A Nonlinear Theory of Atmospheric Blocking: An Application to Greenland Blocking Changes Linked to Winter Arctic Sea Ice Loss

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ABSTRACT

In this paper, the impact of winter Arctic sea ice concentration (SIC) decline over Baffin Bay, Davis Strait, and Labrador Sea (BDL) on Greenland blocking (GB) is first examined. It is found that the GB has a longer duration, a more notable westward movement, and a larger zonal scale in the low SIC winter than in the high SIC winter. In particular, the decay of GB may become slower than its growth in the low SIC winter, but the reverse is seen in the high SIC winter. The GB in the low SIC winter can have a more important impact on cold anomalies over North American midlatitudes than in the high SIC winter because of its slower decay and stronger retrogression. The influence of large BDL SIC loss on the GB mainly through reduced meridional potential vorticity gradient (PV_y) related to reduced zonal winds over the North Atlantic mid- to high latitudes (NAMH) due to BDL warming is further examined by using the nonlinear phase speed and energy dispersion speed formula of blocking based on a nonlinear wave packet theory of atmospheric blocking. In this theory, the preexisting synoptic-scale eddies rather than the eddy straining or deformation is important for the blocking intensification and maintenance, which contradicts the eddy straining theory of Shutts. It is revealed from this theoretical model that under weaker NAMH zonal wind conditions the energy dispersion speed of GB may become smaller due to weaker PV_y during its decaying phase than during the blocking growing phase, in addition to the GB having larger negative phase speed and stronger nonlinearity. The opposite is true when the PV_y is larger. Thus, under a large SIC loss condition the GB shows notable retrogression, large zonal scales, and a long lifetime, which has a slower decay than its growth.

1. Introduction

In last three decades, the winter Arctic sea ice concentration (SIC) has been observed to undergo rapid decline over Barents–Kara Seas (BKS) and Baffin Bay, Davis Strait, and Labrador Sea (BDL) (Parkinson et al. 1999; Rothrock et al. 1999; Wadhams and Davis 2000; Comiso 2006; Comiso et al. 2008; Perovich et al. 2018; Li et al. 2020). While the linkages of winter Eurasian midlatitude cold extremes with the BKS sea ice loss have been widely established in many previous studies (Honda et al. 2009; Petoukhov and Semenov 2010; Inoue et al. 2012; Cohen et al. 2014; Li et al. 2015; Overland et al. 2011, 2015; Luo et al. 2016a,b), these studies cannot allow us to infer their causality (Luo et al. 2019a).

On the other hand, more attention in recent years has been focused on examining whether and how the Arctic SIC decline influences the atmospheric circulation and cold extreme weather over the North America (Outten and Esau 2012; Lee et al. 2015; Kug et al. 2015; Overland and Wang 2018). But such effects are still not sufficiently understood [the review of Cohen et al. (2018a)] because of the lack of a good theoretical model describing the evolution of atmospheric blocking. Lee et al. (2015) noted that low autumn SIC over Bering Sea contributed to the 2013/14 North American cold winter. Chen and Luo (2017, 2019) further found that the change of Greenland blocking (GB) in persistence and movement due to large BDL SIC decline over the west of Greenland can influence cold anomalies and cold extremes over the North American midlatitudes. However, the causal linkage between the GB and BDL SIC changes remains elusive, because they are coupled together and interact with each other (Chen and Luo 2017, 2019; Cohen et al. 2018b). This deserves further investigation.

In past decades, many theoretical models have been established to describe the maintenance of atmospheric

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blocking (Yeh 1949; Egger 1978; Charney and DeVore 1979; Shutts 1983; Haines and Marshall 1987). However, these models failed to describe the lifetime and spatiotemporal evolution of blocking in an unfiltered height field (Berggren et al. 1949; Rex 1950). Thus, developing a new theoretical model to describe the spatiotemporal evolution of blocking has been an important research target of atmospheric dynamics (Charney and DeVore 1979; Shutts 1983; Haines and Marshall 1987; Luo 2000, 2005; Luo and Li 2000; Mu and Jiang 2008; Luo et al. 2014, 2019a,b; Chen et al. 2015; Nakamura and Huang 2018; Screen et al. 2018). Since the North Atlantic blocking or GB is often excited by high-frequency synoptic-scale eddies (Berggren et al. 1949; Illari and Marshall 1983; Holopainen and Fortelius 1987; Mullen 1987), the proposed theoretical model must consider the role of synoptic-scale eddies in the blocking development. The nonlinear multiscale interaction (NMI) model of blocking based on the idea of the interaction among the mean zonal flow, planetary blocking wave, and synoptic-scale eddies proposed and developed by Luo (2000, 2005) and Luo et al. (2014, 2018, 2019b) provides an efficient tool for understanding the causal linkage between the GB and BDL SIC changes. The main purpose of the present paper is to further use the extended NMI model of Luo et al. (2018, 2019b) to reveal how the local change of zonal winds over the North Atlantic mid- to high latitudes affects the evolution of GB from a potential vorticity gradient perspective, because the BDL SIC decline corresponds to the local reduction of zonal winds in the mid- to high-latitude North Atlantic due to reduced meridional temperature gradient even when the GB events are absent. In this paper, we further present a new finding that the decay (growth) of GB becomes slower than its growth (decay) under the BDL SIC decline (increase) condition. Such a GB evolution process is important for the variation of North American cold anomalies and downstream air temperature anomalies over the North Atlantic and Europe. Moreover, our study further provides a theoretical explanation for why the GB has such a slow decay under the influence of BDL SIC decline.

This paper is organized as follows: In section 2, we describe the data and method. Based on the reanalysis data the basic characteristics of the GB change related to the BDL SIC and North Atlantic mid- to high-latitude (NAMH) zonal wind changes are presented in section 3. It is found that the BDL SIC change may influence the GB probably through the NAMH zonal wind change. In section 4, we present an extended nonlinear multiscale interaction (ENMI) model of the blocking wave packet theory proposed by Luo (2000, 2005) and Luo et al. (2014, 2018, 2019b) to describe the life process of eddy-driven

blocking. Using this ENMI model, one can obtain the nonlinear phase speed and nonlinear energy dispersion speed of blocking under some assumptions, which are the generalization of the previous model (Luo et al. 2019b). The theoretical results about the change of GB with the NAMH zonal wind change are given in section 5. The conclusions and discussion are summarized in section 6.

2. Data and method

We use the daily 500-hPa geopotential height (Z500) and zonal wind (U500), surface air temperature (SAT), and SIC taken from the ERA-Interim data (<http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/>) with a horizontal resolution of $2.5^\circ \times 2.5^\circ$ during the period from December 1979–February 1980 to December 2017–February 2018 (1979–2017) (Dee et al. 2011). We further obtain the daily anomalies at each grid by nonseasonalizing and subtracting their seasonal cycle and linear trend at each grid point.

There are alternative measures and indices of Greenland blocking (e.g., Hanna et al. 2013, 2016; Barnes et al. 2014; Woollings et al. 2018). Similar results can be obtained based upon different measures and criteria (Diao et al. 2006). Here, the one-dimensional blocking index of Tibaldi and Molteni (1990, hereafter TM) is used to identify GB events related to the sea ice change over BDL in terms of the reversal of Z500 over the region 90°W – 0° , which is defined by $\text{GHGN} = [\text{Z500}(\lambda, \phi_N) - \text{Z500}(\lambda, \phi_0)]/(\phi_N - \phi_0)$ and $\text{GHGS} = [\text{Z500}(\lambda, \phi_0) - \text{Z500}(\lambda, \phi_S)]/(\phi_0 - \phi_S)$ at given three reference latitudes: $\phi_N = 80^\circ\text{N} + \Delta$, $\phi_S = 40^\circ\text{N} + \Delta$ and $\phi_0 = 60^\circ\text{N} + \Delta$ for $\Delta = -5^\circ, 0^\circ$, or 5° at each longitude λ . In this case, a GB event can be identified if $\text{GHGS} > 0$ and $\text{GHGN} < -10 \text{ gpm } (^\circ\text{lon})^{-1}$ are satisfied at least for contiguous 12.5° of longitude and 3 consecutive days (Chen and Luo 2017, 2019).

3. Linkage of the GB change to the BDL SIC decline and the reduction of associated North Atlantic mid- to high-latitude zonal wind and meridional PV gradient

As noted in previous studies, the strength of meridional potential vorticity gradient (PV_y) (Luo et al. 2018, 2019a,b) related to the background westerly wind is important for the change of atmospheric blocking, because the blocking system has strong nonlinearity and weak dispersion in a weak PV_y environment and then maintains its larger amplitude and longer lifespan (Luo et al. 2019b). In the barotropic atmosphere, there is $\text{PV}_y = \beta - U_{yy} + FU$ for a slowly varying zonal wind,

where β is the nondimensional meridional gradient of the Coriolis parameter, $U_{yy} = \partial^2 U / \partial y^2$, U is the nondimensional background westerly wind and $F \approx 1$ is the Froude number. It is useful to first present the basic characteristics about how the GB change is linked to the variation of the BDL SIC anomaly and associated background westerly wind U and PV_y before examining the physical mechanism of the GB change.

