Winter Midlatitude Cold Anomalies Linked to North Atlantic Sea Ice and SST Anomalies: The Pivotal Role of the Potential Vorticity Gradient

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ABSTRACT

This study establishes a linkage between the North Atlantic sea ice concentration (SIC) or sea surface temperature (SST) and cold anomalies over northern Europe and North America through the Greenland blocking (GB) change. It is revealed that the magnitude of the meridional potential vorticity (PV) gradient in the North Atlantic mid- to high latitudes plays a key role in whether strong cold anomalies occur over the North America (NA) or northern Europe (NE) or both, while it is related to the SIC change observed over Baffin Bay, Davis Strait, and the Labrador Sea (BDL collectively) and the North Atlantic SST anomaly. When the midlatitude Atlantic SST is strongly warm or when the BDL SIC anomaly is largely positive, there is a corresponding large PV gradient over the North Atlantic. In this case, no intense cold anomalies are seen over NA due to less westward movement and the short lifetime of GB. Instead, a relatively strong cold anomaly appears over western and southern Europe. Its prior large-scale atmospheric circulation is the positive phase of the North Atlantic Oscillation (NAO). Moreover, strong cold anomalies can simultaneously occur over NA and NE only when the PV gradient is small under the influence of large SIC decline or intense mid- to high-latitude SST cooling across the Gulf Stream Extension. Its prior large-scale atmospheric circulation is a negative NAO phase. Daily composites show that strong cold anomalies over NA occur along the northwest–southeast direction in the presence of large SIC decline, whereas strong cold anomalies occur in NA midlatitudes even in the absence of large BDL SIC decline when mid- to high-latitude SST cooling is strong.

1. Introduction

Over the past two decades, the most pronounced feature of winter climate change has been the strong cooling trend over East Asia (Overland et al. 2011; Outten and Esau 2012; Screen and Simmonds 2013; Cohen et al. 2014) and enhanced Arctic warming associated with the rapid decline of sea ice concentration (SIC) in the Barents–Kara Seas (BKS) (Mori et al. 2014; Overland et al. 2015; Kug et al. 2015; Shepherd 2016). The physical cause of such a warm Arctic–cold Eurasia (WACE) pattern has been an important research topic (Cohen et al. 2014; Luo et al. 2016a,b; Overland 2016; McCusker et al. 2016; Sun et al. 2016; Yao et al. 2017; Wegmann et al. 2018). However, whether and how the midlatitude cold extremes are linked to Arctic warming or sea ice decline is not still clarified in

previous studies (Overland 2016). In particular, the Arctic–midlatitude weather linkage is difficult to establish (Cohen et al. 2018a,b; Overland and Wang 2018) because midlatitude cold events are dependent not only on the Arctic change (Mori et al. 2014; Kug et al. 2015) but also on the internal variability of atmospheric circulation patterns in mid- to high latitudes such as Ural blocking (Luo et al. 2016a,b; McCusker et al. 2016; Sun et al. 2016).

Recently, Luo et al. (2018, 2019) used the magnitude of the meridional potential vorticity (PV) gradient to establish a linkage between the Arctic change and midlatitude circulation or weather patterns by extending the nonlinear multiscale interaction model of blocking events to include slowly varying basic flow related to Arctic warming. They found that a small PV gradient can increase the persistence of blocking through reducing energy dispersion and enhancing nonlinearity and then favor increased midlatitude cold extremes (Luo et al. 2019). To

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some extent, a strong PV gradient may be considered as a barrier that inhibits blocking and the southward intrusion of cold Arctic air (Luo et al. 2018, 2019). The PV gradient can be reduced when Arctic warming is strong, even when the midlatitude continent is warm. The Arctic warming or SIC decline can significantly influence midlatitude cold extremes when the PV gradient is small. However, cold extremes can still occur in the midlatitudes when the midlatitude continent is cold and when the PV gradient is small, even though the Arctic warming or SIC decline is less strong. Thus, it is likely that the presence of midlatitude cold extremes does not necessarily require a strong Arctic warming or a large SIC decline. In other words, the Arctic warming or SIC decline is not a necessary condition for the generation of midlatitude cold extremes.

In recent years, cold winters have frequently occurred over the eastern United States and continental Europe (Cattiaux et al. 2010; Wang et al. 2010; Overland 2016; Messori et al. 2016). In particular, the co-occurrence of cold extremes over the North American and European continents was also observed in the same winter of 2013/14 (Kendon and McCarthy 2015; Trenary et al. 2015; Messori et al. 2016). Many modeling studies have examined the impact of SIC and SST over the North Pacific on cold winters in North America (Wang et al. 2014; Lee et al. 2015; Hartmann 2015). However, some cold winters over the North America do not involve the effect of SST anomalies or sea ice decline over North Pacific (Overland and Wang 2018). These cold extremes may also be influenced by the SST and SIC anomalies over North Atlantic. However, it is unclear how the cold extremes over the United States depend on the North Atlantic SST or SIC decline or their combination, even though the NAO is influenced by the North Atlantic SST (Czaja and Frankignoul 2002). Some studies have linked the cold extremes over North America to the SIC decline over Baffin Bay, Davis Strait, and the Labrador Sea (BDL collectively) (Chen and Luo 2017; Ballinger et al. 2018; Cohen et al. 2018a) and atmospheric internal variability (Seager and Henderson 2016).

However, some cold events over the eastern United States and Europe do not need a large BDL SIC decline or they do not occur even in the presence of large BDL SIC anomalies (Overland 2016; Cohen et al. 2018b). This raises a question of whether the winter cold extremes over the eastern United States and Europe are likely two sides of the same atmospheric circulation pattern over North Atlantic or whether they are linked to the North Atlantic SIC and SST anomalies. Some studies have linked the cold extremes over North America to the SIC decline over the BDL area through the westward movement of Greenland blocking (GB) (Chen and Luo 2017; Ballinger et al. 2018; Cohen et al. 2018a). However, it is unclear

under what conditions the midlatitude cold extremes are related to the BDL SIC decline or under what conditions they are unrelated. Based on the above consideration, in this paper we try to establish a linkage between the North Atlantic SIC or SST anomalies and cold anomalies over North America (NA) and northern Europe (NE) through examining the GB change related to the magnitude of meridional potential vorticity (PV) gradient. We address the following questions: 1) What is the physical link between the cold anomalies over NA or NE and North Atlantic SIC or SST anomalies? 2) Under what condition is the BDL SIC decline necessary for cold anomalies over NA? Or why do the cold anomalies over NA depend intermittently on the BDL SIC decline?

This paper is organized as follows: in section 2, we describe the data and method. In section 3, we examined the likely connections of cold anomalies over NA and NE to the meridional PV gradient in North Atlantic mid- to high latitudes on interannual and decadal trend time scales. The key role of the meridional PV gradient in the generation of continental cold anomalies is emphasized in section 4. We present a physical explanation for why the SIC or SST anomaly can influence cold anomalies over NA and NE in section 5. The conclusions and discussion are summarized in the final section.

2. Data and method

a. Data

We used daily ERA-Interim reanalysis data from December 1979/February 1980 to December 2016/February 2017 (1979–2016 hereafter) (<http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/>) for $2.5^{\circ} \times 2.5^{\circ}$ grids (Dee et al. 2011), which includes 500-hPa geopotential height (Z500), 500-hPa zonal wind (U500), surface air temperature (SAT), and potential vorticity (PV) on the 315-K isentropic level, which is close to the 200–300-hPa levels of the mid- to high-latitude troposphere. Using $2.5^{\circ} \times 2.5^{\circ}$ grid data is also reasonable because the blocking is a large-scale phenomenon. In fact, using higher-resolution data such as $0.75^{\circ} \times 0.75^{\circ}$ data does not significantly affect the obtained result (not shown). For the daily sea ice concentration and sea surface temperature, we also used the $1^{\circ} \times 1^{\circ}$ grid SIC and SST data from the ERA-Interim reanalysis data. The monthly mean SST we used here is the $1^{\circ} \times 1^{\circ}$ grid HadISST1 data taken from the Hadley Centre (<https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html>). In this study, the anomaly is defined as a deviation from the climatological mean field of each calendar day (1979–2016). The detrending method we applied here is a linear detrending during the time interval 1979–2016.

b. Blocking identification method

We use the one-dimensional (1D) blocking index developed by Tibaldi and Molteni (1990), here called the TM index, to identify individual blocking events over Greenland, although many two-dimensional blocking indices have been developed to examine the spatial distribution and climatological characteristics of the blocking action (Diao et al. 2006; Scherrer et al. 2006; Davini et al. 2012). Because our attention is mainly placed on finding blocking events over Greenland, it is reasonable to apply the 1D TM index to the calculation of Greenland blocking events.

