and the RPR well. Both the spring–summer mean temperature [0.4°C–1°C $(10\ \mathrm{yr})^{-1}$] and the RPR [2%–6% $(10\ \mathrm{yr})^{-1}$] increased significantly in the Canadian Arctic Archipelago from 1979–2015. Moreover, we suggest that, aside from the contribution of climate warming, the North Atlantic Oscillation is probably another key factor influencing temporal and spatial differences in the RPR over northern Canada.

Key words: climate change, rainfall to precipitation ratio, northern Canada, North Atlantic Oscillation

Citation: Han, W., C. D. Xiao, T. F. Dou, and M. H. Ding, 2018: Changes in the proportion of precipitation occurring as rain in northern Canada during spring–summer from 1979–2015. *Adv. Atmos. Sci.*, **35**(9), 1129–1136, https://doi.org/10.1007/s00376-018-7226-3.

1. Introduction

Climate change has amplified in the Arctic (Serreze et al., 2009; Screen and Simmonds, 2010; Pithan and Mauritsen, 2014), and Arctic surface temperatures are rising at a rate more than twice the global average (Bekryaev et al., 2010; Miller et al., 2010). As a result, there have been unprecedented declines in the Arctic sea-ice area, leading to larger open water areas that are exposed for longer periods of time (Cavalieri and Parkinson, 2012) and the Arctic becoming warmer and wetter (Przybylak, 2007; Overland et al., 2014; Boisvert and Stroeve, 2015).

Precipitation is more susceptible to warming trends in high-latitude regions (Irannezhad et al., 2016); however, most studies of precipitation change in high-latitude regions have concentrated on analyses of total precipitation (Serreze et al., 2000; Yao et al., 2012; Irannezhad et al., 2016), as well as heavy and extreme precipitation (Zhang et al., 2001; Groisman et al., 2005). However, the precipitation form is equally as important as the quantity and intensity for understanding the seasonality of hydrological cycles and the health of the ecosystem in the cryosphere (Hasnain, 2002; Putkonen and Roe, 2003; Ye, 2008).

Precipitation falls to the ground as rain, snow, sleet, and other forms, each of which has considerable impacts on the surface runoff and energy balance (Loth et al., 1993; Ding et al., 2014). As the air temperature rises, more precipitation falls as rain instead of snow (Knowles et al., 2006; Screen and Simmonds, 2012; Ye and Cohen, 2013). This rain brings heat to the snow cover and affects the snow morphology and albedo (Stirling and Smith, 2004), causing the surface to absorb more solar energy and accelerating snow cover and sea-ice melt (Perovich and Polashenski, 2012). Conversely, snowfall can stop or reverse the decline in albedo during the initial melting phase (Perovich et al., 2017). This emphasizes the importance of different precipitation forms in the rapidly changing Arctic climate system.

Previous studies of precipitation forms have concentrated on the number of rain days during the cold season (Aanes et al., 2000; Putkonen and Roe, 2003; Cohen et al., 2015). There are relatively few stations in the Arctic, and only a minority of them have recorded liquid and solid precipitation; thus, changes in precipitation forms over the Arctic during the spring–summer transition period are poorly understood. The transition from the cold season (snow accumulation period)

* Corresponding author: Cunde XIAO
Email: cdxiao@bnu.edu.cn

to the warm season (melting period) generally occurs during late-spring through mid-summer in Canada. Thus, the variability and potential trends of the rainfall to precipitation ratio (RPR) in Canada north of 60°N during the spring–summer period were analyzed in the present study.

We used observational data and ERA-Interim data to analyze changes in the proportion of precipitation occurring as rain in northern Canada and discuss its possible causes. The remainder of this paper is organized as follows: Section 2 introduces the data and method used in this study. Section 3.1 compares the ERA-Interim with the observational data during 1979–2007. Section 3.2 presents the variations in surface air temperature and RPR in northern Canada during 1979–2015. Section 3.3 discusses the variability associated with the North Atlantic Oscillation (NAO). Finally, Section 4 summarizes and discusses the conclusions.

2. Data and methods

2.1. *Data*

We obtained the daily data from the Canadian Daily Meteorological database for 11 meteorological stations in northern Canada from Environment and Climate Change Canada (Table 1). All of the selected stations include the daily mean temperature (°C), precipitation (mm), rainfall (mm) and snowfall as water equivalent (mm during March–July from 1979 to 2007. All data were subjected to quality control using "DLY04" daily network programs (DLY is the term used to refer to the Monthly Record of Daily Data) that are based on observations made at manned and automated sites (see http://climate.weather.gc.ca/about_the_data_index_e.html). The stations were culled according to the following steps:

(1) Any station that was missing one or more elements (daily mean temperature, precipitation, rainfall, snowfall) was excluded.