Figure 1a shows the 1990–2013 trend of DJF-mean SIC anomaly as an example. It is seen that large negative SIC anomalies mainly appear over the BKS and BDL. The impact of BKS SIC decline on cold anomalies and cold extremes over Eurasia has been widely investigated in previous studies (Honda et al. 2009; Petoukhov and Semenov 2010; Inoue et al. 2012; Cohen et al. 2018a; Luo et al. 2016a, 2019a; Yao et al. 2017). But not many studies are focused on the impact of BDL SIC anomaly on North American midlatitude cold extremes, although the BDL SIC anomalies can probably affect the SAT change over North America and north Europe through the GB change (Chen and Luo 2017, 2019; Cohen et al. 2018b). Here, we further consider the impact of the BDL SIC anomaly on the GB. The time series of detrended DJF-mean SIC and SAT anomalies averaged over the BDL region (50° – 75° N, 90° – 50° W) are shown in Figs. 1b and 1c, respectively. The calculation reveals that the SAT anomaly over the BDL has a strong negative correlation of -0.84 ($p < 0.01$) with the SIC anomaly. Thus, a large SIC decline in the BDL corresponds to a strong warming that exists even in the absence of GB (Chen and Luo 2017). The regressed Z500 and SAT anomalies against the BDL-averaged DJF-mean SIC time series (Fig. 1d) show that a GB can correspond to a negative BDL SIC anomaly, which resembles a negative North Atlantic Oscillation (NAO[−]) pattern. For this case, a strong cold anomaly mainly occurs over north Europe, whereas a very weak and nonsignificant cold anomaly is only seen over the east part of North American midlatitudes. The midlatitude cold anomaly vanishes if the positive SIC anomaly is present, because the atmospheric circulation linked to the BDL SIC rise shows a positive North Atlantic Oscillation (NAO⁺) pattern. Below, we will examine how the BDL SIC change influences the GB.

a. Changes in the GB and associated SAT anomaly and their link to the BDL SIC change

Recent observational study shows that there is a significant increasing trend of GB in all seasons (Hanna et al. 2016). To see the impact of the BDL SIC on the GB change, we first pick GB events in low and high BDL SIC winters using the TM index. The low (high) SIC winter is defined as the value of normalized DJF-mean

SIC anomaly averaged over the BDL below -0.5 standard deviations (STDs) (above 0.5 STDs) as shown in Fig. 1b. Using this definition, we can detect 14 high and 10 low SIC winters during 1979–2017. It is found that there are 18 GB events in the high SIC winter and 21 GB events in the low SIC winter, whose events per winter correspond to 1.29 GB events in the high SIC winter and 2.1 GB events in the low SIC winter, respectively. Thus, the GB events are more frequent in the low SIC winter than in the high SIC winter. The composite daily Z500 and SAT anomalies of GB events are shown in Fig. 2 in the high and low SIC winters. It is noted that the composite positive Z500 anomaly shows a retrogression for GB events in the high and low SIC winters, but has a longer lifetime in the low SIC winter (Fig. 2b) than in the high SIC winter (Fig. 2a). The positive Z500 anomaly prior to the GB onset is also found to show a larger zonal scale in the low SIC winter than in the high SIC winter. Another interesting result we find here is that a cold anomaly appears in the relatively high latitudes of North America (mainly north of 40° N) especially during the growth and mature phases of GB in the high SIC winter because the westward movement of GB is less evident (Fig. 2a). The cold anomaly can shift toward the midlatitudes along the northwest–southeast direction with the temporal evolution of GB. Overall, the time-mean cold anomaly during the GB life cycle is less strong (not shown). In the low SIC winter, a strong cold anomaly mainly appears in the North American midlatitudes (south of 40° N) especially during the decaying phase of GB (Fig. 2b). These results are new findings different from previous studies (Chen and Luo 2017, 2019), who did not find intense North American cold anomaly mainly during the GB decaying phase.

To characterize the change of the GB under different SIC conditions, we further define the zonal span of the 100-gpm contour of the composite daily Z500 anomaly averaged over the 5° latitude range around the latitude of the maximum Z500 anomaly as the zonal scale L_B of GB and its maximum value as the blocking intensity B_I . The time–longitude evolution of the composite daily Z500 anomaly averaged over the 5° latitude range around the latitude of the maximum Z500 anomaly and the temporal variations of composite daily L_B and B_I are shown in Fig. 3 for GB events in low and high SIC winters. It is found that while the GB has a stronger retrogression in the low SIC winter than in the high SIC winter, the positive Z500 anomaly during the GB life cycle shows slow growth and rapid decay in the high SIC winter (Fig. 3a), but rapid growth and slow decay in the low SIC winter (Fig. 3b). Such a GB change can be clearly seen from Fig. 3d, which is crucial for the

FIG. 1. (a) Linear trend spatial pattern of DJF-mean SIC anomaly during 1990–2013. (b),(c) Time series of normalized detrended DJF-mean (b) SIC and (c) SAT anomalies averaged over the BDL region (50° – 75° N, 50° – 90° W) during 1979–2018 with a negative correlation coefficient of -0.84 . (d) Linear regression of DJF-mean Z500 [contour interval (CI) = $5 \text{ gpm (std dev)}^{-1}$] and SAT [$^{\circ}\text{C (std dev)}^{-1}$] anomalies against the domain-averaged BDL SIC time series (multiplied by -1.0) during 1979–2017. In (d), the dot represents the region above the 95% confidence level for a two-sided Student's t test.

occurrence region of North American cold anomalies. The amplitude of the GB during its prior period (from lag -20 to -10 days) is also larger in the low SIC winter than in the high SIC winter (not shown). Moreover, we can see that the GB has a larger zonal scale in the low SIC winter than in the high SIC winter (Fig. 3c), which is likely related to the zonal scale of the initial blocking and the strength of the background westerly wind or PV_y over the North Atlantic as noted below.

The temporal variations of composite daily anomalies of BDL-averaged SIC (SAT) and domain-averaged PV_y over the North Atlantic (45° – 65° N, 70° W– 0°) of GB events are shown in Fig. 4. It is found that while the long-lived GB occurring in the low SIC winter corresponds to strong warming and large SIC decline over the BDL, the BDL SIC (SAT) is always lower (higher)

in the low SIC winter than in the high SIC winter (Figs. 4a,b). We also note that during the GB life cycle the BDL SIC does not notably change with the temporal evolution of the GB (from lag -10 to 10 days), even though the BDL warming (positive SAT anomaly) is significantly amplified by the GB in the low SIC winter. This suggests that the variation of the BDL SIC is a slower process than that of the BDL SAT. We further see that prior to the GB onset (from lag -20 to -10 days), the BDL SAT (SIC) is higher (lower) for long-lived GB in the low SIC winter (blue line in Figs. 4a and 4b) than for short-lived GB in the high SIC winter (red line in Figs. 4a and 4b). To some extent, a large decline of the BDL SIC during the prior period from lag -20 to -10 days may be considered as a background of long-lived GB. Thus, the large decline

of this prior BDL SIC is a prerequisite for long-lived GB and its slow decay.

In addition, it is noted that while U500 (Fig. 4d) and PV_y (Fig. 4c) over the NAMH also change with the evolution of GB, they are smaller during their prior period from lag -20 to -10 days for long-lived GB in the low SIC winter than for short-lived GB in the high SIC winter. Thus, a comparison with Figs. 4a and 4b reveals that a large decline of the prior BDL SIC corresponds to a small prior U500 and PV_y . In other words, the small prior PV_y that favors long-lived GB is linked to low prior BDL SIC. This point is easily explained in terms of the nonlinear wave packet theory of blocking system by Luo et al. (2019b), who noted that the magnitude of background PV_y determines the lifetime and zonal movement of blocking. According to $PV_y = \beta - U_{yy} + FU$ in Luo et al. (2019b), the reduced prior PV_y (Fig. 4c) is due to the local weakening of the prior westerly wind (Fig. 4d) produced by the prior positive height and temperature anomalies over BDL (Fig. 4a) related to the prior BDL SIC decline (Fig. 4b). Thus, a large BDL SIC decline can lead to a small PV_y over the NAMH. This conclusion also holds over Eurasia for the winter SIC decline in the BKS (Luo et al. 2019a). The above results are based on the classification of the domain-averaged DJF-mean SIC anomaly over BDL, which are also acceptable for the classification of the BDL-averaged winter-mean SIC anomaly for GB events excluded as shown in supplemental files (Fig. S1 in the online supplemental material).

b. Linkages of the GB and cold anomaly changes to the strength of PV_y associated with North Atlantic westerly wind

As revealed from the daily evolution of the BDL SIC or SAT and PV_y or U500 over the NAMH presented above, a large decline of the BDL SIC corresponds to a small PV_y during the GB life cycle and during its prior period. While the BDL SIC is not largely changed with the evolution of the GB, long-lived GB requires that the prior BDL SIC is low. On the other hand, because the temporal evolution of blocking is directly related to the magnitude of background PV_y , it is useful to examine the relationship between PV_y over the NAMH and BDL SIC from a winter-mean perspective. On this basis, the GB events are further classified in terms of the value of domain-averaged DJF-mean PV_y over the NAMH to understand how the variation of DJF-mean PV_y associated with the BDL SIC change influences the GB evolution. Because $PV_y = \beta - U_{yy} + FU$ includes the variation of the NAMH westerly wind in strength and meridional distribution, it is also useful to examine

the variation of the winter-mean U500 and its relationship with PV_y .

To identify the linkage of the DJF-mean PV_y or U500 change over the NAMH to the DJF-mean BDL SIC decline, here we exclude the blocking days of GB events (from lag -10 to 10 days) in winter when the DJF-mean anomaly fields are calculated. In this case, the interannual changes of the obtained DJF-mean NAMH PV_y and BDL SIC may be regarded as important background factors influencing the GB evolution. The regressed DJF-mean U500 and PV_y anomalies against the time series of normalized DJF-mean BDL SIC anomaly (multiplied by -1.0) are shown in Figs. 5a and 5b. It is noted that the BDL SIC decline corresponds to reduced zonal winds (Fig. 5a) and reduced PV_y (Fig. 5b) over the NAMH (45° – 65° N, 70° W– 0°). To examine whether the reduction of DJF-mean NAMH zonal winds or PV_y is due to the DJF-mean SIC decline rather than due to the presence of GB events, it is useful to further show the time series of normalized DJF-mean U500 and PV_y anomalies averaged over the NAMH in Figs. 5c and 5d for GB events excluded. It is easily seen that the correlation coefficient between the DJF-mean U500 (Fig. 5c) and PV_y (Fig. 5d) over the NAMH is 0.977 ($p < 0.01$) for GB events excluded, whereas the DJF-mean BDL SIC has significant positive correlations of 0.49 ($p < 0.05$) and 0.53 ($p < 0.05$) with the NAMH U500 and PV_y . Moreover, it is noted that the DJF-mean U500 (PV_y) over the NAMH for GB events excluded has a significant positive correlation of 0.84 (0.85) ($p < 0.01$) with itself for GB events included as shown in Figs. S2c and S2d. The regression shows that a large decline of the BDL SIC corresponds to a small U500 or PV_y over the NAMH (Figs. 5a,b). Thus, the local reduction of the NAMH PV_y is associated with the BDL SIC decline via the local weakening of zonal winds over the NAMH. In fact, even when the GB events are included in the winter height, temperature and wind fields, the obtained DJF-mean PV_y and U500 anomaly patterns (Figs. S2a,b) resemble those in Figs. 5a and 5b. Thus, the presence of the GB events does not significantly change the DJF-mean BDL SIC and NAMH PV_y (U500) anomaly patterns in that the DJF-mean BDL SIC for GB events excluded in winter has a high positive correlation of 0.96 ($p < 0.01$) with that for GB events included. The above analysis reveals that the reduced DJF-mean U500 or PV_y related to BKS warming in association with the DJF-mean BDL SIC decline are not largely influenced by the GB events. As noted by Chen and Luo (2017) and further revealed by the evidence presented below, the reduction of the NAMH zonal winds is mainly attributed to reduced meridional temperature gradient and positive height anomalies over Greenland due to enhanced