The 1D TM index is defined in terms of the reversal of Z500 gradients: $\text{GHGN} = \{[Z500(\lambda, \phi_N) - Z500(\lambda, \phi_0)]/(\phi_N - \phi_0)\}$ and $\text{GHGS} = \{[Z500(\lambda, \phi_0) - Z500(\lambda, \phi_S)]/(\phi_0 - \phi_S)\}$, at given three 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 . We define a GB event to have taken place in a given region if the following two criteria are satisfied: 1) $\text{GHGS} > 0$; and 2) $\text{GHGN} < -10$ gpm per degree over at least a 15° span of contiguous longitude and three consecutive days (Luo et al. 2018).

Thompson and Wallace (2001) noted that the GB is more common during the negative NAO (NAO^-) phase. However, Luo et al. (2007) and Woollings et al. (2008) found from theoretical and observational aspects that the GB event is essentially identical to the NAO^- event in pattern and time scale with 10–20 days. Thus, in this paper we do not distinguish the different roles of GB and NAO^- events in the cold anomalies over NE and NA, and instead we only emphasize the contribution of the GB change to the variability of continental cold anomalies.

c. PV gradient thinking of the blocking system and its physical meaning

Motivated by the study of Luo et al. (2018, 2019), here we use the meridional PV gradient, $\text{PV}_y = \partial \text{PV} / \partial y$, instead of the PV anomaly used widely in previous studies (Hoskins et al. 1985) to characterize the blocking system. Note that $\text{PV} = -g(f + \mathbf{k} \cdot \nabla_\theta \times \mathbf{v})\partial\theta/\partial p$ is the same as that defined in Hoskins et al. (1985), where g is the gravity acceleration, $\mathbf{k}$ is a unit vertical vector, f is the Coriolis parameter, and ∇_θ is the three-dimensional gradient operator in the $xy\theta$ space, $\mathbf{v}$ is the three-dimensional wind vector, p is the pressure, and θ is the potential temperature. But in a barotropic atmosphere, PV can be approximated as $\text{PV} = \nabla^2\psi - \psi/R_d^2 + f$ (where ψ is the barotropic streamfunction and R_d is the radius of Rossby deformation) or its nondimensional form of $\text{PV} = \nabla^2\psi - F\psi + f$ [f is the nondimensional

Coriolis parameter, ψ is the nondimensional streamfunction, and $F = (\tilde{L}/R_d)^2$] when the characteristic velocity $\tilde{U} \sim 10 \text{ ms}^{-1}$ and length $\tilde{L} \sim 1000 \text{ km}$ are used. This PV is referred to as a quasigeostrophic barotropic PV as in Charney and DeVore (1979). In fact, because the GB is a strongly nonlinear system, its behavior can be characterized by its linear energy dispersion, as denoted by C_{gp} , and its nonlinearity strength, as denoted by δ , in the nonlinear multiscale interaction model of eddy-driven blocking based on the barotropic assumption as in Luo et al. (2018). They obtained $C_{\text{gp}} \propto \text{PV}_y$ and $\delta \propto (1/\text{PV}_y)$ for $\text{PV}_y \neq 0$ as the two most important parameters that influence the behavior of the subsequent blocking system. Such mathematical expressions are also suitable for studying the behavior of blocking related to Rossby wave trains. Specifically, Luo et al. (2018) found that the blocking system involves both weak energy dispersion and strong nonlinearity and hence maintains a long lifetime when the meridional PV gradient (PV_y) prior to the blocking onset is small, especially in high latitudes. In contrast, it has both strong energy dispersion and weak nonlinearity and hence is short-lived if PV_y is large. In parallel with the barotropic idea presented above, we may use the meridional gradient of the PV on the isentropic surface to reflect the behavior of blocking system, and it is found that the results are not sensitive to height level (not shown) because the blocking has an approximate barotropic structure. In this case, the magnitude of the DJF-mean PV_y on the 315-K isentropic surface may be used to classify the 1979–2016 winters and associated GB events. On this basis, the physical cause of why the cold anomalies over NE or NA or both may occur can be elucidated by observing the movement, position, and persistence of the GB and their causal linkages with the PV gradient, SIC, and SST in the North Atlantic. In this paper, we directly use the PV data from ERA-Interim to carry out our investigation.

3. Cold anomalies over North America and Europe and their links to SST and SIC anomalies in the North Atlantic

We first examine whether the cold anomalies over NA and NE are related to the SIC and SST anomalies in the North Atlantic before we establish a causal linkage between the continental cold anomalies and the meridional PV gradient. We investigate this problem from decadal, interannual, and subseasonal time scales. While the GB only exhibits a subseasonal variation and weakly affects the BDL SIC and North Atlantic SST as noted below, we infer that the SIC and SST anomalies can influence the trend and interannual variations of cold anomalies over NE and NA through influencing the

FIG. 1. (a) Linear trend of Z500 [contour interval (CI) = 10 gpm decade⁻¹, solid (dashed) lines indicate positive (negative) values] and surface air temperature (SAT; color shading) anomalies during 1990–2013, where the dots represent the region above the 95% confidence level for a two-sided Student's *t* test. (b) Time series of domain-averaged DJF-mean SAT anomalies over North America (30°–50°N, 120°–75°W; SAT_{NA}) and northern Europe (50°–70°N, 0°–50°E; SAT_{NE}).

trend and interannual variations of the GB in duration, movement, and position. To some extent, the SIC and SST anomalies can be considered as the background states of the GB event because they show a slow time variation. Although the linear trend of SAT during 1990–2013 is presented in the next subsection, in the main sections of this paper our emphasis is placed on examining the link of continental cold anomalies with the SIC or SST, PV_y, and GB changes on interannual and subseasonal time scales.

a. Weak warm Arctic–cold continent pattern over North America and North Europe

Figure 1a shows linear trends of DJF-mean Z500 and SAT anomalies during 1990–2013. We choose to consider the trend for 1990–2013 because the Arctic SIC has been observed to undergo a most rapid decline trend that accompanies a distinct cooling trend over East Asia during 1990–2013 (Cohen et al. 2014). The linear trend of SAT during 1990–2016 also shows a spatial pattern resembling that during 1990–2013, but with a smaller amplitude (not shown). Such a linear trend during 1990–2013 may be referred to as a decadal trend because its significance depends on the time interval. A significant cooling trend in East Asia is also seen to occur together

with Arctic warming in the Barents–Kara Seas, thus showing a warm Arctic–cold Eurasia (WACE) pattern (Overland et al. 2011; Cohen et al. 2014; Luo et al. 2016a; Shepherd 2016). This WACE pattern is related to the presence of persistent Ural blocking occurring together with a positive NAO (NAO⁺) (Luo et al. 2016a,b) and is modulated by the Atlantic multidecadal oscillation (AMO) through the SIC decline in the BKS (Luo et al. 2017). However, no distinct cooling trends are seen over NA and NE, even though there is a positive height anomaly trend of Z500 over Greenland (contour in Fig. 1a) (Yao et al. 2017). As we will reveal below, while the BDL sea ice loss contributes to the trends of cold anomalies over NA and NE, the North Atlantic midlatitude sea temperature anomaly seems to have an opposite effect.

b. Connections of cold anomalies over North America and Europe to the North Atlantic SIC and SST anomalies

It has been recognized that the Greenland blocking or NAO[−] is important for cold anomalies over Europe (Hurrell 1995; Thompson and Wallace 2001). Because the GB or NAO is modulated by the SST (Czaja and Marshall 2001) and SIC (Chen and Luo 2017) anomalies in the North Atlantic, we conclude that the cold