(2) It was considered incomplete in one month if there were missing data for five or more days in this month.

(3) It was considered incomplete in one year if there were missing data for one or more months between March and July in this year.

(4) Any station that was missing more than three years of data in 1979–2007 was excluded.

Because of the sparse distribution of station locations and a lack of recent observational data, we used reanalysis data (ERA-Interim) in this study. ERA-Interim is the global atmospheric reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). When compared with previous reanalysis data, significant advances have been made in the hydrological cycle, the quality of the stratospheric circulation, and the consistency in time of the reanalyzed fields via many model improvements, the use of four-dimensional variational analysis, a revised humidity analysis, the use of variational bias correction for satellite data, and other improvements in data handling (Dee et al., 2011). Screen and Simmonds (2012) found no obvious tendencies or discontinuities in the snowfall-to-precipitation

Table 1. Annual average and correlation of spring–summer mean temperature, total rainfall, total precipitation and the RPR between the observation data and ERA-Interim data from 11 stations during 1979–2007.

Station name	Latitude	Longitude	Spring–summer mean T (°C)			Total rainfall (mm)			Total precipitation (mm)			RPR (%)		
			Obs	ERA-I	Correlation	Obs	ERA-I	Correlation	Obs	ERA-I	Correlation	Obs	ERA-I	Correlation
EUREKA A	79.98°N	85.93°W	−13.15	−11.64	0.775**	19.21	37.9	0.852**	31.86	78.13	0.812**	56.46	47.29	0.746**
RESOLUTE CARS	74.72°N	94.97°W	−11.73	−9.41	0.718**	28.2	46.46	0.771**	63.17	90.69	0.705**	42.77	50.13	0.573**
CAMBRIDGE BAY A	69.11°N	105.14°W	−9.68	−7.39	0.884**	34.27	50.72	0.673**	57.22	83.3	0.702**	57.99	60.29	0.596**
NORMAN WELLS A	65.28°N	126.80°W	2.88	1.92	0.963**	93.87	158.57	0.800**	121.27	192.82	0.802**	76	81.4	0.725**
BAKER LAKE A	64.30°N	96.08°W	−6.72	−6.06	0.940**	67.77	94.82	0.724**	106.34	145.11	0.751**	62.11	64.61	0.618**
CORAL HARBOUR A	64.19°N	83.36°W	−7.14	−6.88	0.976**	66.96	95.6	0.801**	117.88	150.92	0.825**	54.66	62.55	0.671**
MAYO A	63.62°N	135.87°W	6.16	2.04	0.956**	115.65	222.07	0.573**	132.13	255.46	0.610**	87.41	87.08	0.506**
YELLOWKNIFE A	62.46°N	114.44°W	2.52	2.42	0.973**	84.01	75.13	0.721**	111.36	96.3	0.720**	73.9	76.31	0.670**
HAY RIVER A	60.84°N	115.78°W	3.45	5.45	0.953**	100.24	128.28	0.817**	128.39	150.33	0.838**	75.92	84.83	0.848**
WATSON LAKE A	60.12°N	128.82°W	5.29	4.18	0.949**	157.1	194	0.861**	185.86	236.66	0.881**	83.4	81.56	0.749**
FORT SMITH A	60.02°N	111.96°W	5.34	5.11	0.977**	129.5	159.33	0.803**	153.71	182.9	0.812**	83.39	86.54	0.885**

**Statistically significant at the greater than 99% confidence level.

ratio (SPR) difference between observations and ERA-Interim as a function of time. The precipitation and snowfall are the 12-h accumulated totals, while the temperature is the 12-h mean in ERA-Interim. Therefore, we summed the 12-h accumulated totals to have daily accumulated precipitation and snowfall (Screen and Simmonds, 2012) and averaged the two 12-h mean temperatures to have a daily mean temperature. ERA-Interim daily data in March to July from 1979 to 2015, including precipitation (mm d^{-1}), snowfall as water equivalent (mm d^{-1}) and 2-m temperature (K) for midday and midnight, at time step of 12-h, are provided at a horizontal resolution of $0.75^\circ \times 0.75^\circ$ The ERA-Interim data can be downloaded from ECMWF (http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/), while the AO and NAO indices can be downloaded from NOAA/CPC (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/teleconnectiote.shtml).

2.2. *Methods*

We focused in this study on the spring–summer season, defined here as March to July. The spring–summer mean temperature ($\bar{T}$) here is the average of the daily mean temperature (T) from March to July. The spring–summer AO/NAO index (AO: Arctic Oscillation) is also the averaged monthly AO/NAO index during the same period.

The SPR is widely used to reflect the proportion of precipitation occurring as snow. However, here we focus on the rainfall in spring–summer and express the proportion of precipitation occurring as rain as a rainfall-to-precipitation ratio (RPR):

$$\mathrm{RPR} = \frac{R}{P} \times 100\% ,$$

where R is the total rainfall and P is the total precipitation in spring–summer. We obtained the rainfall data by subtracting the snowfall from the total precipitation in the ERA-Interim data.