FIG. 2. Time sequences (2-day interval) of composite daily Z500 (CI = 40 gpm) and SAT anomalies (color shading) of (a) 18 Greenland blocking events in the 14 high SIC winters and (b) 21 Greenland blocking events in the 10 low SIC winters. The dots represent the regions above the 95% confidence level for the two-sided Student's t test.

high-latitude warming related to the BDL SIC loss even when the GB is absent. As a result, the reduced zonal winds can lead to reduced PV_y over the North Atlantic mid- to high latitudes. Of course, the reduction of the NAMH PV_y is also related to the positive sea surface temperature (SST) anomaly north of the Gulf Stream extension (Chen and Luo 2019), even though the winter positive SST anomaly south of the Gulf

Stream can drive wave trains propagating into Eurasia (Sato et al. 2014; Simmonds and Govekar 2014).

As implicated by the above analysis, the different strengths of the DJF-mean BDL SIC or NAMH PV_y anomaly may be considered as the different background conditions of GB events. Here, we further classify the DJF-mean SIC and PV_y for GB events excluded into the high and low winters to examine how the frequency of

FIG. 2. (continued)

GB events depends on the magnitude of the DJF-mean BDL SIC or NAMH PV_y . Based on the ± 0.5 STDs of the DJF-mean BDL SIC and NAMH PV_y time series, we can find 15 high and 11 low PV_y winters during 1979–2017. There are also 14 high and 10 low SIC winters as noted above. Here, we only present composite DJF-mean U500 and PV_y anomalies in the high or low SIC and PV_y winters. Figures 6a–d show the composite DJF-mean U500 and PV_y anomalies in the high and low SIC winters for GB events excluded, whereas Figs. 6e–h correspond to those in the high and low PV_y winters for GB events excluded. It is noted that the changes of U500

and PV_y anomalies are zonally localized between high and low SIC (PV_y) winters. Over the North Atlantic mid- to high latitudes between 40° and 70°N there is a positive U500 anomaly in the high SIC winter (Fig. 6a), but a negative U500 anomaly in the low SIC winter (Fig. 6b). Correspondingly, PV_y is intensified (reduced) over the North Atlantic region between 45° and 65°N in the high (low) SIC winter as seen in Fig. 6c (Fig. 6d). In other words, the SIC decline over BDL can reduce PV_y south of Greenland via reduced zonal winds due to decreased meridional temperature gradient and positive height anomaly related to BDL warming. We further

FIG. 3. (top) Time-longitude evolution of composite daily Z500 anomaly ($CI = 20$ gpm) averaged over a 5° latitude range around the latitude of the maximum Z500 anomaly (thick line denotes the 100-gpm contour) for (a) 18 GB events in 14 high SIC winters and (b) 21 GB events in 10 low SIC winters. The arrow denotes the movement direction. (bottom) Temporal variations of composite daily (c) blocking zonal scale L_B and (d) blocking intensity B_I of GB events in high (solid line) and low (dashed line) SIC winters.

see that while the changes of U500 and PV_y anomalies in the high and low PV_y winters (Figs. 6e–h) are slightly stronger than those in the high and low SIC winters (Figs. 6a–d), they have similar spatial patterns. The above results are also found for the case with GB events included (Fig. S3). That is to say, while the DJF-mean BDL SIC and PV_y (U500) anomalies are not largely changed by the GB evolution, they may significantly influence the temporal variation of the subsequent GB. It is worthy of noting that the DJF-mean PV_y has a similar spatial pattern between the low (high) SIC and PV_y winters, whereas their intensity difference is relatively large. This suggests that the magnitude of the DJF-mean PV_y over the NAMH is not only related to the BDL SIC change, but also related to other many factors such as North Atlantic SST, internal atmospheric variability and so on. Based on the result of Chen and Luo (2019, their Fig. 5), it is estimated that the decline of the winter BDL SIC contributes to the nearly 40% variability of DJF-mean PV_y over the NAMH, whereas

the North Atlantic midlatitude SST anomaly and other factors including internal atmospheric variability contribute to its about 28% and 32% variability, respectively. Thus, the reduction of the NAMH DJF-mean PV_y depends more mainly on the BDL SIC decline than the North Atlantic SST anomaly or internal atmospheric variability or other factors. Below, a further daily analysis is presented to evaluate the likely influence of the SIC change on the GB and North American cold anomaly by examining how the GB change depends on the variation of PV_y in winter.

The calculation reveals that there are 21 and 23 GB events in 15 high and 11 low PV_y winters, which correspond to 1.4 and 2.1 GB events per winter, respectively. Thus, the GB events are more frequent in the low PV_y winter than in the high PV_y winter, similar to the results in the low and high SIC winters as noted above. We show the time-mean composite daily Z500 and SAT anomalies averaged from lag -8 to 0 days and lag 0 to 8 days of GB events in the high and low PV_y winters in Figs. 7a–d. We see that in the high PV_y winter the positive time-mean

FIG. 4. Temporal variations of composite daily (a) SAT (K) and (b) SIC (%) anomalies averaged over the BDL (50° – 75° N, 90° – 50° W) and (c) nondimensional PV_y (scaled by $10^{-11} \text{ m}^{-1} \text{ s}^{-1}$) and (d) U500 (m s^{-1}) anomalies averaged over 45° – 65° N, 70° W– 0° during the blocking life cycle in high (red line) and low (blue line) SIC winters. The high (low) BDL SIC winter is defined as the value of the DJF-mean BDL SIC for GB events included above 0.5 (below -0.5) STDs. Gray shading denotes that the difference between two lines is statistically significant at the 95% confidence level based on a two-sided Student's t test.

Z500 anomaly has shorter zonal scale during the GB decay phase (Fig. 7b) than during the GB growth phase (Fig. 7a). For this case, the GB shows less westward movement during the blocking evolution process because the change of the time-mean position of the positive Z500 anomaly between the blocking growing (from lag -8 to 0 days) and decaying (from lag 0 to 8 days) phases is less distinct. We also see a relatively weak cold anomaly (not statistically significant) over North American high latitudes north of 50° N during the growing (Fig. 7a) and decaying (Fig. 7b) phases of GB. Thus, only a weak, insignificant cold anomaly appears over the North America during the blocking evolution in the high PV_y winter. The positive Z500 anomaly of the GB has a larger zonal scale in the low PV_y winter (Figs. 7c,d) than in the high PV_y winter (Figs. 7a,b), which also shows a larger zonal scale during the blocking decaying phase (Fig. 7d) than during the blocking

growing phase (Fig. 7c). During the GB growing phase, a weak cold anomaly (not significant) is seen in the North American continent north of 40° N (Fig. 7c). But during the blocking decaying phase the cold anomaly is intensified and becomes much stronger and more persistent than during the blocking growing phase and further shifts to lower latitudes (near 20° N) of the eastern United States (Fig. 7d). In the presence of GB, the cold anomaly over the European continent is stronger in the low PV_y winter than in the high PV_y winter, which expands toward higher latitudes of Europe during the blocking decay phase than during the growth phase. This result is different from the finding of Chen and Luo (2017), who did not mention the latitude change of the continental cold anomaly.

In the low PV_y winter, the positive center of the Z500 anomaly is located to the east of Greenland during the GB growing phase, but the west of Greenland during the GB decaying phase, thus exhibiting a marked westward

FIG. 5. (left) Linear regressions of DJF-mean (a) U500 [$\text{m s}^{-1} (\text{std dev})^{-1}$] and (b) PV_y [$\text{m}^{-1} \text{s}^{-1} (\text{std dev})^{-1}$] anomalies for GB events excluded (blocking days from lag -10 to 10 days are removed, where lag 0 denotes the peak day of the GB) against the time series of the DJF-mean BDL SIC anomaly for GB events excluded (multiplied by -1.0) during 1979–2017, where the dotted regions represent the 95% confidence level for an F test. (right) Temporal variations of normalized detrended DJF-mean (c) U500 and (d) PV_y anomalies averaged over $45^\circ\text{--}65^\circ\text{N}$, $70^\circ\text{W--}0^\circ$ for GB events excluded (blocking days from lag -10 to lag 10 days are excluded), where $R(\text{SIC}, \text{U500}) = 0.49$ and $R(\text{SIC}, \text{PV}_y) = 0.53$ represent the correlation coefficients of the domain-averaged DJF-mean U500 and PV_y time series with the DJF-mean BDL SIC anomaly, as well as $R(\text{U500}, \text{PV}_y) = 0.98$ denotes the correlation coefficient between DJF-mean U500 and PV_y time series. The dotted (dot-dashed) line represents the -0.5 (0.5) standard deviations of the U500 and PV_y time series.

shift. Overall, strong cold anomalies are seen to simultaneously occur over the east and central parts of North American midlatitudes and European continents (northern Europe and central Europe) because of enhanced retrogression and increased duration of GB in the low PV_y winter. Because the BDL SIC decline leads to reduced PV_y over the NAMH, it is concluded that cold extremes may simultaneously take place over the eastern United States and European continents under the influence of large BDL SIC decline, as observed in Kendon and McCarthy (2015) and Messori et al. (2016).