FIG. 2. Correlation patterns of detrended (a),(b) SIC and (c),(d) SST anomalies with the domain-averaged DJF-mean SAT (SAT_{NA} and SAT_{NE}) time series over (left) North America and (right) northern Europe. The dots represent the region above the 95% confidence level for a Monte Carlo test with 1000 simulations.

anomalies over NE and NA are likely related to the SST and SIC anomalies in the North Atlantic through the GB change. To understand how the cold anomalies over NA and NE are connected to North Atlantic SIC and SST anomalies, we show the correlation maps of SIC and SST anomalies with the domain-averaged DJF-mean SAT time series over NA and NE in Fig. 2 for detrended data. It is interesting to note that the SAT anomaly over NE shows a significant positive spatially correlation with the BDL SIC anomaly (Fig. 2b), and a significant negative (positive) spatially correlation with the high-latitude (midlatitude) North Atlantic SST anomaly north (south) of the Gulf Stream Extension (GSE) (Fig. 2d). However, the significant correlations of the SAT anomaly over NA with the BDL SIC and midlatitude North Atlantic SST anomalies cover much narrower regions than those of the SAT anomaly over NE (Figs. 2a,c). This suggests that the NE cold anomaly correlates more strongly with the BDL SIC and midlatitude North Atlantic SST anomalies than the NA cold anomaly does. In the next section, we will establish the linkages of cold anomalies over NE and NA with SIC and SST anomalies in the North Atlantic by constructing a meridional PV gradient index that unifies the SIC and SST into a single index.

c. Linkage of the meridional PV gradient with the SIC and SST anomalies in the North Atlantic

We show the time series of the domain-averaged DJF-mean SIC anomaly over the BDL (50° – 70° N, 90° – 30° W)

and SST anomaly over the midlatitude North Atlantic (35° – 50° N, 60° – 20° W; simply SST hereafter) in Fig. 3a for raw or nondetrended data and in Fig. 3b for detrended data. The BDL SIC (midlatitude North Atlantic SST) exhibits a distinct downward (upward) trend with -0.43 standard deviation (STD) decade $^{-1}$ (0.61 STD decade $^{-1}$) [significant at the 99% confidence level for a Student's t test or Mann–Kendall (MK) test] during 1979–2016, thus showing opposite trends (Fig. 3a). While the SIC and SST time series (Fig. 3a) have a negative correlation of -0.31 (not significant at the 90% confidence levels), their detrended time series (Fig. 3b) have almost no correlation (the correlation coefficient is 0.07) (Table 1). This suggests that the BDL SIC change is hardly related to the midlatitude SST south of the GSE. While the BDL SIC (midlatitude SST) exhibits a linear downward (upward) trend for raw data, the SAT anomalies over NA and NE show less evident trends. Although the domain-averaged SAT anomalies in the two continents have significant correlations with the SIC and SST for certain regions of BDL and North Atlantic midlatitudes, no significant domain-averaged correlations with SIC over the BDL and SST over North Atlantic midlatitudes are found. But we can detect a significant positive correlation between the NE cold anomaly and BDL SIC anomalies for detrended data (Table 1). This suggests that the SAT anomaly averaged over NA may be not directly linked to the BDL SIC or midlatitude SST changes, while the air temperature over NE may be linked to the BDL SIC anomaly on interannual time scales.

FIG. 3. (top) Time series of DJF-mean SIC anomalies (blue line) averaged over the BDL (50°–70°N, 90°–30°W) and SST anomaly (red line) over the midlatitude North Atlantic (35°–50°N and 60°–20°W) for (a) nondetrended and (b) detrended data during 1979–2016, and (bottom) regressed Z500 and SAT anomalies against the detrended DJF-mean (c) SIC (multiplied by -1.0) and (d) SST time series in (b). In (c) and (d), the solid (dashed) lines denote the positive (negative) anomalies and the dots represent the region above the 95% confidence level for the F test.

As we can further see from Table 1, the NA and NE SAT exhibit significant positive correlations with the DJF-mean PV gradient, although there are strong positive correlations between the PV gradient and SIC or SST for detrended data. Thus, it is inferred that the air temperatures over NE and NA are indirectly linked to the BDL SIC and midlatitude North Atlantic SST through the winter PV gradient change related to changes in the BDL SIC and midlatitude North Atlantic SST because the PV and SIC or SST show strong correlations. Such a linkage may partly result from GB changes in that the nonlinear behavior of the blocking is directly related to the magnitude of the meridional PV gradient (Luo et al. 2018). We show the regressed DJF-mean Z500 and SAT anomaly fields against the BDL SIC (multiplied by -1) and SST time series in Figs. 3c and 3d. It is noted that the BDL SIC decline corresponds to a strong GB pattern and a cold anomaly over NE

(Fig. 3c), but a cold anomaly is hardly seen in the midlatitude region of the eastern United States. The regressed Z500 field against the DJF-mean SST time series shows a NAO⁺ pattern with a negative height anomaly over Greenland, a warm anomaly over NE, and a weak cold anomaly over southern Europe and the Middle East (Fig. 3d). Also, a warm anomaly is seen over the NA. In other words, the intensified midlatitude SST warming south of GSE favors the NAO⁺ (Czaja and Marshall 2001) and hence a warm anomaly over NE (Hurrell 1995). Because the midlatitude SST warming corresponds to a warm anomaly over NE, the intensified midlatitude SST warming tends to weaken the NE cold anomaly caused by the BDL SIC decline. Thus, the opposite change of the NE SAT anomaly between the BDL SIC decline and midlatitude SST increase leads to a reduced cold anomaly over NE. This result also holds for raw or nondetrended data (not shown).

TABLE 1. Correlation coefficients of domain-averaged SAT_{NA} (30°–50°N, 120°–75°E), SAT_{NE} (50°–70°N, 0°–50°E), BDL SIC (50°–70°N, 90°–30°W), midlatitude SST (30°–50°N, 60°–20°W), and PV_y (45°–65°N, 80°–20°W) during 1979–2016 for nondetrended (detrended) DJF-mean time series by using the effective degree of freedom. The 95% and 99% confidence levels for a Student's t test are denoted by one and two asterisks, respectively.

Nondetrended (detrended)	SAT _{NE}	SIC	SST	PV _y
SAT _{NA}	0.20 (0.22)	$-0.09 (-5.09 \times 10^{-4})$	0.27 (0.21)	0.37* (0.36*)
SAT _{NE}		0.34 (0.58*)	0.38 (0.29)	0.33* (0.33*)
SIC			$-0.31 (0.07)$	0.58* (0.63**)
SST				0.27 (0.42*)

FIG. 4. Time-mean (a),(c) PV and (b),(d) PV_y anomalies (color shading) on the 315-K isentropic surface and Z500 anomalies (contour interval = 40 gpm) averaged (top) from lag -5 to 5 days and (middle) from lag -20 to -10 days prior to the blocking onset for GB events (lag 0 denotes the peak of the GB) during 1979–2016 identified with the TM index. (bottom) DJF-mean PV_y (color shading) on the 315-K isentropic surface with GB events (e) included and (f) excluded (blocking days from lag -10 to 10 days are removed). The solid (dashed) contours in (a) and (b) represent the positive (negative) anomaly of Z500.

