Interannual Variability and Long-Term Changes of Atmospheric Circulation over the Chukchi and Beaufort Seas

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ABSTRACT

The Beaufort Sea high (BSH) plays an important role in forcing Arctic sea ice and the Beaufort Gyre. This study examines the variability and long-term trends of atmospheric circulation over the Chukchi and Beaufort Seas using the ECMWF Interim Re-Analysis (ERA-Interim) for the period 1979–2012. Because of the mobility of the BSH through the year, EOF analysis is applied to the sea level pressure (SLP) field in order to investigate the principal patterns of BSH variability. In each season, the three leading EOF modes explain nearly 90% of the total variance and reflect a strengthened or weakened BSH centered over the western Arctic Ocean (EOF1), a north–south dipole-like SLP anomaly (EOF2), and a west–east dipole-like SLP anomaly (EOF3), respectively. These three EOF modes offer distinct influences on local climate in each season and have different connections with the large-scale climate variability modes in winter. In particular, the second principal component (PC2) associated with EOF2 in the autumn exhibits a tendency toward high-index polarity significant at the 5% level, and is related to strongly reduced sea ice extent.

Further, the authors have detected significant anticyclonic trends among surface wind fields associated with a strengthened BSH during summer and autumn, but significant cyclonic trends associated with a weakened BSH during early midwinter, consistent with significant trends in SLP gradients between western Arctic Ocean and the adjoining landmass. Comparison with forced trends of surface winds from various simulations from the IPCC Fifth Assessment Report (AR5) indicates that summertime changes in atmospheric circulation cannot be explained by natural external forcing or lower boundary forcings and may instead be attributable to external anthropogenic forcing.

1. Introduction

The Beaufort Sea high (BSH) represents both a high pressure center/ridge over the Beaufort Sea during winter, and a relatively weak area of high pressure that covers most of the Beaufort Sea during summer, when it tends to center north of Alaska (Overland 2009; Serreze and Barrett 2011). Serreze and Barrett (2011) define the BSH strength index as the frequency of anticyclonic relative vorticity and, based on this vorticity index, find that a strong BSH in summer is most closely related with the negative phase of the northern annular mode (NAM), although sea level pressure (SLP) anomaly patterns of strong wintertime BSH have little correlation to large-scale natural variability modes, such as the NAM, the
Arctic dipole anomaly, the Pacific–North American (PNA) teleconnection pattern, and the Pacific decadal oscillation (PDO). Gleicher et al. (2011) use a vorticity-based index to investigate the spatial and temporal characteristics of the Beaufort anticyclone. Moore (2012) shows that SLP at the center of the BSH varies on a decadal time scale based on intensity and location of the summer BSH; since the late 1990s, there has been a trend toward a stronger summer BSH. Stegall and Zhang (2012) have found that, along with the changed atmospheric circulation over the Chukchi and Beaufort Seas, there have been increasing trends in areal-averaged monthly mean and 95th percentile wind speeds from July through November. There are also many papers on wind regimes along the Beaufort coast from station observations (e.g., Manson and Solomon 2007; Small et al. 2011).

The primary purpose of this study is to further investigate temporal variability and long-term changes in basin-scale atmospheric circulation over the Chukchi and Beaufort Seas. At the 1-hPa contour interval, a closed BSH appears regionally for all seasons except for winter. The local maximum of the winter and summer BSH are centered over the Chukchi Sea and Beaufort Sea, respectively, with spring and autumn representing transitions between winter and summer conditions (Fig. 1). This movement of the BSH suggests that simply using the frequency of anticyclonic relative vorticity as an index of the strength of the BSH according to Serreze and Barrett (2011) or Gleicher et al. (2011) may not fully capture the dominant variability modes of the BSH throughout the year. An empirical orthogonal function (EOF) analysis of SLP over the Chukchi and Beaufort Seas and northern Alaska is thus performed to investigate the primary features of the BSH variability. In addition, motivated by the results of Moore (2012) and Stegall and Zhang (2012), we examine whether significant changes in atmospheric circulation have occurred in the summer, autumn, and winter seasons over the Chukchi and Beaufort Seas from 1979 onward, and we thus study the role that the atmospheric circulation has played in driving the observed sea ice decline.

Finally, we compare the observed changes in atmospheric circulation with those of simulations generated for the Fifth Assessment Report of the United Nations Intergovernmental Panel on Climate Change (IPCC AR5) in order to assess whether long-term changes in atmospheric circulation over the Chukchi and Beaufort Seas are attributable to either anthropogenic or natural external forcing, or boundary sea surface temperature (SST) and sea ice forcing. The forced trends of SLP and surface winds are calculated from the IPCC AR5 simulations forced by observed atmospheric composition changes (reflecting both anthropogenic and natural

**Fig. 1.** Seasonally averaged SLP and surface winds (m s$^{-1}$, vectors) over the Chukchi and Beaufort Seas for the period 1979–2012, from the ERA-Interim. Contour intervals are 1 hPa for SLP.