The strong (weak) westward movement of positive Z500 anomalies during the GB life cycle in the low (high) PV_y winter can be clearly seen from the time–longitude evolution of the meridionally averaged Z500

anomaly over a 5° latitude range around the latitude of the maximum Z500 anomaly in Figs. 8a and 8b. It is also noted that the blocking zonal scale L_B is significantly enlarged during the GB decaying phase (Fig. 8c), even though it has a larger zonal scale in the low PV_y winter (dashed line in Fig. 8c) than in the high PV_y winter (solid line in Fig. 8c). In particular, in the high PV_y winter the zonal scale L_B of GB can be largely shortened as the blocking decays and is much shorter than during its growth phase (solid line in Fig. 8c). The reversed result is seen in the low PV_y winter (dashed line in Fig. 8c). While PV_y changes with the evolution of GB, PV_y is always smaller during the prior period of GB and its subsequent evolution period for long-lived GB in the low PV_y winter than for short-lived GB in the high

FIG. 6. Composite DJF-mean (a),(b),(e),(f) U500 (m s^{-1}) and (c),(d),(g),(h) PV_y ($\text{m}^{-1} \text{s}^{-1}$) anomalies for GB events excluded (blocking days from lag -10 to 10 days are removed, and lag 0 denotes the blocking peak day) in the (a),(c) 14 high and (b),(d) 10 low SIC winters for GB events excluded and (e),(g) 15 high and (f),(h) 11 low PV_y winters for GB events excluded during 1979–2017. The dot represents the region above the 95% confidence level for the two-sided Student's t test.

PV_y winter. Thus, the low prior PV_y is crucial for the long lifetime and evolution of the subsequent GB event, even though largely influenced by the GB evolution. These results are acceptable for the classifications of the DJF-mean PV_y anomalies over the NAMH even for GB events excluded (Fig. S4) and included (Fig. S5).

We also see from the temporal variation of the amplitude B_I of GB that the UB has both slow growth and rapid decay under the high PV_y condition (solid line in Fig. 8d), but rapid growth and slow decay under the low PV_y condition (dashed line in Fig. 8d). Such a GB evolution is also seen from the time evolution of the

FIG. 7. Time-mean composite daily Z500 (CI = 40 gpm) and SAT (color shading) anomalies (a),(c) from lag -8 to 0 days and (b),(d) from lag 0 to 8 days of (a),(b) 21 GB events in 15 high PV_y winters and (c),(d) 23 GB events in 11 weak PV_y winters. The dot represents the region above the 95% confidence level for the two-sided Student's t test.

composite daily Z500 anomaly during the GB life cycle in low and high SIC winters (not shown). Because the North American midlatitude cold anomaly in the low PV_y winter is much stronger during the blocking decaying phase than during the blocking growing phase, it is inferred that the cold anomaly over the eastern United States is more likely related to the persistent slow decay, increased duration and enlarged zonal scale of GB.

c. NAO patterns as the precursors of the different GB evolutions

As noted in previous studies, the NAO^- is identical to the GB (Luo et al. 2007; Davini et al. 2012), but the NAO^+ corresponds to intensified zonal winds in the North Atlantic midlatitudes and is followed by the European blocking or Ural blocking due to the decay of the NAO^+ (Luo et al. 2007). Nevertheless, there are more Ural blocking (UB) events associated with positive (neutral) North Atlantic Oscillation or NAO^+ (NAO^0) than negative North Atlantic Oscillation or NAO^- [for the definitions of these events, see Luo et al. (2017, 2019)]. However, how the NAO^+ or NAO^- prior to the GB onset affects the GB evolution is unknowable. To find the precursor of the slow decay and rapid growth of GB in the low PV_y

winter, we show time-mean composite Z500 and SAT, U500, PV_y and SIC anomalies averaged over a prior period of GB from lag -20 to -10 days in the high and low PV_y winters in Fig. 9. It is found that prior to the GB onset (from lag -20 to -10 days) a weak NAO^+ pattern appears over the North Atlantic in the high PV_y winter (Fig. 9a), but a weak NAO^- pattern is seen in the low PV_y winter (Fig. 9b). We also see that the U500 and PV_y anomalies are large (small) over the NAMH in the presence of weak prior NAO^+ (NAO^-) patterns, as seen in Figs. 9c and 9e (Figs. 9d,f). The weak prior NAO^+ (NAO^-) is likely related to the high (low) BDL SIC as seen in Fig. 9g (Fig. 9h), which corresponds to North Atlantic high-latitude cooling (warming) as shown in Fig. 9a (Fig. 9b). Moreover, the small (large) zonal scale of the initial GB in the high (low) PV_y winter is also related to the presence of a weak NAO^+ (NAO^-) pattern prior to the GB onset. Similar results are also found in the high (low) SIC winter (Fig. S6). Thus, it is concluded that the BDL SIC change can influence the GB probably through changing the prior NAO, associated background zonal wind and PV_y . In the presence of a weak prior NAO^+ (NAO^-) pattern, the GB has a shorter (larger) initial zonal scale, which favors the slow growth and rapid decay (rapid growth

FIG. 8. (top) Time-longitude evolutions of composite daily Z500 anomalies averaged over a 5° latitude range around the latitude of the maximum Z500 anomaly (CI = 20 gpm; thick line denotes the 100-gpm contour) for GB events in the (a) high and (b) low PV_y winters. The arrow denotes the movement direction. (bottom) Temporal variations of composite daily (c) blocking zonal scale L_B and (d) blocking intensity B_I of GB events in high (solid line) and low (dashed line) PV_y winters.

and slow decay) of the GB. These results are new findings different from previous studies (Chen and Luo 2017, 2019), who did not find that the initial zonal scale of the GB is related to the phase of NAO.

It is noted that the previous theoretical models (Charney and DeVore 1979; Haines and Marshall 1987) cannot be used to explain why the GB shows rapid growth and slow decay under a large BDL SIC loss (small PV_y) condition. However, this problem can be investigated by using the following extended NMI model.

4. Extended NMI model and its analytical solution

a. Extended NMI model

As seen in Fig. 6, the background U500 and associated PV_y anomalies are zonally localized. Thus, we should further extend the NMI model of blocking in

Luo et al. (2019b) to include a zonally nonuniform slowly varying zonal flow or local background zonal winds. Since the GB is often driven by synoptic-scale eddies, in this paper we consider synoptic-scale eddies as a driving of GB. Here, we still used an equivalent barotropic model as in Luo et al. (2014, 2019b) because the blocking has an approximate barotropic structure. Considering a zonally nonuniform slowly varying basic flow $\bar{\psi}(x, y)$, the total barotropic streamfunction ψ_T can be decomposed into three parts: $\psi_T = \bar{\psi}(x, y) + \psi + \psi'$ as made in Luo (2000, 2005) and Luo and Li (2000), where ψ and ψ' denote planetary- and synoptic-scale streamfunction anomalies, respectively. Under a zonal-scale separation assumption, the nondimensional potential vorticity (PV) equations of the planetary- to synoptic-scale coupling in a β plane with a nondimensional width of L_y , scaled by the characteristic length $\tilde{L}$ (~ 1000 km) and velocity $\tilde{U}$ (~ 10 m s⁻¹), can be obtained as

FIG. 9. Time-mean composite daily (a),(b) Z500 (CI = 20 gpm) and SAT (color shading), (c),(d) U500, and (e),(f) meridional PV gradient (PV_y), and (g),(h) BDL SIC anomalies averaged from lag -20 to -10 days in the (a),(c),(e),(g) high and (b),(d),(f),(h) low PV_y winters selected based on the DJF-mean PV_y time series for GB events excluded. The dots represent the regions above the 95% confidence level for two-sided Student's t test.

$$\begin{aligned} & \left(\frac{\partial}{\partial t} + U \frac{\partial}{\partial x} \right) (\nabla^2 \psi - F\psi) + J(\psi, \nabla^2 \psi) + \text{PV}_y \frac{\partial \psi}{\partial x} \\ & = -\nabla \cdot (\mathbf{v}' q')_P - V \frac{\partial \nabla^2 \psi}{\partial y} - \frac{\partial \psi}{\partial y} \left(\frac{\partial^2 U}{\partial x \partial y} - \frac{\partial^2 V}{\partial x^2} \right), \quad (1a) \end{aligned}$$

$$\begin{aligned} & \left(\frac{\partial}{\partial t} + U \frac{\partial}{\partial x} \right) (\nabla^2 \psi' - F\psi') + \text{PV}_y \frac{\partial \psi'}{\partial x} \\ & = -J(\psi', \nabla^2 \psi) - J(\psi, \nabla^2 \psi') - V \frac{\partial \nabla^2 \psi'}{\partial y} \\ & \quad - \frac{\partial \psi'}{\partial y} \left(\frac{\partial^2 U}{\partial x \partial y} - \frac{\partial^2 V}{\partial x^2} \right) + \nabla^2 \psi_S^*, \quad (1b) \end{aligned}$$

where ∇^2 is the two-dimensional horizontal Laplacian operator, $J(a, b) = (\partial a / \partial x)(\partial b / \partial y) - (\partial b / \partial x)(\partial a / \partial y)$ is the Jacobian operator, $(U, V) = (-\partial \bar{\psi} / \partial y, \partial \bar{\psi} / \partial x)$ denotes the zonal and meridional components of the nonuniform basic flow, $\text{PV}_y = \beta + V_{xy} - U_{yy} + FU$ ($V_{xy} = \partial^2 V / \partial x \partial y$ and $U_{yy} = \partial^2 U / \partial y^2$) is the meridional PV gradient of the nonzonal basic flow, $\mathbf{v}' = (u', v') = (-\partial \psi' / \partial y, \partial \psi' / \partial x)$, $q' = \nabla^2 \psi'$, $F = (\tilde{L} / R_d)^2$ is the Froude number, R_d is the radius of Rossby deformation, $\beta = \beta_0 \tilde{L}^2 / \tilde{U}$, β_0 is the meridional gradient of the Coriolis parameter at a given reference latitude φ_0 , $-\nabla \cdot (\mathbf{v}' q')_P$ represents the blocking-scale component of the eddy vorticity flux convergence induced by synoptic-scale eddies ψ' , $\nabla^2 \psi_S^*$ is a synoptic-scale wave maker used to excite and maintain the preexisting synoptic-scale eddies upstream of blocking (Luo 2005) and other notations can be found in Luo et al. (2019b).