As shown in Fig. 3c, the cold anomalies over NE are related to the presence of GB and its change. The westward movement of the GB can influence cold anomalies over NA (Chen and Luo 2017). Thus, it is useful to examine how the lifetime, movement, and position of the GB are changed and how they depend on the SIC and SST anomalies. As noted above, the small value of PV_y is crucial for the persistence of blocking. The SAT variations over NE and NA are likely linked to the SIC and SST anomalies through changes in the PV gradient and associated GB. We perform composites of daily Z500, PV, and PV_y anomalies for all GB events identified with the TM index and show the time-mean Z500, PV, and PV_y anomalies averaged from lag -5 to 5 days (lag 0 denotes the GB peak) in Figs. 4a and 4b. It is seen that the PV anomaly pattern presents a dipole structure with a negative (positive) anomaly within the anticyclonic (cyclonic) region of the GB (Fig. 4a). Correspondingly, during the GB episodes the meridional

PV gradient (PV_y) anomaly exhibits a tripolar structure with a negative anomaly region between the anticyclonic and cyclonic anomalies of the GB (Fig. 4b). Prior to the GB onset, the PV_y anomaly is negative over the North Atlantic between 40° and 60° N (Fig. 4d), even though the PV anomaly is weakly negative over Greenland and its adjacent region (Fig. 4c). In previous studies, many investigators have used the PV anomaly to characterize blocking (Hoskins et al. 1985). In fact, such a PV anomaly is only a manifestation of the blocking anomaly and cannot be regarded as a background condition of blocking onset and maintenance because it vanishes in the absence of blocking. But the background PV gradient can exist when the background PV has a difference between high latitudes and mid-latitudes, even though the blocking is not present. Thus, we can consider the PV gradient rather than the PV anomaly as a background condition of blocking. As noted below, the presence of this prior negative PV_y

anomaly in the North Atlantic mid- to high latitudes is attributed to the SIC and SST changes there because the SIC and SST without the effect of GB events have slower variations than the GB events.

Figures 4e and 4f show DJF-mean PV_y distributions with and without GB events. It is seen that the DJF-mean PV_y is lower in the mid- to high latitudes (45° – 60° N, 70° – 0° W) during the blocking episode (Fig. 4e) than during the nonblocking episode (Fig. 4f). The significant reduction of PV_y is mainly located in the south side of Greenland and can be seen from their difference field (not shown), which is similar to Fig. 4b. While the intensity of the DJF-mean PV_y is affected by the GB events (Figs. 4e,f), the basic spatial structure of the DJF-mean PV_y is not changed considerably. This can be further seen from a comparison between Figs. 4e and 4f. As we have noted above, the magnitude of PV_y prior to the blocking onset is important for the development of subsequent blocking. Crudely speaking, the DJF-mean PV_y may be considered as a background state of the blocking event because it is not significantly affected by individual GB events. Thus, we may use the magnitude of domain-averaged DJF-mean PV_y south of Greenland to measure changes in the GB in intensity, movement, and position. However, the causal linkage between the GB and background PV_y can be established by performing a daily field composite during GB events and examining the values of PV_y prior to the blocking onset. It is also noted that the most pronounced variation of the DJF-mean PV_y is concentrated between the anticyclonic and cyclonic centers of the GB (Fig. 4b). This motivates us to choose 45° – 65° N, 80° – 20° W as an average region of the domain-averaged DJF-mean PV_y to examine how the behavior of the GB is linked to the magnitude of DJF-mean PV_y . A similar result is also found even when we examine a larger area 45° – 75° N, 80° – 20° W that includes the positive anomaly of PV_y during the blocking episode as an average domain of PV_y (not shown). For this reason, in this paper we can use the value of DJF-mean PV_y averaged over the region 45° – 65° N, 80° – 20° W as an indicator to describe a linkage between the GB and winter PV gradient changes.

We show the time series of domain-averaged DJF-mean PV_y over the region 45° – 65° N, 80° – 20° W in Fig. 5a (red curve). It is clearly seen that the DJF-mean PV gradient shows no significant trend during 1979–2016 even during 1990–2013 (both their slopes are not significant). Although the winter-mean BDL SIC and midlatitude SST anomalies have no significant correlation (Figs. 3a,b), they likely contribute to the variation of the winter-mean PV gradient in North Atlantic mid- to high latitudes through changing mean air temperature and zonal wind distributions (Luo et al. 2018). Thus, the

FIG. 5. (a) Time series of domain-averaged DJF-mean PV_y over the region (45° – 65° N and 80° – 20° W) (red line) and its regressed curve (black line) based on $PV_y = \alpha_1 \text{SST} + \alpha_2 \text{SIC}$, $\alpha_1 = 0.47$, and $\alpha_2 = 0.66$, where the red dotted (dashed) line denotes the linear trend of PV_y from 1979 to 2016 (1990–2013) and the value of p represents its probability value for the Mann-Kendall trend test; and the regressed DJF-mean PV_y anomaly against the domain-averaged DJF-mean (b) SIC (multiplied by -1.0) and (c) SST time series in Fig. 3b. The dots denote the same as that in Fig. 3.

PV gradient in the North Atlantic mid- to high latitudes depends not only on the BDL SIC, but also on the midlatitude or high-latitude SST anomaly. They have significant correlations besides the correlation between the PV gradient and midlatitude SST anomaly being insignificant for raw data (Table 1). To examine the contributions of the winter BDL SIC and midlatitude SST anomalies to the change in winter PV gradient, one needs to estimate the values of α_1 and α_2 through using the multiple linear regression method when $PV_y = \alpha_1 \text{SST} + \alpha_2 \text{SIC}$ is assumed (Wilks 2011). The magnitudes of the regression coefficients α_1 and α_2 reflect the relative roles of midlatitude SST and BDL SIC in the PV gradient change. We find that $\alpha_1 = 0.47$ and $\alpha_2 = 0.66$

FIG. 6. Number of low (blue) and high (red) PV_y , SST, and SIC winters; low PV_y winters with low SST (orange); low PV_y winters with nonlow SST (yellow); low PV_y winters with low SIC (dark green); low PV_y winters with nonlow SIC (light green); high PV_y winters with high SST (dark purple); high PV_y winters with nonhigh SST (light purple); high PV_y winters with high SIC (black); and high PV_y winters with nonhigh SIC (gray) during 1979–2016.

during 1979–2016. The calculation shows that the regressed PV_y time series has a positive correlation of 0.68 ($p < 0.01$) with the reanalysis data result in Fig. 5a (red curve). Thus, the effects of BDL SIC and midlatitude SST can be unified into this PV gradient index. This index serves as a useful index for examining how the winter BDL SIC and midlatitude North Atlantic SST changes combine to influence extreme cold events over NE and NA. We show the regressed fields of DJF-mean PV gradient anomalies against the DJF-mean BDL SIC and midlatitude SST time series in Figs. 5b and 5c. It is seen that while the PV gradient in the North Atlantic mid- to high latitudes (45° – 65° N) can be reduced by the BDL SIC decline, it can be intensified by the midlatitude SST warming. Thus, it is likely that the less distinct trend of the PV gradient in the recent decades can be attributed to the opposing changes of the PV gradient between the BDL SIC and midlatitude SST anomalies.

4. Key role of the winter PV gradient in continental cold anomalies

Here, we define the normalized value of detrended domain-averaged DJF-mean PV_y above (below) 0.5

(−0.5) STD as a high (low) PV_y winter. Similar definitions can be made for the normalized time series of domain-averaged DJF-mean BDL SIC and midlatitude SST anomalies. It is easily found that there are 13 low (13 high) PV_y , 11 low (14 high) SST, and 9 low (11 high) SIC winters during 1979–2016 (Fig. 6).

To understand the roles of the different combinations among PV_y , SIC, and SST in the cold anomalies over NE and NA, PV_y , SIC, and SST winters can be further classified into eight combinations (Fig. 6). We find 6 low PV_y winters (below −0.5 STD) with low SST (below −0.5 STD), 7 low PV_y winters with nonlow SST (above −0.5 STD), 6 low PV_y winters with low SIC, 7 low PV_y winters with nonlow SIC, 6 high PV_y winters (above 0.5 STD) with high SST (above 0.5 STD), 7 high PV_y winters with nonhigh SST (below 0.5 STD), 6 high PV_y winters with high SIC, and 7 high PV_y winters with nonhigh SST during 1979–2016. In these classifications, some winters of small or large PV gradient are overlapped because the magnitude of the PV gradient depends not only on the BDL SIC, but also on the midlatitude North Atlantic SST, although the BDL SIC change is also related to the SST in the North Atlantic high latitudes.