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sources, and termed Historical), as well as in those forced by greenhouse gas only (HistoricalGHG), by natural forcing only (HistoricalNAT), and from both observed SST and sea ice and atmospheric radiative forcing [Atmospheric Model Intercomparison Project (AMIP)] (Taylor et al. 2012). Note that in this study we are interested in the basin-scale variability in the Beaufort–Chukchi Seas, and this is distinguished from Small et al. (2011) and others who examine the patterns of winds along the Beaufort coast from the station observations, in which the local topography plays an important role.

The remainder of this paper is organized as follows: section 2 describes the data sources and analysis methods used, the results of which are presented in section 3, and a summary is given in section 4.

2. Data and analysis methods

The atmospheric datasets used in this study are monthly geopotential height, SLP, and near-surface wind and temperature on a 1.5° latitude–longitude grid taken from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim, hereafter ERA-I; Dee et al. 2011). As in Rigor et al. (2002), the seasons are defined in this study as the cold months of winter [November–March (NDJFM)], the early winter [November–January (NDJ)], the warm summer months [June–August (JJA)] when surface air temperature over the Arctic Ocean tends to be uniform and close to the freezing point, and the transitional seasons of spring [April–May (AM)] and fall [September–October (SO)]. EOF, regression, and linear-trend analyses are performed for each season. Before conducting these analyses, the long-term mean seasonal cycle was subtracted from all atmospheric data. In the EOF analysis, area weighting is accomplished by multiplying each field by the square root of the cosine of latitude before computing the covariance matrix. The statistical significance of regression coefficients, correlation coefficients, and the linear trends of the normalized principal component (PC) indices in the analyses are assessed using
the $t$ statistic. Because of autocorrelations in the atmospheric fields, the effective sample sizes in the $t$ tests are estimated using the relationship outlined in Bretherton et al. (1999):

$$N_{\text{eff}} = N \left( \frac{1 - r_1 r_2}{1 + r_1 r_2} \right),$$  

where $N_{\text{eff}}$ is the effective sample size, $N$ is the sample size, and $r_1$ and $r_2$ are the lag-1 autocorrelations of the time series being correlated. Note that the lag-1 autocorrelations of the PC indices and other circulation indices are on the order of 0.1–0.2; the effective sample size is about 90% of the sample size in our study.

Interannual variability of the BSH is studied using EOF analysis, within which dominant patterns of the SLP field and their associated PC time series over the Chukchi and Beaufort Seas and northern Alaska (64.5°–87°N, 160°E–110°W), as given in the ERA-I, are determined. Regressions of the three leading normalized PCs (the PC indices) of SLP in the above domain are used to characterize the leading BSH variability modes. Impacts of the leading BSH variability modes for surface winds over the Beaufort Sea and northern Alaska are investigated using data from ERA-I, along with monthly sea ice concentration (SIC) data (Comiso and Nishio 2008). Monthly SIC is in units of percent on
a 25-km-resolution grid, and is taken from the newly updated homogeneous monthly high-resolution data derived from passive microwave satellite data with the bootstrap algorithm for the period 1979–2011.

Connections between the leading BSH variability modes and large-scale atmospheric variability modes (NAM and PNA) are examined. As in Serreze and Barrett (2011), we use the index values for NAM defined by Ogi et al. (2004) (downloaded from http://wwwoa.ees.hokudai.ac.jp/people/yamazaki/SV-NAM/index.html), and for PNA prepared by the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) (downloaded from http://www.cpc.ncep.noaa.gov/). Note that Ogi et al. (2004) defined the NAM separately for each calendar month through an EOF analysis of National Centers for Environmental Prediction (NCEP) geopotential height fields from 1000 to 200 hPa, poleward of 40°N. This contrasts with the NAM index from CPC, which is based on a single EOF analysis of geopotential height fields for all calendar months (Thompson and Wallace 1998). The NAM index from the CPC is most strongly expressed in winter, while the NAM as defined by Ogi et al. (2004) is strong throughout the year. Also note that the procedure used to calculate PNA teleconnection indices by the CPC is based on the rotated principal component analysis (RPCA) used by Barnston and Livezey (1987), and the RPCA technique is applied to 500-hPa geopotential height anomalies for all calendar months. The PNA index is thus most strongly expressed in winter, but also observed in other seasons.
At each season, the linear trends of three leading PC indices and their significance are examined, and the contribution of long-term change in each PC index to sea ice is investigated. At each season, the observed linear trends of SIC over the western Arctic are first calculated throughout the period 1979–2012. The component of the trends that is linearly congruent with each monthly PC index is then estimated at each grid point by using the method of Wu and Straus (2004): 1) regressing monthly SIC values of that grid point’s time series onto the PC index simultaneously (at lag 0), and then 2) multiplying the resulting regression coefficient by the linear trend in the PC index.