In Eq. (1), the slowly varying nonzonal basic flow (U, V) has been assumed to be maintained by the external source. When the basic flow $\bar{\psi}(x, y)$ is purely zonal, one can have $U(x, y) = U(y)$ and $V(x, y) = 0$ for $\bar{\psi}(x, y) = \bar{\psi}(y)$. In this case, Eq. (1) reduces to the extended NMI

model of blocking obtained by Luo et al. (2018, 2019b) because PV_y can be written as $\text{PV}_y = \beta - U_{yy} + FU$.

b. Model solution

To obtain the analytical solution of Eq. (1), we need to make some simplifying assumptions. We first assume that the zonal nonuniform variation of the basic flow is more slowly varying to allow $U(x, y) = U(X_2, y)$ and $V(x, y) = V(X_2, y)$ ($X_2 = \varepsilon^2 x$ is slowly varying zonal coordinate, ε is a positive small parameter and $0 < \varepsilon \ll 1.0$). Then we used an assumption that the meridional variation of the nonzonal basic flow is slowly varying compared to the carrier wave of the blocking anomaly similar to that used in Luo et al. (2019b). Using the Wentzel–Kramers–Brillouin (WKB) method, the total streamfunction ψ_T solution of an eddy-driven blocking event having zonal and meridional wavenumbers of k and m , respectively, from Eq. (1) in a fast variable form can be obtained as

$$\psi_T = \bar{\psi} + \psi + \psi' = \psi_P + \psi', \quad (2a)$$

$$\psi_P = - \int_0^y U(x, y') dy' + \int_{-L_x}^x V(x', y) dx' + \psi_B + \psi_m, \quad (2b)$$

$$\psi_B = B \sqrt{\frac{2}{L_y}} \exp[i(kx - \omega t)] \sin my + \text{cc}, \quad (2c)$$

$$\psi_m = -|B|^2 \sum_{n=1}^{\infty} q_n g_n \cos(n + 1/2)my, \quad (2d)$$

$$\psi' \approx \psi'_1 + \psi'_2, \quad (2e)$$

$$\begin{aligned} \psi'_1 &= f_0(x) \{ \alpha_1 \exp[i(\tilde{k}_1 x - \tilde{\omega}_1 t)] + \alpha_2 \exp[i(\tilde{k}_2 x - \tilde{\omega}_2 t)] \} \\ &\quad \times \sin\left(\frac{m}{2}y\right) + \text{cc}, \quad (2f) \end{aligned}$$

$$\begin{aligned} \psi'_2 &= -\frac{m}{4} \sqrt{\frac{2}{L_y}} B f_0 \sum_{j=1}^2 Q_j \alpha_j \exp\{i[(\tilde{k}_j + k)x - (\tilde{\omega}_j + \omega)t]\} \left[p_j \sin\left(\frac{3m}{2}y\right) + r_j \sin\left(\frac{m}{2}y\right) \right] \\ &\quad + \frac{m}{4} \sqrt{\frac{2}{L_y}} B^* f_0 \sum_{j=1}^2 Q_j \alpha_j \exp\{i[(\tilde{k}_j - k)x - (\tilde{\omega}_j - \omega)t]\} \left[s_j \sin\left(\frac{3m}{2}y\right) + h_j \sin\left(\frac{m}{2}y\right) \right] + \text{cc}, \quad (2g) \end{aligned}$$

$$\begin{aligned} & i \left(\frac{\partial B}{\partial t} + C_g \frac{\partial B}{\partial x} \right) + \lambda \frac{\partial^2 B}{\partial x^2} + \delta |B|^2 B \\ & + G f_0^2 \exp[-i(\Delta k x + \Delta \omega t)] = 0, \quad (2h) \end{aligned}$$

where $\omega = Uk - \text{PV}_y k / (k^2 + m^2 + F)$, $C_g = \partial \omega / \partial k = U - \text{PV}_y (m^2 + F - k^2) / (k^2 + m^2 + F)^2$ is the group velocity, $\Delta \omega = \tilde{\omega}_2 - \tilde{\omega}_1 - \omega$, $\Delta k = k - (\tilde{k}_2 - \tilde{k}_1)$,

$\tilde{\omega}_j = U \tilde{k}_j - \text{PV}_y \tilde{k}_j / (\tilde{k}_j^2 + m^2 / 4 + F)$ ($j = 1, 2$), $|B|^2 = BB^*$, $k = 2k_0$, $k_0 = 1 / (6.371 \cos \varphi_0)$ (6.371 is Earth's radius scaled by 1000 km), φ_0 is the reference latitude, $\lambda = [3(m^2 + F) - k^2] \text{PV}_y k / (k^2 + m^2 + F)^3$ is the linear dispersion term, $\delta = \delta_N / \text{PV}_y$ ($\delta_N > 0$) is the nonlinearity strength, $q_n = q_{Nn} / \text{PV}_y$, $m = -2\pi / L_y$, $\alpha_1 = 1$, $\alpha_2 = \alpha$, and $\alpha = -1$; $\tilde{k}_j = n_j k_0$ ($j = 1, 2$), n_j is the positive integer (e.g., $n_1 = 9$ and $n_2 = 11$),

B^* is the conjugate of B , L_x is the zonal wavelength of the blocking anomaly ψ_B , $f_0(x) = a_0 \exp[-\mu \varepsilon^2(x + x_T)^2]$ is the slowly varying eddy amplitude distribution located at $x = -x_T$, $\mu > 0$, a_0 is the constant eddy amplitude, cc denotes the complex conjugate of its preceding term, and p_j , r_j , s_j , h_j , Q_j , q_{Nn} , δ_N , g_n , G , and other parameters can be found in Luo et al. (2019b).

Equation (2) is an analytical solution of the extended NMI model in Eq. (1), which considers the blocking wave packet as a nonlinear initial-value problem as noted in Luo (2000, 2005). While it has the same form as in Luo et al. (2019b), U and PV_y in Eq. (2) are replaced by $U(x, y)$ and $PV_y = \beta + V_{xy} - U_{yy} + FU$. In Eq. (2), ψ_T is the total streamfunction field of eddy-driven blocking in a slowly varying basic flow and ψ_P represents the planetary-scale streamfunction field of blocking. Also note that ψ_B is the streamfunction of the blocking wavy anomaly and ψ_m denotes the mean zonal wind change due to the feedback of intensified blocking. The decomposition of ψ' into two parts: ψ'_1 and ψ'_2 in the form of $\psi' = \psi'_1 + \psi'_2$ has been used to derive the analytical solution in Eq. (2) from Eq. (1), where ψ'_1 represents the preexisting synoptic-scale eddies with two zonal wavenumbers k_1 and k_2 corresponding to the frequencies of $\tilde{\omega}_1$ and $\tilde{\omega}_2$ prior to the blocking onset. But ψ'_2 denotes the deformed eddies produced by the intensified blocking because it includes the amplitude B of blocking.

Under the assumption $J(\bar{\psi} + \psi, q + PV) \approx 0$, one can obtain $\partial q / \partial t \approx -\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ during the initial stage ($t \sim 0$) of the blocking onset because of the approximations $\psi'_2 \approx 0$, $\mathbf{v}' \approx \mathbf{v}'_1 = (-\partial \psi'_1 / \partial y, \partial \psi'_1 / \partial x)$ and $q' \approx q'_1 = \nabla^2 \psi'_1$. Here, we refer to $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ as the preexisting eddy forcing induced by preexisting synoptic eddies ψ'_1 or eddy forcing as in Luo (2000) and Luo et al. (2019b). In other words, $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ induced by preexisting synoptic-scale eddies must match the downstream incipient blocking structure so that the initial blocking dipole can be amplified into a typical dipole blocking when $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ has the same wavenumber as that of the blocking anomaly ψ_B (Luo et al. 2014, 2019b). This means that the presence of preexisting synoptic-scale eddies is a precursor of blocking formation in winter (Luo 2000, 2005) and in summer (McLeod and Mote 2015, 2016). In contrast, the eddy deformation or straining is unimportant for the blocking intensification and decay, which contradicts the eddy straining theory of Shutts (1983) as noted in Luo et al. (2014, 2019b). Because $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ has the frequency of $\tilde{\omega}_2 - \tilde{\omega}_1$, $\Delta \omega$ in $Gf_0^2 \exp[-i(\Delta kx + \Delta \omega t)]$ represents the difference between the frequencies of the blocking wavy anomaly ψ_B and preexisting eddy forcing $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$. Moreover, B is the complex envelope amplitude of blocking as a nonlinear wave packet, which is described by a nonlinear Schrödinger equation forced by the preexisting eddy forcing $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$.

c. Nonlinear phase speed and energy dispersion speed of blocking

According to Luo (2000) and Luo et al. (2019b), the nonlinear phase speed of blocking can be approximately obtained as

$$C_{NP} = C_p + C_N = U - \frac{PV_y}{k^2 + m^2 + F} - \frac{\delta_N M_0^2}{2kPV_y}, \quad (3)$$

where $C_p = \omega/k$, $C_N = -\delta_N M_0^2 / (2kPV_y)$, and $M_0 = |B|_{\max}$ is the maximum amplitude of blocking at each latitude. We further note that C_N reflects the blocking amplification-induced westward movement because of $C_N < 0$. The westward speed becomes largely negative when the blocking amplitude is large. Thus, C_N represents the nonlinear effect of the blocking amplification on the movement speed of blocking. In the real calculation, the domain-averaged height anomaly over the anticyclonic anomaly region of blocking may be considered as the value of M_0 .