FIG. 7. Composite fields of DJF-mean (a),(b) SIC, (c),(d) SST, (e),(f) PV_y , and (g),(h) Z500 [CI = 20 gpm, the solid (dashed) lines denote the positive (negative) anomaly] and SAT (color shading) anomalies for (left) 6 low PV_y winters (below -0.5 STD) with low midlatitude SST (below -0.5 STD) and (right) 6 low PV_y winters with low SIC (below -0.5 STD) during 1979–2016. The dots represent the area above the 95% confidence level based on a 5000 simulation Monte Carlo test.

a. Do the cold anomalies over North America and North Europe need large SIC decline?

We first show the composite DJF-mean SIC, SST, PV_y , and Z500 and SAT anomalies for low PV_y winters with low SST and low SIC in Fig. 7. It is apparent that when a strong SST cooling appears over North Atlantic mid- to high latitudes across the GSE (Fig. 7c), strong cold anomalies can be seen over NA and northern Eurasia

including NE (Fig. 7g) because of the presence of a winter-mean GB resembling a NAO^- , even if the negative BDL SIC anomaly is not large (Fig. 7a). We also find that the high SST anomaly is weakened in the south of GSE and its warming center is mainly located in the Gulf Stream region (Fig. 7c). For this case, the SST cooling is intense and covers the whole North Atlantic mid- to high latitudes. Although the BDL SIC decline is less strong (Fig. 7a), the meridional PV gradient may be small

FIG. 8. As in Fig. 7, but for (left) 7 low PV_y winters with nonlow midlatitude SST (above -0.5 STD) and (right) 7 low PV_y winters with nonlow SIC (above -0.5 STD).

(Fig. 7e) as a result of mid- to high-latitude SST cooling. However, when the warm SST anomaly is intensified and shifted northward into the Labrador Sea (Fig. 7d), the high SST anomaly corresponds to a large BDL SIC decline (Fig. 7b) and a small meridional PV gradient (Fig. 7f). Thus, a winter-mean GB anomaly resembling a NAO^- and cold anomalies over NE and NA are still seen in this case (Fig. 7h). In Figs. 7g and 7h, a warm west/cold east SAT dipole pattern appears over the North American continent. Thus, one can see a warm west (near California and Arizona) and cold east (mainly over

eastern parts of United States) temperature dipole anomaly when the BDL SIC is low or when the mid-latitude SST cooling is strong. In other words, the North Atlantic change is also able to result in the warm west/cold east temperature dipole over the North America. This result is different from previous findings of Wang et al. (2014), Lee et al. (2015), and Hartmann (2015), who emphasized the role of the Pacific basin in the North American warm west/cold east temperature dipole.

For low PV_y winters with nonlow midlatitude SST, the results in Figs. 8a, 8c, 8e, and 8g are found to be similar to

those for low PV_y winters with low SIC (Figs. 7b,d,f,h). We also see that while the PV gradient is small for low PV_y winters with nonlow SIC (Fig. 8f), it is larger than that for low PV_y winters with low SIC because the mid- to high-latitude SST cooling (Fig. 8d) and BDL SIC anomaly (Fig. 8b) are weak. Thus, a relatively weak cold anomaly can be seen over NA for this case (Fig. 8h). The above results suggest that while the large negative SIC anomaly in BDL favors the cold anomaly over the NA, it is not a necessary condition for the NA cold anomaly to occur. This result is different from the finding of Chen and Luo (2017). Specifically, when there is a strong and widespread SST cooling in the North Atlantic mid- to high latitudes, a strong cold anomaly can still be seen over NA even in the absence of a large negative BDL SIC anomaly.

In the barotropic atmosphere, because there is $PV_y \approx \beta - U_{yy} + FU$ for a slowly varying zonal flow U , the sea ice and SST anomalies can modulate PV_y through changing the meridional background temperature gradient and associated background westerly wind U . This mechanism can be applied to the North Atlantic basin and other regions.

b. Key role of the magnitude of the meridional PV gradient

For a comparison, it is useful to show the composite DJF-mean SIC, SST, PV_y , and Z500 and SAT anomalies in Fig. 9 for high PV_y winters with high SST or SIC. It is seen that when the BDL SIC anomaly is positive (Figs. 9a,b) or when the positive midlatitude SST anomaly south of the GSE is strong (Figs. 9c,d), the meridional PV gradient is large in the North Atlantic mid- to high latitudes (Figs. 9e,f) and corresponds to a NAO^+ pattern with warm anomalies over NA and NE (Figs. 9g,h). Similar results are found for high PV_y winters with nonhigh SST (Figs. 10a,c,e,g) or nonhigh SIC (Figs. 10b,d,f,h). The above results clearly indicate that the small PV gradient in the North Atlantic mid- to high latitudes is crucial for whether cold anomalies appear over NA and NE, while it is closely related to the BDL SIC and midlatitude SST anomalies.

5. Generation mechanisms of continental cold anomalies

a. Links of the GB and cold anomaly changes to the magnitude of PV gradient

The above results indicate that the cold anomalies over NA and NE are closely linked to the magnitude of the DJF-mean PV gradient in the North Atlantic mid- to high latitudes associated with the BDL SIC and midlatitude SST anomalies. However, they do not allow us to infer a

causal link. Below, this problem will be examined in detail based on the evolution of a composite daily field.

To understand the basic physics of the small background PV gradient affecting the cold anomalies over NA and NE, it is useful to examine the relationship between the GB change and the magnitude of the PV gradient from a daily evolution perspective. We find that during 1979–2016 there are 53 GB events identified with the TM index, in which the high (low) PV_y winters correspond to 11 (25) GB events. Thus, there is a considerable increase ($\sim 127\%$) in the number of GB events as PV_y changes from a high to a low value. As we will further see below, the position, movement, and persistence of the GB are significantly related to the magnitude of the background PV_y related to the slow SIC and SST changes over a longer period than the time scale of GB.

We show the time-mean composite Z500 and SAT, PV_y , SST, and SIC anomalies averaged from lag -5 to 5 days (lag 0 denotes the day of the GB peak) for GB events in the high and low PV_y winters in Fig. 11. It is noted that the anomaly center of GB is located in the south of Greenland (near $60^\circ N$) for the high PV_y winter (Fig. 11a), but is mainly found at higher latitudes of Greenland near $65^\circ N$ for the low PV_y winter (Fig. 11b). In other words, the small (large) PV gradient favors the generation of a high-latitude (low-latitude) GB. Because the high-latitude GB is long-lived and moves westward in the low PV_y winter (Fig. 11j), it can produce strong cold anomalies to influence the NE and NA (Fig. 11b). But because the low-latitude GB is short-lived and quasi-stationary in the high PV_y winter (Fig. 11i), it has a weak effect on NA and NE (Fig. 11a).

The movement of the GB can crudely be explained in terms of the nonlinear phase speed of blocking as $C_{NP} = U - PV_y / (k^2 + m^2 + F) - \delta_N M_0^2 / (2kPV_y)$ in a barotropic atmosphere (Luo et al. 2019) according to Luo (2000), where $\delta_N > 0$ and is a constant [δ_N can be found in the appendix of Luo et al. (2018)], $PV_y = \partial \xi_a / \partial y = \xi_{ay}$ (ξ_a is the absolute vorticity or barotropic PV), $F \approx 1$, k and m are the nondimensional zonal and meridional wavenumbers of the GB, respectively, U is the nondimensional U500, and M_0 is the amplitude of the GB whose 500-hPa streamfunction anomaly is scaled with $2\tilde{U}\tilde{L}\sqrt{2/L_y}$, where $L_y = 5$. Note that in the mathematical expression of C_{NP} , ξ_a is the nondimensional 500-hPa absolute vorticity scaled with $\tilde{U}/\tilde{L}$, where $\tilde{U} = 10 \text{ m s}^{-1}$ and $\tilde{L} = 10^6 \text{ m}$. Although C_{NP} was obtained based on the barotropic model, we may use C_{NP} to estimate the movement of the GB.