The significance of the 33-yr linear trends of the monthly atmospheric fields is assessed by examining whether the trend in each grid box can be explained by natural variability at the 5% significant level. We estimate the natural variability of the linear trends for the atmospheric fields from the preindustrial control simulations generated for the IPCC AR5 and determine whether the observed trend in each grid box can be explained by this natural variability. A total of 3000 yr of preindustrial control simulations from six AR5 models listed in Table 1 (CNRM-CM5, GFDL CM3, GISS-E2-H, HadCM3, MPI-ESM-LR, and MPI-ESM-MR; expansions of these model names are given in appendix B) are used to estimate the natural variability of the 33-yr trends. These models are chosen for the fact that the seasonal climatology of the BSH (see Fig. 1) is reasonably captured in their preindustrial control and historical simulations (see appendix A). Data from these models are originally available on grid projections different from the ERA-I and are thus interpolated into the ERA-I grid prior to analysis.

The climatic drifts of the control simulations in each grid box are removed separately for each model. To increase the sample size, we calculate the natural variability of the 33-yr linear trends for the season JJA from overlapping segments in the control simulations beginning 17 yr apart. The result is about 180 realizations of estimated linear trends for the JJA season for the 33-yr period. Within each grid box, we count the percentage of realizations whose amplitudes of trends for JJA are larger than the corresponding observed trend. If such a percentage in a grid box is less than 5% (10%), then we reject the null hypothesis that the observed summer JJA trend can be explained by natural variability at the 5% (10%) significance level. This same method is applied for the trends during SO and other seasons.

The variances of the detrended observed and control model winds and SLP are calculated at each of the grid boxes. A one-tailed $F$ test (Snedecor and Cochrans 1989) is used to determine whether the amplitude of the modeled variability is significantly larger or smaller than observed at each grid box at the 5% significant level. It is found that the control simulations in the above six modes generally have similar spatial distributions of variability as the observations (not shown). Each model simulates comparable or substantially larger variability than observed over most grid boxes. The spatially averaged variances of SLP in the Historical simulations during 1973–2005 from six models are also compared with that in the ERA-I during 1979–2011 for four seasons (see Table A1; the table indicates that all models simulate comparable or larger total variance than observed for each season). Therefore, it is believed that the above significance tests here do not overestimate the significance the observed trends.

To determine the possible roles that external anthropogenic and natural radiative forcings, and boundary forcing play in the long-term atmospheric circulation, the trends of surface winds are examined from the IPCC AR5 Historical, HistoricalGHG, and HistoricalNAT simulations for 1979–2005, and from AMIP simulation prescribed for 1979–2008 from the these models listed in Table 1. We have data from these Historical, HistoricalGHG, HistoricalNAT, and AMIP runs for 36, 14, 14, and 10 different multimodel ensemble members, respectively (Table 1). As these forced simulations conclude in 2005 in Historical, HistoricalGHG, and HistoricalNAT simulations, in 2008 in AMIP, only the 26-yr trends for the period 1979–2005 or the 29-yr trends for the period 1979–2008 for the surface winds are calculated in these experiments. The cumulative significance of these 26-yr (29-yr) linear trends of monthly surface wind fields in each grid box over the western Arctic Ocean is assessed using the same method as for the observations.

3. Results

a. Variability modes of SLP and their relationships with large-scale circulation modes

Figure 1 shows that the center of the BSH moves throughout the year. The spatial distribution of the monthly mean surface wind is generally anticyclonic under the influence of the monthly SLP pattern. With changes in the strength and location of the BSH, the surface wind field surrounding the BSH varies from season to season in terms of its speed and direction. To examine the interannual variability of the above mobile BSH, an EOF analysis is performed to determine the dominant patterns and their associated PC time series in the SLP field over the Chukchi and Beaufort Seas and to the north of Alaska (64.5°–87°N, 160°E–110°W) for each season. As shown in Table 2, the leading four EOF SLP
modes explain 57.8%, 21.0%, 11.3%, and 3.1% of the total variance in spring; 61.3%, 17.3%, 10.5%, and 3.7% in summer; 49.8%, 24.9%, 12.9%, and 5.0% in fall; 66.1%, 16.7%, 9.9%, and 2.3% in early winter (NDJ); and 60.7%, 20.1%, 11.2%, and 2.3% in the full winter (NDJFM). The first three leading EOF modes for all seasons (Fig. 2) account for more than 90% of their corresponding total variances. Applying the guidelines from North et al. (1982) here, we find that all first, second, and third EOFs are adequately separated both from one another and from the fourth EOF mode in each season. The EOFs for the full 5-month winter (NDJFM) are very similar to those for the 3-month early winter (NDJ), and thus only EOFs for the early winter are shown.