To validate the performance of the ERA-Interim data, we compared the station value with the nearest reanalysis grid that covers the station site. The reanalysis–observation comparison of the mean value and standard deviation were conducted based on the two-tailed t-test and F-test, respectively. The trends of the spring–summer mean temperature, rainfall, total precipitation and RPR were determined based on least-squares linear regression. The trends were tested using the ordinary Student's t-test. A confidence level above 95% was deemed statistically significant.

3. Results

3.1. *Performance of ERA-Interim*

To validate the performance of the ERA-Interim data, we compared the spring–summer mean temperature, rainfall, total precipitation and RPR between observations and ERA-Interim data from 1979–2007. Analysis of the mean value, trend and standard deviation of the data was carried out to estimate whether the station observations could be simulated well by the ERA-Interim data. The results showed that ERA-Interim can reasonably simulate the annual variation of the spring–summer mean temperature. The observed mean value, trend and standard deviation of the spring–summer mean temperature were comparable with ERA-Interim. For the mean value, except for the MAYO A, RESOLUTE CARS, CAMBRIDGE BAY A and HAY RIVER A stations, the spring–summer mean temperature showed a small deviation between the observational data and ERA-Interim (Table 1, Fig. S1 in electronic supplementary material). The spring–summer mean temperature showed an increasing trend at 11 stations, despite a discrepancy in the slope value (Fig. S1). The increasing trends were significant at the EUREKA A, RESOLUTE CARS and CORAL HARBOUR A stations in the observational data and significant at the EUREKA A, CAMBRIDGE BAY A, RESOLUTE CARS, CORAL HARBOUR A, and BAKER A stations in ERA-Interim. The changes in standard deviation were generally consistent at 11 stations (Fig. S1). The correlation coefficients of the spring–summer mean temperature ranged from 0.72–0.98 between ERA-Interim and observations (Table 1). The correlations were significant at the 99% confidence level. Figure 1a shows the trends in the spring–summer mean temperature during 1979–2007, and the observations used to validate ERA-Interim. We can see that the trend in the spring–summer mean temperature at each station was consistent with the surrounding reanalysis data. The mean temperature increased significantly in northern Canada during 1979–2007, except in Yukon Territory, which is located in the Rocky Mountains. The temperature in this region showed a discrepancy between the observational data and ERA-Interim, as did the MAYO A station. The impact of topography may be the main reason for this discrepancy.

The amounts of total precipitation and rainfall were overestimated in ERA-Interim (Table 1). This is mainly because the estimates of precipitation are produced by the forecast model, based on temperature and humidity information derived from the assimilated observations. Approximations used in the model's representation of moist processes strongly affect the quality and consistency of the hydrological cycle. Though imperfect, the overestimated precipitation in ERA-Interim is less pronounced than in ERA-40 (Dee et al., 2011). These data are widely used in Arctic research (Screen and Simmonds, 2012; Cohen et al., 2015).

We focused in this study on changes in the proportion of precipitation occurring as rain. The reanalysis–observation comparison showed that the mean value of the RPR is reproduced well by ERA-Interim, except at the HAY RIVER A, EUREKA A and CORAL HARBOUR A stations (Table 1, Fig. S2), and the trend of RPR in ERA-Interim is comparable with the observations, except at the HAY RIVER A and EUREKA A stations (Fig. S2). There was a significant increasing trend in RPR at RESOLUTE CARS station in ERA-Interim, while other stations showed no significant trend. The changes in standard deviation were smaller in ERA-Interim than in the observational data, except at YELLOWKNIFE A station (Fig. S2). The correlation coefficient of RPR be-