We also see that during the GB episode the time-mean (from lag -5 to 5 days) PV_y south of Greenland is smaller in the low- PV_y winter (Fig. 11d) than in the high- PV_y winter (Fig. 11c). The negative BDL SIC

FIG. 9. As in Fig. 7, but for (left) 6 high PV_y winters (above 0.5 STD) with high midlatitude SST (above 0.5 STD) and (right) 6 high PV_y winters with high SIC (above 0.5 STD).

anomaly is large (Fig. 11f) or correspondingly there is a midlatitude SST cooling south of the GSE (Fig. 11h). The reverse is found for the high- PV_y winter (Fig. 11e or Fig. 11g). While the PV_y is small during the mature phase (from lag -5 to 5 days) of the GB in the high- PV_y winter, it does not mean that PV_y is small prior to the blocking onset. To interpret how the GB moves with the magnitude of PV_y , we calculate C_{NP} to examine this problem. In fact, U and PV_y vary with time as a response to the generation of the GB. In this case, we may calculate the large-scale zonal wind (U500) and associated

PV_y in C_{NP} during the GB life cycle to examine how the GB moves with time in the zonal direction. We first show the time-mean U500 anomalies averaged from lag -5 to 5 days and from lag -20 to -10 days in Figs. 12a–d and nondimensional U500, PV_y and M_0^2/PV_y for GB events in high and low PV_y winters in Figs. 12e–g. It is found that the zonal wind within the anticyclonic region of the GB is stronger in the high- PV_y winter (Fig. 12a) than in the low- PV_y winter (Fig. 12b), even though its prior zonal wind south of Greenland is also intense (Figs. 12c,d). While M_0^2/PV_y (PV_y) of the GB is larger

FIG. 10. As in Fig. 7, but for (left) 7 high PV_y winters (above 0.5 STD) with nonhigh midlatitude SST (below 0.5 STD) and (right) 7 high PV_y winters with nonhigh SIC (below 0.5 STD).

(smaller) in the low- PV_y winter (blue line in Figs. 12f and 12g) than in the high- PV_y winter (red line in Figs. 12f and 12g), their difference is not significant during the mature phase of the GB. This suggests that the westward movement of the GB in the low- PV_y winter (its less movement in the high- PV_y winter) is more likely due to the weakening (strengthening) of zonal winds or a large (small) increase of M_0^2/PV_y within the anticyclonic region of the GB or over Greenland.

We show time-mean composite Z500 and SAT, PV_y , SST, and SIC anomalies averaged from lag -20 to -10 days

prior to the blocking onset in Fig. 13 for GB events in the high- and low- PV_y winters. The time-mean prior PV_y , SST and SIC anomalies may be considered as the background condition of GB events. It is found that while PV_y is small during the mature period of GB in the high- PV_y winter (Fig. 13c), it is large prior to the GB onset (Fig. 13c) because the prior BDL SIC anomaly is largely positive (Fig. 13e) or because the prior SST anomaly shows a low-over-high dipole with a strong warm anomaly south of the GSE (Fig. 13g). Correspondingly, the prior height anomaly shows a NAO^+ pattern. To some extent, it is believed

FIG. 11. Time-mean composite daily (a),(b) Z500 [$CI = 40$ gpm; solid (dashed) lines denote positive (negative) anomalies] and SAT (color shading), (c),(d) PV_y , (e),(f) SIC, and (g),(h) SST anomalies averaged from lag -5 to 5 days for GB events in (left) high (11 GB events) and (right) low PV_y (25 GB events) winters during 1979–2016. Also shown is the time–longitude evolution of Z500 anomalies averaged over 50° – 70° N for (i) high PV_y winter and over 60° – 80° N for (j) low PV_y winter. The thick black line indicates contour with 60 gpm and the dot represents the area above the 95% confidence level for a Monte Carlo test based on 5000 simulations.

FIG. 12. Time-mean fields of composite daily U500 anomalies averaged (a),(b) during the mature period from lag -5 to 5 days and (c),(d) during the prior period from lag -20 to -10 days of GB events for high- PV_y winters in (a) and (c) and low- PV_y winters in (b) and (d). Time series of composite daily nondimensional (e) U500, (f) PV_y , and (g) M_0^2/PV_y for PV_y and M_0 averaged over 50° – 70° N and 80° – 20° W in high- PV_y winters (red line) and over 60° – 80° N and 80° – 20° W in low- PV_y winters (blue line), where M_0 denotes the nondimensional domain-averaged Z500 anomaly. In (a)–(d), the dots have the same meaning defined as in Fig. 7. In (e)–(g), the gray shading denotes the difference between two lines is above the 95% confidence level for a Monte Carlo test with 5000 simulations.

that the NAO^+ or large positive BDL SIC anomaly or midlatitude warming south of the GSE prior to the GB onset is a precursor of a short-lived and quasi-stationary GB through strengthening the prior PV gradient. In the low- PV_y winter, we see that prior to the GB onset the PV gradient is small (Fig. 13d), the BDL SIC decline is large (Fig. 13f), and the midlatitude SST anomaly south of the GSE is cold (Fig. 13h) as opposed to those in the high- PV_y winter. For this case, the prior height anomaly shows a NAO^- pattern. Thus, the NAO^- pattern associated with the large negative BDL SIC anomaly or midlatitude SST

cold anomaly prior to the GB onset is a precursor of long-lived and retrograde GB by reducing the meridional PV gradient.

b. Causal linkages of the PV gradient and associated GB with the SIC and SST anomalies

To strengthen our understanding of the causal link among the GB, PV_y , SIC, and SST, it is useful to show the time series of domain-averaged composite daily Z500, PV_y , SIC, and SST during the GB episode in Fig. 14. It is seen that the GB is long-lived (short-lived)

FIG. 13. Time-mean composite daily (a),(b) Z500 [$CI = 40$ gpm, the solid (dashed) lines denote the positive (negative) anomaly] and SAT (color shading), (c),(d) PV_y , (e),(f) SIC, and (g),(h) SST anomalies averaged from lag -20 to -10 days for GB events in (left) high- PV_y (11 GB events) and (right) low- PV_y (25 GB events) winters during 1979–2016. The dots represent the area above the 95% confidence level for a Monte Carlo test based on 5000 simulations.

in the small (large) PV gradient winter (Fig. 14a). A large change in PV_y is seen in Fig. 14b as the GB grows and decays (Fig. 14a). However, PV_y prior to the GB onset is smaller in the low- PV_y winter than in the high- PV_y winter (Fig. 14b). Thus, a small (large) prior PV_y is a favorable background condition of the long-lived (short-lived) GB. On the other hand, we see that the BDL SIC and midlatitude SST anomalies do not significantly change with the GB evolution (Figs. 14c,d). In other

words, the responses of SIC and SST to the presence of GB are so weak that the winter mean SIC and SST depend not strongly on the occurrence or evolution of GB (Figs. 14c,d). To some extent, the magnitude of the prior PV_y is closely associated with the background SIC and SST over the North Atlantic in winter. During the prior period of GB (from lag -20 to -10 days), the negative anomaly of BDL SIC or cold midlatitude SST anomaly tends to reduce the PV gradient to favor the persistence

FIG. 14. Time series of daily (a) Z500 anomalies averaged over the regions 50° – 70° N, 90° – 30° W (red line) and 60° – 80° N, 90° – 30° W (blue line), (b) PV_y anomaly averaged over 45° – 65° N, 80° – 20° W, (c) SIC anomaly averaged over 50° – 70° N, 90° – 30° W, and (d) SST anomaly averaged over the midlatitude North Atlantic (35° – 50° N, 60° – 20° W) for GB events in high- PV_y (red line) and low- PV_y (blue line) winters. Gray shading denotes the difference of the GB amplitude, PV_y , SIC, and SST between high- and low- PV_y winters being significant at the 95% confidence level for a Monte Carlo test based on 5000 simulations.

of subsequent GB by producing a positive height anomaly over Greenland and a weakened zonal wind south of Greenland. Thus, we infer that the magnitude of the PV gradient prior to the GB onset is mainly dominated by the BDL SIC and midlatitude SST anomalies prior to the GB onset. This can be explained in terms of the zonal wind change related to the meridional air temperature gradient prior to the blocking onset (Fig. S2 in the online supplemental material). We also find that a long-lived GB (blue line in Fig. 14a) requires a small prior PV gradient (blue line in Fig. 14b), which is related to the large negative anomaly of BDL SIC (blue line in Fig. 14c) or cold midlatitude SST anomaly (blue line in Fig. 14d) prior to the GB onset. In fact, the large BDL SIC decline or cold midlatitude SST anomaly can correspond to a reduced mean westerly wind in the North Atlantic mid- to high latitudes because of reduced meridional temperature gradient (not shown). Thus, the prior large BDL SIC decline or strong prior midlatitude SST cooling south of the GSE can maintain the subsequent GB for longer lifetimes and induce its westward movement through reducing the PV gradient. This may explain why the cold anomalies over NA depend intermittently on the SIC decline.