The first leading EOFs (Fig. 2, top) show positive anomalies centered over the central Arctic Ocean, primarily reflecting a strengthened or weakened BSH for all four seasons. Table 2 shows that there exist statistically significant correlations between the first PC (PC1, associated with EOF1) and the NAM index values for all five seasons (from −0.50 in NDJFM to −0.77 in JJA), indicating that a strengthened BSH is closely allied with the negative phase of the NAM through the year. In summertime, PC1 is also significantly correlated with the PNA index ($r_{0.36}$). The above summertime result is consistent with Serreze and Barrett (2011), who find that a strong BSH based on the vorticity index has combined features of the negative phase of the NAM mainly, and the positive phase of the PNA, to a weaker extent.

The second EOFs (Fig. 2, middle) show north–south dipole-like SLP anomalies in the Chukchi and Beaufort Seas and to the north of Alaska, characterizing the northern or southern shift of the BSH for all four seasons. The positive northern BSH node is located over the Arctic during all seasons. Strong and significant correlation between PC2 associated with Fig. 2h (Fig. 2k) and the PNA index in SO (NDJ) is about 0.43 (0.59), indicating a stronger-than-normal Aleutian low in the North Pacific, associated with the positive phase of the PNA, tends to cause a positive departure in SLP over the northern Beaufort Sea.

The patterns of the third leading EOFs (Fig. 2, bottom) for all four seasons are associated with the west–east dipole-like variability of the BSH, with a positive SLP anomaly for the Beaufort Sea and a negative SLP anomaly for the Chukchi and East Siberian Seas. PC3, associated with the west–east dipole-like variability mode, is not significantly correlated with the NAM and PNA time series for all five seasons—implying that the regional mode of the BSH differs from the large-scale circulation patterns.

At the 5% significance level, only PC2 indices associated with EOF2 during SO and NDJ have exhibited a significant tendency toward high-index polarity with linear trends of 0.35 and −0.23 decade$^{-1}$, respectively (Fig. 3). At the 10% level, the linear trends of the PC2 index during JJA and the PC3 index for NDJ are also found to be statistically significant (not shown). Other PCs for these four seasons, and three PCs in NDJFM (not shown), did not exhibit significance at the 10% level.

**b. Impacts of the BSH’s variability modes on local climate changes**

Simultaneous regression (lag 0) maps of surface wind fields on the corresponding normalized PC indices associated with the leading three EOFs (Fig. 2) are shown in Fig. 4 for each season, with their significance determined by the Student’s $t$ tests. Associated with the strengthened BSH in Fig. 2 (top), anticyclonic wind anomalies are primarily found over the ocean and northeasterly wind anomalies are found over Alaska for all seasons in Fig. 4 (top). Related to the EOF2-like SLP anomalies, easterly wind anomalies are prevalent over the ocean in Fig. 4 (bottom). Such anomalous meridional wind associated with the third leading EOFs blows from the northern North Pacific to the western Arctic and then to the eastern Arctic, favorable
to the Transpolar Drift Stream (Wu et al. 2006). The PC3 in AM has an insignificant linear trend (Table 2) but has been in extreme negative values during recent years, which are associated with persistent northerly winds in the Bering Sea that advect the ice southward to the melt edge. This may explain why the sea ice along the Bering Sea coast of Alaska has been more extensive in the spring in recent years (Matthewman and Magnusdottir 2011).

The atmospheric forcing on SIC associated with the BSH variability in Fig. 2 is also shown in Fig. 4. The most prominent feature here is that the strongly reduced western Arctic sea ice anomalies during SO share association with the north–south dipole-like SLP anomalies of the Chukchi–Beaufort Seas shown in Fig. 2h, implying anomalous easterly flow over the Beaufort and Chukchi Seas and Alaska (Fig. 4h), and a strong warm temperature anomaly over the Arctic Ocean (not shown).
During SO, there is a strong tendency toward high index polarity of PC2 indices, as shown in Fig. 3a. The extreme positive value of PC2 in autumn 2007 is clearly seen, with a standard deviation of 2.5, which directly contributed to the major reduction in September sea ice extent in that year. The same positive phase, but with a smaller magnitude, was also present in the years 1995, 1998, 2005, 2006, 2010, and 2011. The observed linear trends of SIC over the western Arctic that are linearly congruent with respect to the PC2 index have the same patterns as SIC in Fig. 4h, but with a different magnitude of 0.35 decade\(^{-1}\) (the linear trend of the PC2 index). Therefore, the significantly reduced autumn sea ice extent over the western Arctic Ocean during recent decades is most likely related to the long-term changes characterized by a northward-shifted and strengthened BSH associated with the EOF2 shown in Fig. 2h.