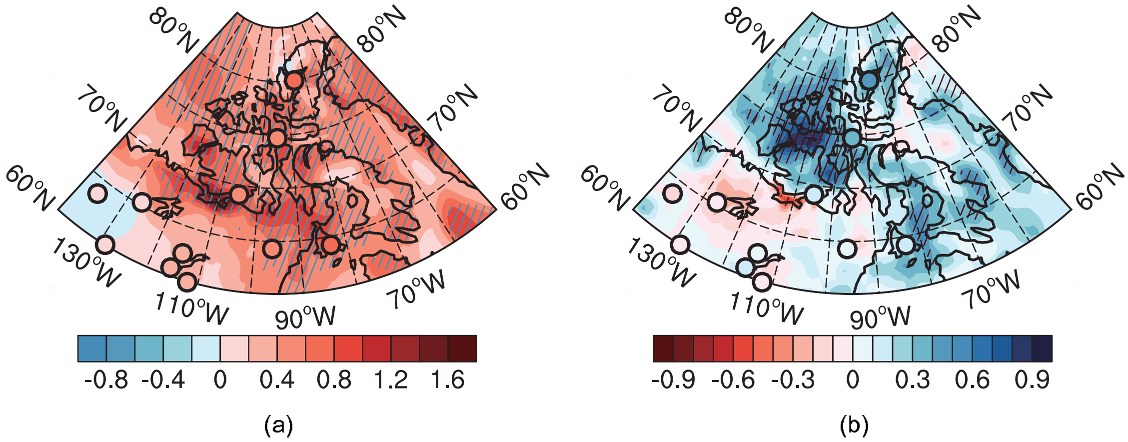

Fig. 1. Trend of (a) mean temperature [units: °C (10 yr)$^{-1}$] and (b) RPR [units: % (10 yr)$^{-1}$] in ERA-Interim during spring–summer, 1979–2007. The color of dots denotes the value of the trend from the Canadian meteorological stations. Dashed shading indicates statistical significance at the greater than 95% confidence level.

tween ERA-Interim and observations was calculated to be about 0.51–0.89, all values of which were significant at the 99% confidence level (Table 1). ERA-Interim agreed well with the station observations in terms of the changes in the RPR. The trends of RPR at the 11 stations were consistent with the surrounding reanalysis data (Fig. 1b). The spring–summer RPR increased significantly in the Canadian Arctic Archipelago from 1979–2007. In general, ERA-Interim can be used to analyze the changes in spring–summer temperature and the RPR in northern Canada.

3.2. *Changes in surface air temperature and RPR*

When it comes to the reasons for the changes in precipitation forms, the changes in surface air temperature are considered first, and we find that the RPR has a good connection with the surface air temperature (Fig. 2). As the temperature rises, the RPR increases obviously. Below, we analyze both the changes in surface air temperature and RPR over northern Canada.

Analysis based on ERA-Interim showed that the spring–summer mean temperature is about −20°C to 7.5°C in northern Canada (Fig. 3a). The spring–summer mean temperature

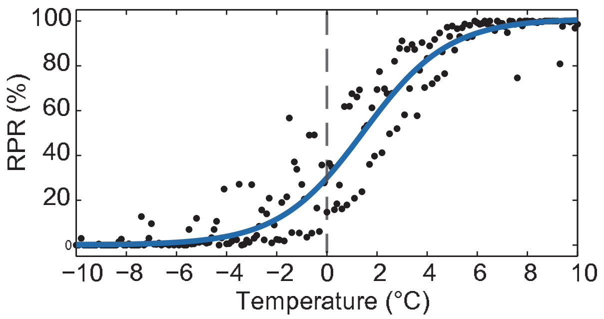

Fig. 2. RPR as a function of mean surface air temperature at 11 stations during spring–summer, 1979–2007. Dashed line means the proportion of precipitation occurring as rain when the surface air temperature is 0°C.

increased significantly over most of northern Canada from 1979 to 2015, in addition to some areas in Yukon Territory (Fig. 3b). An increase of 0.4°C–1°C (10 yr)$^{-1}$ is apparent in the Canadian Arctic Archipelago, which is concurrent with the results of previous studies showing that the temperature has increased rapidly in these areas, based both on observational data (Przybylak, 2007) and other reanalysis data (Overland et al., 2014).

The analysis based on ERA-Interim showed that the RPR ranges from 25% to 85% from northeast to southwest (Fig. 3c). The spatial distribution of the RPR is closely related to the distribution of surface air temperature. Against the background of warming, the RPR also increases over most of northern Canada, except near the Mackenzie Mountains, Bake Lake, and the north coast of Ellesmere Island (Fig. 3d, Fig. S3). We find that the total precipitation trend is not significant, but the rainfall increases significantly over Baffin Island, Banks Island, M'Clure Strait, and other islands (Figs. S3 and S4). The change in rainfall is greater than that of total precipitation over most of the Canadian Arctic Archipelago. The RPR increases by 2%–6% (10 yr)$^{-1}$ (significant at the 95% confidence level) in the Canadian Arctic Archipelago, except near Queen Elizabeth Island and Baffin Bay.

The spring–summer mean temperature shows a significant positive correlation with the RPR in northern Canada (60°–70°N, 70°–140°W), and a significant negative correlation in the area near Queen Elizabeth Islands (Fig. 3e). There is a significant regional difference in the relationship between air temperature and the RPR in the Canadian Arctic Archipelago, due to the complicated impacts of topography.