To further understand how the cold anomalies over NA and NE are modulated by the DJF-mean PV_y , BDL SIC, and midlatitude SST anomalies through the GB change, it is useful to perform composites of daily Z500

and SAT anomalies during GB episode for various combinations of low- and high- PV_y winters for different BDL SIC and midlatitude SST anomalies. Figure 15 shows the time-mean composite daily Z500 and SAT anomalies averaged from lag -10 to 10 days of GB events for eight combinations: low- PV_y winters with low SST or low SIC (Fig. 15a or Fig. 15b), low- PV_y winters with nonlow SST or nonlow SIC (Fig. 15c or Fig. 15d), high- PV_y winters with high SST or high SIC (Fig. 15e or Fig. 15f), and high- PV_y winters with nonhigh SST or nonhigh SIC (Fig. 15g or Fig. 15h). The time-longitude variations of the latitude-averaged Z500 anomalies of the corresponding GB events are shown in Fig. 16. It is noted that the GB is located in the high latitudes near Greenland (Figs. 15a–d) as PV_y is small, but in the lower latitudes south of Greenland as the PV gradient is large (Figs. 15e–h). The cold anomalies are seen over NE and NA (Figs. 15a–d) because the GB is long-lived and shows a westward movement (Figs. 16a–d) when PV_y is small. However, less evident cold anomalies are observed over NE and NA (Figs. 15e–h) as PV_y is large, because the GB is short-lived and located in the south of Greenland, and shows less evident retrogression (Figs. 16e–h).

c. Precursors of the different GB patterns affecting the different continental cold anomalies

While the cold anomalies can appear over NE and NA (Figs. 15a–d) for the DJF-mean PV_y being small, the

FIG. 15. Time-mean 500-hPa geopotential height [$CI = 20$ gpm, the solid (dashed) lines denote the positive (negative) anomaly] and SAT anomalies averaged from lag -10 to 10 days of GB events for (a) low- PV_y winter with low SST (12 GB events); (b) low- PV_y winter with low SIC (13 GB events); (c) low- PV_y winter with nonlow SST (13 GB events); (d) low- PV_y winter with nonlow SIC (12 GB events); (e) high- PV_y winter with high SST (5 GB events); (f) high- PV_y winter with high SIC (6 GB events); (g) high- PV_y winter with nonhigh SST (6 GB events); and (h) high- PV_y winter with nonhigh SIC (5 GB events). The dots have the same meaning as in Fig. 14.

North American cold anomalies show different spatial patterns. They are mainly classified into northwest-southeast (NW-SE) and midlatitude types. The NW-SE type presents a strong cold anomaly over NA from high latitudes to midlatitudes along the NW-SE direction and a relatively weak cold anomaly over NE, which corresponds to a NW-SE-oriented GB dipole pattern (Figs. 15b,c). The midlatitude-type cold anomaly is relatively weak and mainly located over the North

American midlatitudes (Figs. 15a,d). A strong cold anomaly appears over northern Eurasia as well. Thus, we can infer that these different cold anomaly patterns have different precursors, although they are caused by the GB patterns under the small PV gradient conditions. To identify the precursors of the two types of cold anomalies, it is useful to show the time-mean composite daily PV_y , $U500$, SIC, and SST anomalies averaged during the prior period of the GB events from lag -20

FIG. 16. Time-longitude evolution of composite daily Z500 anomalies for GB events averaged over 60° – 80° N in the low PV_y winters with (a) low SST, (b) low SIC, (c) nonlow SST, and (d) nonlow SIC and over 50° – 70° N in the high- PV_y winters with (e) high SST, (f) high SIC, (g) nonhigh SST, and (h) nonhigh SIC. The black contour indicates 60 gpm, and the meaning of the dots is the same as in Fig. 14.

to -10 days (see Figs. S1–S4) for the same classifications as in Fig. 15.

It is noted that while the PV gradient in the North Atlantic mid- to high latitudes (Figs. S1a–d) is small prior to the GB onset in the low-PV_y winters, the GBs have different spatial patterns (Figs. 15a–d). In general, the spatial shape of the large-scale height anomaly is mainly dominated by the meridional distribution of background zonal winds (Luo et al. 2010). Thus, it is likely that the spatial structures of the GB patterns are influenced by the spatial distributions of the prior zonal winds (Figs. S2a–d). Below, we will explain why the GB dipoles show a NW–SE orientation in Figs. 15b and 15c, but a northeast–southwest (NE–SW) orientation in Figs. 15a and 15d. Because the SIC anomaly prior to the GB onset is largely negative in the BDL as shown in Figs. S3c and S2b, the prior high-latitude North Atlantic zonal wind is inevitably weakened due to the prior BDL warming (Chen and Luo 2017). Such a high-latitude zonal wind change can produce a horizontal orientation of the subsequent GB dipole as seen in Figs. 15a–d. Because the anticyclonic center of the GB in Fig. 15b or Fig. 15c is located at 50° – 80° N with the cyclonic center at 30° – 45° N, the weakening (strengthening) of the North Atlantic mean zonal winds over the region 45° – 65° N (30° – 45° N) in Figs. S2b and S2c can lead to the NW–SE orientation of the GB dipole. We also see that the SIC has a large negative anomaly in the BDL and a positive anomaly in the BKS, thus showing an evident east–west seesaw between the BDL and BKS (Figs. S3b,c). Associated with the large BDL SIC decline (Figs. S3b,c), the SST warming covers a broad North Atlantic region and extends to the BDL (Figs. S4b,c). Such an SST distribution corresponds to a weakened midlatitude warming south of the GSE. Because the NW–SE-oriented GB dipole pattern corresponds to a strong NW–SE-type cold anomaly over NA, the large prior BDL decline associated with the strong positive SST anomaly in the North Atlantic mid- to high latitudes is likely a precursor of the generation of strong North American cold anomalies along the NW–SE direction.

On the other hand, we see that because the anticyclonic (cyclonic) center of the GB dipole is located at 55° – 80° N (35° – 50° N), the weakening of the North Atlantic zonal winds between 45° and 65° N spurs the NE–SW orientation of the GB dipole as seen in Figs. 15a and 15d. This NE–SW-oriented GB dipole corresponds to relatively weak midlatitude cold anomaly over NA (color shading in Figs. 15a and 15d). In this case, the SIC decline is limited in the BDL but large in the BKS. Associated with this BDL SIC decline, the midlatitude SST warming south of the GSE is weakened and a cold SST anomaly intensifies and extends to high latitudes

such that a strong widespread cooling appears in the North Atlantic mid- to high latitudes (Figs. S4a,d). Thus, the strong prior North Atlantic mid- to high-latitude SST cooling (Fig. 7c) is a precursor of the midlatitude cold anomaly over NA (Fig. 7g). Such strong mid- to high-latitude North Atlantic cooling prior to the GB onset can lead to a small prior PV gradient even if the large BDL SIC decline is absent (Fig. 7a). While the cold anomalies can occur over NA, they are different between both forced by the large BDL SIC decline and North Atlantic mid- to high-latitude cooling across the GSE. In other words, the NW–SE and midlatitude-type cold anomalies over NA require different precursors, although the PV gradients are small in both cases. It is further revealed that the small PV gradient plays a key role in the increased persistence of GB, whereas the movement and horizontal tilting of the GB is dominated by its meridional position and the meridional distribution of the prior North Atlantic mid- to high-latitude zonal winds.