Another prominent feature is the summertime atmospheric forcing in Fig. 2d upon the Arctic SIC variability shown in Fig. 4d, reflecting that the sea ice anomalies are mainly caused by Ekman drift associated with summertime NAM-like anticyclonic anomalous surface winds in Fig. 4d, also seen in Ogi et al. (2008).
Recall that summertime PC1 and the NAM indices are significantly correlated ($r \approx -0.77$) (Table 2).

c. **Long-term change in atmospheric circulation in the ERA-I**

**Figure 5** shows trends in the surface wind field and SLP in the summer, fall, and early winter (NDJ) seasons. Figure 5 clearly indicates that there are large-scale, statistically significant, anticyclonic circulation trends in summer and autumn, and cyclonic gyre trends in early winter, indicating the strengthening of the BSH in summer and fall but weakening in early winter. The trend of the SLP field during JJA for the 33-yr period, shown in Fig. 5b, is similar to that for the 16-yr period 1996–2011 shown in Moore (2012), although a significant increase in the SLP (about 1.2 hPa decade $^{-1}$) is only found over the northeast Arctic Ocean and Queen Elizabeth Islands. The SLP field does not exhibit significant changes over most of the western Arctic Ocean basin in autumn, with an insignificant increase (about 0.4 hPa decade $^{-1}$) in the SLP over the northern Chukchi–Beaufort Seas, although a significant decrease ($-1.4$ hPa decade $^{-1}$) is seen over the southern Chukchi–Beaufort Seas, East Siberian Sea, and the adjoining landmass (Fig. 5d). During early winter, a large but insignificant SLP reduction is found over most of the central Arctic Ocean (with a maximum of about $-1.4$ hPa decade $^{-1}$), with an insignificant increase over the southern Chukchi–Siberian Arctic and the adjoining landmass (Fig. 5f).
As shown in Fig. 5, there are strong contrasts between large-scale, coherent, significant changes in surface winds and small-scale or no significant changes in SLP from summer to early winter (consistent with that few PCs associated with EOFs in Fig. 2 show significant liner trends). In each season, there are substantial trends in SLP gradients between the Arctic Ocean and the surrounding landmass. A trend toward a significant anticyclonic (cyclonic) circulation over the western Arctic in summer and autumn is in agreement with the positive (negative) SLP gradients between the ocean and the landmass. Significant anticyclonic wind trends in summer and autumn induce an anomalous Ekman drift of the sea ice toward the central Arctic and increase the areal coverage of open water over the marginal seas (Ogi et al. 2008), as well as an increase in advection of warm air toward the ocean in fall—both of which have acted to decrease the amount of ice in the Beaufort and Chukchi Seas, as seen in Fig. 4h.

d. Attribution of long-term change in atmospheric circulation

To explore the possible roles of external forcings on the observed long-term changes in the BSH in Fig. 5, the linear trends in monthly JJA, SO, and NDJ surface winds over the 26-yr period 1979–2005 in the multimodel ensemble-mean Historical, HistoricalGHG, and HistoricalNAT simulations produced by the AR5 models in Table 1 are shown in Fig. 6. During JJA, the anticyclonic pattern in the ERA-I (Fig. 5a) is reasonably reproduced in both Historical and HistoricalGHG experiments (Figs. 6a,d), while a cyclonic pattern is simulated in HistoricalNAT (Fig. 6g). In addition, the magnitudes of the anticyclonic wind trends in both the Historical and HistoricalGHG experiments are much lower than that in the ERA-I, indicating that the AR5 models in Table 1 may underestimate the response of the atmospheric circulation over the Chukchi and Beaufort Seas to anthropogenic forcing.

During SO, the only significant southeast wind is over the Chukchi Sea while a weak but insignificant anticyclonic trend over the Beaufort and Arctic Seas is simulated in the Historical experiment (Fig. 6b). An insignificant trend pattern with anticyclonic wind trend over the Arctic and weak westerly wind over the southern Chukchi–Beaufort Seas is simulated in the HistoricalGHG experiment (Fig. 6e). Significant southwest wind trends are found over the southern Chukchi–Beaufort Seas in the HistoricalNAT experiment (Fig. 6h). On the other hand, the cyclonic pattern in the ERA-I of increasing westerly wind during the early-winter NDJ is not found in the Historical and HistoricalGHG experiments (Figs. 6c,f) but is generally captured in the HistoricalNAT experiment but with weak amplitude (Fig. 6i).