In Eq. (2h), C_g represents the group velocity of the blocking wave packet and does not change with the blocking evolution. But it varies with the blocking change when the blocking is forced by the strong eddy forcing. In this case, one can obtain the modified group velocity C_{gm} of eddy-driven blocking if Eq. (2h) can be analytically solved. Using the perturbed inverse scattering transform (PIST) method (Okamawari et al. 1995) to solve Eq. (2h), the modified group velocity C_{gm} and phase speed C_{pm} of eddy-driven blocking and its energy dispersion speed C_{ND} can be obtained as

$$C_{gm} = C_g + \frac{dZ}{dt}, \quad (4a)$$

$$C_{pm} = \left[\omega - \frac{d(KZ)}{dt} - \frac{d\Theta}{dt} - \frac{C_g}{\sqrt{2\lambda}} \left(K + t \frac{dK}{dt} \right) \right] / \left(k - \frac{K}{\sqrt{2\lambda}} \right), \quad (4b)$$

$$C_{ND} = C_{gm} - C_{pm}, \quad (4c)$$

where Z , K and Θ represent the change parts of the group velocity, wavenumber and phase of the blocking soliton during its life cycle. The parameter equations of Z , K , Θ , and the blocking amplitude $M_0(t)$ can be found in the appendix.

When the eddy forcing strength is relatively weak, we have $Z \approx 0$ and $K \approx 0$. In this case, one can obtain $d\Theta/dt \approx (\delta/2)M_0^2$ from the blocking parameter equations given in the appendix. It is also noted that C_{pm} in Eq. (4b) reduces to the nonlinear phase speed of the form $C_{pm} \approx C_{NP} = \omega/k - \delta M_0^2 / (2k) = U - PV_y / (k^2 + m^2 + F) - \delta_N M_0^2 / (2kPV_y)$ as found in Luo et al. (2019b). For this case, because $C_{gm} \approx C_g$, the energy dispersion speed C_{ND} can be simplified to $C_{ND} \approx [2k^2 PV_y / (k^2 + m^2 + F)^2] + \delta_N M_0^2 / (2kPV_y)$, which

is also obtained in Luo et al. (2019b). On the other hand, we can have $C_{gm} = C_g + dZ/dt$ and $d\Theta/dt \approx (\delta/2)M_0^2$, if $K \approx 0$ even in the presence of strong eddy forcing. Thus, it is thought that Eq. (4) is a generalization of the nonlinear phase speed and energy dispersion speed derived by Luo et al. (2019b). It is further mentioned that the above simplified nonlinear phase speed and energy dispersion speed formula of blocking can be applied to the case of UB because the UB mainly results from the propagation of large-scale wave trains generated in the North Atlantic and because the eddy forcing over Eurasia is relatively weak (Luo et al. 2016a,b). But Eq. (4) can be used to reveal the physical cause of the GB evolution (movement, duration and energy dispersion changes) under different PV_y conditions because the GB is mainly driven by preexisting synoptic-scale eddies and because the eddy forcing is less weak. Of course, C_{NP} and C_{ND} may be used to approximately estimate the movement speed and energy dispersion speed of GB under different PV_y conditions.

5. Theoretical results

In this section, we present theoretical results about the evolution of dipole blocking under the different background westerly winds or PV_y conditions based on Eq. (2) by solving Eq. (2b). The same numerical scheme as used in Luo et al. (2019b) is used to obtain the solution of the blocking amplitude B for given parameters in Luo et al. (2019b, their Table 1) if the initial value of blocking and the background zonal wind are prespecified.

a. Impact of intensified and reduced background westerly winds on dipole blocking

Here, we consider $U_S(x, y) = U_0 + \Delta U_S(x, y)$ and $U_W(x, y) = U_0 - \Delta U_W(x, y)$ to approximate strong and weak NAMH background westerly winds under high and low SIC conditions, respectively, where $\Delta U_S(x, y) = \Delta u_1 e^{\gamma_1(y-y_3)^2} \cos[(\pi/4)y] + \Delta u_2 e^{\gamma_2(x-x_1)^2} e^{-(y-y_2)^2}$ and $\Delta U_W(x, y) = \Delta u_1 \arctan[\gamma_3(y - y_3)] + \Delta u_1 e^{\gamma_2(x-x_1)^2} e^{\gamma_4(y-y_3)^2}$. Correspondingly, the meridional wind $V_S(x, y)$ and $V_W(x, y)$ in PV_y can be derived from $U_S(x, y)$ and $U_W(x, y)$. For $U_0 = 0.55$, $\Delta u_1 = 0.2$, $\Delta u_2 = 0.1$, $\gamma_1 = -0.1$, $\gamma_2 = -0.05$, $\gamma_3 = 0.2$, $\gamma_4 = -0.8$, $y_1 = 1.5$, $y_2 = 3.3$, $y_3 = 3$, and $x_1 = -1.5$, we show the horizontal distributions of $U_S(x, y)$, $U_W(x, y)$, and associated $PV_y = \beta + V_{xy} - U_{yy} + FU$ for $F \sim 1$ in Fig. 10. In this figure, while the given background westerly winds and PV_y are highly idealized, they reflect a fact that the background zonal winds and PV_y are stronger in the North Atlantic midlatitudes than in the North Atlantic high latitudes. Figures 10a and 10c crudely represent intensified U500 and PV_y over North Atlantic high latitudes south of

Greenland in the high BDL SIC winter (Figs. 6a,c), whereas Figs. 10b and 10d show reduced DJF-mean U500 and PV_y anomalies over the North Atlantic high latitudes in the low BDL SIC winter as seen from the reanalysis data (Figs. 6b,d). Here, we also refer to the reduced (intensified) background westerly wind mainly in North Atlantic high latitudes as the weak (strong) background westerly wind.

On the other hand, because the BDL SIC decline or small PV_y corresponds to an initial blocking having a large zonal scale (Fig. 2b), it is assumed that the initial amplitude $B(x, y, 0)$ of GB has the form of $B(x, y, 0) = B_0 e^{-\kappa(y-y_B)^2} e^{-\rho x^2}$, which represents the zonally uniform initial amplitude for the BDL SIC decline for $\rho = 0$. But it corresponds to $\rho > 0$ for the BDL SIC increase (Fig. 2a), which reflects the zonally non-uniform initial amplitude. On this basis, we may use the above extended NMI model to examine how the background westerly wind change influences the behavior of blocking for its different initial values. Here, we fix $\kappa = 0.15$ and $y_B = 3.75$, but vary the value of ρ . The parameters of preexisting synoptic-scale eddies used here are chosen to be the same as in Table 1 of Luo et al. (2019b). We show the planetary-scale streamfunction ψ_P , wavy anomaly streamfunction ψ_B , synoptic-scale streamfunction anomaly ψ' and total streamfunction ψ_T fields of an eddy-driven blocking in Figs. 11 and 12 for $\rho = 0.05$ and $\rho = 0$ with $B_0 = 0.4$. It is found that the evolution of the blocking wavy anomaly ψ_B better captures the spatial structure and life cycle of GB (Fig. 2). Since the basic zonal wind is larger in the lower latitudes than in the higher latitudes, the blocking wavy anomaly ψ_B shows an asymmetric dipole with a strong anticyclonic anomaly to northwest and a weak cyclonic anomaly to southeast (Figs. 11b and 12b) due to a large retrograde movement of the blocking anticyclone in the high latitudes (Figs. 11b and 12b). While the planetary-scale field ψ_P exhibits a life cycle of dipole structure with the period of 10–20 days (Figs. 11a and 12a), the total field ψ_T behaves as a meandering westerly jet stream (Figs. 11d and 12d) because the preexisting synoptic-scale eddies are intensified and split into two branches around the blocking region during the blocking growing phase (Fig. 11c and 12c). Thus, the extended NM model here can be used to reveal the physical reason of the GB change in the high and low SIC or PV_y winters.

To make a comparison, we further quantify the main features of blocking dipole in Figs. 11 and 12. We show the time–zonal evolution of blocking streamfunction anomaly ψ_B averaged over $2.5 \leq y \leq 5.0$, the temporal variations of the blocking scale L_B and intensity B_I as defined by the maximum value ψ_N of the blocking

FIG. 10. Horizontal distributions of (a),(b) background westerly wind and (c),(d) associated PV_y for (a),(c) a strong westerly wind of $U = U_0 + \Delta u_1 e^{\gamma_1(y-y_1)^2} \cos[(\pi/4)y] + \Delta u_2 e^{\gamma_2(x-x_1)^2} e^{-(y-y_2)^2}$ and (b),(d) a weak westerly wind of $U = U_0 - \Delta u_1 \arctan[\gamma_3(y-y_3)] - \Delta u_1 e^{\gamma_2(x-x_1)^2} e^{\gamma_4(y-y_3)^2}$, where $U_0 = 0.55$, $\Delta u_1 = 0.2$, $\Delta u_2 = 0.1$, $\gamma_1 = -0.1$, $\gamma_2 = -0.05$, $\gamma_3 = 0.2$, $\gamma_4 = -0.8$, $y_1 = 1.5$, $y_2 = 3.3$, $y_3 = 3$, and $x_1 = -1.5$.

anticyclone and time-mean blocking wavy anomaly ψ_B from days 3 to 12 in Fig. 13 for large (U_S) and small (U_W) background zonal winds. It is found that the blocking dipole exhibits a weak retrogression in the strong high-latitude background westerly wind (Fig. 13a), but a strong westward movement in the weak high-latitude background westerly wind (Fig. 13b). This blocking becomes more long lived in the weak westerly wind case than in the strong case owing to smaller PV_y in high latitudes. We further see that the blocking intensity B_I (or ψ_N) can have a slow decay under a weak westerly wind condition (dashed line in Fig. 13d), but a rapid decay under a strong westerly wind condition (solid line in Fig. 13d). The time variation of the blocking scale L_B further shows that the blocking dipole has a larger zonal scale in a weak westerly wind state (dashed line in Fig. 13c) than in a strong westerly wind state (solid line in Fig. 13c). This feature can

also be seen from the time-mean blocking wavy anomaly ψ_B in Figs. 13e and 13f, which shows that the positive time-mean ψ_B anomaly has a longer zonal scale in the weak zonal wind case (Fig. 13f) than in the strong case (Fig. 13e).