3.3. *Variability associated with the NAO*

Temperature has important impacts on the form of precipitation, with increasing temperatures causing the precipitation form to change from snow to rain earlier (Knowles et al., 2006; Ye, 2008). Against the background of warming in the Arctic, the rising temperatures have a greater impact on changing the form of precipitation (rain/snow) than altering

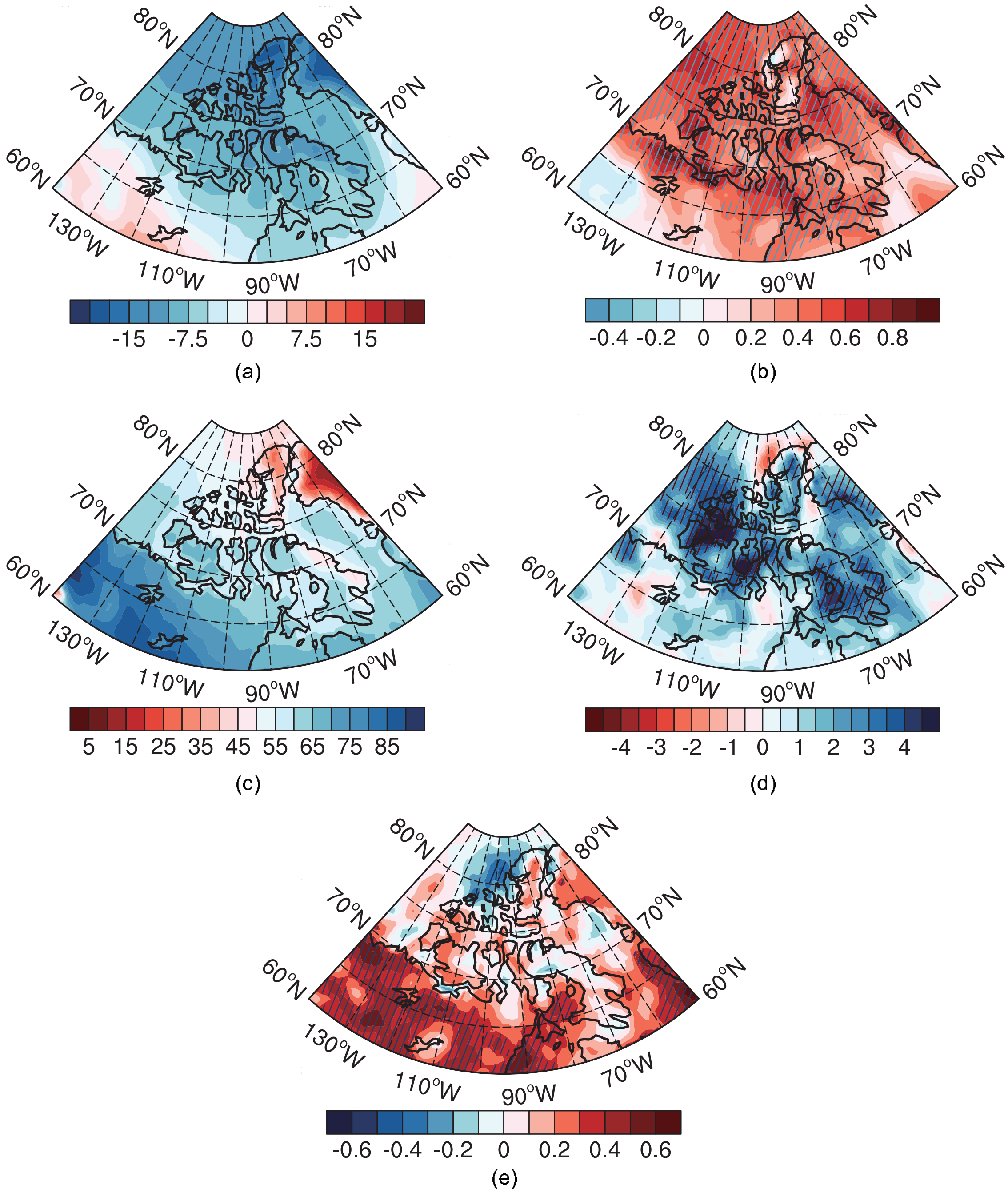

Fig. 3. Climatology of (a) mean temperature (units: °C) and (c) RPR (units: %), along with trends of (b) mean temperature [units: °C $(10\ \mathrm{yr})^{-1}$] and (d) RPR [units: % $(10\ \mathrm{yr})^{-1}$], and (e) the correlation between mean temperature and RPR, in northern Canada during spring–summer, 1979–2015. Dashed shading indicates statistical significance at the greater than 95% confidence level.

the total precipitation amount (Screen and Simmonds, 2012).

Further analysis indicated a consistent temperature increase over northern Canada, while there are regional differences for the RPR. Aside from the impact of topography, precipitation can also be influenced by atmospheric circulation and moisture supply. Thus, teleconnection patterns may also contribute to changes in the RPR in spring–summer. The AO and NAO are the most important atmospheric teleconnection patterns in the Arctic. Based on correlation analysis of both the AO and NAO indices with the RPR in northern Canada over the past few decades, the NAO shows identifiable regional signatures in the changes in the RPR.