Recent observational studies have shown that summer to autumn sea ice anomalies in the Arctic significantly impact the following wintertime atmospheric circulation in the Northern Hemisphere (Honda et al. 2009; Wu and Zhang 2010). The influence of Arctic sea ice on the atmosphere has been investigated in many modeling studies (e.g., Deser et al. 2007). The multimode ensemble-mean trend maps of monthly surface winds in the 29-yr period of 1979–2008 in the AMIP experiment are shown in Figs. 6j–l. Obviously, neither the observed enhanced easterly surface wind in JJA and SON seasons nor the westerly surface wind in NDJ shown in Fig. 5 is simulated in the AMIP experiments, suggesting that the observational trends in surface winds might not be attributed to the boundary forcings.

Overall, the above results suggest that the observed enhanced easterly surface wind over the Beaufort Sea during summer season may be attributable to anthropogenic external forcing, but neither to natural external forcing nor to boundary SST and sea ice forcings. There are considerable uncertainties in attributing observed long-term atmospheric circulation to anthropogenic or natural external forcing during the fall and early winter seasons.

4. Summary

Many important features of variability in the BSH have been revealed in previous studies (Overland 2009; Serreze and Barrett 2011; Gleicher et al. 2011; Moore 2012). This paper presents some further efforts that contribute to our knowledge of the BSH and its impact on regional climate change. With an EOF analysis, we show three leading temporal variability modes of the BSH at each season, which explain nearly 90% of their corresponding total variances. The first leading EOF primarily reflects a stationary BSH for all seasons, and explains about 50%–60% of the corresponding total variance. Consistent with Serreze and Barrett (2011), we also find that this EOF1-like variability of the BSH is related to the NAM and PNA in summer. But our results suggest that EOF1-like variability is closely allied with the negative phase of the NAM in all seasons. The second and third leading EOFs are not shown in previous studies, and characterize the north–south and west–east dipole-like variability of the BSH respectively during all four seasons, and significant correlations exist between PC2 and the PNA index in fall and winter.

We have also detected a statistically strengthened summer and autumn BSH and a weakened early-middle winter BSH going back to 1979. Although no basin-scale
significant trends are found in the total SLP field during any season, consistent with the findings in Serreze and Barrett (2011) and Gleicher et al. (2011), the PC2 index in fall (early winter) SLP field has exhibited a significant positive trend (negative) at the 5% significance level, resulting in the increasing pressure gradient between the Arctic Ocean and the surrounding landmass. Basin-scale significant changes in the BSH is clearly seen in the surface wind fields with significant anticyclonic trends from summer to autumn, and significant cyclonic trends in NDJ. There exists a shift from more frequent anomalous surface westerly winds to easterly winds since the mid-1990s during SO, and such changes in the atmospheric circulation over the western Arctic Ocean are identified as being a strong contributor to the observed downward trend in autumn sea ice extent. Combined with the findings that there have been increasing trends in areal-averaged monthly mean and 95th percentile wind speeds over the Beaufort Sea from July through November in Stegall and Zhang (2012), our results suggest that significant changes in atmospheric circulation over the western Arctic have occurred from July through November since 1979, and have strongly contributed to the observed sea ice decline.

Ensemble-mean forced trends in surface winds taken from multimodel historical all-forcing runs and simulations using greenhouse gas–only forcing from IPCC
AR5 models suggest that the increased greenhouse gas concentration since the late twentieth century may be a cause of the abovementioned significant changes in atmospheric circulation over the Chukchi and Beaufort Seas during summertime. However, the significant trends of surface winds in fall and early winter are not reproduced in AR5 simulations, indicating that there might not be significant impacts of external forcings and boundary forcings on the long-term changes in the BSH during these two seasons, or that the physical processes associated with the BSH are not well represented in the current AR5 models.

Moore (2012) shows that the enhanced BSH since the late 1990 is coincident with a statistically significant trend toward lower Eady growth rates (Hoskins and Valdes 1990) over much of the Beaufort Sea, with the largest negative values occurring in the vicinity of the climatological center of the BSH. Such coincident trends in the enhanced BSH and lower Eddy growth rates are also found during summertime for 1970–2011, but not during fall and early wintertime in our study (not shown). During 1970–2011, there are no significant trends of Eddy growth rates in SO and NDJ seasons. However, Eddy growth rates are mostly used as a diagnostic for cyclogenesis. To our knowledge, there are few diagnostics for anticyclogenesis. While we have examined the roles of external and boundary forcings on long-term trends in the BSH here, further study is needed to investigate the associated mechanisms behind those trends.