While the above theoretical results are consistent with the reanalysis data, it is useful to further clarify why the GB shows slow decay and strong retrogression under a weak background wind condition. We may use C_{NP} to estimate the movement speed of the blocking dipole. In our following calculation, when U and V in PV_y are replaced by U_S and V_S (U_W and V_W) for the strong (weak) background westerly wind, we can calculate C_{NP} under different westerly wind conditions if M_0 is known. Here, for each y at each day the maximum value $(\psi_B)_{\max}$ of the blocking wavy anomaly ψ_B in the zonal direction is chosen as the value of $M_0 = (\psi_B)_{\max}/(2\sqrt{2/L_y})$. The nonlinear energy dispersion speed C_{ND} of the blocking

FIG. 11. Temporal evolutions of instantaneous (a) planetary-scale streamfunction ψ_P (CI = 0.15), (b) blocking wavy anomaly ψ_B (CI = 0.2), (c) synoptic-scale streamfunction ψ' (CI = 0.3), and (d) total streamfunction ψ_T (CI = 0.3) fields during the life cycle of an eddy-driven dipole blocking for the zonally nonuniform initial amplitude $B(x, y, 0) = B_0 e^{-\kappa(y-y_B)^2} e^{-\rho x^2}$, where $B_0 = 0.4$, $\kappa = 0.15$, $y_B = 3.75$, and $\rho = 0.05$ for a given strong background westerly wind in Fig. 10a.

dipole can also be calculated from Eq. (4) by solving Eq. (A2) in the appendix for $M_0(0) = 0.4$, $K(0) = 0$, $Z(0) = 0$ and $\Theta(0) = 0$.

We show time–meridional evolutions of nonlinear phase speed C_{NP} and energy dispersion speed C_{ND}

averaged over $-3.0 \leq x \leq 2.0$ of eddy-driven blocking in Figs. 14a–d for strong and weak background westerly winds. It is seen from Figs. 14a and 14b that while the nonlinear phase speed C_{NP} of blocking dipole is negative, its negative value over North Atlantic high

FIG. 12. As in Fig. 11, but for the weak background westerly wind in Fig. 10b and zonally uniform initial amplitude of GB in the form of $B(x, y, 0) = B_0 e^{-\kappa(y-y_B)^2}$ where $B_0 = 0.4$, $\kappa = 0.15$, and $y_B = 3.75$.

latitudes is larger in the weak background westerly wind (Fig. 14b) than in the strong background westerly wind (Fig. 14a). This means that the anticyclonic anomaly of blocking can show a stronger westward movement in the weak wind case than in the strong wind case. This point is easily explained. Because PV_y over high latitudes is much smaller in the weak

background wind than in the strong background wind (Figs. 10c,d), $C_N = -\delta_N M_0^2 / (2kPV_y)$ becomes a more largely negative value (dashed line in Fig. 14e), even though the blocking intensity ψ_N or B_I is slightly strong. Thus, it is inevitable to see a strong westward movement of GB under the weak zonal wind or small PV_y condition as shown in Fig. 9. Of course, such a

FIG. 13. (top) Time–zonal evolutions of blocking streamfunction anomaly ψ_B averaged over $2.5 \leq y \leq 5.0$ in (a) strong and (b) weak background zonal winds, where the arrow denotes the movement direction and a thick line represents the 0.35 contour. (middle) Temporal variations of (c) blocking scale L_B and (d) blocking amplitude ψ_N in the strong (solid line) and weak (dashed line) background zonal winds. (bottom) Time-mean blocking wavy anomaly streamfunctions ψ_B from days 3 to 12 for (e) strong and (f) weak background zonal winds.

retrogression becomes more notable if the blocking amplitude is larger (not shown).

Over the blocking region the nonlinear energy dispersion speed C_{ND} of dipole blocking decreases with the blocking growth (Figs. 14c,d). But the temporal

variation of C_{ND} depends on the strength and spatial distribution of background westerly winds and associated PV_y . It is noted that C_{ND} is small over the high-latitude region of blocking (Figs. 14c,d), which is decreased rapidly during the blocking growth phase

FIG. 14. (top),(middle) Time–meridional evolutions of (a),(b) nondimensional nonlinear phase speed C_{NP} ($CI = 0.04$) and (c),(d) energy dispersion speed $C_{ND} = C_{gm} - C_{pm}$ ($CI = 0.05$) averaged over $-3.0 \leq x \leq 2.0$ of eddy-driven blocking dipole in (a),(c) strong and (b),(d) weak background zonal winds, where C_{gm} and C_{pm} are obtained from Eq. (4) and the blocking evolution equations by using the perturbed inverse scattering transform method given in the appendix. (bottom) Temporal variations of (e) C_{NP} and (f) C_{ND} averaged over $-3.0 \leq x \leq 2.0$ and $3.0 \leq y \leq 4.5$ during the blocking life cycle for strong (solid line) and weak (dashed line) background zonal winds.

and remains be small during the time from days 3 to 10, but becomes relatively large after day 10 for a strong background wind case (solid line in Fig. 14f). This process leads to the slow growth and rapid decay

of dipole blocking. In contrast, C_{ND} is larger during the time from days 1 to 9 in the weak wind case than in the strong wind case, but still smaller after day 10 (dashed line in Fig. 14f). Such a nonlinear energy dispersion

FIG. 15. Temporal variation of blocking amplitude ψ_N of eddy-driven dipole blocking in the (a),(c) strong background zonal wind of $U_S(x, y) = U_0 + \Delta U_S(x, y)$ with $U_0 = 0.55$ (red line), $U_0 = 0.65$ (blue line), and $U_0 = 0.75$ (black line) and (b),(d) weak background zonal wind of $U_W(x, y) = U_0 - \Delta U_W(x, y)$ with $U_0 = 0.55$ (red line), $U_0 = 0.45$ (blue line), and $U_0 = 0.35$ (black line). (a),(b) Zonally uniform initial amplitude in Fig. 12 and (c),(d) zonally nonuniform initial amplitude in Fig. 11.

speed change can lead to a relatively fast growth of dipole blocking and its persistent slow decay.

b. Factors influencing the pace of blocking evolution

Although the above results reveal that the zonal scale, evolution speed, and movement of dipole blocking are related to the strength and spatial distribution of background westerly winds, it is also useful to further examine what factors influence the evolution of blocking dipole. Here, we examine two cases: 1) slowly varying background westerly wind $U_S(x, y) = U_0 + \Delta U_S(x, y)$ and $U_W(x, y) = U_0 - \Delta U_W(x, y)$ and 2) uniform westerly wind $U(x, y) = U_0$. We fix $\Delta U_S(x, y)$ and $\Delta U_W(x, y)$ as in Figs. 10a and 10b, but vary the value of U_0 . The temporal variations of the blocking intensity or amplitude ψ_N are shown in Figs. 15a–d for the different values of U_0 in the strong westerly wind $U_S(x, y) = U_0 + \Delta U_S(x, y)$ (Figs. 15a,c) and weak westerly wind

$U_W(x, y) = U_0 - \Delta U_W(x, y)$ (Figs. 15b,d) for the zonally nonuniform (Figs. 15a,b) and uniform (Figs. 15c,d) initial amplitudes.

It is found from the time variation of ψ_N in Fig. 15 that the growth and decay paces of the blocking depend strongly on the spatial structure of the initial blocking related to the phase of NAO. While the slow growth and rapid decay of the blocking amplitude are less distinct for the zonally uniform initial amplitude even under the strong background westerly wind condition (Fig. 15a), the rapid growth and slow decay of the blocking amplitude are distinct under the weak wind condition especially for a small U_0 (Fig. 15b). Thus, it is inferred that the large zonal scale of initial blocking, as denoted by the zonally uniform initial amplitude, tends to suppress the slow growth and rapid decay of blocking, instead favor the rapid growth and slow decay of blocking. For the zonally nonuniform

FIG. 16. Temporal variation of eddy forcing $[-\nabla \cdot (\mathbf{v}'_1 q'_1)_p]$; CI = 0.05] for (a) strong background zonal wind $U_S(x, y) = U_0 + \Delta U_S(x, y)$ and (b) weak background zonal wind $U_W(x, y) = U_0 - \Delta U_W(x, y)$ with $U_0 = 0.55$. The solid (dashed) lines represent the cyclonic (anticyclonic) vorticity forcing.

initial amplitude the slow growth and rapid decay of blocking can be clearly seen for a strong background westerly wind case, while slightly influenced by the strength of U_0 (Fig. 15c). Under a weaker background westerly wind condition, the rapid growth and slow decay of the blocking evolution are still seen (Fig. 15d). Such a blocking evolution asymmetry almost vanishes even when U_0 is larger. Thus, a strong background westerly wind and small zonal scale of the initial blocking as denoted by the zonally nonuniform initial amplitude favor the slow growth and rapid decay of blocking, but inhibit the rapid growth and slow decay of blocking. Similar results are found for the uniform westerly wind (Fig. S7), even though the blocking evolution is somewhat influenced by the strength of uniform westerly wind. Thus, it is inferred that the local weakening of the NAMH zonal wind or PV_y due to the BDL SIC decline is important for the long lifetime, strong retrogression, rapid growth and slow decay of the GB.

c. The physical cause of why the GB evolution has different paces under different westerly wind conditions

The different pace of the blocking evolution can be approximately explained using the matching mechanism between the blocking PV and preexisting eddy forcing $-\nabla \cdot (\mathbf{v}'_1 q'_1)_p$ according to $(\partial q / \partial t) \approx -\nabla \cdot (\mathbf{v}'_1 q'_1)_p$ (Luo et al. 2014). To validate this hypothesis, we calculate the spatial structure of $-\nabla \cdot (\mathbf{v}'_1 q'_1)_p$ during the blocking life cycle in Fig. 16 for the same eddy parameters as in Table 1 of Luo et al. (2019b).