The NAO is characterized by a north–south dipole, with its centers located in the area of the Icelandic low and the Azores high, respectively (Wallace and Gutzler, 1981; Barnston and Livezey, 1987). During positive (negative) phases of the NAO, the Azores high is strengthened (weakened) and the Icelandic low is deepened (shallowed). The meridional circulation is strong during positive NAO. Dry tropical conditions are much more common with positive NAO throughout much of North America in spring (Sheridan, 2003). Figure 4a shows the NAO dominates the variations in RPR over northern Canada. Negative NAO resulted in a larger RPR in northern Canada from 1979–2015, and was significant in most areas of northern Canada (−0.6 to −0.3), except the region near Baffin Bay (Fig. 4a). We find that the NAO has different effects on rainfall and snowfall (Figs. 4b and c). Positive (negative) NAO results in less (more) rainfall and a slight change in total precipitation in south Nunavut, overall leading to a small (large) RPR. In the Northwest Territories, meanwhile, positive (negative) NAO results in more (less) snowfall and a slight change in total precipitation, also leading to a small (large) RPR. However, the NAO has little influence on the rainfall and total precipitation in areas near Baffin Bay. In general, the changes in RPR in northern Canada may include the effects of both warming and the NAO.

4. Discussion and conclusions

Few studies of different precipitation forms have been conducted in Arctic regions. In view of the important effects of the local hydrology and ecology in the Arctic climate system, we analyzed the changes in the spring–summer RPR over northern Canada and discussed the potential causes.

The precipitation form is determined by the vertical temperature of the atmosphere, particle size distribution and the microphysics scheme. Among these factors, the temperature is dominant (Sankaré and Thériault, 2016). In this study, we focused on the changes in the RPR in northern Canada and its link with surface air temperature changes. ERA-Interim agrees well, qualitatively, with observations of the spring–summer mean temperature and RPR. Evaluation of the ERA-Interim data indicated that the spring–summer mean temperature increased significantly [0.4°C–1°C (10 yr)$^{-1}$], as did the rainfall and the RPR [2%–6% (10 yr)$^{-1}$], in the Canadian Arctic Archipelago from 1979 to 2015.

We also paid attention to the regional differences in the RPR. Correlation analysis of both the AO and NAO indices with the RPR in northern Canada over the past few decades indicated the NAO plays a dominant role in the variations of RPR over northern Canada. Positive (negative) NAO resulted in a small (larger) RPR in northern Canada from 1979–2015. Specifically, positive (negative) NAO resulted in less (more) rainfall in south Nunavut, but more (less) snowfall in the Northwest Territories, all leading to a small (large) RPR. Therefore, aside from the contribution of climate warming, the NAO is probably another key factor resulting in the temporal and spatial variations in the RPR over northern Canada.

The results presented herein improve our understanding of climate change over the Canadian Arctic and the potential impacts of precipitation phase changes on the cryosphere. It