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FIG. A1. Ensemble-mean seasonally averaged SLP and surface winds over the Chukchi and Beaufort Seas from the Historical runs in the CNRM-CM5 model for the period 1979–2005.

FIG. A2. As in Fig. A1, but from the GFDL CM3 model.
Fig. A3. As in Fig. A1, but from the GISS-E2-H model.

Fig. A4. As in Fig. A1, but from the HadCM3 model.
reviewers. This work is funded by the Bureau of Ocean Energy Management (BOEM) of the U.S. Department of the Interior (DOI) under Contract M06PC00018 and NSF Grant ARC-1023592. This work is also supported by the National Natural Science Foundation of China (Grant 41075052) and the National Key Scientific Research Plan of China (Grant 2012CB956002). Computing resources were provided by the Arctic Region Supercomputing FIG. A5. As in Fig. A1, but from the MPI-ESM-LR model.

FIG. A6. As in Fig. A1, but from the MPI-ESM-MR model.
Center at the University of Alaska Fairbanks. We also wish to acknowledge the efforts of the many scientists involved in developing, running, and archiving the model simulations that have been used in this study. Without their efforts, this analysis would not have been possible.

APPENDIX A

Simulated Seasonal Climatology and Variability of Atmospheric Circulations over the Chukchi and Beaufort Seas from Six AR5 Models

In this study, control simulations from six AR5 models listed in Table 1 (CNRM-CM5, GFDL CM3, GISS-E2-H, HadCM3, MPI-ESM-LR, and MPI-ESM-MR; see appendix B for expansions) are used to estimate the natural variability of the linear trends of SLP and surface winds over the Chukchi and Beaufort Seas, and forced trends from these models are compared with that in the ERA-I in the attribution studies. Other AR5 models with available control and simulations are also examined. As noted in section 2, these six models are chosen for the fact that the seasonal climatology of the BSH and its interannual variability modes in the ERA-I (see Fig. 1) are reasonably simulated in preindustrial control and forced simulations from these models, but not in other AR5 models. To prove this, the mean seasonal cycles and leading EOF modes of interannual variability in the SLP field over the Chukchi and Beaufort Seas in the Historical runs from models in Table 1 are compared to that in the ERA-I here. The period of 1973–2005 from the Historical runs is used to have the same length of SLP as that from the ERA-I (33 yr) in the EOF analysis. This is done for all six models, as their results are comingled to come up with the conclusion regarding the detection and attribution of the recent changes over the Chukchi and Beaufort Seas. For comparison, simulated SLP values from these models are interpolated into the ERA-I grids, and the same EOF analysis as in the ERA-I is applied to each ensemble of the interpolated SLP fields during 1973–2005. The spatial pattern correlation (Taylor 2001) is examined to provide a way of summarizing how closely an EOF from the model matches the corresponding one in the ERA-I, and the ensemble-mean spatial pattern correlation is reported.

FIG. A7. Ensemble averaged percentages (%) of the total variance of SLP explained by EOF modes in the 33-yr Historical simulations from six models vs spatial pattern correlations of EOF modes in the ERA-I and the Historical simulations from six models for the season of (a) AM, (b) JJA, (c) SO, and (d) NDJ. In each season, the plot is grouped in three colors, each indicating a leading EOF mode in the simulation. The observed percentages in the ERA-I are marked by colored triangles in each plot for comparison.
First, ensemble-mean seasonally averaged SLP and surface winds over the Chukchi and Beaufort Seas for the period 1979–2005 in the Historical simulations from six models are shown in Figs. A1–A6. Although three models (CNRM-CM5, GISS-E2-H, and HadCM3) simulate a higher peak central pressure of the BSH in summer than the ERA-I, many features of the Beaufort Sea high are well simulated by these models. At a 1-hPa contour interval adopted, there is a closed BSH in the mean field of SLP for AM, JJA, and NDJFM in CNRM-CM5 (Fig. A1); for AM and NDJFM in GFDL CM3 (Fig. A2); for AM, JJA, and SO in GISS-E2-H (Fig. A3); for all seasons in HadCM3 (Fig. A4); for JJA in MPI-ESM-LR (Fig. A5); and for AM, JJA, and NDJFM in MPI-ESM-MR (Fig. A6). All models show that the local maximum of the winter over the Chukchi Sea and summer BSH are centered over the Beaufort Sea. The center of the simulated BSH moves throughout the year, indicating that the mobile feature of the BSH is captured in the models. The range in mean winter and summer SLP in the ERA-I (about 7 hPa in Fig. 1) is also reflected by these models (Figs. A1–A6), although CNRM-CM5 simulates a weaker one (about 5 hPa in Fig. A1) and HadCM3 reproduces a stronger one (about 11 hPa in Fig. A4). The spatial distribution of the seasonal mean surface wind is generally anticyclonic under the influence of the seasonal BSH pattern in all models (Figs. A1–A6).