In Fig. 16, the negative (positive) anomaly of $-\nabla \cdot (\mathbf{v}'_1 q'_1)_p$ marks an anticyclonic (cyclonic) vorticity forcing. The spatial pattern of $-\nabla \cdot (\mathbf{v}'_1 q'_1)_p$ evolves differently, even though its initial pattern (on day 0) is the same in the strong and weak background westerly wind cases. The blocking anticyclone is intensified (weakened) when the positive anomaly of blocking over the north side of $y = 2.5$ is close to or within the preexisting eddy forcing-induced anticyclonic (cyclonic) vorticity region. Such a role may be referred to as the “positive (negative) feedback” of preexisting eddy forcing on the blocking. In the strong background westerly wind case, a strong preexisting eddy forcing-induced cyclonic anomaly north of $y = 2.5$ after day 9 can reappear rapidly and the westward movement of blocking is prohibited so that the blocking anticyclone is close to the region of cyclonic vorticity forcing (Fig. 17a). This process can significantly reduce the anticyclonic anomaly of blocking. In other words, the negative feedback of preexisting eddy forcing on the blocking is enhanced to result in a rapid decay of blocking, especially when the background westerly wind is strong or when the initial blocking has a short zonal scale.

In a weak westerly wind background, the preexisting eddy forcing has a relatively long duration. The eddy forcing-induced cyclonic vorticity anomaly north of $y = 2.5$ after day 9 is weakened (Fig. 16b) and the retrogression of the blocking anticyclone is enhanced (Fig. 12b) when the initial blocking has a large zonal scale. In this situation, the blocking anticyclone is farther away from the eddy-induced cyclonic vorticity forcing

FIG. 17. Schematic diagram of whether the blocking decay is a slow process under the eddy forcing $[-\nabla \cdot (\mathbf{v}'_1 q'_1)_P]$ in (a) strong and (b) weak background zonal winds. The color shading represents the horizontal distribution of $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$ after the later (decay) period of $-\nabla \cdot (\mathbf{v}'_1 q'_1)_P$, the contour lines denote the blocking streamfunction anomaly, and + (–) represents the preexisting eddy-induced cyclonic (anticyclonic) vorticity.

region (Fig. 17b). As a result, the negative feedback of preexisting eddy forcing on the blocking is weakened to cause a slower decay of blocking than in the strong zonal wind case.

d. Comparison with the reanalysis data result

For a comparison, we calculate the composite daily nondimensional nonlinear phase speed C_{NP} of GB events to understand whether the theoretical result is consistent with the reanalysis data result. Here, we only quantify the movement speed of the blocking anticyclone during the GB life cycle. In C_{NP} , DJF-mean U500 and PV_y without the inclusion of GB events averaged over 50° – 80° N, 20° – 70° W can be considered as the values of the background U and PV_y because the blocking anticyclone mainly appears in this high-latitude region. For our calculation of C_{NP} , the daily Z500 anomaly averaged over 55° – 65° N, 30° – 60° W is chosen as the value of daily M_0 .

We show the temporal variations of nondimensional composite daily nonlinear phase speed C_{NP} for GB events in Fig. 18 in the high and low PV_y and SIC winters. It is found that the GB shows a stronger retrogression in the low PV_y winter (dashed line in Fig. 18a) than in the high PV_y winter (solid line in Fig. 18a). Such a westward movement is more distinct when the blocking amplitude is larger. The same result can be detected in the high and low SIC winters (Fig. 18b). Obviously, the estimate value of the phase speed of GB from Eq. (3) is consistent with the model and reanalysis data results. Thus, while the small PV_y favors the long life of GB, it also promotes the westward movement of GB. Thus, the small prior PV_y is a prerequisite for the long lifetime, strong

retrogression and slow decay of GB that significantly influences the North American midlatitude cold anomaly. Although the DJF-mean PV_y change is modulated by the North Atlantic SST anomaly, its reduction can be mainly due to the BDL SIC decline (Chen and Luo 2019). The above results lead us to infer that the BDL SIC decline can influence the speed of the GB evolution through changing the basic westerly wind and PV_y .

6. Conclusions and discussion

In this paper, we first examined the influence of the winter sea ice concentration (SIC) decline over Baffin Bay, Davis Strait, and Labrador Sea (BDL) on the Greenland blocking (GB) and North American cold anomaly in winter. It is revealed that the GB can have a short lifetime, less strong retrogression, and a short zonal scale in the high SIC winter, which also shows a slow growth and rapid decay. For this type of blocking, cold anomalies mainly occur during the blocking growing phase and are located over North American high latitudes. In the low SIC winter, the GB has a long duration, strong westward movement and large zonal scale. At the same time, the GB can behave as a relatively rapid growth and a slow decay. For this case, strong cold anomalies mainly occur during the GB decaying phase and are located over the eastern United States south of 40° N because of notable retrogression and slow decay of the GB. It is further found that the rapid growth and slow decay of the GB in the low SIC winter are associated with the large zonal scale of the initial blocking related to the weak precursor NAO^- . But the slow growth and rapid

FIG. 18. Temporal variations of nondimensional composite daily nonlinear phase speed C_{NP} for GB events in the high (solid line) and low (dashed line) (a) PV_y and (b) SIC winters, where domain-averaged values of U and PV_y are calculated over 50° – 80° N, 20° – 70° W, and M_0 is the blocking amplitude averaged over 55° – 65° N, 30° – 60° W.

decay of GB in the high SIC winter are linked to the small zonal scale of the initial blocking related to the weak precursor NAO^+ . The results are new findings different from the previous results of [Chen and Luo \(2017, 2019\)](#), who did not mention that the slow decay of the GB and the zonal scale of its initial amplitude are linked to the BDL SIC decline and important for the North American midlatitude cold anomaly.

A nonlinear multiscale interaction (NMI) model of blocking is extended to include zonally localized slowly varying background westerly winds related to the SIC decline (increase) to identify the physical cause of the different speed of the GB evolution under different SIC backgrounds. It is shown that the BDL SIC change may influence the GB through changing the movement speed, energy dispersion and nonlinearity of blocking system due to changes in the background westerly wind and PV_y . In the low (high) SIC winter, the zonal wind and PV_y are reduced (intensified) over the NAMH south of Greenland so that the nonlinearity of the GB system is intensified (weakened). Then, the nonlinear phase speed and energy dispersion speed formula of the nonlinear blocking wave packet theory are used to explain why the GB has rapid growth, slow decay, long lifetime and strong retrogression in the low SIC winter. It is revealed that in the low SIC winter the nonlinear phase speed of blocking anticyclone is largely negative as the reduced zonal wind and PV_y appear in the south of Greenland due to the large SIC decline. At the same time, we also find that the energy dispersion speed is rapidly decreased during the blocking growing phase and maintains a small value even during the decay phase, which is still small before the blocking disappearance. In this case, the GB can show a slow decay. In contrast, in the high SIC winter the decrease of the

energy dispersion speed of blocking system is more rapid so that it maintains a slower growth than in the low SIC winter. But the energy dispersion speed can be rapidly recovered to a large positive value during the blocking decay phase. As a result, the GB may show a rapid decay in this case.

The theoretical results presented above reveal that the BDL SIC decline can induce the long lifetime and notable retrogression of GB as well as its rapid growth and slow decay due to enhanced retrograde speed and reduced energy dispersion through reducing NAMH zonal winds and PV_y . While the changes in winter NAMH westerly winds and PV_y are related to North Atlantic SST anomalies and other factors, it seems that the reduced NAMH zonal winds and PV_y are mainly due to the BDL SIC decline ([Chen and Luo 2019](#)). Thus, the above results are acceptable even though we did not consider the role of SST anomalies. In fact, the NAMH zonal wind and PV_y changes obtained from the reanalysis data include the joint role of NAMH SST, BDL SIC anomalies, North Atlantic SSTs, internal atmospheric variability, and other factors. Nevertheless, exactly distinguishing the different roles of the BDL SIC decline, North Atlantic SST anomalies and internal atmospheric variability in the reduction of NAMH zonal winds and PV_y is difficult. This problem deserves a further investigation, which will be reported in another paper.

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APPENDIX

The Parameters Equations of Blocking Evolution under the Preexisting Eddy Forcing

According to the perturbed inverse scattering transform (PIST) method (Okamawari et al. 1995), We may assume that Eq. (2h) has an envelope soliton solution like (Luo 2000)

$$\begin{aligned} \psi_B = & 2M_0 \operatorname{sech} \left\{ \sqrt{\frac{\delta}{2\lambda}} M_0(t) [x - C_g t - Z(t)] \right\} \\ & \times \cos \left\{ kx - \frac{K(t)}{\sqrt{2\lambda}} [x - C_g t - Z(t)] + \Theta(t) - \omega t \right\} \\ & \times \sin(my). \end{aligned} \quad (\text{A1})$$

The blocking parameters M_0 , $K(t)$, $Z(t)$, and $\Theta(t)$ satisfy the following equations:

$$\frac{dM_0}{dt} = Ga_0^2 M_0 \sqrt{\frac{\delta}{2\lambda}} \int_{-\infty}^{\infty} \Re \sin \Phi \operatorname{sech} \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) dx', \quad (\text{A2a})$$

$$\begin{aligned} \frac{dK}{dt} = & -Ga_0^2 \frac{\delta}{\sqrt{2\lambda}} \int_{-\infty}^{\infty} \Re \cos \Phi \operatorname{sech} \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) \\ & \times \tanh \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) dx', \end{aligned} \quad (\text{A2b})$$

$$\begin{aligned} \frac{dZ}{dt} = & -\sqrt{2\lambda} K + Ga_0^2 \sqrt{\frac{\delta}{2\lambda}} \int_{-\infty}^{\infty} \Re \sin \Phi \operatorname{sech} \\ & \times \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) x' dx', \end{aligned} \quad (\text{A2c})$$

$$\begin{aligned} \frac{d\Theta}{dt} = & -\frac{1}{2}(K^2 - \delta M_0^2) - \frac{K}{\sqrt{2\lambda}} \frac{dZ}{dt} + Ga_0^2 \sqrt{\frac{\delta}{2\lambda}} \int_{-\infty}^{\infty} \Re \cos \Phi \\ & \times \left[1 - \sqrt{\frac{\delta}{2\lambda}} M_0 x' \tanh \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) \right] \\ & \times \operatorname{sech} \left(\sqrt{\frac{\delta}{2\lambda}} M_0 x' \right) dx', \end{aligned} \quad (\text{A2d})$$

where $\Re = \exp[-2\mu\epsilon^2(x + x_T)^2]$, $x' = x - C_g t - Z$, and $\Phi = \Delta kx + \Delta\omega t - Kx'/\sqrt{2\lambda} + \Theta$.

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