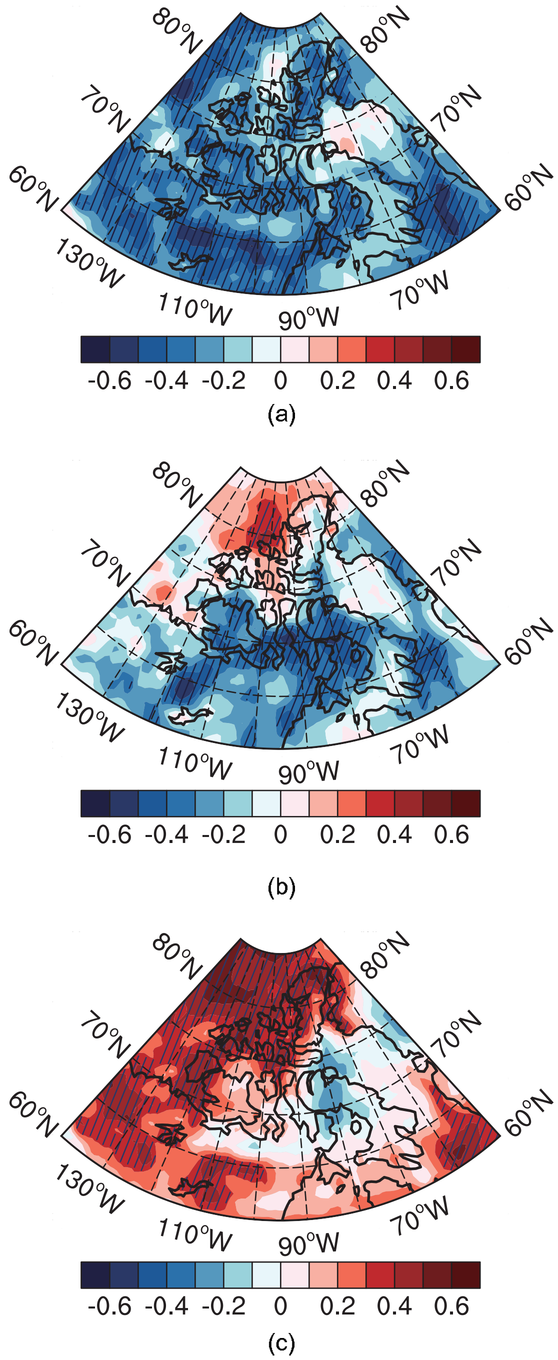

Fig. 4. Correlation of (a) RPR, (b) rainfall, and (c) snowfall with the NAO during spring–summer, 1979–2015. Dashed shading indicates statistically significant correlations at the greater than 95% confidence level.

is, however, important to note that the temporal coverage of the ERA-Interim data employed here is not long enough to capture the interdecadal signal of the NAO. The changes in the RPR in northern Canada may include the effects of both warming and the NAO. Further studies are needed to better understand the ecological environment in the Arctic climate system.

Acknowledgements. This study was supported by the National Key Basic Research Program of China (Grant No. 2013CBA01804), the National Science Foundation of China (Grant Nos. 41425003 and 41401079), the State Oceanic Administration of the People's Republic of China Project on Climate in Polar Regions (Grant No. CHINARE2016-2020), the Key Research Program of the Chinese Academy of Sciences (Grant No. KJZD-EW-G03), and the Opening Founding of the State Key Laboratory of Cryospheric Sciences (Grant No. SKLCS-OP-2016-03). The authors would like to express their appreciation to Julian Morales for providing the daily meteorological data from Environment and Climate Change Canada.

Electronic supplementary material Supplementary material is available in the online version of this article at https://doi.org/10.1007/s00376-018-7226-3.

REFERENCES

Aanes, R., B. E. Sæther, and N. A. Øritsland, 2000: Fluctuations of an introduced population of Svalbard reindeer: The effects of density dependence and climatic variation. *Ecography*, **23**, 437–443, https://doi.org/10.1111/j.1600-0587.2000.tb00300.x.

Barnston, A. G., and R. E. Livezey, 1987: Classification, seasonality, and persistence of low-frequency atmospheric circulation patterns. *Mon. Wea. Rev.*, **115**, 1083–1126, https://doi.org/10.1175/1520-0493(1987)115<1083:CSAPOL>2.0.CO;2.

Bekryaev, R. V., I. V. Polyakov, and V. A. Alexeev, 2010: Role of polar amplification in long-term surface air temperature variations and modern Arctic warming. *J. Climate*, **23**, 3888–3906, https://doi.org/10.1175/2010JCLI3297.1.

Boisvert, L. N., and J. C. Stroeve, 2015: The Arctic is becoming warmer and wetter as revealed by the Atmospheric Infrared Sounder. *Geophys. Res. Lett.*, **42**, 4439–4446, https://doi.org/10.1002/2015GL063775.

Cavalieri, D. J., and C. L. Parkinson, 2012: Arctic sea ice variability and trends, 1979-2010. *The Cryosphere*, **6**, 881–889, https://doi.org/10.5194/tc-6-881-2012.

Cohen, J., H. C. Ye, and J. Jones, 2015: Trends and variability in rain-on-snow events. *Geophys. Res. Lett.*, **42**, 7115–7122, https://doi.org/10.1002/2015GL065320.

Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, https://doi.org/10.1002/qj.828.

Ding, B. H., K. Yang, J. Qin, L. Wang, Y. Y. Chen, and X. B. He, 2014: The dependence of precipitation types on surface elevation and meteorological conditions and its parameterization. *J. Hydrol.*, **513**, 154–163, https://doi.org/10.1016/j.jhydrol.2014.03.038.

Groisman, P. Y., R. W. Knight, D. R. Easterling, T. R. Karl, G. C. Hegerl, and V. N. Razuvaev, 2005: Trends in intense precipitation in the climate record. *J. Climate*, **18**, 1326–1350, https://doi.org/10.1175/JCLI3339.1.

Hasnain, S. I., 2002: Himalayan glaciers meltdown: Impact on south Asian rivers. *Proc. Fourth International FRIEND Conf. Held at Cape Town*, South Africa, IAHS, **274**, 417–423.

Irannezhad, M., H. Marttila, D. L. Chen, and B. Kløve, 2016: Century-long variability and trends in daily precipitation characteristics at three Finnish stations. *Advances in Climate Change Research*, **7**, 54–69, https://doi.org/10.1016/j.accre.2016.04.004.