Second, Fig. A7 suggests that there are remarkable similarities between simulated and observed EOF1 in SLP with ensemble-mean spatial pattern correlations above 0.90 in each season for all the models. This can also be shown by the EOF1 for each model displaying positive anomalies centered over the central Arctic Ocean (not shown). Figure A7 also shows that spatial pattern correlations between observed and simulated north–south dipole-like EOF2 are about 0.6–0.9 at JJA, SO, and NDJ and 0.6–0.9 at AM, indicating that the northern or southern shifted variability of the BSH is captured in the models. The range in mean winter and summer SLP in the ERA-I (about 7 hPa in Fig. 1) is also reflected by these models (Figs. A1–A6), although CNRM-CM5 simulates a weaker one (about 5 hPa in Fig. A1) and HadCM3 reproduces a stronger one (about 11 hPa in Fig. A4). The spatial distribution of the seasonal mean surface wind is generally anticyclonic under the influence of the seasonal BSH pattern in all models (Figs. A1–A6).

Third, as shown in Fig. A7, the leading three EOF modes explain 50%–70%, 15%–20%, and 10%–15% of the total variance in each season, respectively, which account for more than 90% of their corresponding total variances of SLP for all models. This suggests that the simulated SLP variability over the Chukchi and Beaufort Seas from these models is also dominated by these three EOF modes, as in the ERA-I (Table 2).

At last, the ensemble-mean spatial-averaged variances of SLP in the EOF analysis from the 33-yr Historical simulations during 1973–2005 from six models are compared with that from the ERA-I during 1979–2011 for four seasons. Table A1 indicates that all models reproduce comparable total variance of summer and winter SLP, but larger total variance of spring and autumn SLP. As noted in section 2, we found that these six models simulate comparable or substantially larger variability than observed over nearly all grid boxes. Overall, the significance tests taken in sections 3c and 3d do not overestimate the significance of the observed trends.

**APPENDIX B**

**Expansions for Model Names**

<table>
<thead>
<tr>
<th>Models</th>
<th>Names</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNRM-CM5</td>
<td>Centre National de Recherches Météorologiques Coupled Global Climate Model, version 5</td>
</tr>
<tr>
<td>GFDL CM3</td>
<td>Geophysical Fluid Dynamics Laboratory Climate Model, version 3</td>
</tr>
<tr>
<td>GISS-E2-H</td>
<td>Goddard Institute for Space Studies Model E2, coupled with the Hybrid Coordinate Ocean Model (HYCOM)</td>
</tr>
<tr>
<td>HadCM3</td>
<td>Hadley Centre Coupled Model, version 3</td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>Max Planck Institute Earth System Model, low resolution</td>
</tr>
<tr>
<td>MPI-ESM-MR</td>
<td>Max Planck Institute Earth System Model, medium resolution</td>
</tr>
</tbody>
</table>

**Table A1.** Ensemble-mean spatial-averaged SLP variance (hPa$^2$) in the EOF analysis from the Historical simulations during 1973–2005 and that from the ERA-I during 1979–2011 (with the ratio between simulated and observed variances in parentheses) from six models for four seasons.

<table>
<thead>
<tr>
<th>Obs/models</th>
<th>AM</th>
<th>JJA</th>
<th>SO</th>
<th>NDJ</th>
</tr>
</thead>
<tbody>
<tr>
<td>ERA-I</td>
<td>220</td>
<td>214</td>
<td>218</td>
<td>772</td>
</tr>
<tr>
<td>CNRM-CM5</td>
<td>438  (1.99)</td>
<td>273  (1.27)</td>
<td>472  (2.16)</td>
<td>782  (1.01)</td>
</tr>
<tr>
<td>GFDL CM3</td>
<td>710  (3.22)</td>
<td>276  (1.29)</td>
<td>620  (2.84)</td>
<td>954  (1.24)</td>
</tr>
<tr>
<td>GISS-E2-H</td>
<td>412  (1.87)</td>
<td>204  (0.95)</td>
<td>416  (1.91)</td>
<td>824  (1.07)</td>
</tr>
<tr>
<td>HadCM3</td>
<td>446  (2.03)</td>
<td>198  (0.93)</td>
<td>298  (1.36)</td>
<td>676  (0.87)</td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>661  (3.00)</td>
<td>279  (1.30)</td>
<td>453  (2.08)</td>
<td>760  (0.99)</td>
</tr>
<tr>
<td>MPI-ESM-MR</td>
<td>667  (3.03)</td>
<td>270  (1.26)</td>
<td>570  (2.61)</td>
<td>977  (1.27)</td>
</tr>
</tbody>
</table>