As mentioned above, the cold anomaly mainly appears over western and southern Europe as the PV gradient is large in the North Atlantic mid- to high latitudes, whereas the North American cold anomaly almost disappears. Here, we will further examine what precursors dominate the disappearance of North American cold anomalies. While the GB can still occur in the large prior PV gradient regions (Figs. S1e–h), the GB is infrequent, short-lived, and immobile or eastward-moving (Figs. 16e–h). Thus, it mainly influences Europe in this case. We also see that the large prior PV gradient in the North Atlantic mid- to high latitudes is closely related to the high prior SST anomaly south of the GSE, although the SST anomaly shows a tripolar structure in Figs. S4f and S4g and a basin warming with a strong center located in the south of the GSE in Figs. S4e and S4h. Such SST distribution corresponds to a prior NAO⁺ state (Fig. 13a) with large prior PV gradient (Figs. S1e–h) that is associated with an intensified prior mid- to high-latitude zonal wind (Figs. S2e–h) or a large positive prior BDL SIC anomaly even a small BDL SIC decline (Figs. S3e–h). Hence, it is inferred that the strong midlatitude warming south of the GSE or small BDL SIC decline (or large positive BDL SIC anomaly) is a precursor of the disappearance of North American cold anomalies, although the SST anomaly shows a tripolar structure across the GSE or a wide basin warming.

d. Linear trends of Greenland blocking and PV gradient and their relationship

While the cold anomaly shows less distinct trends over NA and NE, it is useful to examine whether the PV

gradient in the North Atlantic mid- to high latitudes and GB have pronounced trends. Before examining this problem, we first calculate the DJF-mean PV_y averaged over the region 45° – 65° N and 80° – 20° W for GB events excluded (blocking days from lag -10 to 10 days are removed) and show its result in Fig. 17a. The calculation reveals that the DJF-mean PV_y for GB events excluded has a significant positive correlation of 0.81 ($p < 0.01$) with that for GB events included. This means that the main portion of the winter PV_y change is not attributed to the presence of GB events, but more likely attributed to the combined BDL SIC and North Atlantic SST changes prior to the GB onset through background westerly wind changes. That is to say, the magnitude of the DJF-mean PV_y provides a background condition for the GB change. It is also noted that the DJF-mean PV_y shows a downward trend which is not significant at the 90% confidence level for GB events excluded.

We also calculate the durations of individual GB events in terms of the values of domain-averaged Z500 anomalies and show corresponding results in Figs. 17b and 17c. Here, we identify the duration of the GB event as the number of days during which the daily Z500 anomalies averaged over the region 55° – 75° N, 90° W– 0° exceed 80 or 60 gpm. It is seen that while the GB duration has no clear trend during 1979–2016 but an increased trend during 1990–2013. This increasing trend has a slope of 2.17 days decade $^{-1}$ and is not significant at the 90% confidence level for the 80-gpm threshold, but is significant ($p < 0.1$) for the 60-gpm threshold. Thus, we infer that the weak increased trend of the GB duration during 1990–2013 is very likely due to the weak downward trend of the PV gradient in the North Atlantic mid- to high latitudes, while the magnitude of the PV gradient is influenced by the GB and mediated by the opposite change between the BDL SIC and North Atlantic midlatitude SST anomalies.

6. Conclusions and discussion

In this paper, we have proposed a new linkage pathway through which cold anomalies over NA and NE are linked to the BDL SIC and North Atlantic SST changes. In this linkage, the magnitude of meridional PV gradient in the North Atlantic mid- to high latitudes is crucial for whether cold anomalies occur over NA or NE or both, while it is related to the BDL SIC and North Atlantic SST changes. This PV gradient index reflects the combined effect of the BDL SIC and North Atlantic SST anomalies on cold anomalies over NE and NA through its impact on the GB change in terms of position, movement, and persistence.

FIG. 17. (a) Time series of domain-averaged DJF-mean PV_y for GB events included (red) and excluded (blue), where the excluded case represents the removal of blocking days from lag -10 to lag 10 days. Time series of the durations (dot) of GB events during 1979–2016 for the (b) 80- and (c) 60-gpm thresholds, where the dots denote the duration of each GB event in low (blue), high (red), and neutral (green) PV_y winters (the neutral PV_y is defined to be between -0.5 and 0.5 STDs). The dashed line denotes the linear trend of the PV gradient in (a) and the average duration of GB events in (b) and (c) during 1990–2013.

It is shown that the GB is more frequent in the small PV gradient winter than in the large winter. When the PV gradient is large, the GB shows less movement, and is short-lived and mainly located in the south of Greenland. This GB cannot produce a strong North American cold anomaly, but can lead to a cold anomaly over the south and west of Europe. We find that the NAO^+ with a large PV gradient in the North Atlantic mid- to high latitudes is a precursor of short-lived GB and closely related to a tripolar North Atlantic SST pattern with

strong midlatitude SST warming south of the GSE or large positive BDL SIC anomaly. However, a long-lived and retrograde GB is easily observed when the PV gradient is small, which occurs in higher latitudes over Greenland. Strong cold anomalies are seen to occur over NE and NA for this type of GB. A North American warm west/cold east temperature dipole is also easily seen in this case. It is further found that while the NAO[−] with a small PV gradient is related to the midlatitude SST cooling south of the GSE and negative BDL SIC anomaly, it is a prior condition of the long-lived and retrograde GB.

By examining composite daily fields, we found that the cold anomaly can still occur over NA even in the absence of large negative BDL SIC anomaly when the GB occurs in a small PV gradient region. While an intense cold anomaly can occur over NA for a small prior PV gradient, the spatial pattern of the NA cold anomaly may be different, which depends on the extent of the BDL SIC decline prior to the GB onset. A strong cold anomaly is seen in the mid- to high latitudes of NA along the NW–SE direction as the negative BDL SIC anomaly prior to the GB onset is large and comprises an east–west seesaw with the positive BKS SIC anomaly. Such a NA cold anomaly requires the presence of a prior midlatitude SST cooling south of the GSE and a large positive SST anomaly over North Atlantic mid- to high latitudes. However, when the midlatitude SST warming is weakened so that a strong and widespread SST cooling dominates the North Atlantic mid- to high latitudes, the cold anomaly mainly occurs in the midlatitudes of NA because of the NE–SW orientation of GB. Such a midlatitude cold anomaly does not need a large prior BDL SIC decline. Instead, a large North Atlantic basin SST cooling from high latitudes to midlatitudes is required as a prior condition of the GB for this case. The above results suggest that the large negative BDL SIC anomaly or strong mid- to high-latitude North Atlantic SST cooling can significantly reduce the PV gradient in North Atlantic mid- to high latitudes to cause the lengthening and retrogression of GB and then produce intense cold anomalies over NA. In particular, a warm west/cold east SAT dipole can be easily seen over NA in the small PV gradient winter with low BDL SIC. On the other hand, while the BDL SIC and midlatitude SST anomalies are not largely influenced by the GB change, they can determine the magnitude of the PV gradient prior to the blocking onset and then influence the evolution of the GB with a 10–20-day time scale. Thus, it is thought that the North Atlantic SIC and SST anomalies as background states can influence cold anomalies over NA and NE through changing the magnitude of the PV gradient prior to the blocking onset and then the

subsequent GB. These results are different from previous findings (Messori et al. 2016; Ballinger et al. 2018; Cohen et al. 2018a; Overland and Wang 2018).

Moreover, it is revealed that the BDL SIC and midlatitude SST show opposite trends. These trends can counteract each other to result in a weak trend of the meridional PV gradient in that the BDL SIC decline (midlatitude North Atlantic SST warming) leads to the reduced (enhanced) meridional PV gradient. For this reason, the duration of the GB exhibits a weak upward trend. Thus, the trends of cold anomalies over NE and NA are inevitably weak. However, it must be pointed out that although we presented a new linkage pathway about how the BDL SIC or North Atlantic SST anomaly affects the midlatitude weather patterns, a numerical model study is needed to further understand whether our PV gradient theory of the Arctic–midlatitude linkage can apply to other regions and seasons. This deserves a further investigation. Of course, the cold winters over NA also depend on the upstream large-scale wave trains coming from the North Pacific related to the Arctic SIC over the North Pacific side and tropical forcing. This problem warrants a further investigation in the future work.

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