Knowles, N., M. D. Dettinger, and D. R. Cayan, 2006: Trends in snowfall versus rainfall in the western United States. *J. Climate*, **19**, 4545–4559, https://doi.org/10.1175/JCLI3850.1.

Loth, B., H.-F. Graf, and J. M. Oberhuber, 1993: Snow cover model for global climate simulations. *J. Geophys. Res.*, **98**, 10 451–10 464, https://doi.org/10.1029/93JD00324.

Miller, G. H., R. B. Alley, J. Brigham-Grette, J. J. Fitzpatrick, L. Polyak, M. C. Serreze, and J. W. C. White, 2010: Arctic amplification: Can the past constrain the future? *Quaternary Science Reviews*, **29**, 1779–1790, https://doi.org/10.1016/j.quascirev.2010.02.008.

Overland, J. E., M. Y. Wang, J. E. Walsh, and J. C. Stroeve, 2014: Future arctic climate changes: Adaptation and mitigation time scales. *Earth's Future*, **2**, 68–74, https://doi.org/10.1002/2013EF000162.

Perovich, D., C. Polashenski, A. Arntsen, and C. Stwertka, 2017: Anatomy of a late spring snowfall on sea ice. *Geophys. Res. Lett.*, **44**, 2802–2809, https://doi.org/10.1002/2016GL071470.

Perovich, D. K., and C. Polashenski, 2012: Albedo evolution of seasonal Arctic sea ice. *Geophys. Res. Lett.*, **39**, L08501, https://doi.org/10.1029/2012GL051432.

Pithan, F., and T. Mauritsen, 2014: Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nature Geoscience*, **7**, 181–184, https://doi.org/10.1038/NGEO2071.

Przybylak, R., 2007: Recent air-temperature changes in the Arctic. *Annals of Glaciology*, **46**, 316–324, https://doi.org/10.3189/172756407782871666.

Putkonen, J., and G. Roe, 2003: Rain-on-snow events impact soil temperatures and affect ungulate survival. *Geophys. Res. Lett.*, **30**, 1188, https://doi.org/10.1029/2002GL016326.

Sankaré, H., and J. M. Thériault, 2016: On the relationship between the snowflake type aloft and the surface precipitation types at temperatures near 0°C. *Atmos. Res.*, **180**, 287–296, https://doi.org/10.1016/j.atmosres.2016.06.003.

Screen, J. A., and I. Simmonds, 2010: The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature*, **464**, 1334–1337, https://doi.org/10.1038/nature09051.

Screen, J. A., and I. Simmonds, 2012: Declining summer snowfall in the Arctic: Causes, impacts and feedbacks. *Climate Dyn.*, **38**, 2243–2256, https://doi.org/10.1007/s00382-011-1105-2.

Serreze, M. C., A. P. Barrett, J. C. Stroeve, D. N. Kindig, and M. M. Holland, 2009: The emergence of surface-based Arctic amplification. *The Cryosphere*, **3**, 11–19, https://doi.org/10.5194/tc-3-11-2009.

Serreze, M. C., and Coauthors, 2000: Observational evidence of recent change in the northern high-latitude environment. *Climatic Change*, **46**, 159–207, https://doi.org/10.1023/A:1005504031923.

Sheridan, S. C., 2003: North American weather-type frequency and teleconnection indices. *Int. J. Climatol.*, **23**, 27–45, https://doi.org/10.1002/joc.863.

Stirling, I., and T. G. Smith, 2004: Implications of warm temperatures and an unusual rain event for the survival of ringed seals on the coast of southeastern Baffin Island. *Arctic*, **57**, 59–67, https://doi.org/10.14430/arctic483.

Wallace, J. M. and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the northern hemisphere winter. *Mon. Wea. Rev.*, **109**, 784–812, https://doi.org/10.1175/1520-0493(1981)109<0784:TITGHF>2.0.CO;2.

Yao, T. D., and Coauthors, 2012: Different glacier status with atmospheric circulations in Tibetan Plateau and surroundings. *Nature Climate Change*, **2**, 663–667, https://doi.org/10.1038/nclimate1580.

Ye, H. C., 2008: Changes in frequency of precipitation types associated with surface air temperature over northern Eurasia during 1936–90. *J. Climate*, **21**, 5807–5819, https://doi.org/10.1175/2008JCLI2181.1.

Ye, H. C., and J. Cohen, 2013: A shorter snowfall season associated with higher air temperatures over northern Eurasia. *Environ. Res. Lett.*, **8**, 014052, https://doi.org/10.1088/1748-9326/8/1/014052.

Zhang, X. B., W. D. Hogg, and É. Mekis, 2001: Spatial and temporal characteristics of heavy precipitation events over Canada. *J. Climate*, **14**, 1923–1936, https://doi.org/10.1175/1520-0442(2001)014<1923:SATCOH>2.0.CO;2.